Geology of East-Central Idaho: Geologic Roadlogs for the Big and Little Lost River, Lemhi, and Salmon River Valleys

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INTRODUCTION

This article contains a discussion of the regional geology and historical geography of east-central Idaho, north of the Snake River Plain and southeast of the Salmon River, and four roadlogs for the main highways in this huge (15,000-km^2) area. The roadlogs progress from south to north and west to east. They include:

1) Big Lost River Valley,
2) Little Lost River-Pahsimeroi Valley,
3) Birch Creek-Lemhi Valley, and
4) Stanley to Salmon along the Salmon River.

This article covers much of the same country described by Ross (1963) in the first geologic roadlog of east-central Idaho, and is a companion to field trip descriptions of south-central Idaho to the west, from Arco to Ketchum (Link et al., 1988), and of the central Lost River Range (Janecke, 1995a). With those papers, it completes an updated set of geologic descriptions of east- and south-central Idaho. Our intention is to provide more precise and fully referenced information than available in the Roadside Geology of Idaho (Alt and Hyndman, 1989). Our intended audience includes all manner of geologists and those interested in the interaction of Idaho geology, geography and history.

Regional geologic map references for the present discussions include Challis National Forest (Wilson and Skipp, 1994), Idaho National Engineering Laboratory (Kuntz et al., 1994); Challis 1° x 2° sheet (Fisher et al., 1992); Dillon 1° x 2° sheet (Ruppel et al., 1993); Hailey and western Idaho Falls 1° x 2° sheets (Worl et al., 1991); and the surficial map of the eastern Snake River Plain (Scott, 1982). One of our agendas is to promote use of those maps. Without a geologic map, many of our descriptions lose context. Another important standard reference are stratigraphic correlation charts for Idaho (Isaacson et al., 1983 and Ballard et al., 1983).

As part of this introduction we present several summary figures. First is a regional geographic map (Fig. 1), and a simplified geologic map on the same base (Fig. 2). The Proterozoic and Paleozoic stratigraphic correlation diagram (Fig. 3) also shows thrust fault relations. Figure 4 shows major structures and basins of the Trans-Challis fault zone and Figure 5 shows the Challis Volcanic field. Photographs in the road log section illustrate many of the field localities. In the introductory summary we cite (out-of-sequence with their position in the text) several photographs in which specific stratigraphic features are present.

East-central Idaho contains diverse and economically important, mineral deposits. Specific areas of mineralization are discussed in the roadlogs. More general discussions include Hall (1985), Ruppel and Lopez (1988), Fisher and Johnson (1995a), and Worl et al., (1995), and papers therein.

GEOLOGIC HISTORY OF EASTERN IDAHO
Summary

East-central Idaho (Figs. 1 & 2) lies within the Cordilleran fold and thrust belt, and in the Basin-and-Range province. Rocks and structures within this region reflect a long and complex history of deformation. Strata were deposited here in the Mesoproterozoic Belt intracratonic rift basin, and episodically in the late Neoproterozoic and Paleozoic Cordilleran miogeocline. The Early Neoproterozoic Beaverhead Impact caused shock deformation and may have controlled subsequent tectonic instability in the southern Beaverhead and Lemhi Mountains. During the Paleozoic, marginal basins and uplifts formed, rather than the regionally extensive Paleozoic passive-margin succession that is present south of the Snake River Plain. Paleozoic tectonostratigraphic events include: transpressional latest Devonian and Mississippian Antler deformation, Early Mississippian
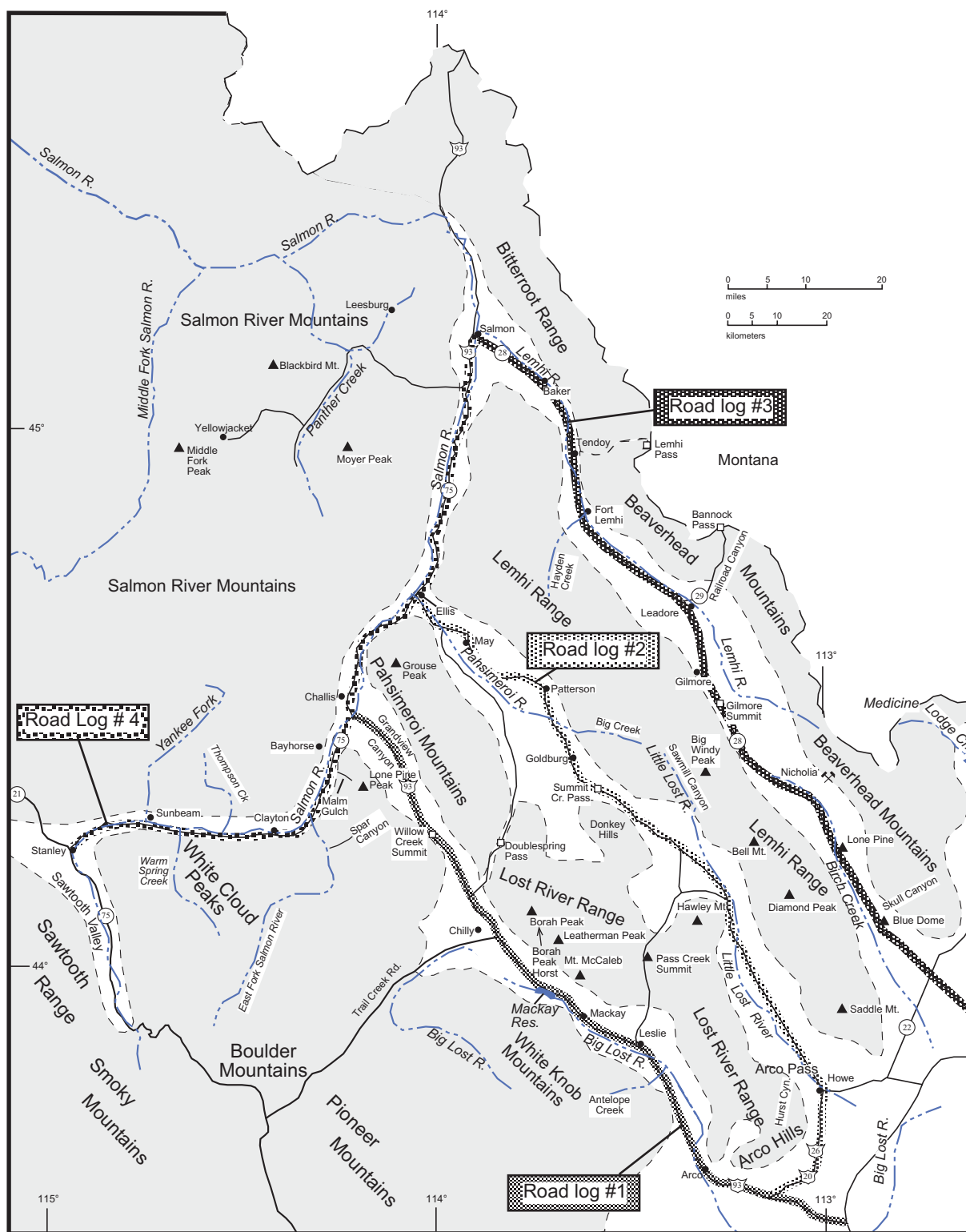
Figure 1. Map showing geography of east-central Idaho, with some of the localities mentioned in this article. Route of segments 1,2,3, and 4 are shown. Figure 2 is at the same scale.

faulted foreland-basin deposition east of the Antler belt, and inversion tectonics during the Pennsylvanian and Permian Ancestral Rockies orogeny. Large carbonate bank systems were present in Silurian, Late Devonian, and Late Mississippian time. In the central-Idaho black-shale mineral belt, syngenetic sedimentary-exhalative base-metal deposits of the Devonian Milligen Forma-

tion formed in normal-fault bounded marginal-basins. The Early Mississippian Madison Group carbonate bank did not prograde west into east-central Idaho.

Deformation and intrusion of the Mesozoic Cordilleran orogenic belt produced regional northeast-vergent thrust faults, numerous folds (Ross, 1947), and, in the western part of the area,
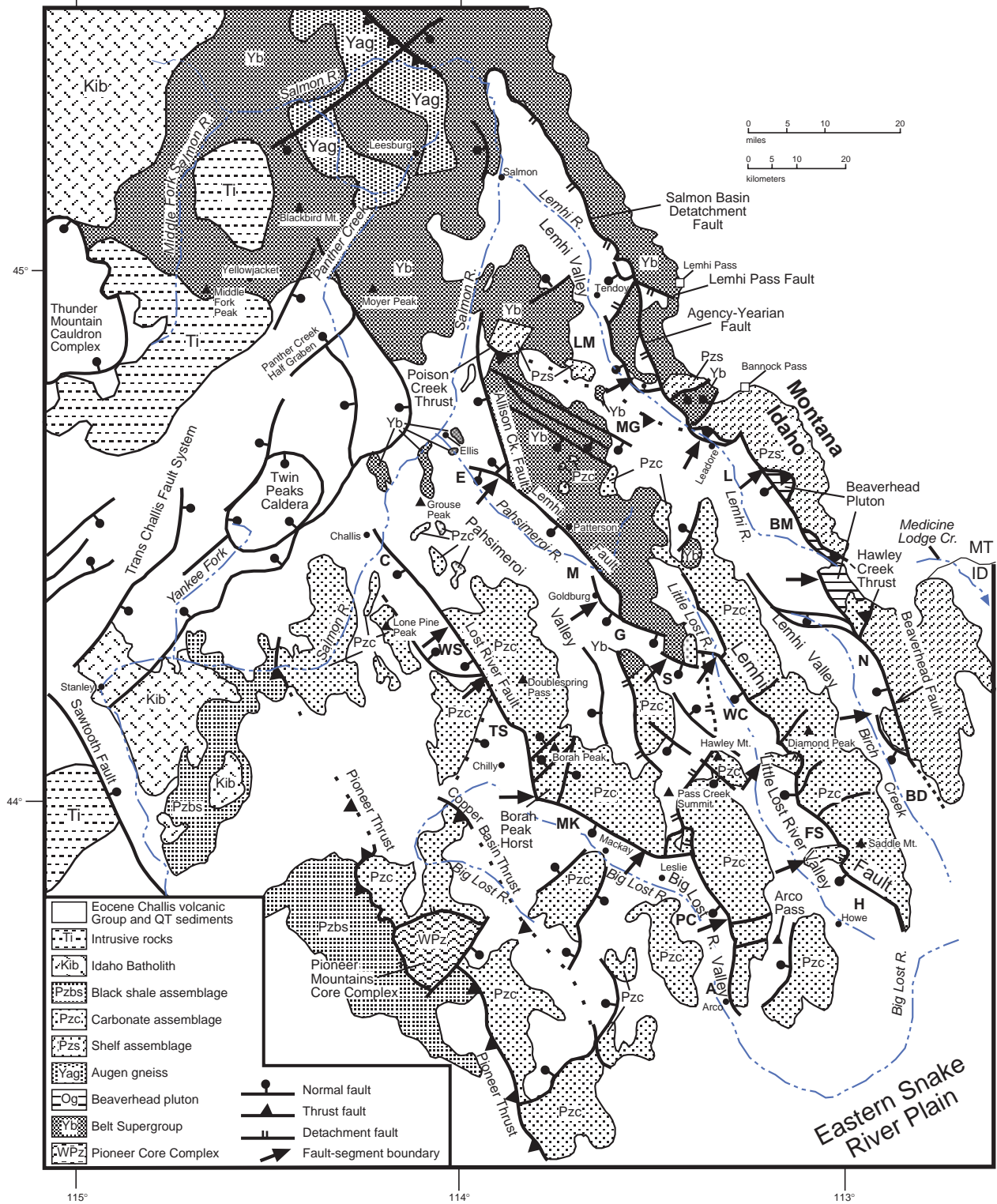
Figure 2. Regional geological map of pre-Tertiary rocks of east-central Idaho. Map compiled from Bond (1978) and modified after Wilson and Skipp (1994), Schmidt et al. (1994) and Janecke et al. (1997; 1998). Segment boundaries on normal faults are indicated by heavy arrows and are named as follows (from Janecke, 1993 and references therein): Lost River fault; A=Arco; PC=Pass Creek; MK=Mackay; TS=Thousand Springs; WS=Warm Spring; C=Challis; Lemhi Fault, H=Howe; FS=Fallert Springs; S=Sawmill Gulch; G=Goldburg; P=Patterson; M=May; Beaverhead fault, BD=Blue Dome; N=Nicholia; BM=Baldy Mountain; L=Leadore; MG=Mollie Gulch; LM=Lemhi.
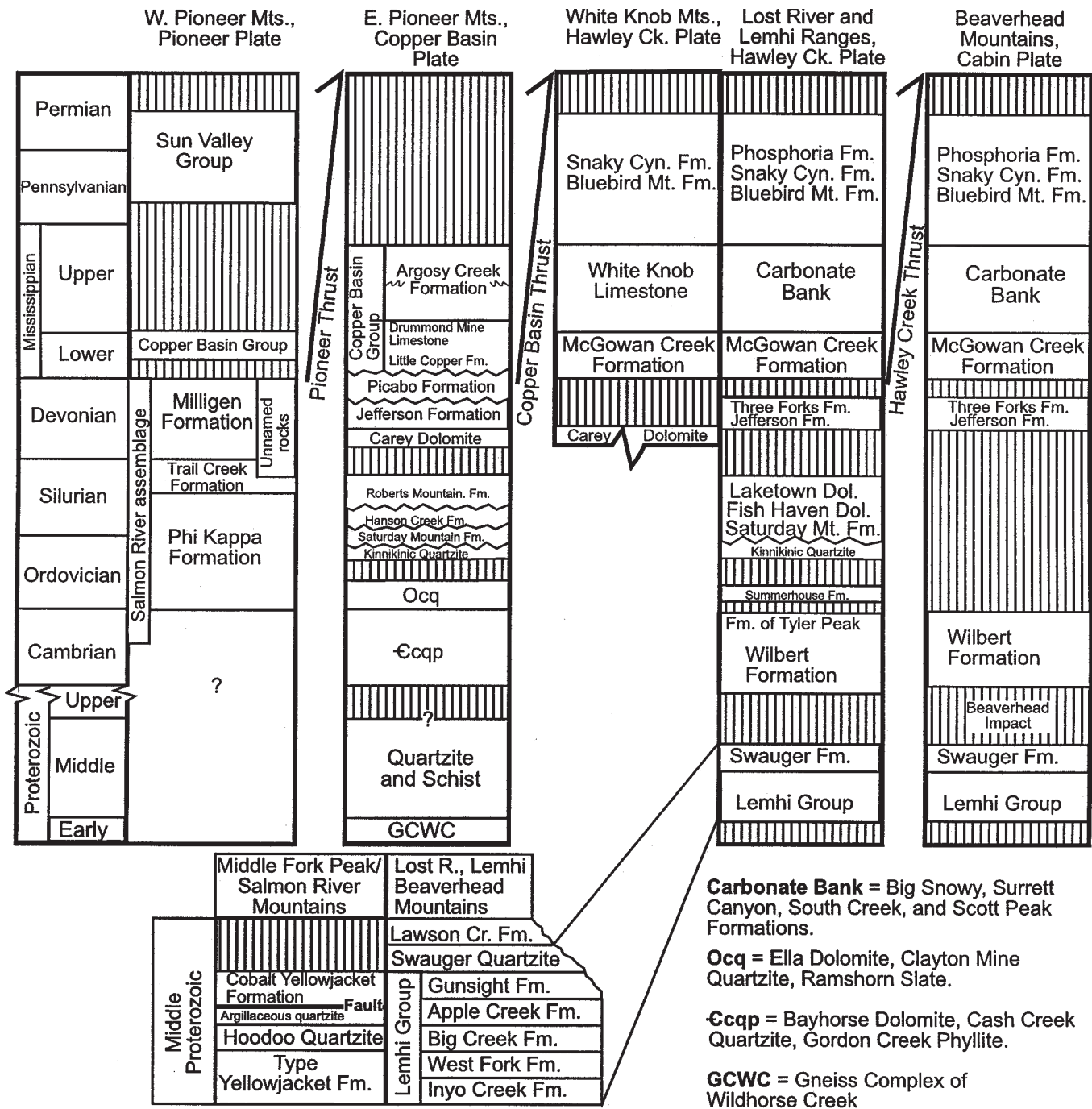
Figure 3. Stratigraphic units present in several thrust sheets and mountain ranges of east-central Idaho, after Skipp (1988a), Wilson and Skipp (1994), Link et al. (1995; 1996).

the extensive, mainly Late Cretaceous, Atlanta lobe of the Idaho batholith. Extension along several sets of normal faults began before Middle Eocene Challis volcanism, exhumed the Pioneer metamorphic core complex, and produced numerous Tertiary half-grabens in a system of north-trending Paleogene basins. The Challis Volcanic Group and associated shallow plutons covered and intruded much of the northern and western parts of the area, and produced diverse mineral deposits.

The area is part of the northern Basin-and-Range province,

and is actively extending along a system of dominantly north-northwest-striking normal faults. The magnitude 7.3 Borah Peak earthquake in 1983 provoked interest in the neotectonics and earthquake hazards of the area. Today, east-central Idaho is on the northern flank of the late Cenozoic track of the Yellowstone-Snake River Plain hotspot, which has produced bimodal volcanic rocks along the plain and an east-northeast-trending topographic bulge.
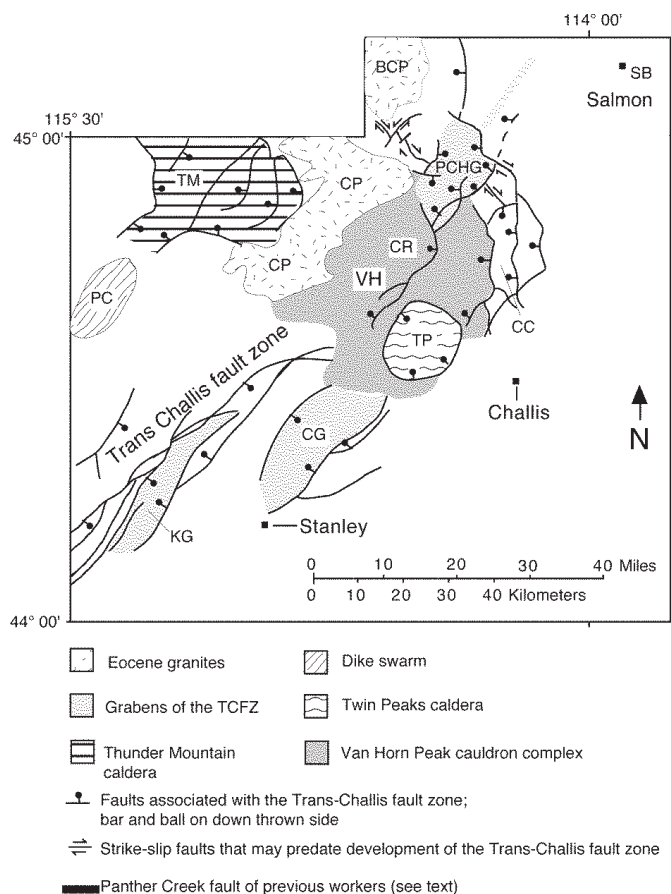
Figure 4. Challis Volcanic Field and Trans-Challis Fault Zone. Regional map of the main calderas of the Challis Volcanic field, the Trans-Challis fault zone (TCFZ) and younger normal faults. BCP=Bighorn Crag pluton; CC=Corral Creek cauldron segment; CG=Custer graben; CP=Casto pluton, CR=Castle Rock cauldron segment; KG=Knapp Creek graben; PC=Pistol Creek dike swarm; PCHG=Panther Creek half graben; TM=Thunder Mountain cauldron complex; TP=Twin Peaks caldera; VH=Van Horn Peak cauldron complex. Sources: Shockey (1957); Bennett (1977); Harrison (1985); Ekren (1988); Janecke (1994); Evans and Connor (1993); Fisher et al. (1992); Janecke and Snee (1993); Modreski (1985). (Modified from Janecke et al., 1997, Figure 1).

Proterozoic and Paleozoic Events

Paleogeography

Proterozoic and Paleozoic rocks of east-central Idaho are part of three major Cordilleran thrust plates, from west to east the Pioneer, Hawley Creek, and Cabin plates. Horizontal separation on the bounding thrust faults is not well known, but there are no exotic terranes within this area. All the Paleozoic rocks were deposited in basins marginal to or connected with the Wyoming craton. As we discuss the stratigraphic units we make specific mention, when necessary, to the presence of a feature in a specific thrust plate, which we define and describe in the succeeding section. As demonstrated in the central and southern Basin-and-Range province (Levy and Christie-Blick, 1989), Neogene extension generally acted to reverse Mesozoic shortening. Thus, a palinspastic map of east-central Idaho would likely reflect no more than 50 km of northeast shortening from depositional sites of Pro-

terozoic and Paleozoic rocks. This rough estimate is more conservative than some older, more mobilist estimates (Skipp and Hait, 1977; Ruppel, 1978)

Belt Supergroup

Middle Proterozoic metasedimentary rocks in the Salmon River Mountains, and northern parts of the Lost River, Lemhi and Beaverhead ranges include meta-sandstones and siltites of the Yellowjacket Formation, Lemhi Group, and Swauger Formation of the 1470-1390 Ma Belt Supergroup (Figs. 3, 23, 48-50) (Ruppel, 1975; Aleinikoff et al., 1996; Sears et al., 1998; Winston et al., 1999, this volume). The Belt Supergroup rift basin formed across much of east-central Idaho and western Montana (Winston and Link, 1993). Its irregular but roughly north-trending eastern margin lies about 150 mi east of the Beaverhead Mountains in southwest Montana. The southwest margin is less well defined, but must lie northeast of the Pioneer metamorphic core complex and southwest of the Lost River Range.

In general, the Belt Supergroup contains quartzose sandstone and mudrock derived from highlands to the southwest, that are now part of a Pacific continent, possibly Australia or Siberia (Sears and Price, 1978; Ross et al., 1991a; 1992). The Lemhi Group and Yellowjacket Formation were generally deposited in alluvial aprons and marginal mudflats on the edges of the great Belt intracratonic sea or lake (Winston and Link, 1993; Winston et al., 1999, this volume). These rocks are generally poor in carbonate, but the upper part of the Apple Creek Formation, of the Lemhi Group (Fig. 48), contains the same depositional cycles as the Middle Belt carbonate (Helena Formation) of southwestern Montana.

Northwest of Salmon, Idaho, a bimodal suite of granitic plutons and mafic rocks invaded the Belt basin about 1370 Ma (Evans and Zartman, 1990; Doughty and Chamberlain, 1996), a time that is now thought to immediately post-date Belt deposition (Aleinikoff et al., 1996). Shearing and metamorphism of these rocks northwest of Salmon, Idaho increases northward (Doughty and Chamberlain, 1996).

Modern paleogeographic analyses of the Belt basin have not identified a medial Belt Island or Salmon River Arch (Skipp and Link, 1992; Link et al., 1993; Winston et al., 1999, this volume). The concept that during deposition of the Belt Supergroup there was a highland between the Lemhi Group in Idaho and the Belt Supergroup in Montana, plus the notion that the regional Medicine Lodge thrust fault structurally separated the Lemhi Group from the Yellowjacket Formation and Belt Supergroup (Armstrong, 1975; Ruppel and Lopez, 1978; Ruppel, 1986; Ruppel et al., 1993), are stratigraphically and structurally unnecessary, and were based on inaccurate K-Ar geochronology (Aleinikoff et al., 1996; Evans, 1999).

Beaverhead Impact

A large bolide struck east-central Idaho in early Neoproterozoic time (850-900 Ma) and produced a crater 75-150 km in diameter (Hargraves et al., 1990; Kellogg et al., 1999). The Beaverhead Impact produced shatter cones, breccia, and pseudotachylite in Mesoproterozoic quartzite and Archean gneiss on the east side of the Beaverhead Mountains in southwest Mon-
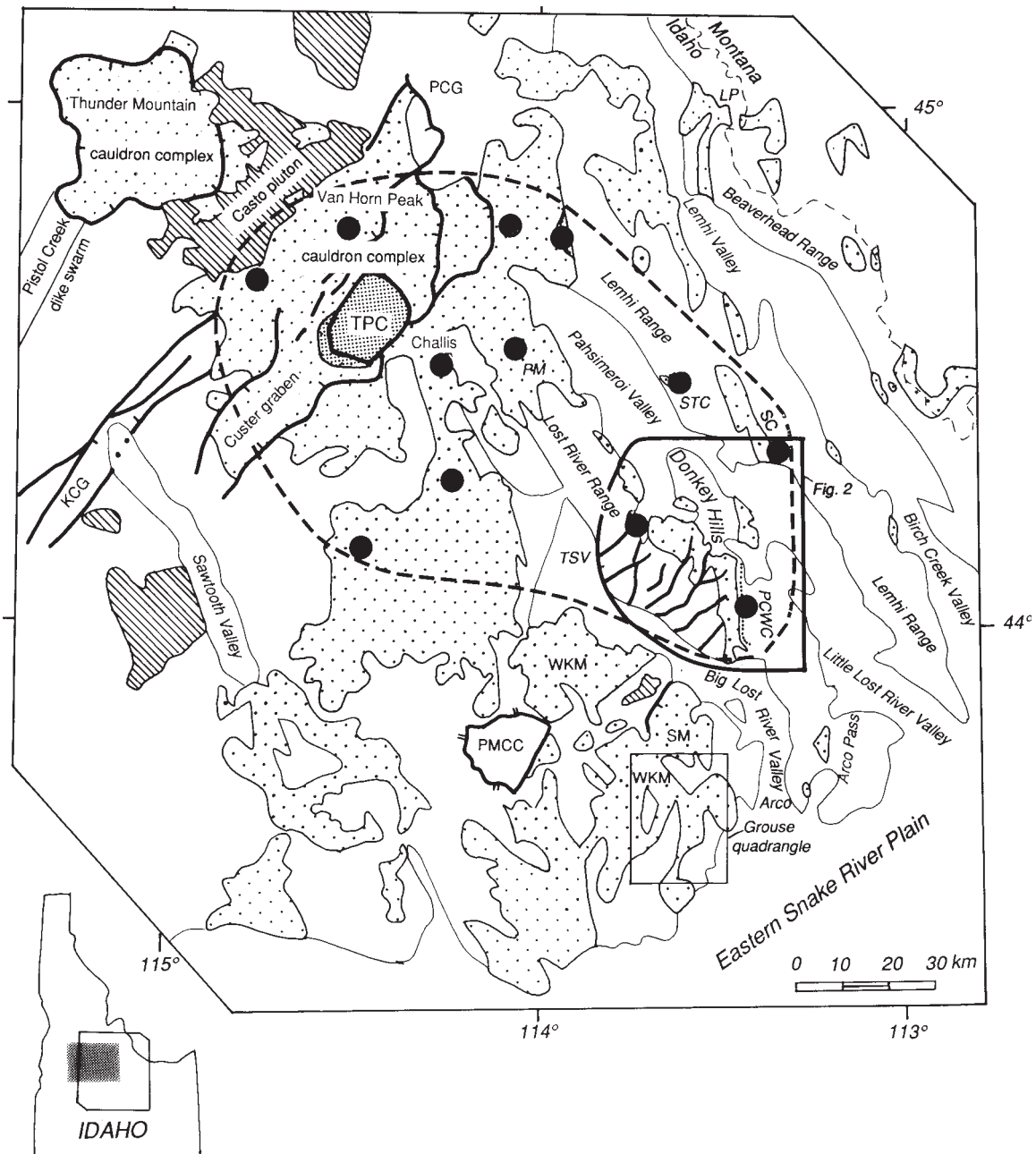
Figure 5. Map of Challis volcanic field, Idaho, showing the distributions of volcanic rocks (coarse dots) and selected large plutons (diagonal rule). The tuff of Challis Creek in the Twin Peaks caldera (TPC) is shown with fine dots and exposures of the tuff outside the caldera are large solid dots. The presently known maximum extent of the tuff of Challis Creek is dashed. For clarity, only selected Eocene normal faults are shown. Abbreviations: PMCC=Pioneer metamorphic core complex; WKM=White Knob Mountains; TSV=Thousand Springs Valley; LP=Lemhi Pass; PCG=Panther Creek graben; PM=Pahsimeroi Mountains; STC=Stinking Creek; SC=Sawmill Canyon; KCG=Knapp Creek graben; SM=Sheep Mountain; and PCWC=Pass Creek-Wet Creek reentrant. Challis 1° x 2° sheet is shaded on index map. Sources: Bond (1978); Fisher (1985); Fisher et al. (1992); Hardyman (1985); Hardyman and Fisher (1985). (Reproduced from Janecke and Snee, 1993, Figure 1). Area labeled as Fig. 2 shows the outline of Figure 8.

tana (Hargraves et al., 1990; 1994). Farther west, overlapping, midcrustal geophysical anomalies are centered south of Challis, Idaho in the central Lost River Range and are attributed to the impact (McCafferty, 1995). If reconstructions are correct, this is one of the eight largest impacts known in earth history. Carr and Link (1999, this volume) describe breccia and conglomerate in the Neoproterozoic formation of Leaton Gulch south of Challis, Idaho, which they connect with the Beaverhead Impact. Long-term effects of the Beaverhead Impact may include Neoproterozoic

and Paleozoic recurrent uplift of the Lemhi Arch (Hargraves et al., 1994).

Wilbert Rift Basin and the Missing Miogeocline

In latest Neoproterozoic and Cambrian time, the Wilbert Formation and overlying formation of Tyler Peak (McCandless, 1982) were deposited in an interpreted depositional rift basin which broadened southward into southeastern Idaho (Skipp and Link,

1992). Pebbly sandstones of the Wilbert Formation are present in the Borah Peak horst of the central Lost River Range, and in the southern Lemhi and southernmost Beaverhead ranges, across several thrust sheets (Fig. 3). Areas farther north in the Beaverhead Mountains that contain shatter cones of the Beaverhead Impact were inferred by Skipp and Link (1992) to be part of the Wilbert Formation but are now correlated to the Lemhi Group (Belt Supergroup) (Ruppel, 1998; Kellogg et al., 1999).

Thick (up to 6 km) basal strata of the Cordilleran miogeocline, present near Pocatello south of the Snake River Plain, include Neoproterozoic diamictite and quartzose rocks (Pocatello Formation and Brigham Group) and overlying Cambrian carbonate rocks (Link et al., 1993). However, north of the eastern Snake River Plain, this time interval is represented only by 150 m of the Wilbert and Tyler Peak formations in the southern Lemhi Range (McCandless, 1982) and at most several hundred meters of Wilbert Formation in the southern Beaverhead Range (Skipp and Link, 1992). In the Bayhorse area west of Challis, however, thick siliciclastic Cambrian rocks are present (Hobbs et al., 1991; Hobbs, 1995). These thick Cambrian rocks are far west of their expected positions based on facies patterns to the south.

Scattered small occurrences of Neoproterozoic rocks are present in west-central Idaho, as pendants in the Idaho batholith (Lund, 1999), but relations and correlations are sketchy due to subsequent metamorphism and intrusion. Pope and Sears (1997) suggest that some of the Neoproterozoic and Cambrian miogeoclinal rocks that are missing from central Idaho may have been translated north to British Columbia during Mesozoic continental truncation.

Lemhi Arch and Beaverhead Mountain Pluton

In the Lost River, Lemhi and southern Beaverhead Ranges, pebbly Lower Ordovician sandstones (Summerhouse Formation) unconformably overlie Mesoproterozoic or Neoproterozoic to Cambrian strata. In the southern Beaverhead Mountains, this unconformity was attributed to the Skull Canyon disturbance by Scholten (1957). This disturbance and the sub-Ordovician unconformity are manifestations of the Salmon River/Lemhi Arch, a westward projecting promontory or uplift within the Cordilleran miogeocline, whose original concept as a recurrent Paleozoic feature (Sloss, 1954) is still sound. This feature may also explain the near absence of Neoproterozoic to Cambrian rocks in east-central Idaho (Burchfiel et al., 1992). Extension of the feature into Mesoproterozoic time (Armstrong, 1975; Ruppel, 1986) is incorrect however, mainly because such extension was based on inaccurate K-Ar geochronology (Armstrong, 1975) that suggested there was intrusion and metamorphism northwest of Salmon Idaho near 1500 Ma, prior to, or during formation of the Belt basin (Skipp, 1987; Skipp and Link, 1992; Winston et al., 1999, this volume). These intrusive rocks have been redated at 1370 Ma and clearly postdate much or all of the Belt Supergroup (Evans and Zartman, 1990; Doughty and Chamberlain, 1996). The absence of thick Neoproterozoic and Cambrian strata in the Lost River, Lemhi, and Beaverhead ranges, and the unique Ordovician Beaverhead Mountains pluton (see below) are regional anomalies within the miogeocline whose origin remains elusive.

Temporary tectonic stability and depositional continuity between east-central Idaho and the miogeocline were first established in Middle Ordovician time, with deposition of the north-derived shallow marine quartz arenite sheet of the Kinnikinic Quartzite (Figs. 27 and 44), whose equivalents (Swan Peak and Eureka quartzites) are present south to southern Nevada (Ross, 1977; Oaks et al., 1977). The Kinnikinic Quartzite thins eastward to a feather edge west of the Cabin thrust; it is missing in the Cabin plate (Fig. 3) (James and Oaks, 1977; Skipp, 1988). Oaks et al. (1977) interpreted Kinnikinic deposition to have been controlled by the exposed, northwest-trending, Lemhi Arch along the Idaho/Montana border, and a mainly submerged, northeast-trending, Arco Arch along the present eastern Snake River Plain.

Shortly after Middle Ordovician deposition of the Kinnikinic Quartzite, granite and syenite of the Beaverhead Mountains pluton invaded the western Beaverhead Mountains (~483 Ma, Rb-Sr) (Scholten, 1957; Scholten and Ramspott, 1968; Lucchitta, 1966; Evans and Zartman, 1988). Two small intrusions in the Salmon River Mountains, west of Salmon, also date from this time (Evans and Zartman, 1988). The relationship, if any, between these early Paleozoic plutons and the unstable and unconformity-riddled early Paleozoic section of east-central Idaho is uncertain. In the Lost River Range and in most of southeastern Idaho, distant from the early Paleozoic intrusions, Ordovician to Silurian dolostones of the Saturday Mountain, Fish Haven and Laketown formations conformably to disconformably overlie the Kinnikinic Quartzite (Oaks et al., 1977; James and Oaks, 1977). The "western assemblage" fine-grained Phi Kappa and Trail Creek formations of the Pioneer thrust plate contain Ordovician graptolites of an open-ocean warm-water Pacific province, significantly different from coeval western North American assemblages (Berry, 1998).

Silurian and Devonian Carbonate Banks and Basins

During Silurian time the Lost River and Lemhi Ranges (Hawley Creek thrust plate) were part of the regional carbonate platform now represented by the Laketown Dolomite and equivalents (Figs. 3 and 14) (Poole et al., 1977; Sheehan et al., 1991). Silurian strata thin to the east to a feather edge below Devonian carbonates in the southern Lemhi Range. The thickest Silurian carbonate platform succession is in the Lone Pine Peak area south of Challis (Fig. 2) where the Lone Mountain dolomite is about 800 m thick (Hays et al., 1978; 1980, Hobbs et al., 1991). In the Hawley Creek-Copper Basin thrust plate of the Pioneer Mountains, Silurian platy calcareous siltstones are assigned to the Roberts Mountains Formation (600 m) (Skipp and Sandberg, 1975; Dover, 1983). Areas to the west of the Pioneer thrust expose "western assemblage" Silurian Trail Creek Formation which contains slope-facies cherty argillite and siltite (Dover, 1983; Link et al., 1995).

Lower and Middle Devonian strata of the Hawley Creek and Copper Basin plate include restricted shallow marine Carey Dolomite and incised valley deposits of the estuarine Beartooth Butte Formation that contain a rich fossil-fish fauna (Johnson et al., 1988; Grader and Dehler, 1999, this volume). A regional upper Devonian carbonate ramp was deposited during the eustatic sea level rise manifested in the Taghanic onlap (Johnson et al., 1991). It extended to western Montana and is represented by the Jefferson

and Three Forks formations (Sandberg et al., 1988; Dorobek, 1991). The Grandview Member of the Jefferson contains bioherms at Grandview Canyon south of Challis (Fig. 15) (Isaacson et al., 1988; Isaacson and Dorobek, 1988). Late Paleozoic uplift of the Lemhi Arch (*sensu* Sloss, 1954) is manifested in the Lemhi Range by southward thinning of the Jefferson Formation from 900 m in the central part to 100 m at the south end of the range (Grader and Dehler, 1999, this volume). Dorobek et al. (1991), in contrast, argued that such differential subsidence was due to reactivation of preexisting structures during vertical and horizontal loading by the Antler orogenic belt.

On the Pioneer thrust sheet, Devonian carbonaceous, argillaceous and locally calcareous strata of the Milligen Formation and equivalent (?) Salmon River assemblage make up the central-Idaho Black Shale Mineral Belt (Fig. 42, 43) (Link et al., 1995). In the southern Wood River Valley, and at the head of Slate Creek in the White Cloud Mountains, these strata contain syngenetic Ag-Pb-Zn mineral deposits formed in inferred exhalative rift-basins (Hall, 1985; Turner and Otto, 1995). Rocks of the black-shale belt were deformed and uplifted by the Antler orogen, though specific non-reactivated Antler structures are elusive (Skipp and Sandberg, 1975; Skipp and Hait, 1977; Skipp and Hall, 1980; Dover, 1980; Hall, 1985; Wilson et al., 1994; Link et al., 1996).

Antler Flysch Trough

Early Mississippian rapid subsidence and normal faulting of the Antler orogen is manifested in thick conglomeratic foreland-basin deposits of the Copper Basin Group in the White Knob and Pioneer Mountains, and the coeval fine-grained McGowan Creek Formation in the Lost River and Lemhi Ranges (Fig. 9) (Paull et al., 1972; Paull and Gruber, 1977; Nilsen, 1977). The anomalous thickness of the Copper Basin Group, the coarse and proximal recycled sedimentary detritus, and evidence for syn-sedimentary growth faults suggest a transpressional or transtensional setting for Antler deformation in Idaho (Wilson et al., 1994; Rodgers et al., 1995; Link et al., 1995; 1996).

Mississippian strata are largely missing from the Pioneer thrust plate, which was presumably part of the uplifted Antler highland and a source terrane for Copper Basin strata. Age populations of detrital zircons in the Argosy Creek Formation, Copper Basin Group are dominated by 2000 to 1700 Ma grains (Preacher et al. 1995; Link et al., 1996). This signature is similar to that in the Middle Ordovician Valmy Formation of the Roberts Mountains allochthon, Nevada (Smith and Gehrels, 1994). The data suggest that Ordovician sandstones of the Basin Gulch Member, Phi Kappa Formation (Valmy equivalent in the Pioneer thrust plate) or Kinnikinic Quartzite (Hawley Creek thrust plate) were a major source for sand and quartzite cobbles in the Copper Basin Group. The ultimate source for the ~1800 Ma Paleoproterozoic sand grains was the Peace River-Athabasca Arch in northern Canada (Ketner, 1968; Ross et al., 1991b).

Upper Mississippian and Pennsylvanian Carbonate Bank

East of the Copper Basin-McGowan Creek flysch trough, two major Mississippian shallow-subtidal to intertidal carbonate bank complexes prograded westward from the Wyoming-Montana shelf (Poole and Sandberg, 1977; 1991; Skipp, Sando, and Hall, 1979).

The Lower Mississippian (Kinderhookian and Osagean) Madison Group comprises the lower complex, but these well-known limestones of the western Wyoming area did not extend westward across the Montana border to east-central Idaho. In the Cabin thrust plate of the southern Beaverhead Mountains, thin (150 m) argillaceous strata of the McGowan Creek Formation represent this time interval (Skipp, 1988a). The McGowan Creek Formation thickens westward and makes up huge talus slopes in the Lost River Range (Fig. 9, right side of Fig. 10), where it was initially incorrectly mapped as the Devonian Milligen Formation (Ross, 1947; Sandberg, 1975).

The upper carbonate bank complex, of Late Mississippian (Meramecian and Chesterian) age, prograded rapidly across what had been the Antler flysch trough, and is represented by thick and prominent carbonate bank strata (Middle Canyon, Scott Peak, South Creek, and Surrett Canyon formations, Huh, 1967; Skipp, Sando and Hall, 1979) of the Beaverhead, Lemhi and Lost River Ranges (Figs. 9, 13, 21, 25 and 26) and the White Knob Limestone of the White Knob Mountains (Link et al., 1996).

In the Hawley Creek, Fritz Creek, and Cabin thrust plates, Upper Mississippian, Pennsylvanian, and Permian mixed carbonate and siliciclastic rocks of the Snaky Canyon, Bluebird Mountain, Arco Hills, and locally, the Phosphoria formations are the youngest exposed strata (Skipp, Hoggan et al., 1979; Skipp, 1988a). The Snaky Canyon Formation was deposited on a carbonate platform that at times received fine-grained, likely aeolian, quartzose silt from the craton to the north and east, and which developed local *Palaeoaplysina* bioherms near Howe (Fig. 16, 17)(Breuninger et al., 1988; Canter and Isaacson, 1991).

Wood River Basin

On the Pioneer thrust plate, the Oquirrh-Wood River basin, a regional north-northwest-trending, Middle Pennsylvanian to Early Permian mixed carbonate-siliciclastic turbidite basin, formed above deformed strata of the Antler highland. It has been interpreted as the most distant manifestation of the Ancestral Rockies orogen, and its subsidence has been attributed to crustal loading by reactivation of a segment of the Antler thrust load in southwestern Idaho (Geslin, 1998), or to inversion of the Antler foreland basin in the Pioneer Mountains (Skipp and Hall, 1980). The mixed carbonate-siliciclastic Sun Valley Group filled this basin, with sources to the west for chert-pebbles, north for quartzose sand and east for carbonate material (Hall et al., 1974; Hall, 1985; Mahoney et al., 1991; Burton and Link, 1992). The central part of the basin was periodically oxygen-starved at the sediment-water interface, and organic-rich mudrocks of the Dollarhide Formation were deposited (Hall, 1985; Link et al., 1995).

Missing Mesozoic Record

The early Mesozoic history of east-central Idaho is little known, as no rocks are preserved, except for isolated Mesozoic strata in the southern Lemhi Range and Beaverhead Mountains (Lucchitta, 1966; Skipp, Hoggan et al., 1979). Mesozoic strata were either never deposited or, more likely, have since been eroded from the rest of east-central Idaho.

Mesozoic Deformation: Cordilleran Orogenic Belt

In the Cretaceous, east-central Idaho shortened in an east-northeast to west-southwest direction as part of the Sevier fold-and-thrust belt (Ruppel, 1978; Skipp, 1988a; Allmendinger, 1992). Rocks of the Pioneer thrust sheet on its western edge were invaded by the Late Cretaceous Atlanta lobe of the Idaho batholith (Lewis and Kiilsgaard, 1987).

Many contacts previously mapped as regional thrust faults in east-central Idaho (e.g. Skipp and Hait, 1977; Ruppel, 1978; Hall, 1985; Hobbs, 1995) are stratigraphic contacts or small faults (Rodgers and Janecke, 1992). The Glide Mountain thrust in the Pioneer Mountains (Dover, 1981), for example, was shown to be a mix of sheared stratigraphic contacts and Eocene normal faults (Wilson et al., 1994). Other "major" thrust faults (e.g. Dover, 1983; Ruppel et al., 1993), have been reinterpreted as decollements associated with folds, ductility contrasts in the stratigraphic package, or deformation zones produced during bedding-parallel shear (Burton and Link, 1995; Rodgers et al., 1995; Winston et al., 1999, this volume). Small local thrust faults duplicate the Paleozoic strata throughout this region (Beutner, 1968; Hait, 1987; Susong et al., 1990; Fisher and Anastasio, 1994).

Folds in east-central Idaho occur at a range of scales from outcrop to mountain. They deform Middle Proterozoic to Triassic rocks, trend generally north-northwest, and are upright to east-northeast-vergent (Ross, 1947; Beutner, 1968; Dover, 1981; 1983; Garmezy, 1981; Skipp, 1984; 1998; Mapel et al., 1965; Mapel and Shropshire, 1973; Fisher and Anastasio, 1994; Rodgers et al., 1995; Jeppson and Janecke, 1995).

Pioneer Thrust Plate

The west-dipping Pioneer thrust fault strikes northward through the central Pioneer Mountains (Dover, 1983) (Fig. 2). Rocks west of the fault are lower Paleozoic "western-facies" cherty and argillitic strata, unconformably overlain by the Pennsylvanian and Permian Sun Valley Group deposited in the southwestward-deepening Wood River Basin (Skipp and Hall, 1980; Mahoney et al., 1991; Link et al., 1995; Geslin, 1998) (Fig. 3). These rocks are intruded along the Salmon River and in the White Cloud Peaks by Late Cretaceous granitoids of the Idaho batholith (Fisher et al., 1992; Worl et al., 1991). The Pioneer thrust is exposed north and south of the Pioneer metamorphic core complex (Dover, 1981) (Figs. 1 and 2).

Hawley Creek-Copper Basin Thrust Plate

East of the Pioneer thrust, the Hawley Creek thrust plate comprises the rocks exposed in the eastern White Knob Mountains, Lost River, central and southern Lemhi and westernmost Beaverhead ranges (Skipp, 1987; Tietbohl, 1986; Rodgers and Janecke, 1992; Tysdal and Moye, 1996; Tysdal, 1996a). The tectonic significance of the Copper Basin thrust fault, in the White Knob Mountains (Skipp and Hait, 1977) is uncertain because the fault is high-angle and juxtaposes rocks of different facies but of essentially the same age (Skipp and Bollmann, 1992). The Hawley Creek-Copper Basin thrust plate contains Mesoproterozoic Belt rocks, pre-upper Devonian stratigraphic thinning associated with the Lemhi Arch, the Lower Mississippian Antler flysch trough, and the Upper Mississippian carbonate bank (Fig. 3). The Hawley

Creek thrust fault is exposed in the western Beaverhead Mountains and the northern Lemhi Range (Lucchitta, 1966; Skipp, 1987; 1988a)(Fig. 2).

Fritz Creek and Cabin Thrust Plates

East of the Hawley Creek thrust fault in the Fritz Creek and Cabin thrust plates of the Beaverhead Mountains (Fig. 3) are strata of the Mesoproterozoic Belt Supergroup rift basin that contain shattercones of the Beaverhead Impact Event (Hargraves et al., 1994; Lucchitta, 1966), the eastern edge of the Neoproterozoic Wilbert Formation rift basin (Skipp, 1988a; Skipp and Link, 1992), and eastward-thinning strata of the Paleozoic miogeocline. Ordovician rocks are missing from the Cabin Plate, and Devonian carbonates overlie Neoproterozoic to Cambrian, Mesoproterozoic, or Archean rocks (Fig. 3) (Skipp, 1988a). This unconformity is attributed to Ordovician uplift of the Lemhi Arch (Oaks et al., 1977).

Paleogene Uplift and Pre-Challis Tectonism

Crustal thickening in the western part of the Sevier thrust belt accompanied intrusion of the Idaho batholith, mainly in Late Cretaceous time (Lewis et al., 1987; Johnson et al., 1988; Lewis and Kiilsgaard, 1991; Kiilsgaard et al., 1995). The Atlanta lobe of the Idaho batholith was tilted eastward, exhumed and uplifted 12-19 km after emplacement in the Late Cretaceous but before Middle Eocene time (Jordan, 1994; Rodgers et al., 1995).

Analysis of the pre-Tertiary subcrop in the Lost River and Lemhi Ranges shows that progressively older rocks crop out beneath the Eocene unconformity northward from the eastern Snake River Plain (Rodgers and Janecke, 1992). This pattern is in part due to uplift of hanging wall rocks over frontal and lateral ramps in underlying thrust faults (Skipp, 1987; 1988a; Rodgers and Janecke, 1992), but Neoproterozoic to Devonian recurrent uplift along the Lemhi arch may also contribute (Sloss, 1954; Scholten, 1957; Burchfiel et al., 1992; Ruppel, 1986; Skipp and Link, 1992).

Locally preserved beneath the Challis Volcanic Group are conglomeratic deposits (Smiley Creek conglomerate and equivalent units), many of which grade upsection into the volcanic rocks (Anderson, 1961; Ross, 1962; Staatz, 1979; Dover, 1983; Burton and Blakley, 1988; Janecke, 1992d; 1995b; VanDenburg, 1997; Anastasio and Schmitt, 1998, Blankenau, in press). Clasts typically reflect a local provenance, or western or northern source areas (Vandenburg, 1997; Blankenau in press).

Tertiary Extensional Deformation and Magmatism

The number of well-documented geometrically and temporally distinct episodes of extension in eastern Idaho and adjacent parts of Montana has grown in recent years from two (Wust; 1986; O'Neill and Pavlis, 1988), to three (Janecke et al. 1991; Janecke 1992), to four (Huerta and Rodgers, 1996; Tysdal 1996a; 1996b) or five (Sears and Fritz, 1998; VanDenburg et al., 1998). It is clear that the directions of extension in eastern Idaho and southwest Montana have changed dramatically during the Cenozoic. The initial, pre-volcanic episode of extension, the main post-volcanic, Paleogene basin-forming episode and the youngest Basin-and-Range episode of extension, extended the crust in a northeast-southwest direction, but intervening syn-volcanic and early

Miocene extension was in a northwest-southeast direction (McIntyre et al., 1982; O'Neill and Pavlis, 1988; Janecke, 1992a; 1995a; Fritz and Sears, 1993; Sears and Fritz, 1998; VanDenburg et al., 1998). Whereas gravitational collapse can readily account for northeast-southwest extension, the causes of northwest-southeast extension are less certain (c.f. VanDenburg et al., 1998).

Pre-Challis volcanic extensional to transtensional deformation was localized along southwest-dipping normal and normal-oblique faults, and affected the Pioneer Mountains, and the Lemhi and Beaverhead ranges. In the Pioneer Mountains, younger-on-older strike-slip deformation on a southwest-dipping low-angle fault predates emplacement of Challis-related dikes (Burton and Link, 1995; Rodgers et al., 1995; Huerta and Rodgers, 1996). In the southern Beaverhead Mountains, Skipp (1984, 1985) first interpreted the younger-on-older Divide Creek fault as a pre-Challis normal fault. She traced the normal fault 35-40 km to the northwest into the Hawley Creek area, and suggested a similar origin for normal faults in the Goat Mountain area of the Beaverhead Mountains, north of Leadore (Staatz, 1973, 1979) (Fig. 2). VanDenburg (1997; VanDenburg et al., 1998) confirmed that one of the normal faults north of Leadore predates the Challis Volcanic Group, and showed that the folded fault originally dipped gently to the southwest. Strikingly similar southwest-dipping normal faults in the northern Lemhi Range are also interpreted to predate the Eocene Challis Volcanic Group (Tysdal, 1996a, 1996b, Tysdal and Moye, 1996). Southwest-directed shearing in the southwest part of the Pioneer metamorphic core complex is kinematically like these pre-Challis structures, but isotopic analyses indicate a slightly younger age (49 to 45 Ma, syn-Challis)(Silverberg, 1990)

Trans-Challis Fault System

Extension on the Trans-Challis system was synchronous with Middle Eocene Challis volcanism (McIntyre et al., 1982; Kiilsgaard et al., 1986; Janecke, 1992a). Evidence for synchronous magmatism and deformation includes growth fault relationships along the Trans-Challis fault zone (Hammond, 1994; Janecke et al., 1997), intrusive rocks emplaced along northeast-striking normal faults (Hardyman and Fisher, 1985; Ekren, 1985), cross-cutting relationships between northeast-striking faults and volcanic rocks in the Lost River Range (Janecke, 1992a), and the consistent northeast strike of Eocene dikes in the region (Fisher et al., 1992; Nelson and Ross, 1968; Doughty and Sheriff, 1992). In the Lost River and Lemhi Ranges, the Trans-Challis system includes both high-angle and low-angle normal faults, with displacements of a few hundreds of meters to a few kilometers (Ross, 1947; Baldwin, 1951; Janecke, 1992a). Although the dominant structural trend is northeast during Challis volcanism, coeval northeast and north-northwest-striking normal faults have been documented in the White Knob Mountains (Snider, 1995) and the Panther Creek half graben (Janecke et al., 1997) near the end of Challis volcanism.

Some of the ductile deformation and uplift of the Pioneer core complex (Fig. 1), dates from this syn-Challis episode of extension (Wust, 1986; O'Neill and Pavlis, 1988; Silverberg, 1990) but the kinematics and timing of this event are controversial (compare Silverberg and Wust).

Challis Magmatic Episode

Beginning about 50 Ma lava flows, tuffs and intrusive rocks of the Challis Volcanic Group (Figs. 4 and 5) were deposited across central Idaho during an intense and voluminous volcanic episode that ended abruptly about 5 m.y. later (McIntyre et al., 1982; Janecke and Snee, 1993; Fisher et al., 1992; Snider, 1995; M'Gonigle and Dalrymple, 1996). The volcanic vents, dikes, mineralization, and normal faults during Challis magmatism were focused along the northeast-trending Trans Challis fault zone (Kiilsgaard et al., 1986). Eocene dacite porphyry and pink granite intrusive rocks are most voluminous along the eastern margin of the Idaho batholith (Lewis and Kiilsgaard, 1991; Stewart et al., 1992; Mahoney, 1992; Mahoney and Link, 1992; Schmidt, 1997)

Explosive volcanism was focused in large calderas northwest of Challis, Idaho (Fig. 5), whereas lava flows erupted from fissures and a few central vents scattered across central Idaho. A small caldera south of Salmon, Idaho (Ruppel et al., 1993; Blankenau, in press) and a volcanic-tectonic depression in the White Knob Mountains west of Mackay (Snider and Moye, 1989; Snider, 1995) are the only documented source areas of ash-flow tuffs outside the large Van Horn Peak cauldron complex north of Challis (Figs. 2, 4, 41 and 46)(McIntyre et al., 1982).

In the Lost River and Lemhi ranges, and the Pioneer Mountains, andesitic and dacitic lava flows dominate the volcanic stratigraphy and tuffs are mostly small-volume and derived from distant sources (Ross, 1947; Mapel and Shropshire, 1973; Janecke and Snee, 1993). Farther north in the Salmon River Mountains, and the White Knob Mountains, ash flow tuffs dominate the volcanic stratigraphy (McIntyre et al., 1982; Fisher et al., 1992; Ekren, 1988; Skipp, 1988b; 1989; Skipp et al., 1990; Skipp and Bollmann, 1992; Snider, 1995; Palmer, 1997; Palmer and Shawkey, 1997; Blankenau, in press).

By the end of the Challis volcanic episode, most of the locally significant prevolcanic topography had been buried beneath volcanic rocks in the central part of the field (Janecke and Snee, 1993). Only in those areas where large offsets occurred on synvolcanic normal faults, such as along the eastern margin of the Panther Creek half-graben (Fig. 4, 5) (Hammond, 1994; Janecke et al., 1997), were pre-Tertiary rocks exposed at the end of Challis volcanism.

Paleogene Basin-Forming Event

Paleogene sedimentary rocks deposited after the end of Challis volcanism are localized in north-northwest-trending half grabens (Fig. 2) (Janecke, 1994) and range in age from about 46 to less than 30 Ma (Harrison, 1985; Janecke and Snee, 1993; Axelrod, 1998, VanDenburg et al., 1998; Blankenau, in press). We refer to this phase of east-northeast to west-southwest extension as the "Paleogene basin-forming event". Faults of this third Tertiary episode of extension are more widely spaced than the older syn-Challis normal faults, and they accommodated much more slip. The Tertiary sedimentary rocks in the hanging walls of the west-southwest-dipping normal faults are coeval to slip along the faults (Janecke, 1992a, 1994; VanDenburg, 1997; Blankenau, in press). In east-central Idaho and southwest Montana, the presence of thin silicic tuffs beneath or interbedded in the lowest part of many

basin-fill sequences shows that extension in the Eocene-Oligocene rift zone began during the final phases of Challis volcanism (Janecke, 1994; Janecke et al., in press).

Faults known or inferred to date from this time include, from south to north (Fig. 2, 4), Arco Pass fault system, Pass Creek-Wet Creek fault system, Donkey fault, Sawmill Canyon fault, Allison Creek fault (Janecke, 1992a, 1993, 1994)(it was later called the Salmon River fault by Tysdal and Moye, 1996), Agency-Yearian fault, Lemhi Pass fault, and Salmon basin detachment fault (Janecke, 1994; 1995a; VanDenburg, 1997; VanDenburg et al., 1998; Janecke et al., 1998; Blankenau, in press). Each dips west or west-southwest and places east-tilted Tertiary sedimentary rocks in the hanging wall over Proterozoic to Paleozoic bedrock in the footwall. Slip estimates range up to ~ 11 km for portions of the Pass Creek and Donkey faults (Janecke, 1992a).

Evidence presented in Janecke (1994) suggests that the half grabens formed during the Paleogene basin-forming event were laterally restricted to a 100 ± 25 km wide corridor of extension. The western boundary of the rift is west of the Pioneer and Bitterroot core complexes. To the east, in southwest Montana, coeval dominantly fine-grained tuffaceous sedimentary rocks of the Renova Formation were being deposited in a broad basin (Fritz and Sears, 1993; Sears and Fritz, 1998). Janecke (1994, 1995a) hypothesized the presence of an uplifted rift shoulder in the Tendoy Range of southwest Montana between this broad Renova basin and the coeval half grabens of east-central Idaho. More recent work in the Salmon basin and the adjacent Horse Prairie basin of southwest Montana shows that the Paleogene basin-forming event was protracted and characterized by distinct sub-phases of deformation along west-southwest to south-southwest-dipping low-angle normal faults (VanDenburg, 1997; VanDenburg et al., 1998; Blankenau, in press).

The north-northwest trends of many of the known and inferred Paleogene half grabens parallel the regional north-northwest trending folds and thrusts north of the eastern Snake River Plain. Some normal faults in the eastern part of the rift zone probably sole into preexisting thrust faults (Constenius, 1982; Ponton, 1983; Janecke et al., in press). These characteristics are consistent with gravitational collapse models for the origin of the rift zone (Janecke, 1994).

Miocene Northeast-Striking Normal Faults

In southwest Montana, late Early Miocene normal faults with northeast strikes postdate the Paleogene basin-forming event and predate active Basin-and-Range extension (Fritz and Sears, 1993; VanDenburg et al., 1998; Sears and Fritz, 1998). To date, only one such fault has been identified in eastern Idaho (the Little Eightmile Creek fault north of Leadore; Staatz, 1979; VanDenburg et al., 1998) but others are likely.

Basin-and-Range Extension

Basin-and-Range faulting is the final episode of extension north of the eastern Snake River Plain. Its initiation is poorly dated and may range from about 17 to <5 Ma (Rodgers and Anders, 1990; Janecke, 1992a, 1993, 1994; Fritz and Sears, 1993; Anders et al., 1993; Sears and Fritz, 1998). This episode of extension produced the majestic Sawtooth, Lost River, Lemhi and

Beaverhead ranges in east-central Idaho, and spawned the 1983 Ms 7.3 Borah Peak earthquake along the Lost River fault (Crone et al., 1987; Crone and Haller, 1991). Basin-and-Range faults generally strike north-northwest to northwest and accommodated northeast-southwest extension (Stickney and Bartholomew, 1987), but locally the faults strike north and west (Scott et al., 1985; Janecke, 1993). Round Valley south of Challis and the Big Lost River Valley separate mostly northeast-dipping Basin-and-Range normal faults to the west from southwest-dipping normal faults to the east (Fig. 2).

Late Cenozoic extension of east-central Idaho along a northeast-southwest cross section through the center of the Lost River, Lemhi and Beaverhead faults is about 10 to 20% (Janecke, 1992a; Anders et al., 1993). Dip slip across the Beaverhead fault at Blue Dome has been estimated at 2.5 km (Rodgers and Anders, 1990). Heave (the horizontal component of slip) across segments of the Lost River fault ranges from 1.8 to 3.9 km (Janecke et al., 1991). Tilted late Cenozoic volcanic rocks suggest cumulative horizontal slip components across the center of the Lost River, Lemhi and Beaverhead faults between 6.5 and 10.3 km (Anders et al., 1993).

Active Basin-and-Range Faulting: Seismic Parabola

Active Basin-and-Range normal faults in eastern Idaho lie in the northern arm of a southwest-opening parabola centered on the Yellowstone volcanic plateau, which is characterized by a marked concentration of seismicity, a topographic rim, and active normal faults (Scott et al., 1985; Anders et al., 1989; Pierce and Morgan, 1992). The Basin-and-Range faults are divided into segments (Fig. 2) 11 to 43 km long (averaging 14 km) that rupture during individual large earthquakes (Crone and Haller, 1991). Boundaries between segments coincide with bends, bedrock ridges, preexisting faults, or relay ramps along the faults (Crone and Haller, 1991; Turko and Knuepfer, 1991; Janecke, 1993)(Fig. 2). Locations and or/existence of persistent segment boundaries along the southern Lost River and Lemhi faults may be reconsidered in light of recent trenching studies (Hemphill-Haley et al., 1994). The central segments of the normal faults in the parabola have been active more recently than adjacent segments, and normal faulting appears to young away from the eastern Snake River Plain (Anders et al., 1989; Crone and Haller, 1991; Pierce and Morgan, 1992). A drainage divide in the hanging walls of the Beaverhead, Lemhi and Lost River faults coincides roughly with the axis of the seismic parabola.

The Borah Peak earthquake in 1983 ruptured the Thousand Springs and part of the Warm Springs segment of the Lost River fault (Scott et al., 1985; Crone et al., 1987). The earthquake nucleated at the southern end of the Thousand Springs segment, in a segment boundary (Susong et al. 1990) and ruptured to the north-northwest. Aftershocks defined a planar southwest-dipping fault to a depth of 16 km (Richins et al., 1987). Geodetic data showed that the hanging wall subsided 1.2 m and the footwall was uplifted 0.2 m (Stein and Barrientos, 1985). The complex fault patterns documented in the segment boundary between the Thousand Springs and Mackay segments of the Lost River fault are typical of many segment boundaries (Susong et al., 1990; Machette et al., 1991; Janecke, 1993).

ROADLOG 1: INEEL TO ARCO TO CHALLIS: BIG LOST RIVER VALLEY

The simplified geologic map of Wilson and Skipp (1994) is an essential reference for this road log. Set odometer to zero at the junction between Highway 26 and Highway 20 near the Central Facilities Area of the Idaho National Engineering and Environmental Laboratory, (INEEL, formerly INEL and NRTS). This area is shown on the geologic maps of Scott (1982) and Kuntz et al. (1994).

Snake River Plain

In general the INEEL is located on a Pleistocene basalt lava plain (Hackett and Smith, 1992; Hughes et al., 1997; 1999; McCurry et al., 1999, this volume), locally pierced by Pleistocene rhyolite domes, and locally covered by a thin veneer of sediment derived from streams and deposited as channel, overbank, aeolian, and playa/lacustrine deposits (Geslin et al., 1999). To the west are Big Southern Butte and the route of the Goodale Cutoff of the Oregon Trail (Link and Phoenix, 1996). Big Southern Butte is a Pleistocene composite rhyolite dome that tilts Snake River Plain basalt flows northward (Spear, 1979, Spear and King, 1982; Fishel, 1993; Kuntz et al., 1994; Hackett and Smith, 1992).

The Big Lost River Rest Area is reached about seven miles north of the highway junction. In most years the Big Lost River here is dry, due to irrigation diversions. However it flowed during the mid-late 1990s, due to a wet climatic cycle. The Big Lost River flood (Rathbun, 1993; Cerling et al., 1994) scoured this area during the late Pleistocene, and produced scabland topography near Box Canyon (Fig. 6). The flood originated about 50 mi upstream along the East Fork of the Big Lost River, as a jökulhlaup (Rathbun, 1993).

Heading northward from the Big Lost River the road crests a lava rise and the view in all directions is grand. Folded Paleozoic rocks of the Arco Hills are to the north. The flat area to the north and east is the Big Lost Trough, a Pliocene and Pleistocene sedimentary basin on the north side of the Axial Volcanic Zone of the eastern Snake River Plain (Gianniny et al., 1997; Geslin et al., 1997; 1999). On a clear day one can see northeast to the west side of the Teton Mountains.

Figure 6. Photograph of Box Canyon, area of Big Lost River flood, looking northwest up the Big Lost River. The canyon is carved into basalts of the eastern Snake River Plain.

Butte City Area

Leave INEEL and reset mileage to zero at junction between Highway 26 to Arco and Highways 22 and 33 to east toward Howe and Rexburg. Stay on Highway 26 to the west. The Little Lost River roadlog (Roadlog #2) starts here and heads northeast to Howe and then north to the Pahsimeroi Valley. Butte City, population 59, is 3.9 miles west of the junction.

Near Butte City, in the immediate hills to the north are tight folds in thin-bedded Pennsylvanian and Permian Snaky Canyon Formation. The old Blackfoot to Mackay railroad grade is crossed just west of Butte City. Appendicitis Hill is in the middle distance to the northwest. West of it are folded upper Paleozoic sedimentary rocks in the southern Pioneer Mountains, with prominent Blizzard Mountain (site of a ski lift), where the Copper Basin thrust fault is exposed (Nilsen, 1977; Skipp, 1988b; 1989; Skipp et al., 1989, Skipp and Bollmann, 1992). Immediately to the south of Blizzard Mountain are the basaltic cinder cones of Craters of the Moon, which follow a northwest-trending rift system. The parallelism of rift zones on the eastern Snake River Plain and Basin-and-Range normal faults to the north shows that both areas are extending in a roughly northeast-southwest direction (Kuntz, 1992).

Arco Hills

The southern end of the Lost River Range and the Arco Hills are north of the road (Fig. 1). These consist of folded, but generally eastward-younging middle and upper Paleozoic strata, Eocene Challis volcanic rocks, Paleogene sedimentary rocks and Neogene ashflow tuffs cut by both east-northeast striking and west to west-northwest dipping normal faults (Fig. 7) (Kuntz et al., 1994; Janecke, 1995a). The type section of the Mississippian Arco Hills Formation (Skipp, Hoggan, et al., 1979) is at the southernmost end of the range.

The east-tilted Arco Pass half-graben (see map Fig. 7) occupies the low pass between the Arco Hills to the east and the south end of the Lost River Range to the west (Janecke, 1992e; 1992f; 1995a). It is part of the north-trending middle Eocene to early Miocene extensional system (the Paleogene basin-forming event) (Janecke, 1994).

Arco

The intersection with Highway 20 and 26 in Arco is at mile 7.7. Mississippian Scott Peak Formation forms vertical cliffs northeast of town painted with the graduation year of all the high school classes of the 20th century. A measured section of the upper part of this sequence is described by Skipp, Hoggan, et al. (1979). The field trip of Link et al. (1988) describes the geology from here west to Sun Valley.

Arco is near the southern end of the Lost River Range. The active Lost River fault bounds the west-southwest side of the range and extends about 90 mi northward to Challis. Six segments along the Lost River fault are identified (Scott et al., 1985; Crone and Haller, 1991; Janecke, 1993). Locations of segment boundaries and names of segments are shown on Figure 2. The central segments of the Lost River fault, like those of the adjacent Lemhi and Beaverhead faults, were more recently active (Pierce and
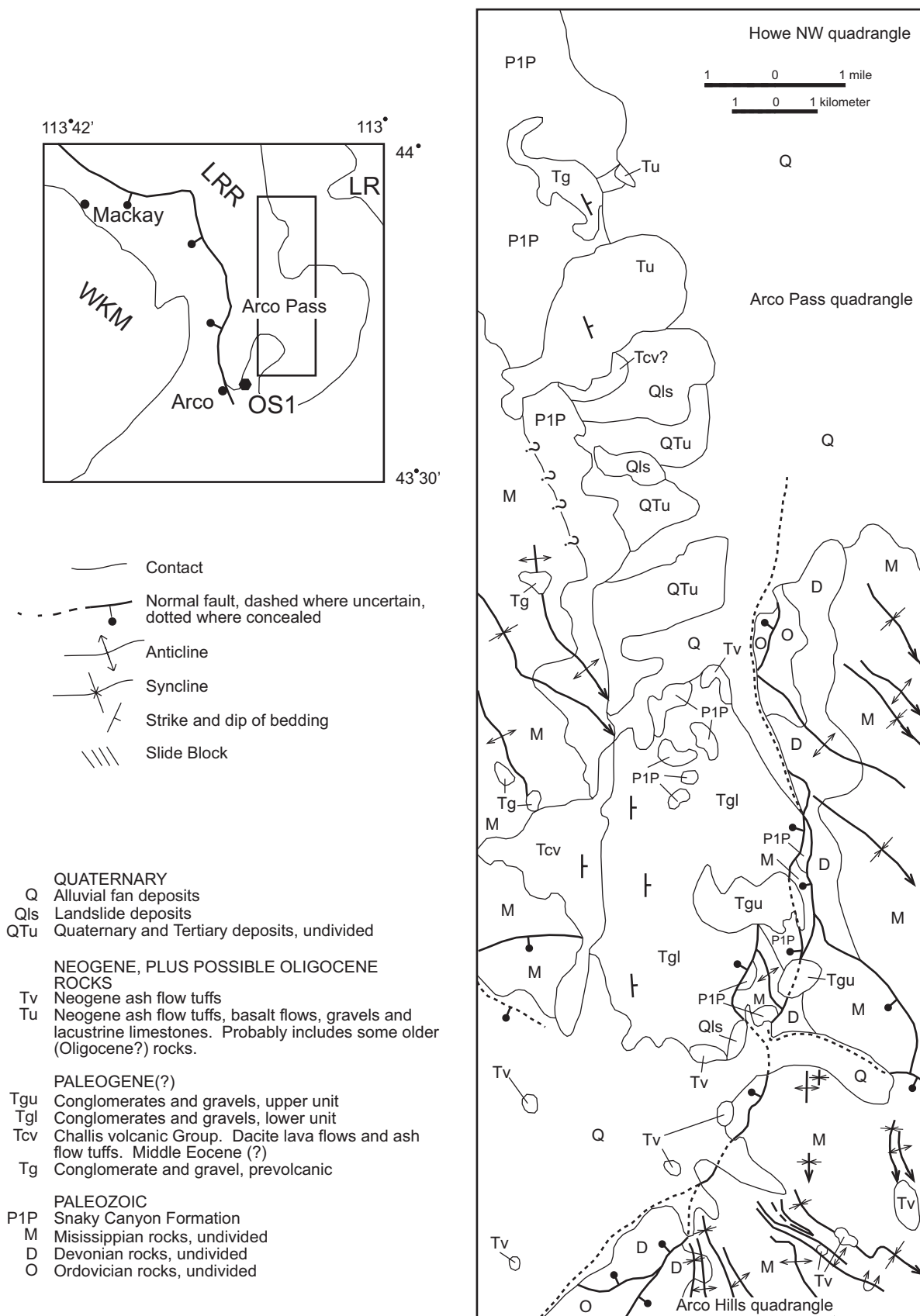
Figure 7. *Simplified geologic map of the Arco Pass, and parts of the Howe NW and Arco Hills 7.5' quadrangles, Idaho. Sources of data: Janecke, 1992c and unpublished mapping; B. Skipp, unpublished mapping in extreme northern and southern part of area. Some northeast-striking faults in the west-central part of the map are not shown. (Reproduced from Janecke, 1995a, Figure 5).*

EXPLANATION

- fault scarp
- low-angle normal fault
- normal fault
- thrust fault
- segment boundary
- Tuff of Challis Creek
- Dike

Morgan, 1992). The southernmost, Arco segment produced more than seven surface faulting earthquakes between 140,000 and 20,000 years ago and shows strong temporal clustering of events (Olig et al., 1995). The Pass Creek segment, to the north, had a cluster of activity around 17 to 18 ka (Olig et al., 1995). Janecke et al. (1991) document 4.5 to 6 km of dip separation across this part of the Lost River fault.

Figure 8 is a simplified geologic map of the central and southern Lost River Range, whose western side we will follow all the way to Challis. The range contains many north northwest-trending folds cut by a complex three-dimensional array of normal faults. The King Mountain Hang glider launch site turnoff is at mile 15.5. At mile 16.9 is a nice view of an anticline-syncline pair in Carboniferous limestones of the southern Lost River Range to the east. Moore (formerly Lost River) is reached at mile 18.8.

The usually-dry Big Lost River is crossed at mile 20.3. Cottonwood trees along the river died in the late 1980s due to dry years and groundwater withdrawal for irrigation. Water-related resentment remains rife in the valley.

Antelope Creek and Pioneer Mountains

The Antelope Creek Road is on the left at mile 18.9. South of Antelope Creek is Appendicitis Hill, which contains folded but generally eastward-younging Mississippian, Pennsylvanian and Permian limestones overlain by east-dipping lava flows of the Eocene Challis Volcanic Group (Skipp et al., 1979; Skipp et al., 1979; Skipp and Hall, 1980; Link et al., 1988; Snider, 1995). Skipp and Hait (1977) inferred a thrust fault between Appendicitis Hill and the Lost River Range but Rodgers and Janecke (1992) showed that a thrust in this position would have less offset than the thickness of the Mississippian rocks. Cirque exposures in the high Pioneer Mountains, above Iron Bog Campground, 18 miles to the west up the Antelope Creek road, expose conglomeratic turbidites of the Mississippian Copper Basin Group (Paull et al., 1972; Nilsen, 1977; Wilson et al., 1994; Link et al., 1996). These rocks were deposited in a fault-bounded, rapidly subsiding basin east of the Antler highland.

Lost River Range

At mile 24.1 the road crosses the Big Lost River. To the east is the Lost River Range. Figures 9 to 14 are photographs and cross sections in this area.

The Butte County line is crossed at mile 21.6 as the road enters the village of Darlington. King Mountain is the high point to

the east, with Beaverland Pass south of it, and the southern end of the Lost River Range farther southeast.

Leslie and Pass Creek

The Pass Creek road is on the right, just north of the village of Leslie, and 26.8 miles south of Mackay. A west-dipping low-angle normal fault cuts the range east of Pass Creek (Fig. 2, 8) (Janecke, 1992a; 1992d; 1995a). Challis lava flows were erupted from a vent complex west of Pass Creek (Janecke and Snee, 1993). Northwest of there at the mouth of Crows Nest Canyon is an enormous debris flow cut by the Pass Creek segment of the Lost River fault. Concentric pressure ridges are developed in the toe of this landslide. About mile 30.5, southeast of Mackay, prominent dip slopes in Mississippian Scott Peak Formation (Fig. 3) can be seen in Mahogany Gulch to the north (Janecke, 1992e).

The Leslie Hills to the south of the road contain folded Carboniferous limestones (Davis, 1983). Farther to the south are the hills north of Antelope Creek, including Sheep Mountain. This was a Challis eruptive center in middle Eocene time (Moye et al., 1988; Snider, 1995).

Mackay and White Knob Mountains

Reset odometer to zero at the main intersection in downtown Mackay, population 547. To the northeast the prominent cliffed summit is Mt. McCaleb (Fig. 9). The peak is held up by subhorizontal and cliff-forming limestones of the Mississippian Scott Peak Formation (Janecke, 1992d).

To the southwest in the White Knob Mountains is the Empire Copper mine, whose copper skarn deposits, related to the Eocene White Knob granite, were productive from 1907 into the 1970s (Nelson and Ross, 1969; Wilson et al., 1995). The original townsite here was Houston, located south of Mackay on the Big Lost River at the mouth of Alder Creek.

Mackay is near the northeast end of a northeast-trending horst of Paleozoic rocks in the White Knob Mountains (Fig. 2). The Eocene White Knob intrusive suite and northeast-trending dike swarms occupy the central axis of the horst and invade the Mississippian McGowan Creek Formation and White Knob Limestone (Nelson and Ross, 1968; 1969; Snider, 1995). Eocene Challis volcanic rocks which underlie most of the lower foothills in the hanging wall of the horst south of the Empire Mine, were deposited across an irregular erosion surface cut into folded upper Paleozoic carbonate rocks (Snider, 1995).

At mile 7.5 is the north end of Mackay Reservoir on the left. The dam is anchored in White Knob Limestone (Nelson and Ross, 1969). East-dipping tuffs in the Challis Volcanic Group crop out

Figure 8. Simplified geologic map of the central part of the Lost River Range and adjoining areas. Hingelines of the numerous folds in the area are omitted for clarity. LPF = Leatherman Pass fault, HCF = Hell Canyon fault. The large black dots are sample locations for fault slip data in Janecke (1992a). Strike and dip symbols show the attitude of Tertiary rocks in the area. Figure reproduced from Janecke (1995a, Figure 3). Map units are as follows:

Post-Eocene rocks: *Q=Quaternary deposits; QT=Quaternary and Tertiary deposits; Qtb=Late Cenozoic cinder cone; Ts=Paleogene sedimentary rocks.*

Challis Volcanic Group-Eocene: *Tt=Tuffaceous rocks; Tc=Andesitic lava flows; Td=Dacitic lava flows and tuffs; Tdc=Lava flows.*

Pre-Tertiary rocks: *lPM=Mississippian and Pennsylvanian rocks; lP=Snaky Canyon Formation; M=Three Forks, McGowan Creek, Middle Canyon, Scott Peak, South Creek, Surrett Canyon, Arco Hills and Bluebird Mountain formations; D=Jefferson Formation; DO=Ordovician and Devonian rocks; DY=Devonian to Proterozoic rocks; SO=Laketown and Fish Haven formations; O=Kinnikinic and Summerhouse formations; OY=Ordovician to Proterozoic rocks; C=Tyler Peak and Wilbert Formations; Y=Swauger Formation; YZ=Proterozoic Rocks.*

Figure 9. Aerial view looking east at the Lost River Range from north of Mackay. Mt. McCaleb is the high peak. Low hills in front center are Pennsylvanian and Permian Snaky Canyon Formation. Cross section thorough here shows 4.5 to 4.9 km of dip separation and 3.5 to 3.8 km of heave (Janecke et al., 1991). Trees are on Ordovician to Devonian rocks. Smooth slopes above treeline are underlain by the Mississippian McGowan Creek Formation, grading upsection into the Middle Canyon Formation. Multiple generations of alluvial fans are cut by strands of the Lost River fault in this area (Janecke, 1992e). The Mt. McCaleb normal fault dips northwest and projects though the lowest saddle on the right shoulder (south) of Mt. McCaleb (Janecke, 1992e).

Figure 10. Aerial view looking northeast toward the trace of Leatherman Pass fault which bounds the east side of the Borah Peak horst. Leatherman Peak, underlain by Mississippian limestone, is the pointed peak just right of center, and Leatherman Pass is immediately to the left (west) of it. The Mackay segment of the Lost River fault forms the mountain-front scarp to the right of this view. The segment boundary to the Thousand Springs segment to the north is near the left edge of the photograph. Susong et al. (1990) discuss the complex array of normal faults in the segment boundary zone. Wilson et al., (1990) detail the geology of the lower part of the mountain front. Janecke and Wilson (1992) detail the geology of the whole area.

west of the reservoir (Snider, 1995).

At mile 10.1 is the turnoff to the west that leads to the Fish Hatchery and the Burma Road, which affords a spectacular drive over the White Knob Mountains to Copper Basin, with views to the east of the central Lost River Range (Figs. 9 and 10). This road climbs through a Challis volcanic center that includes lava flows, tuffs, and intrusive domes (Moye et al., 1988; Snider, 1995).

Leatherman Peak

At mile 14.5 is a view to the northeast of Leatherman Peak (Fig. 10), a sharp horn carved from the Mississippian Scott Peak Formation (Ross, 1947; Janecke and Wilson, 1992). The Leatherman Pass normal fault, a major east-dipping normal fault of syn- to pre-Challis age, places Mississippian rocks on Ordovician rocks at Leatherman Pass west of the peak. The prominent white patches in the limestones on Leatherman Peak are baked zones around Eocene dikes. Flourite veins are associated with the dikes. Other mineral deposits in the Lost River Range include stratabound barite in the Devonian Jefferson Formation (Wilson et al., 1990). To the north is the segment boundary between the Mackay and Thousand Springs segments of the Lost River fault (Susong et al., 1990; Bruhn et al., 1991; Janecke, 1993). Figure 11 is a cross section of the range north of here.

East Fork, Big Lost River

The Trail Creek road, on the left at 16.3 miles follows the East Fork of the Big Lost River over the Pioneer Mountains and down to Sun Valley on their western side. The primary source for alluvium in half-graben country like this is the hanging wall of the master normal fault (Leeder and Gawthorpe, 1987, 1993). The extensive Big Lost River drainage system to the west provides much more sediment to this valley than the alluvial fans and short, ephemeral streams draining the steep west front of the Lost River Range.

Borah Peak

The road to the Birch Springs trailhead, that provides access to Borah Peak, is on the right at mile 20.9. The steep west front of the Lost River Range is bordered by the Thousand Springs segment of the Lost River fault, that last moved in October, 1983 (Fig. 12). The Borah Peak earthquake prompted many geologic and geophysical studies (Stein and Barrientos, 1985; Crone et al., 1987; Richins et al., 1987; Susong et al., 1990; Bruhn et al., 1991; Turko and Knuepfer, 1991; Crone and Haller, 1991). The earthquake produced scarps up to 5m high (most were 2 m) and accomodated both left and normal slip (Crone, 1988).

The Borah Peak horst, folded and internally thrusted (Fisher and Anastasio, 1994), contains generally northeast-younging Neoproterozoic to Mississippian rocks (Ross, 1947; Skipp and Harding, 1985; Janecke and Wilson, 1992). This Eocene horst is uplifted on the east-dipping Leatherman Pass normal fault on the east side, and on the northwest-dipping Mahogany Gulch fault on the northwest side. The northwest-dipping Elkhorn Creek fault further extends the horst block. Borah Peak itself is on the east limb of a map-scale syncline and in the hanging wall of a small thrust fault (Figs. 8, 11 and 13).

The Doublespring Pass road is on the right at mile 23.0. This intersection is the site of the (former) town of Chilly. To the north over Doublespring Pass is the route to the east side of the Lost River Range, the Upper Pahsimeroi River, and the towns of Patterson and May. The road crosses several folds in upper Paleozoic rocks of the Lost River Range (Fig. 13). Fisher and
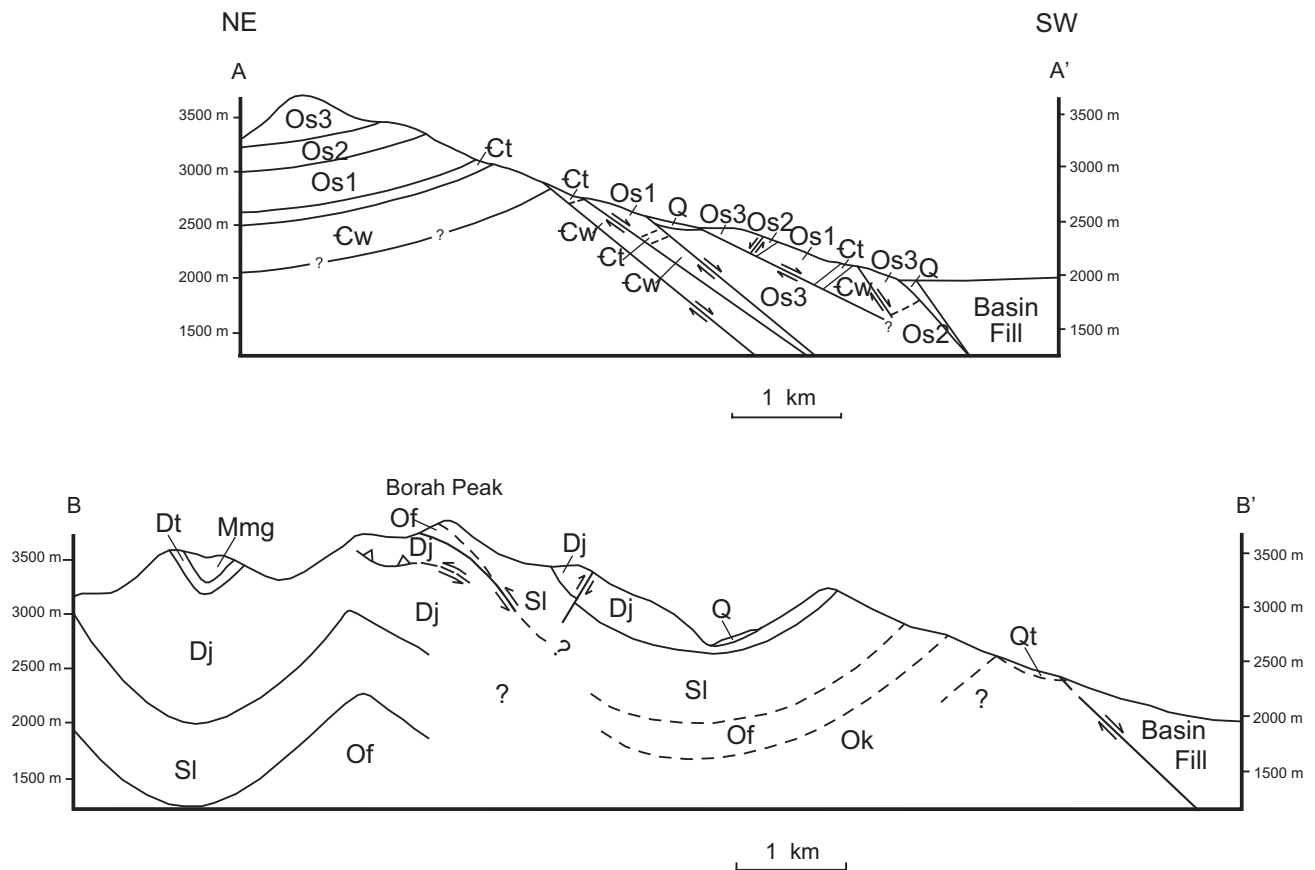
Figure 11. Cross sections through the Borah Peak horst. Compare the complex geometry of the Lost River fault in the segment boundary between the Mackay and Thousand Springs segments (A-A') with the single fault strand along the southernmost part of the Thousand Springs segment (B-B'). Reproduced from Susong et al. (1990, Figure 3). See Figure 8 for locations of cross sections.

Anastasio (1994) and Anastasio et al. (1997) performed detailed strain analyses of these folds.

Pioneer Mountains

Coming up out of the Lost River Valley at mile 28.2 the view to the southwest is of the high peaks of the Pioneer Mountains. The Devil's Bedstead is the prominent peak on the north side of the metamorphic core of the range, and Hyndman Peak is the prominent peak in the central part of the range. The Pioneer Mountains contain a metamorphic core complex which contains Paleoproterozoic gneiss, metamorphosed Paleozoic sedimentary rocks, and foliated Eocene intrusive rocks in the footwall of the Wildhorse detachment fault (Dover, 1981; 1983; Wust, 1986; O'Neill and Pavlis, 1988; Wust and Link, 1988; Silverberg, 1990). The upper plate of the detachment contains folded and thrust faulted Paleozoic sedimentary rocks and Eocene Challis volcanic rocks (Worl et al., 1991; Link et al., 1995; 1996; Rodgers et al., 1995). The Mississippian Copper Basin Group forms many of the slopes on the south and east sides of the metamorphic core. Mixed fine-grained sandstone and carbonate strata of the Pennsylvanian and Permian Sun Valley Group makes up the summits north of the complex. The lower eastern and northern part of the range and the Boulder Mountains to the north are mantled by Challis Volcanic Group. To the south and east are the White Knob Mountains described above (Nelson and Ross, 1968; Wilson et al., 1995).

Willow Creek Summit

Willow Creek Summit, elev. 7160, is reached at mile 30.4. This is a bedrock ridge in the hanging wall of the Lost River fault. It corresponds to a persistent segment boundary between the Thousand Springs and Warm Springs segments (Fig. 2). The 1983 Borah Peak earthquake ruptured through the segment boundary into the southern end of the Warm Springs segment (Crone et al., 1987). The geology of this area is shown on the Challis 1° x 2° sheet (Fisher et al., 1992) and at larger scale on the Bayhorse area map of Hobbs et al. (1991) and the Challis 1° x 2° quadrangle of McIntyre and Hobbs (1987).

Figure 14 is the view to the south toward Borah Peak and the syncline to the west (Figs. 11 and 14) (Ross, 1947; Janecke and Wilson, 1992). The view to the north is of talus-strewn bare slopes of the Pahsimeroi Mountains (Lost River Range north of Doublespring Pass). These contain a large north-northwest trending syncline with Mississippian to Permian rocks in the core and Ordovician to Devonian rocks along the flanks (Mapel et al., 1965; Rodgers and Janecke, 1992). The syncline may reflect a footwall ramp in an underlying thrust fault (Rodgers and Janecke, 1992). Many closely spaced Sevier-aged folds deform the core of this regional-scale syncline (Fig. 13).

Figure 12. View of the Thousand Springs segment of Lost River fault scarp, looking southeast toward Cedar Creek (large canyon on right side of view). Rock Creek is the narrow canyon on the left side of view.
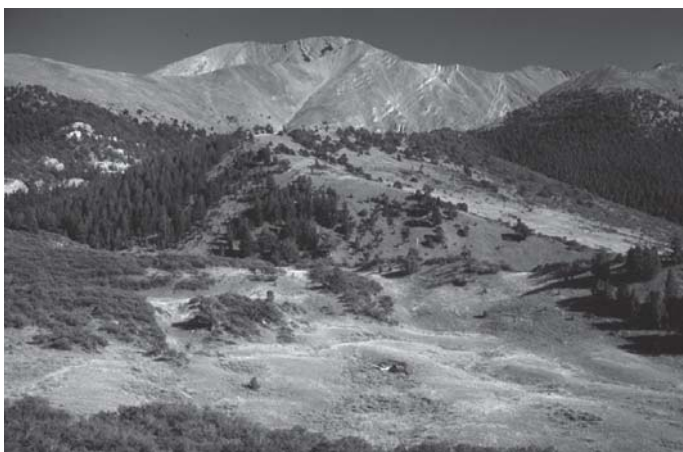

Figure 13. Syncline (at highest peak) and anticline (at right edge) in Mississippian and Pennsylvanian limestones, view to the southeast from road, just northeast of Doublespring Pass. Ribbed outcrops on skyline belong to the Mississippian, Pennsylvanian and Permian Snaky Canyon Formation. Below it are talus-covered outcrops of the Mississippian Bluebird Mountain Formation and massive limestone cliffs of the Surrett Canyon Formation (canyon on left side of view) (Janecke and Wilson, 1992). Glacial deposits mantle the foreground (Cluer, 1989). Fisher and Anastasio (1994) and Anastasio et al. (1997) studied folds in this area.

Antelope Flat, Grandview Canyon

Heading down the north side of Willow Creek Summit, at mile 34.8 the Lost River fault cuts alluvial fans to the east. The highway joins the main canyon of Dry Gulch and Broken Wagon Creek to the southwest. Outcrops on the west of the road at mile 33.9 are Mississippian White Knob Limestone (Hobbs et al., 1991; Link et al., 1996).

To the west, at mile 37.0, is the Spar Canyon road. This affords access to extensive exposures of the Bayhorse assemblage of Paleozoic carbonate rocks in the Bradshaw Basin and Spar Canyon areas (Hays et al., 1980; McFaddan et al., 1988; Hobbs et al., 1991; Grader and Dehler, 1999, this volume). The lacustrine and fluvial deposition of Eocene Challis Group synvolcanic sedimentary rocks along the Spar Canyon road have been described by Palmer (1997) and Palmer and Shawkey (1997).

The road to McGowan Creek is on the east at mile 38.9. This road affords access to the tightly folded Mississippian rocks of

Figure 14. View southeast from Willow Creek Summit of Borah Peak (high peak on left skyline). The two bands on west end of flat ridge to the west (right) of Borah Peak are sandstones at the base of the Jefferson Formation. The underlying Silurian Laketown Dolomite is light in color. At Borah Peak light-colored Ordovician and Silurian rocks are thrust over darker Devonian Jefferson Formation (Janecke and Wilson, 1992) (Fig. 11). A northwest-trending syncline coincides roughly with the lowest pass on the skyline.

Figure 15. Grandview Canyon is cut in the Grandview Dolomite Member of Jefferson Formation. The view is to the north along Highway 93.

the Pahsimeroi Mountains (Mapel et al., 1965). The type section of the McGowan Creek Formation, Lower Mississippian distal flysch formerly mapped as Milligen Formation (Ross, 1947), is up this road (Sandberg et al., 1967; Sandberg, 1975).

At mile 38.6 the highway enters superposed Grandview Canyon (Fig. 15), cut in the Grandview Member of the Devonian Jefferson Dolomite. The road to a bioherm/biostrome complex in the dark dolomite member of the Jefferson Dolomite is on the left at mile 42.2 (Isaacson et al., 1988; Isaacson and Dorobek, 1988; Grader and Dehler, 1999, this volume). This Frasnian buildup is dominated by corals, and overlies a thick sequence of slope to deep ramp deposits. It has been proposed to represent, among other things, shallowing related to an early phase of Antler tectonism (Dorobek et al., 1991).

Lone Pine Peak

Past Grandview Canyon the highway proceeds northward down Warm Springs Creek. The topography is controlled by northnorthwest striking basin and range faults. Lone Pine Peak ridge to the west is on the footwall of an east-dipping normal fault, antithetic to the main Lost River fault. On the Lone Pine Peak range, Silurian to Mississippian strata strike northwestward, parallel to the range, and dip northeast and southwest. Folding and thrusting complicate the geology (Hays et al., 1978; 1980; Hobbs et al., 1991; Fisher et al., 1992).

Figure 16. Howe Peak and ribbed outcrops of Pennsylvanian and Permian Snaky Canyon Formation, view to north from Highway 22 southwest of Howe. Beds at left and right edge dip west, whereas the central spur contains some east-dipping beds.

Figure 17. View westward from just north of Howe toward folded Snaky Canyon Formation in the Howe Hills in the foreground and the Lost River Range in the background. King Mountain is the highest peak in this part of the Lost River Range.

Round Valley and Challis Area

The Challis Hot Springs turnoff is at on the right at mile 51.0. The road to the east over Grouse Peak to the Pahsimeroi Valley begins here. The Neoproterozoic breccias described by Carr and Link (1999, this volume) are in the canyon of Leaton Gulch, below Grouse Peak. The view to the west is of the Salmon River Canyon (see Figs. 45 and 47). To the east is the north end of the Lost River Range; to the west and south is the Paleozoic carbonate ridge of Lone Pine Peak. Challis Volcanic Group lavas form the low hills east of road below Grouse Peak (McIntyre and Hobbs, 1987). The northern and east-central parts of the Lost River Range contain thick sections of Challis volcanic rocks.

Roadlog #1 ends at the intersection of Highways 93 and 75 just south of Challis, near the Yankee Fork Mining Museum. Roadlog #4 along the Salmon River joins here.

ROADLOG #2: GEOLOGIC GUIDE UP THE LITTLE LOST RIVER VALLEY.

This road log begins at the northern border of the INEEL, east of Butte City, at the intersection of U.S. Highway 20/26 with Highway 22/33. Take Idaho Highway 22/33 to the northeast to Howe.

Arco Hills

The Arco Hills, northwest of the road, are a roughly north-northeast trending fault block in the hanging wall of the Lemhi fault and the footwall of the Arco Pass fault (Fig. 2). Visible on their southern front are an anticline-syncline-anticline fold train in Mississippian McGowan Creek and carbonate bank strata (Figs. 2 and 7) (see maps of Skipp and Hait, 1977; Wilson and Skipp, 1994 and Kuntz et al., 1994). McQuarrie and Rodgers (1998) argue that the south-southeast plunge of the folds in this area and elsewhere adjacent to the eastern Snake River Plain is due to flexure toward the plain. Many folds distant from the eastern Snake River Plain also plunge to the southeast (Rodgers et al., 1995; Jeppson and Janecke, 1995), making it difficult to uniqely determine the origin of the southeast plunges.

The large canyon on the left is Deadman Canyon, with Pennsylvanian and Permian Snaky Canyon strata in its headwaters. There is a west-dipping normal fault on the east side of the drainage. Snaky Canyon Formation rocks are in the hanging wall of

this fault. The Mississippian carbonate bank sequence is brought up in the footwall. Upsection in the footwall are spectacular north northwest-trending folds in Snaky Canyon strata under Howe Peak (5 miles from intersection) (Fig. 16). The highway goes northeast for 7 miles from the intersection and then bends to the north, parallel to the strike of the Snaky Canyon Formation (Cantor and Isaacson, 1991; Kuntz et al. 1994).

Howe

Howe is reached at 15 miles. To the west are the Snaky Canyon beds, that contain *Palaeoaplysina* reefs studied by Breuninger et al. (1988) (Figs. 16 and 17). To the east is the southern end of the Lemhi Range, which also contains a folded and thrusted eastward-younging section of Mesoproterozoic to Permian rocks locally overlain by Tertiary Challis Volcanic Group, younger sedimentary rocks, and Neogene volcanic rocks (Hait, 1987; Kuntz et al., 1994). Strata under the pre-Tertiary unconformity become younger to the south in the Lemhi Range (Rodgers and Janecke, 1992). The prominent peak is Saddle Mountain (Fig. 18). Howe Point, at the south end of the Lemhi Range, exposes several of the rhyolite ash-flow tuffs produced by the Snake River Plain-Yellowstone hotspot (Morgan, 1992; Pierce and Morgan, 1992). These are cut by older east-west striking normal faults and the younger Lemhi fault (Kuntz et al., 1994).

Little Lost River Valley

Reset mileage to zero at Howe. Go north on Little Lost River Road. To the west is the Lost River Range, with many tight folds in Pennsylvanian and Permian Snaky Canyon Formation (Fig. 17). The southern Lost River Range contains a folded but generally eastward-younging succession of Ordovician to Permian carbonate rocks overlain by Eocene volcanic rocks, Paleogene sedimentary rocks and Neogene tuffs and basalt flows of the eastern Snake River Plain (Skipp and Hait, 1977) (Fig. 18). At least three generations of Eocene to Recent normal faults cut the range (Janecke, 1992a; 1992c; 1995a). To the west and north proceeding clockwise in the Lost River Range are the Arco Hills, King Mountain, and Hawley Mountain (Fig. 1).

Lemhi Range

To the east is the steep faulted front of the Lemhi Range (Fig. 18). The South Canyon bedrock block juts out into the Little Lost River Valley about ten miles north of Howe. The block contains east-dipping Neoproterozoic to Upper Mississippian strata (Wil-

Figure 18. Aerial view of south end of Lemhi Range. Saddle Mountain is high peak. Geology of this area is described by Hait (1987), and includes strata from Mesoproterozoic to Permian age, cut by the Howe segment of the Lemhi Fault. The South Creek "slide block" of Beutner (1972) is barely visible at the extreme left edge of the photograph.

son and Skipp, 1994). The type section of the Neoproterozoic and Cambrian Wilbert Formation is reached up South Creek Canyon, south and east of the block (Ruppel et al., 1985; McCandless, 1982).

Beutner (1968, 1972) interpreted this block as a landslide that slid westward from the Lemhi Range to the east. A second similar block is present about 15 miles northward along the front of the Lemhi Range, west of Diamond Peak (Ross, 1961). Alternatively, the blocks may be underlain by low-angle normal faults that persist beneath the Little Lost River valley rather than being landslide blocks whose lower bounding surface ultimately intersects the earth's surface at both the up- and down-dip ends.

The southern Lemhi Range contains a generally eastward-younging package with Mesoproterozoic Lemhi Group at the base, on the west side of the range, and Permian rocks on the northeast side (Fig. 3) (Ross, 1961; Beutner, 1968; 1972; Skipp and Hait, 1977; Hait, 1987). The section contains local thrust faults and folds, but is basically a coherent block within the Hawley Creek thrust plate (Rodgers and Janecke, 1992). West-vergent normal-fault blocks formed during tectonic denudation of the Lemhi Range in both Paleogene and Late Miocene to Recent time (Janecke, 1992a).

To the northeast in the Lemhi Range one can see Saddle Mountain and the Wilbert Mine, which was a major silver and lead producer from 1906 to 1931 (Ross, 1961). The ore bodies are sulfide replacements in quartzitic dolomite of the Mesoproterozoic Swauger Formation, likely formed during Eocene Challis magmatic activity (Ruppel and Lopez, 1988).

Paleogene Half-Graben

A huge alluvial fan drains north from Hurst Canyon to the left, out of the low pass between the Arco Hills and the Lost River Range. The drainage follows a west-dipping middle Eocene to Oligocene fault which defines a half-graben through Arco Pass (Fig. 7). Sawmill Canyon in the Lemhi Range about 40 miles north of Howe and the Pass Creek-Wet Creek half graben in the Lost River Range to the west are dismembered parts of a coeval middle Eocene to Oligocene half-graben system (Fig. 8). (Janecke and Snee, 1993; Janecke, 1992a, 1992d; 1994; 1995a). The Donkey fault system, which cuts the Donkey Hills (Janecke, 1992b), also formed during this extensional phase.

Diamond Peak

At 25 miles from Howe the road crosses the Little Lost River. Diamond Peak can be seen in the Lemhi Range to the east (Fig. 25 is a view from the other, east, side of the range), and Hawley Mountain can be seen to the northwest. The rocks west of Diamond Peak are a landslide (?) block of Mississippian carbonates, dropped down to the west against Mesoproterozoic Swauger Formation in the footwall (Beutner, 1972). Earlier studies suggested that this block, and the South Canyon block, to the south, coincided with segment boundaries along the Lemhi fault (Crone and Haller, 1991) but more recent analysis shows that they may not be persistent barriers to rupture (Hemphill-Haley et al., 1991, 1994).

Wet Creek-Hawley Mountain

The Wet Creek Road is reached on the left at mile 34. This is the north end of the Pass Creek-Wet Creek half graben (Fig. 5, 8) documented by Janecke (1992a; 1992b; 1994). There is a spectacular view to the northwest of the high Lost River Range. The road turns to gravel.

Hawley Mountain, the prominent peak to the southwest in the Lost River Range, exposes a complete southeast-dipping stratigraphic section of Mesoproterozoic to Pennsylvanian and Permian rocks (Mapel and Shropshire, 1972). Hawley Mountain is near the southeast end of a small horst block (Fig. 2 and 8). The northeast-dipping Barney fault bounds the horst on the northeast side, and the Hawley Creek fault is on the southwest side. (Figs. 2 and 8)(Janecke, 1993). The Barney fault is antithetic to the Lemhi fault. The Custer County line is crossed at mile 38.

Sawmill Canyon

The Sawmill Canyon area in the Lemhi Range is directly north (Fig. 19). At mile 40 there is a major road junction to Sawmill Canyon on the right. Sawmill Canyon coincides with a prominent north-trending reentrant in the Lemhi Range, and with a major bend in the range-front fault (Fig. 2). Similar sharp bends also occur in the Lost River and Beaverhead faults, and may have a similar origin (Janecke, 1993). At Sawmill Canyon the bend appears to coincide with a segment boundary along the Lemhi fault (Turko and Knuepfer, 1991; Janecke, 1993). A side trip up Sawmill Canyon affords a view of the Lemhi fault scarp cutting alluvial fans along the range front (Fig. 19). The Sawmill Canyon fault, a west-dipping middle Eocene to Oligocene(?) normal fault controls the trend of Sawmill Canyon (Janecke, 1992a, 1992c). The geologic map of the Sawmill Canyon area of Janecke (1992c) documents that Challis lava flows and ash flow tuffs of the Sawmill Canyon half graben are overlain and intercalated with syntectonic conglomerates and gravels (Janecke and Snee, 1993; Janecke, 1995a). Unroofing sequences of clasts in these sedimentary rocks show that they were shed from the footwall of the Sawmill Canyon fault (Janecke, 1994).

Pre-Tertiary strata also dip east in the central Lemhi Range, with Mesoproterozoic rocks making the high peaks west of Sawmill Canyon and Ordovician to Devonian strata making up the crest of the range east of Sawmill Canyon (Ross; 1947; Ruppel and Lopez, 1981). Eocene porphyritic intrusive rocks intrude the Paleozoic limestones. Tertiary rocks overlie a southeast-dipping

Figure 19. View south of the mouth of Sawmill Creek. Scarps along the Sawmill Gulch-East segment of the Lemhi Fault cut alluvial fan surfaces in from the center of the photograph to the farthest right extent of shadow in the upper left. A prominent terrace scarp in central distance is a fluvial scarp. Bedrock at the right edge of photograph is the Ordovician Kinnikinic Quartzite cut by a northeast-striking normal fault that dips northwest, parallel with shadow line near middle right edge. It places Tertiary sedimentary rocks down against the Kinnikinic Quartzite (Janecke, 1992c) (See Fig. 8 for more details).

lateral ramp in the footwall of an underlying thrust fault (Rodgers and Janecke, 1992). Challis strata lie on Pennsylvanian and Permian Snaky Canyon Formation in the southern Lost River and Lemhi Ranges, but overlie uplifted Mesoproterozoic Lemhi Group north-northwest of Sawmill Canyon.

Summit Creek-Goldburg Area

Two miles north of the turnoff to Sawmill Canyon (mile 40) the road crosses a prominent gravel terrace (Unit Qtf1 of Janecke, 1992b) dotted with small mounds and circles of lush vegetation around a barren center. The origin of these features ("mima mounds") is uncertain but their density and degree of development appears to measure the antiquity of Holocene and Quaternary surfaces. Some coincide with large ant colonies.

At mile 45 is the Summit Creek recreation site. Fault scarps developed in alluvial fan deposits northeast and southwest of the road reflect recent surface-faulting earthquakes along the Lemhi fault and the antithetic Barney fault (Janecke, 1993; Janecke 1992c and 1992d). The pattern of faulting is quite complex in the narrow bedrock ridge two miles northwest of here, and probably reflects the presence of a segment boundary along the Lemhi fault (Janecke, 1993).

A large old alluvial fan comes from the Donkey Hills on the left, draining the footwall of the Donkey fault (Figs. 5 and 8). Outwash from Late Pleistocene glaciers along Dry Creek likely contributed to this fan (Mapel and Shropshire, 1973; Janecke, 1992d).

Ordovician Kinnikinic Quartzite is exposed at the dam, with Mesoproterozoic Swauger Formation to the north (Ruppel, 1980). At mile 48, at Goldburg Summit, pink quartzite of the Mesoproterozoic Swauger Formation is exposed on the right below an alluvial fan. The Swauger Ranch, for which the unit is named (Ross, 1947) is 5 miles north of here.

Pahsimeroi Valley

At mile 53 the road on the left goes to the Upper Pahsimeroi River (Figs. 20 and 21), and to the right is the Swauger Ranch. The buried north-trending trace of the Donkey fault is crossed here. To the south the hills in the hanging wall of the west-dipping Donkey fault are underlain by Eocene to lower Miocene(?) Donkey Fanglomerate, which filled a basin that formed during the Paleogene basin forming event (Janecke, 1992a; 1994; 1995a). These deposits are > 2 km thick and are syntectonic to slip on the low-angle Donkey fault.

At mile 55 is the Goldburg townsite and a major road junction. Doublespring Road is to the left. There is a panoramic view to the west here of the Lost River Range, Borah Peak, Doublespring Pass, and the Pahsimeroi Mountains (Fig. 20).

Big Creek to Patterson

The turnoff to the Big Creek Campground is on the right at mile 59. Big Creek is a major access route to the high Lemhi Range to the east (Fig. 22). Crone and Haller (1991) place the segment boundary between the Goldburg and Patterson segments of the Lemhi fault at Big Creek. The presence of an orthorhombic array of fault scarps in this area (Janecke, 1993) supports this placement. There is a big view to the left to central Pahsimeroi Valley and Pahsimeroi Mountains beyond. To the east is the Lemhi Range and type sections of several of the Lemhi Group units (Ruppel, 1975; Winston et al., 1999, this volume).

Continuing north toward Patterson (reached at mile 64), former site of a smelter and mining complex, the west side of the Lemhi Range is east of the road. Triangular facets are particularly well-developed along this northern half of the Lemhi fault (Fig. 23). The rocks at the range front are Mesoproterozoic Big Creek and Apple Creek formations, cut and mineralized by the concealed Eocene Ima stock (Ruppel, 1980). The tree- and scree-covered slopes do not reveal the structure, but up Patterson Creek can be seen the generally east-dipping Mesoproterozoic strata of the Lemhi Group. The Lemhi fault cuts alluvial fans at the mouth of Patterson Creek (Fig. 23) (Crone and Haller, 1991; Turko and Knuepfer, 1991). Scarps up to 1.7 m high are preserved in this area (Haller, 1988).

Lower Pahsimeroi Valley Near May

The road heads northwest, down off the alluvial fan at mile 69, five miles north of Patterson. The irrigated part of the Pahsimeroi Valley is to the north (Fig. 24). Mahogany Creek road is on the left. To the northwest, Grouse Peak can be seen at the north end of Pahsimeroi Mountains. Here the Mesoproterozoic Lawson Creek Formation is locally overlain by the Neoproterozoic formation of Leaton Gulch and Eocene Challis Volcanic Group (Hobbs, 1980; McIntyre and Hobbs, 1987). Southward in the Pahsimeroi Mountains, a west-southwest-dipping panel of Ordovician to Mississippian rocks forms Mahogany Hill, at the head of Grouse Creek. To the south of a south-dipping normal fault are tightly folded Mississippian to Pennsylvanian limestone, which underlie Grouse Creek Mountain and continue southward to Doublespring Pass (Mapel et al., 1965).

At mile 75, eleven miles north of Patterson, Hooper Lane is on the left. Doublespring Pass is 24 miles up Hooper Lane. His-

Figure 20. View southwest of Upper Pahsimeroi River, with Borah Peak on right skyline, and Idaho's only glacier below, in shadow. The prominent low point on the center skyline ridge is Leatherman Pass, cut by the east-dipping Leatherman Pass normal fault, with Leatherman Peak east (left) of it. Figure 10 shows this saddle from the other side of the Lost River Range. The Goldburg Hills are in left front and the northernmost end of the Donkey Hills in left center. The Goldburg normal fault follows the west edge of the Donkey and Goldburg hills (Figs. 2 and 8). Photograph taken from high in the Lake Fork of the Big Creek drainage in the Lemhi Range.

Figure 22. Looking north across the mouth of Big Creek at the Lemhi Fault. The fault runs up the small canyon in the middle of the view and then up to the left edge of photograph. Hanging wall rocks to the southwest of the fault are Tertiary sedimentary rocks of the Donkey Fanglomerate. Names and interpretation of locations of segments and boundaries are slightly different between Janecke (1993) and Crone and Haller (1991). Fault pattern is very complex in this area and includes a network of west to west-southwest-dipping normal faults (Fig. 8) (Janecke, 1992b).
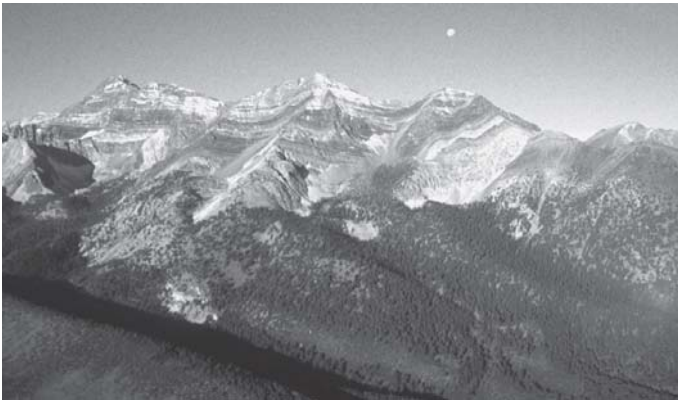

Figure 21. Aerial view of folded Ordovician to Mississippian rocks in the Borah Peak horst west of the headwaters of the Pahsimeroi River. The peak on the left skyline is Borah Peak. The Borah Peak thrust, which here places Ordovician to Silurian carbonates on the Devonian Jefferson Formation, dips to the left, or south, and is found in the bench to the right of Borah Peak (Janecke and Wilson, 1992). Some apparent folds are due to interaction of topography and structures. The beds are less folded than they appear in this view. Tree covered Pinedale lateral moraines in the lower part of the photograph slope down to the right. The area in the left-hand (south) side of the view was mapped by Skipp et al. (1988b).

Figure 23. View north of the west front of the Lemhi Range at Patterson, the town at the mouth of the narrow canyon in the center of the view. White sand dunes accumulated at the mouth of Patterson Creek. The Patterson segment of the Lemhi fault here cuts Pleistocene fan deposits and bounds well-developed faceted spurs (Haller, 1988). Type sections of Mesoproterozoic Lemhi Group rocks are found at the head of Inyo Creek and West Fork of Big Creek, south and east of Patterson.

toric May is on the left at mile 76.

The northern Lemhi Range to the east contains an essentially northeast-dipping section of Lemhi Group and Swauger Formation. The Patterson quadrangle map of Ruppel (1980) shows a maze of thrust faults. Some of them may exist, but regionally, they do not significantly disrupt the stratigraphic sequence (Winston et al., 1999, this volume). Tysdal (1996a and b; Tysdal and

Moye, 1996) show that southwest-dipping normal faults may be more important structures in this part of the range. Figure 24 is a view to the northern Lemhi Range and the Allison Creek (Salmon River) fault, that bounds a Paleogene half-graben east of the Salmon River (Janecke, 1995a; Janecke et al., 1998).

On the left at mile 80 are outcrops of the Swauger Formation, Lawson Creek Formation, and Kinnikinic Quartzite along a road that snakes up and over the summit of Grouse Peak in the northern Pahsimeroi Range to the breccia outcrops in the formation of Leaton Gulch (Hobbs, 1982; Carr and Link, 1999, this volume).

Figure 24. View north from north of May of the northern Pahsimeroi Valley, the northern Lemhi Range, and a large alluvial fan at mouth of Morse Creek. The range-front fault includes the May and Patterson segments of the Lemhi fault. Janecke (1993, Figure 6) shows a map of this area. The Allison Creek normal fault, a west-southwest-dipping high angle normal fault that bounded a Paleogene half-graben, intersects the Lemhi fault at the first major canyon left of the center. The Allison Creek fault separates the tree-covered bedrock in the main part of the Lemhi Range from the sage-covered Tertiary rocks in the lower left part of the range. Tysdal and Moye (1996) mapped the Allison Creek fault to the north. Note the large size of the alluvial fan issuing from the footwall on the right. In this part of the Pahsimeroi Valley fans derived from the footwall of the Lemhi fault rival those shed from the hanging wall, contrary to the predictions of most rift models.

Ellis and the Salmon River

Ellis is reached 82 miles north of Howe, 25 miles north of Patterson. East of the road is a mudcracked outcrop of the Mesoproterozoic Apple Creek Formation (Fig. 48). This is the end of the roadlog. Roadlog #4, along the Salmon River joins here.

ROAD LOG #3: MUD LAKE TO SALMON: BIRCH CREEK AND THE LEMHI VALLEY

Mud Lake Area

Head north to Salmon at the Junction of Idaho Highway 28 and 33 west of Mud Lake (Fig. 1) and set odometer to 0.0. We are heading northwest across the northeastern Snake River Plain. Mud Lake is to the northeast, and the Circular Butte eruptive center to the west. The Lemhi Range is to the west, the Beaverhead Mountains to the east, and the continental divide to the north. The Lost River Range is visible west of the Lemhi Range (Figs. 1 and 2). Price et al (1999, this volume) describe the geology of the southern Beaverhead Range at Reno Point. The regional geology is shown on the maps of Scott (1982) and Kuntz et al. (1994).

At the junction with Highway 22, Howe Point can be seen at the extreme southern tip of the Lemhi Range, to the left (northwest). Diamond Peak is to the northwest (Fig. 25), and the Beaverhead Mountains are to the right.

At mile 18.2 is the northern INEEL boundary. The road to the west is to Kyle Canyon and Snaky Canyon in the southeastern Lemhi Range. Gently east-dipping ash-flow tuffs of the Upper Miocene and Pliocene Heise Group are exposed up the canyon (Morgan, 1992). Farther up the canyon is the type section of the Snaky Canyon Formation plus outcrops of the Permian Phosphoria Formation and Triassic Dinwoody Formation (Skipp, Hoggan, et al., 1979).

Birch Creek Valley

Long Canyon and Southern Beaverhead Mountains

At mile 27.3 is the turnoff to Long Canyon to the east. The southern Beaverhead Mountains have been mapped by Garmezy (1981) and Skipp (1984, 1988a), and are composed of Mesoproterozoic to Permian rocks of the Hawley Creek, Fritz Creek and Cabin thrust plates. Pennsylvanian to Permian rocks of the Snaky Canyon Formation are faulted onto the Neoproterozoic and Cambrian Wilbert Formation along the Beaverhead normal fault north of here. This intrabasinal high coincides with a segment boundary along the Beaverhead fault (Crone and Haller, 1991)(Fig. 2). The Neoproterozoic and Cambrian Wilbert Formation is exposed up Long Canyon, and thickens eastward from the southern Lemhi Range to the Beaverhead Mountains (Skipp and Link, 1992).

At mile 29.3 on the left is a Prehistoric Man Historic Site. Paleo Indians camped and hunted along Birch Creek (Dereg, 1996). At mile 30.5 is the road to Skull Canyon on the right (folded Mississippian Scott Peak Formation, Figure 26). This was the site of Blue Dome, a now-demolished hotel that had a blue roof visible for miles around. Garmezy (1981) mapped the folded but eastward-younging rocks of the southernmost Beaverhead Mountains.

At mile 31.5 is the Pass Creek road to the left, providing access to the southern Lemhi Range, including the area south of Diamond Peak (Fig. 25). This prominent horn on the crest of the Lemhi Range to the west is underlain by Upper Mississippian carbonate-bank strata. The area is shown on the geologic map of Ross (1961). Beutner (1968) mapped the numerous north northwest-trending, east-vergent folds of the southern Lemhi Range in great detail. The origin and the age of several northeast-striking faults (Fig. 2) that cross the range is uncertain. They have up to 3 miles of left separation and may be tear faults to the adjacent thrust faults (Beutner, 1968).

Lone Pine Area

North of Blue Dome the road passes through the Birch Creek Canyon near Lone Pine (mile 32.8), named for a solitary grand limber pine south of town. In the Lone Pine area, the "Medicine Lodge" volcanics (probably Challis equivalent) are overlain by the 6-16 Ma Medicine Lodge limestone, the basalt of Lone Pine, and the 6.5 Ma tuff of Blacktail (Rodgers and Anders, 1990).

Figure 25. View west to Diamond Peak and the southern Lemhi Range from Blue Dome. Birch Creek is in the middle distance, lined by the only trees for miles around. Diamond Peak is underlain by Mississippian limestone.
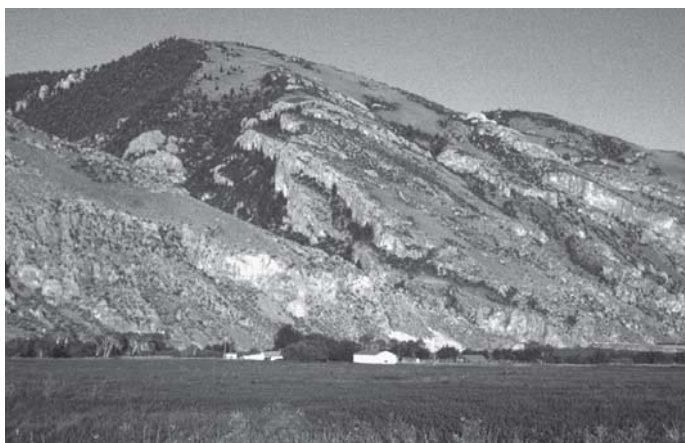
Figure 26. View of the mouth of Skull Canyon, looking southeast from south of Lone Pine. The cliffs are composed of a dip slope of Mississippian limestone of the Hawley Creek thrust plate (Garmezy, 1981; Skipp, 1988a). This bedrock ridge is in the hanging wall of the Beaverhead fault and coincides with the boundary between the Nicholia and Blue Dome segments (Crone and Haller, 1991).

Figure 27. Westward view of Bell Mountain and the central Lemhi Range from north of Lone Pine. The glaciated horn of Bell Mountain (the southern high peak) is underlain by Ordovician Kinnikinic Quartzite, repeated on a thrust fault (Beutner, 1968; Wilson and Skipp, 1994). Rocks on the skyline north of Bell Mountain are tightly folded but eastward younging Devonian to Mississippian carbonate strata.

These rocks dip about 7 to 15° to the east-northeast (Rodgers and Anders, 1990; Anders et al., 1993).

Nicholia and Viola Mine

The highway ascends to the marshy headwaters of Birch Creek in the five miles north of Lone Pine. The historic town of Reno is east of the road at mile 39. Reno was a settlement during mining booms at Nicholia in the Beaverhead Range to the north. At Nicholia a rich silver deposit of the Viola mine operated between 1882 and 1887 supporting a lead smelter and in 1886, a town of 1,000 people (Dereg, 1986). The smelter closed in 1889. Ore was produced from stratabound deposits in Devonian dolostone (Skipp et al., 1988). Price et al. (1999, this volume) describe a side trip to this area.

North of Nicholia the road skirts along the southwest edge of Middle Ridge, a north-northwest trending high in the middle of the basin between the Lemhi Range and Beaverhead Mountains. The ridge consists of Miocene to Pliocene sedimentary rocks of the Medicine Lodge beds (Scholten and Ramspott, 1968). Tephras within the section had sources in the eastern Snake River Plain (M. Perkins, University of Utah, pers. comm.). Northwest-trending extensional folds deform the sequence, and a strand of the Beaverhead fault has uplifted these rocks in its footwall (Scholten and Ramspott, 1968; Crone and Haller, 1991). Spotty exposures of the interbedded gravel, limestone and tephras persist much of the way to Leadore.

At mile 45 is the charcoal kilns road to the left into the Lemhi Range. The prominent horn to the west is Bell Mountain, carved in Ordovician Kinnikinic Quartzite (Fig. 27). A small thrust at Bell Mountain repeats lower Paleozoic rocks (Beutner, 1968). At mile 51 is the Hahn townsite, the site of a lead smelter built in 1908. At mile 52 is the turnoff to the left to Spring Mountain Canyon, draining Big Windy Peak (Fig. 28). Price et al. (1999, this volume) describe a field trip to Spring Mountain Canyon where skarn mineralization and an incised Lower Devonian valley fill system can be seen (Grader and Dehler, 1999, this volume) (Fig. 28). The area is on the Gilmore 15 minute quadrangle map of Ruppel and Lopez (1981).

Lemhi River Valley

Gilmore Area

Gilmore Summit is reached at mile 56. This intrabasinal drainage divide is roughly aligned with similar, anomalous drainage divides in the hanging wall of the Lemhi and Lost River faults. These divides are probably related to the passage of North America over the Yellowstone hotspot because the topographic highs coincide with a seismic and neotectonic parabola centered on the present position of the hot spot (Anders et al., 1989; Pierce and Morgan, 1992). Two miles past the summit (mile 58) is the road to Gilmore. The Texas mining district caused Gilmore to be a thriving mining community, reached by the Gilmore and Pittsburgh Railroad in 1910 (Figs. 29 and 30). The district contains lead-silver veins and replacements related to Eocene stocks intruding Paleozoic limestone (Ruppel and Lopez, 1988). Stratabound Mississippi Valley type lead-silver-zinc deposits in the Devonian Jefferson Formation may also be present, and may provide the source of the metals (Skipp et al., 1983; 1988).

Figure 28. Aerial view to the southwest of Big Windy Peak on the eastern side of the Lemhi Range south of Gilmore. Bedrock is Paleozoic carbonate. Extensive glacio-fluvial deposits in the foreground are derived from Spring Mountain Canyon (see Price et al., 1999, this volume). The origin of the light specks on the alluvial/outwash fans is uncertain.

The Gilmore area grew rapidly after 1902. Most mining activity was finished by 1925. The Gilmore and Pittsburgh railroad allowed transportation of ore from the several mines in the area, but never was solvent. Service was stopped in 1939 and the tracks were ripped up for salvage in 1940. The railroad was built too late into the 20th Century, too close to the depression that followed World War I, and into an area too topographically severe and too limited in mineral and lumber resources to support a railroad (Myers, 1981; Dereg, 1996). The latest phase of mining activity in the Texas district was from 1943 to 1961. Lead-silver-gold replacement orebodies at the Hilltop mine were developed (Ruppel and Lopez, 1988).

Eighteenmile Road

At mile 66.5 is the Eighteenmile Road on right. This affords access to the high Beaverhead Mountains to the south, that were mapped and studied by Skipp (1984; 1985; 1988a). The Clear Creek gypsum deposit, in hydrothermally altered Scott Peak Formation, is reached up this road (Skipp et al., 1988a). This area

Figure 29. Aerial view of abandoned Gilmore and Pittsburgh railroad grade curving up toward townsite of Gilmore in middle right center. The canyon in right center is Meadow Lake Creek.

Figure 30. Wye in abandoned Gilmore and Pittsburgh railroad grade immediately southeast of Leadore. The view is to the west. The grade heads east at bottom of photograph toward view shown in Figure 31. The grade headed south on left side of view toward Gilmore (Fig. 29).

contains the southern part of the Hawley Creek thrust fault (Lucchitta, 1966; Skipp, 1988a) the first major thrust east of the Pioneer Mountains (Rodgers and Janecke, 1992). The red rocks at Eighteenmile Peak are the granite and syenites of the Ordovician (about 483 Ma) Beaverhead Pluton (Ramspott, 1962; Evans and Zartman, 1988).

East-southeast of Leadore, at Hawley Creek, the Hawley Creek thrust fault places Ordovician Beaverhead Mountains Pluton over Permian and Triassic rocks (Lucchitta, 1966; Scholten and Ramspott, 1968; Skipp, 1984, 1988a; Skipp et al., 1988). East of Leadore, the fault trace places Beaverhead Mountains pluton over upper Paleozoic to Mesozoic rocks (Lucchitta, 1966; Ruppel, 1969; and Skipp, 1988a). The concealed trace heads northwest across the Lemhi Valley into the northern Lemhi Range. The Poison Creek thrust of Tysdal (1996a, b) is on strike to the northwest in the northern Lemhi Range (Skipp, 1987; 1988a). Probable westward continuations into the Salmon River Mountains are discussed in Skipp (1987) and Evans (1999).

Leadore

At mile 65 is the village of Leadore, open, windy, cold, and hunkered-down. The Gilmore and Pittsburgh railroad reached here in 1910 and times were good. The Junction and Little Eightmile mining district on Grizzly Hill to the north (Fig. 31) produced lead and silver ores from Paleozoic limestones. The Texas district to the southwest near Gilmore was far more productive (Ruppel and Lopez, 1988). North of Leadore the highway descends the flood plain of the Lemhi Valley. Tertiary volcanic and sedimentary rocks become increasingly more apparent northward from Leadore.

A low pass in the Beaverhead Mountains about half way between Leadore and Lemhi exposes a major southeast-dipping low-

Figure 31. View east-northeast toward Railroad Canyon in the Beaverhead Mountains. The skyline ridge is on the continental divide and along the Montana-Idaho border. Slices of the Hawley Creek thrust sheet are preserved on the lower part of the mountain (Skipp, 1988a). An abandoned railroad grade is the thin line on the right middle of view. The mountain front is bounded by the Beaverhead fault. Crone and Haller (1991) place the segment boundary between the Leadore and Mollie Gulch segments just west (left of this view). A west-dipping normal fault in the footwall of the Beaverhead fault intersects the range front here and probably helped to localize the ruptures along it (Janecke, 1993, Figure 1).

Figure 32. Bulldozer cuts in Mesoproterozoic Apple Creek Formation east of the highway, at the southwest edge of the Lemhi Pass Thorium District, between Lemhi and Tendoy. The Salmon basin and Beaverhead Mountains are visible in the left distance.

angle normal fault (the Goat Mountain thrust of Staatz, 1979) (Figs. 1 and 2). VanDenburg (1997; VanDenburg et al., 1998) showed that the thrust is a major pre-Challis normal fault. Rollover in the hanging wall of the younger Miocene northwest-dipping, Little Eightmile Creek normal fault folded the pre-Challis normal fault (VanDenburg et al., 1998). Grizzly Peak exposes another segment of the pre-Challis normal fault (Staatz, 1973; VanDenburg et al., 1998).

Northern Lemhi Range

At mile 93 is the Hayden Creek road. Hayden Creek affords access to the northern Lemhi Range (Fig. 33), that has been mapped recently by Tysdal (1996a; b). The Hayden Creek diamictite, once thought to represent a fault breccia (Ruppel, 1978) is a sedimentary unit in the Apple Creek Formation (Tietbohl, 1986; Tysdal, 1996a; 1996b). At mile 94 is the village of Lemhi. The Apple Creek Formation is exposed in roadcuts north of Lemhi (Fig. 32). The Poison Creek thrust, which puts Mesoproterozoic Apple Creek Formation over Ordovician Saturday Mountain Formation trends northwesterly through the northern headwaters of Hayden Creek (Tysdal, 1996a).

Lemhi Pass Area

The village of Tendoy is at mile 100. The mountains to the east of Tendoy are part of the Lemhi Pass thorium district (Staatz, 1972; 1973; 1979) (Fig. 32). The area is underlain by Mesoproterozoic Apple Creek Formation (shown by Ruppel et al., 1993 as Yellowjacket Formation) (Winston et al., 1999, this volume), and cut by the south-southwest dipping low-angle Lemhi Pass normal fault (Fig. 2) (VanDenburg et al., in press). Some right-separation occurred across this fault. A thick and variable sequence of Eocene volcanic rocks, interbedded sedimentary rocks, and at least two unconformity-bounded sequences of Eocene to Oligocene sedimentary rocks overlie the Mesoproterozoic rocks in this region (Blankenau, in press; VanDenburg, 1997; Axelrod, 1998).

Salmon Basin Detachment and Extensional Folds

The Salmon basin detachment fault bounds the east side of the Salmon basin north of here and ends southward at the Lemhi Pass fault (Fig. 2). The Salmon basin fault and associated half-graben were active during the Paleogene basin-forming event

Figure 33. View west-southwest up the valley of Hayden Creek with the northern Lemhi Range in the background. Foreground contains a very thick section of mostly northeast-dipping Challis Volcanic Group lava flows overlain by a thin sequence of gently folded Tertiary sedimentary rocks. One or more major west- to southwest-dipping normal faults repeat the Tertiary rocks. Some of the northernmost Paleozoic rocks at this longitude are in horst blocks to the right of Hayden Creek (Tietbohl, 1986). The high peaks are mainly Mesoproterozoic Lemhi Group (Tysdal, 1996b).

(Tucker, 1975; Harrison, 1985; Janecke, 1994; Blankenau, in press). Notable within the half graben are large, map scale anticlines and synclines (with fold heights > 2 km) and locally overturned folds that formed during extension (Janecke et al., 1998; Blankenau, in press). The paleo-elevations of fossil flora in these sedimentary rocks are controversial (Wolfe and Wehr, 1987; Axelrod, 1998).

At least three generations of normal faults extended the southeast margin of the Salmon basin, around Tendoy during the Cenozoic (Blankenau, in press). The history of the Horse Prairie half graben to the east in Montana, is even more complex, and reflects at least five generations of normal faults (VanDenburg et al., 1998). East northeast-west southwest extension dominated over northwest-southeast extension in both areas. The Beaverhead normal fault does not persist north of Tendoy (Crone and Haller, 1991; Blankenau, in press).

Lewis and Clark Trail

The Lewis and Clark trail came down the mountains from Lemhi Pass to near the village of Tendoy. Lewis and Clark, led by Sacagawea, whose brother was the chief of the local Shoshone tribe, came through here in 1805. Among many other versions, Robertson (1998) describes this place and event succinctly.

Fort Lemhi

Two miles north of Tendoy, at mile 102, is the Historical Marker for Fort Lemhi (Fig. 34), which, in 1855 was the first Mormon settlement in Idaho (Snook et al., 1992). The site, at the mouth of Pattee Creek, was abandoned in 1858, and the Mormons retreated southward to Zion. This was a retrenchment of the grand plans for a Mormon State of Deseret that would have spanned much of the interior western United States (Morgan, 1987).

At mile 111 is Baker. The steep front of the Beaverhead Mountains looms to the north (Fig. 35). The mountains are bounded by

the Paleogene Salmon basin detachment fault (Janecke et al., 1998; Blankenau, in press), and are composed of a generally east-dipping section of Apple Creek, Gunsight, and Swauger Formations. Late Pleistocene alpine glaciers sculpted the range. The maze of thrust faults shown on the Dillon 1° x 2° sheet geologic map (Ruppel et al., 1993) is probably not present (see Umpleby, 1913; Anderson, 1956; 1961; Winston et al., 1999, this volume). Freeman Peak is the prominent horn north of Salmon (Figs. 35 and 36). Wimpey Creek, northeast of Baker provides a view of the Salmon basin detachment, whereas a drive up Withington Creek, south of Baker, provides access to the Eocene Withington Creek caldera (Ruppel et al., 1993; Blankenau, in press). This caldera is one of the few calderas outside the central cauldron complex of the Challis volcanic field. Quartzite-bearing, ash flow tuffs filled the caldera in Middle Eocene time (S.U. Janecke, unpublished data). Harrison (1985) described the lithology and depositional environment of the fluvial to lacustrine basin-fill deposits of the Salmon area in detail. Excellent exposures of these rocks are present along the Lemhi River between Baker and Salmon. Access is provided by frontage roads on both sides of the highway. Studies in the southeast part of the Salmon basin (Blankenau and Janecke, 1997) suggest a far more complex stratigraphic and tec-

tonic evolution than envisioned by Harrison (1985).

Salmon

Salmon is reached at mile 119 (Fig. 37). The city was organized in 1869, after discovery of gold at Leesburg in 1866. George L. Shoup, Idaho's last territorial governor, was one of the leading County Commissioners of the new Lemhi County (Snook et al., 1992). Sheared 1370 Ma porphyritic granites northwest of Salmon intruded the Yellowjacket Formation (Evans and Zartman, 1990; Doughty and Chamberlain, 1996) (Fig. 2). This metamorphism

Figure 36. View north up Carmen Creek to Freeman Peak, underlain by Mesoproterozoic Lemhi Group and Swauger Formation. The Miner Lake-Beaverhead Divide fault zone, an enigmatic, steeply southwest-dipping fault, is directly behind Freeman Peak.

Figure 34. Historical marker for site of Fort Lemhi, first Mormon settlement in Idaho, located along Lemhi Creek in middle distance.

Figure 35. View to the northeast of the front of the Beaverhead Mountains just south of Salmon. The high peaks are composed of Mesoproterozoic Lemhi Group and Swauger Quartzite. Freeman Peak is the prominent peak in the left middle. White rocks in foothills are Paleogene basin-fill deposits (probably lacustrine deposits). The Salmon basin detachment fault separates the tree-covered mountains in the distance from the sage-covered foothills in the middle distance.

Figure 37. View looking northeast across the Salmon basin from south of Salmon. The Salmon airport is left of center. The Lemhi River flows from right to left and joins the Salmon River (bottom) just left of the edge of the photograph. Harrison (1985) most recently described the Paleogene basin-fill deposits in low eroded hills between tributaries of the Lemhi and Salmon rivers. She described conglomerates along the basin margin that interfinger with fine tuffaceous lacustrine deposits in the center of the basin. Blankenau and Janecke (1997) found at least two unconformity-bounded sequences of sedimentary rocks in the southeast part of this basin, several miles to the right of this photograph. The sedimentary rocks here are not well dated but may be as old as Middle Eocene to younger than 30 Ma (Axelrod, 1998; Blankenau, in press). Trees cover Mesoproterozoic rocks of the Beaverhead Mountains.

was initially thought to be 1500 Ma (K-Ar, Armstrong, 1975), and formed the incorrect basis for the concept of a Mesoproterozoic Salmon River Arch. This is the end of Roadlog #3. Roadlog #4 joins here.

ROADLOG #4: STANLEY TO CHALLIS ALONG THE SALMON RIVER

The Salmon River Canyon

Stanley to Yankee Fork

This road guide starts in Stanley, Idaho, and heads north and east along the Salmon River on Idaho Highway 75 toward Challis. As such, it crosses the strike of the orogenic belt rather than going more-or-less parallel to strike as do Roadlogs #1-3. The Challis 1° x 2° geologic map of Fisher et al. (1992) is a critical reference for the first two-thirds of this road log. Set odometer to 0.0 miles at the intersection of Highway 75 and 21 in Upper Stanley. Johnson et al. (1988) describe the geology along this road between Stanley and Thompson Creek. From Stanley eastward, the Salmon River first passes through Late Cretaceous granodiorite of the Idaho batholith, then into Paleozoic wallrocks of the batholith, including the Pennsylvanian and Permian Grand Prize Formation (Sun Valley Group), and the lower Paleozoic Salmon River Assemblage (Milligen Formation equivalent) (Fisher et al., 1992; Link et al., 1995).

The Stanley Basin, a west-tilted Basin-and-Range half-graben, is south and west of Stanley, bordered by the Sawtooth Mountains on the west (Fig. 38) and the White Cloud Peaks to the east. The Sawtooth normal fault is antithetic to the dominantly west-southwest dipping Basin-and-Range normal faults to the east. The floor of Stanley basin is almost completely mantled by till. The glacial geology of this area is described by Breckenridge et al. (1988) and Borgert et al. (1999, this volume). Eocene granite is exposed in the Sawtooth Range south of Redfish Lake (Fisher et al., 1992).

Radiometric dates (mainly K-Ar) in biotite granodiorite of the Idaho yield ages of 85 to 70 Ma (Kiilsgaard and Lewis, 1985; Lewis et al., 1987; Criss and Fleck, 1987; Johnson et al., 1988). Hot springs along Idaho Highway 75 coincide with normal faults of the northeast-striking Trans-Challis system. These Eocene faults cut the granitic rocks, and controlled mineralization and topography during Challis volcanism (Kiilsgaard et al., 1986; Bennett, 1986). Placer workings in older alluvium can be seen at several points. About 8 miles from Stanley is the Basin Creek road to the left. Basin Creek is near the southwest end of the Custer graben, a northeast-trending synvolcanic basin (Fig. 4).

Sunbeam and Yankee Fork

Thirteen miles past Stanley is the town of Sunbeam and the mouth of the Yankee Fork river. The Yankee Fork was site of a major gold rush in the 1870's. Peak production, from Challis-related epithermal deposits and their placers, was in the 1880s.

A power dam built on the Salmon River here in 1910 was sabotaged in 1934 (Fig. 39). "It is uncertain.who is entitled to an accolade for dynamiting the south abutment. However, there is no limit to what can be accomplished if no one cares who gets the credit" (Carrey and Conley, 1978, p. 66). This sabotage allowed

salmon, at least until dams on the lower Snake River dealt the species-fatal blow (Fig. 40), to return to Redfish Lake and the Stanley Basin.

The historic gold-placer dredge ten miles up Yankee Fork creek now is a mining museum. Hecla Mining Co. operated the Grouse Peak gold mine near the western headwaters of the Yankee Fork, in the early 1990s (Fig. 41). The Grouse Peak deposit is located in Eocene lake beds, and contains gold deposited by shallow hydrothermal systems (Allen and Hahn, 1994). The upper course of the Yankee Fork follows the southern boundary of the syn-Challis Custer Graben (Fig. 4) (Fisher et al., 1992).

Paleozoic Black-Shale Mineral Belt

The mouth of Warm Spring Creek, a major tributary from the south is reached about 16 miles beyond Stanley. East of here the uplands above road level pass into steeply west-dipping Paleozoic sedimentary rocks of the Black Shale Terrane of Hall and Hobbs (1995). An apophysis of the Idaho batholith persists at road level for about 5 more miles. The black-shale belt includes the structurally complex Cambrian, Devonian, and Mississippian Salmon River Assemblage (Hall, 1985; Fisher et al., 1992), that forms dark-colored talus-covered shaley outcrops (Fig. 42). Part of it probably correlates with the Devonian Milligen Formation that contains syngenetic stratiform silver-lead-zinc deposits near Sun Valley (Hall, 1985; Turner and Otto, 1988; Turner and Otto, 1995).

North of the road and river about 23 miles from Stanley is a major steeply west-dipping Mesozoic(?) gouge zone in the Salmon River Assemblage. Slate Creek, at 24 miles from Stanley coincides with the eastern edge of the Idaho batholith. It is a major drainage of the glacially sculpted White Cloud Peaks area to the south, which contain Grand Prize Formation overlying Salmon River Assemblage, intruded by Cretaceous granodiorite. About ten miles up Slate Creek is the Livingston Mine, that contains syngenetic stratiform mineral deposits in the Salmon River Assemblage (Hall, 1985). Eocene volcanic rocks cap the highest ridges eastward from Slate Creek on both sides of the Salmon River (Fisher et al., 1992).

Twenty-six miles from Stanley is the Thompson Creek Road. The Thompson Creek molybdenum mine is 4.1 miles up the road to the north. It contains a low-fluorine stockwork molybdenum deposit with a few base-metal lead and silver veins hosted by the Salmon River Assemblage (Fig. 43). The deposit formed during Late Cretaceous hydrothermal alteration associated with the Idaho batholith. Extensive associated potassic alteration produced muscovite mica with [40]Ar-[39]Ar ages of 87.4 to 87.6 Ma (Hall; 1985; Johnson et al., 1988).

The Yankee Fork Ranger station is one mile east of Thompson Creek, about 28 miles from Stanley. Iron-stained Salmon River Assemblage forms cliffs on both sides of road. The geologic map of the Bayhorse area (Hobbs et al., 1991) shows the geology between here and Challis in more detail than Fisher et al. (1992).

Bayhorse Assemblage

East of Sullivan Hot Springs, 31 miles from Stanley, the Salmon River Assemblage structurally overlies Cambrian to Ordovician strata of the Bayhorse district (Hobbs et al., 1991) (Fig.

Figure 38. Aerial view to the west of front of Sawtooth Mountains in distance, Stanley Basin in middle distance and lower part of Salmon River Canyon in right foreground. Glaciated northern Sawtooth Mountains are made of mainly Cretaceous intrusive rocks on north and Eocene granite on south (left side of view).

Figure 41. Aerial view looking north up the Yankee Fork to the Grouse Creek open-pit gold mine. Peaks on skyline are Twin Peaks, part of the Eocene Twin Peaks Caldera. The mine is located in a northeast-trending Eocene graben bounded by faults of the Trans-Challis fault zone (Fig. 4).

Figure 39. Dam across the Salmon River at the mouth of the Yankee Fork. It was sabotaged in 1934.

Figure 42. Talus-covered slopes of dark argillaceous rocks of the Paleozoic Salmon River Assemblage on the north side of Salmon River at Clayton. The age and tectonic significance of these rocks is obscure.

Figure 40. The Salmon River was closed to floating on a short stretch between Stanley and Sunbeam in August 1998, due to (futile?) efforts to protect salmon spawning beds.

Figure 43. Aerial view looking northeast at the Thompson Creek open-pit molybdenum mine. The bare ridge in the foreground is underlain by Paleozoic Salmon River assemblage, which hosts the mineral deposit, near a Cretaceous intrusive body. The skyline behind mine is underlain by dacite lava flows of the Eocene Challis Volcanic Group.

Figure 44. Flatirons of lower Paleozoic rocks of the Bayhorse sequence, looking northeast, below Clayton. Kinnikinic Quartzite makes up the massive cliff.

Figure 45. Aerial view looking south up the Salmon River canyon west of Challis. High skyline peak is Castle Peak in the White Cloud Peaks (the "heart of Idaho"). Most of the bare slopes west of the river are underlain by Challis Volcanic Group. Paleozoic rocks underlie the tree-covered ridge above the bare slopes, and are also found as inliers along the Salmon River canyon (Hobbs et al., 1991).

44). The fault here carries tightly folded lower Paleozoic and Milligen-equivalent Salmon River Assemblage over Ordovician Saturday Mountain Formation (see Hobbs et al., 1991, west end of cross section D-D'). Depending on the age of the Salmon River Assemblage immediately above the west-southwest dipping fault, the contact could be a low-angle normal fault. East of (below) this fault, whose movement history has been much debated (Dover, 1978; Hall, 1985; Link et al., 1995) the road enters the miogeoclinal Carbonate Terrane of Hobbs (1995).

Hobbs' (1995) Carbonate Terrane contains the Ordovician and Cambrian Bayhorse Dolomite, the Ordovician Ella Dolomite, Kinnikinic Quartzite and Saturday Mountain Formation, overlain by thick Silurian to Mississippian carbonate strata (Fig. 3). Within these rocks north of the Salmon River is the Bayhorse mining district (Ross, 1937; Hobbs et al., 1991).

Thirty-three miles from Stanley, east of Sullivan Hot Springs, the road crosses through a map-scale anticline with Ordovician Clayton Mine Quartzite in the core and at mile 34 enters Clayton, population 26. Below Clayton is a big vista of Kinnikinic Creek on the north, with Ordovician Kinnikinic Quartzite overlain by orange-weathering carbonate of the Saturday Mountain Formation (Fig. 44). To the east is open country underlain by the Ordovician Ramshorn Slate along the flanks of another major, north-south trending anticline (Hobbs et al., 1991). Challis lava flows, tuffs and intrusive rocks are south of the river high on the ridge tops.

Several outcrops of Jurassic gabbro are just west of the mouth of the East Fork of the Salmon River, and near the Clayton Mine, up Kinnikinic Creek (Hobbs et al., 1991; Fisher et al., 1992). This is the only Jurassic intrusive rock in eastern Idaho. The gabbro intruded preferentially into the Ramshorn Slate (Hobbs et al., 1991).

East Fork Salmon River

Just east of Clayton, thirty seven miles from Stanley, the East Fork of the Salmon River enters from the south. The East Fork road provides access to a thick succession of Eocene Challis Volcanic Group (Fisher and Johnson, 1995b) including sedimentary rocks along Spar Canyon (Palmer, 1997; Palmer and Shawkey, 1997). It also affords access to the Spar Canyon Paleozoic carbonate succession.

The river and highway head off to the north, changing from their eastward course. From here to Challis the canyon is mainly

in Challis Volcanic Group with hills of folded Ordovician or older Clayton Mine Quartzite poking through (Hobbs et al., 1991). The bedrock ridge of Lone Pine Peak, east of the roads, contains Silurian to Mississippian Carbonate Terrane strata studied by Hays et al. (1980) and McFaddan et al. (1988). Notable are reef complexes in the Jefferson Formation and underlying Devonian channel-fill facies (Isaacson and Dorobek, 1988; Isaacson et al., 1988; Grader and Dehler, 1999, this volume).

Forty-six miles from Stanley is Malm Gulch on the right. Up Malm Gulch are extensive outcrops of Challis Group volcaniclastic sediments, including ash-flow tuff, volcanic sandstone, conglomerate and diamictite. A petrified forest of *in situ* *Metasequoia* stumps in Challis sandstones and conglomerates here was described by Ross (1937).

Fifty-two miles from Stanley the road comes out of the Salmon River canyon (Fig. 45). The country opens up to Basin-and-Range country of Round Valley, with the Pahsimeroi Mountains uplifted to the east at the northern end of the Lost River Range. Along the road between here and the northernmost end of the Pahsimeroi Mountains are Challis Volcanics. The Junction with US Highway 93 is reached 56 miles from Stanley. The Land of the Yankee Fork Mining Museum is just north of the road. The Big Lost River roadlog #1 joins here.

Challis to Salmon

Start at intersection of US 93 and Idaho Highway 75 on the south edge of Challis, Idaho. Proceed north on US 93. Set odometer to zero.

The city of Challis overlooks Round Valley, between the Salmon River Mountains and the north end of the Lost River Range (Figs. 45 and 46). Challis was laid out in 1876, and settled by cattle ranchers and then miners supplying the Yankee Fork gold mines (Carrey and Conley, 1978; Dereg, 1996). The rocks south and west of town belong to the Challis Volcanic Group, and consist of dacite and andesite lavas and ashflow tuffs, erupted mainly from centers in the Twin Peaks Caldera and Van Horn Peak caul-

Figure 46. Aerial view of city of Challis, looking north. Densely welded Tuff of Challis Creek is repeated by west-dipping normal faults in the low hills north of town. The underlying white unit is the Tuff of Pennal Gulch (McIntyre and Hobbs, 1987). Janecke et al. (1997) report an $^{40}Ar/^{39}Ar$ date for the tuff of Challis Creek from these exposures of 45.70 ± 0.08 Ma. The tuffs are very accessible from a side street in Challis. The owners nearby don't want rocks rolled on their houses, however.

Figure 47. View of pyroclastic rocks at the northwest edge of the Pahsimeroi Mountains from north of Challis. The dark-colored rocks that underlie the skyline in the middle distance are densely welded Tuff of Challis Creek sitting on top of the bedded white tuff of Pennal Gulch (McIntyre and Hobbs, 1987). There was erosion between deposition of the two units so the base of the Tuff of Challis Creek is inset into paleolows in the underlying tuffaceous sedimentary rocks of Pennal Gulch. Janecke (1992a) interpreted the erosion to reflect initial slip on Paleogene normal faults. Subsequent small normal faults disrupted the once continuous tuff of Challis Creek here. The Twin Peaks caldera northwest of Challis was the source for the tuff of Challis Creek (Hardyman, 1985).

dron complex north and west of town (Fig. 4) (McIntyre et al., 1982; Hardyman, 1985; McIntyre and Hobbs, 1987; Fisher et al., 1992; Janecke and Snee, 1993). North-south side roads off the main east-west street through Challis provide access to a section of east-dipping ashflow tuffs on the northwest side of town (McIntyre and Hobbs, 1987).

Pahsimeroi Mountains

The north end of the Pahsimeroi Mountains, at Grouse Peak east of the Salmon River and east of town (Fig. 47), is underlain by Eocene lava flows and Neoproterozoic strata of the formation

of Leaton Gulch (McIntyre and Hobbs, 1987; Carr and Link, 1999, this volume). Southward in the Pahsimeroi Mountains progressively younger Paleozoic strata crop out, until Pennsylvanian rocks form the ridges near Doublespring Pass (Mapel et al., 1965; Janecke and Wilson, 1992; Fisher and Anastasio, 1994). Challis volcanics extensively cover the northern portion and east side of the range (Janecke and Snee, 1993).

Highway 93 runs north-south, following the Salmon River, and crossing the local northeasterly strike of the Neoproterozoic bedrock in the Pahsimeroi Mountains. North of Challis the road crosses through the northern end of the Lost River Range, Pahsimeroi Valley, and the northern end of the Lemhi Range (Fig. 4, 50). The Salmon River alternately cuts through narrow bedrock canyons and more open country underlain by Challis Volcanic Group.

On the left, seven miles north of Challis is the mouth of Challis Creek and ten miles north is Morgan Creek. A few miles up Morgan Creek Mesoproterozoic Swauger Formation forms the steep cliffs west of the road (McIntyre and Hobbs, 1987).

Canyon North of Challis

In general the Salmon River canyon cuts thick and variable sections of Eocene Challis Volcanic Group rocks overlying Mesoproterozoic strata, including the Lemhi Group and Swauger Quartzite (Hobbs and Cookro, 1995; Winston et al., 1999, this volume). These Mesoproterozoic rocks are locally tightly folded but form an overall south-west dipping homocline (McIntyre and Hobbs, 1987). Repetition by gently southwest-dipping normal faults is likely. The volcanic rocks are cut by many north to northwest-striking Basin-and-Range normal faults and vary in their attitude. Gentle tilts on the volcanic rocks are most common (McIntyre and Hobbs, 1987). The first bedrock canyon is reached 16 miles north of Challis, north of Shep Creek on the south. Here the Mesoproterozoic Swauger Formation forms south-southwest-dipping flatirons through which the river flows. Some exposures of pre-Tertiary rocks between Challis and Salmon are uplifted along Cenozoic normal faults but many bedrock ridges reflect exhumed pre-Tertiary paleotopography.

Mouth of the Pahsimeroi Valley

The mouth of the Pahsimeroi River is about 20 miles north of Challis. The Little Lost River roadlog #2 joins here.

East of the road at Ellis are mud-cracked beds of the Mesoproterozoic Apple Creek Formation (Fig. 48). To the north is the north end of the Lemhi Range (Fig. 49). Two miles to the north the river cuts through younger west-dipping Swauger Formation (Fig. 50). A buried southwest-dipping thrust fault probably separates these exposures and continues north toward an older-on-younger fault along Moyer Creek that Ekren (1988) mapped as a normal-right slip fault (S.U. Janecke, unpublished data). The southern extent of this thrust fault is not known.

Northern Lemhi Range

Twenty five miles from Challis the Swauger Formation is unconformably overlain by Challis Volcanic Group strata, dropped down on the west-dipping Salmon River fault (=Allison Creek fault) that cuts the northwest end of the Lemhi Range (Janecke,

Figure 48. Large mud cracks on bedding plane of Mesoproterozoic Apple Creek Formation on east side of Salmon River at Ellis. Traces of prominent spaced cleavage dip to the right.

Figure 49. West-dipping flatirons of Mesoproterozoic Swauger Formation on the north end of Lemhi Range, east of the Salmon River.

Figure 50. Northeast-dipping Mesoproterozoic Swauger Formation along Salmon River north of Ellis. The light colored rocks in the distance are east-dipping tuff of Ellis Creek overlain by lava flows (McIntyre and Hobbs, 1987). This ridge of Swauger Formation was a paleohigh prior to Challis volcanism.

1993, 1994; Tysdal and Moye, 1996) (Fig. 24). An east-tilted Paleogene half graben formed in the hanging wall (Janecke, 1994). From Allison Creek (26 miles from Challis) north to McKim Creek (29 miles from Challis) the greenish dacite ash-flow tuff of Ellis Creek on both sides of the river. This ashflow tuff is thick and widespread in the Challis volcanic field and its eruption produced the enormous Van Horn Peak cauldron complex (Fig. 4; McIntyre et al., 1983). The canyon of the river parallels the Salmon River fault and stays mainly in Challis volcanic rocks (Fig. 2).

North of McKim Creek the canyon enters again into Apple Creek Formation, including the Apple Creek diamictite unit, that contains cleaved argillite and siltite with pebbles of coarse siltite and coarse-grained metasandstone (Tysdal and Moye, 1996). The high Lemhi Range to the east contains steeply northeast-dipping, locally overturned beds of the Lemhi Group and Swauger Formation. These are repeated by southwest-dipping normal faults of pre-Challis age (Tysdal and Moye, 1996; Tysdal, 1996a and 1996b). South of Ringle Creek (34 miles from Challis) is a cliff of Apple Creek Formation, folded to a north-trending syncline, east of the road. To the north at Ringle Creek the Apple Creek is

unconformably overlain by dacite and basalt lava flows of the Challis Volcanic Group (Tysdal and Moye, 1996).

On the right about 35 miles from Challis, is Poison Creek. The lower 2.5 miles of Poison Creek cross Challis volcanic rocks in the hanging wall of the Salmon River (=Allison Creek) normal fault. In the footwall, the south-dipping Poison Creek thrust places the Mesoproterozoic Apple Creek Formation on Ordovician Saturday Mountain Formation and older rocks (Soregaroli, 1961; Tysdal and Moye, 1996; Tysdal, 1996a; Evans, 1999). This thrust is the westward continuation of the Hawley Creek thrust and may extend to the west of the Salmon River into the Rattlesnake Creek area (Starr, 1955; Landreth, 1964; Ekren, 1988; Skipp, 1987).

The road remains in Challis Group to about a mile north of the turnoff up Rattlesnake Creek to the Twin Peaks Guest Ranch (41 miles from Challis) where it again enters the Apple Creek Formation, dipping vertically, and making big talus cones east of the road. The 45th parallel is in Apple Creek Formation. The Challis Volcanic Group overlies it unconformably about one mile to the north.

At 12 mile road (50 miles from Challis) the rocks are Challis volcanics, but then the road enters Apple Creek Formation again, forming very steeply dipping flatirons. The last bedrock is reached near 10 Mile Creek. From here north to Salmon, the rocks are northeast-dipping Challis volcanics and overlying gently folded Tertiary sedimentary rocks of the Salmon basin (Ruppel et al., 1993). See Roadlog #3 for a description of this Paleogene rift basin.

Salmon Valley

Fifty-five miles from Challis is the Williams Creek road on the west. This main Forest Service road affords access to the Panther Creek area (Janecke et al., 1997), including the Meridian Gold Corporation BearTrack gold mine.

Coming into the Salmon Valley the Beaverhead Mountains loom to the northeast, in the footwall of the Salmon Basin detachment (Figs. 35 and 37). These peaks are made of Lemhi Group and Swauger Formation of the Belt Supergroup (Umpleby, 1913; Anderson, 1956; 1961; Tucker, 1975; Winston and Link, 1993). Tertiary lacustrine and fluvial beds occupy the low country east of Salmon, with Challis volcanics west and south of Williams Creek and in the northern Lemhi Range. Northwest of there is Yellowjacket Formation intruded by 1370 Ma granite (Evans and Zartman, 1990; Ruppel et al., 1993; Doughty and Chamberlain, 1996).

Panther Creek Half-Graben

The geology of the Challis Volcanic Group in the Panther Creek half graben has recently been reviewed by Janecke et al. (1997). They conclude that the half graben formed between about 47.7 and 44.5 Ma, during which time as much as 6.5 km of rocks were deposited. Rates of volcanic activity kept up with basin subsidence, so that the basin was mainly filled with ash-flow tuffs. Post-volcanic conglomerate and reworked ash make up a small part of the fill of the half graben.

Salmon to North Fork

Reset odometer to zero at intersection of US Highway 93 and Idaho Highway 28 in Salmon. Head north on 93. North of Salmon the road cuts through Tertiary rocks until just north of Tower Creek (11 miles from Salmon). There the road and river enter a canyon cut in siltites of Apple Creek (Yellowjacket) Formation, with Tertiary volcanic rocks and conglomerate unconformably overlying it. Harrison (1985) showed that these conglomerates had a source to the west in the Salmon River Mountains and interfinger to the east and south with lacustrine deposits. Excellent exposures of cobble to boulder conglomerates of the Kriley Gulch Formation (the proximal Tertiary basin fill) in the vicinity of Tower Creek show that the unit depositionally overlies the Apple Creek Formation (Harrison, 1985). This, and the irregular trace of the contact between the Tertiary and Mesoproterozoic rocks to the south suggests that the western boundary of the Salmon basin is an angular unconformity. Most previous workers mapped a major normal fault in this location (Ruppel et al., 1993; Harrison, 1985). The road remains in Yellowjacket (likely Apple Creek, Winston et al., 1999, this volume) Formation to North Fork, where the Salmon River heads west across the River of No Return Wilderness.

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