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# Cenozoic Stratigraphy and Tectonic Evolution of the Raft River Basin, Idaho

by

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## ABSTRACT

The Raft River basin, a north-trending structure in the Basin and Range province south of the Snake River Plain, Idaho, is filled with upper Cenozoic sediments to a depth of about 1,700 meters. The basin fill is mainly the Salt Lake Formation of Miocene to Pliocene age, which consists of four members: (1) a lower tuffaceous member of shale, siltstone, and sandstone; (2) a wedge of calc-alkalic rhyolite flows with associated volcanic domes here named the Jim Sage Volcanic Member; (3) an upper tuffaceous and conglomeratic member; and (4) olivine basalt flows in the northern Cotterel Mountains. Isotope dating of volcanic rocks and paleomagnetic data reveal that the age of the Salt Lake Formation here ranges from more than 11 to about 2-3 million years, with nearly continuous deposition in the center of the basin. In Quaternary time, basalts erupted from vents in the northern part of the basin, and main stream alluvial deposits, alluvial fans, and successive loess blankets indicate changes in Pleistocene climate. Subsurface and geophysical studies show that the basin fill is considerably faulted, yet it rests on a low-relief surface of lower Paleozoic(?) and Precambrian metamorphic rocks. The Cenozoic structure of the basin is primarily the result of mid-Tertiary uplift of gneiss domes in the Albion Mountains west of the basin. Uplift gave rise to low-angle detachment faults involving transport of large masses of upper Paleozoic rocks eastward across the low-relief surface. Numerous listric faults, their lower parts merging with the detachment surface, and subsidiary fractures formed in the basin fill, and these became the conduits for the southern Raft River Valley geothermal system, although they are of insufficient depth to account for the high water temperatures observed.

## INTRODUCTION

The purpose of this report is to summarize the Cenozoic stratigraphy and structure of the Raft River basin, as it is now understood in the light of new data, and to present some of the reinterpretations of geologic history and basin geometry that form the basis for reports now in progress. We conclude that deposition in the basin was nearly continuous from early Miocene time and that gravity faulting related to uplift of mid-Tertiary metamorphic core complexes (gneiss domes) is the dominant tectonic mode.

The Raft River Valley occupies a north-trending intermontane tectonic basin about 60 kilometers long and 20 to 24 kilometers wide, with an average valley floor elevation of about 1,400 meters (Figure 1). The valley opens northward onto the broad Snake River Plain. On the west, south, and east sides, the basin is flanked by mountain ranges of varied heights. On the west flank, the Cotterel and Jim Sage Mountains, making up the Malta Range, rise to maximum elevations of about 2,500 meters. Directly west of the Cotterel and Jim Sage Mountains, and separated from them by a narrow valley, lie the Albion Mountains (3,000 meters). South of the basin are the Raft River Mountains (3,000 meters), one of the few east-trending mountain ranges in the North American Cordillera. On the east side of the basin are the Sublett and Black Pine Ranges (2,000 meters and 2,900 meters).

The geology of the basin and surrounding ranges was first mapped by Anderson (1931), who described rocks ranging in age from Precambrian to Quaternary. The U. S. Geological Survey has carried out several ground-water studies of the Snake River Plain region (Russel, 1902; Stearns and others, 1938) and of the Raft River basin (Nace and others, 1961; Mundorff and Sisco, 1963; Walker and others, 1970).

Designation in 1971 of the Raft River KGRA (Known Geothermal Resource Area) (Godwin and others, 1971) and the discovery in 1975 of a major intermediate temperature, liquid-dominated geothermal

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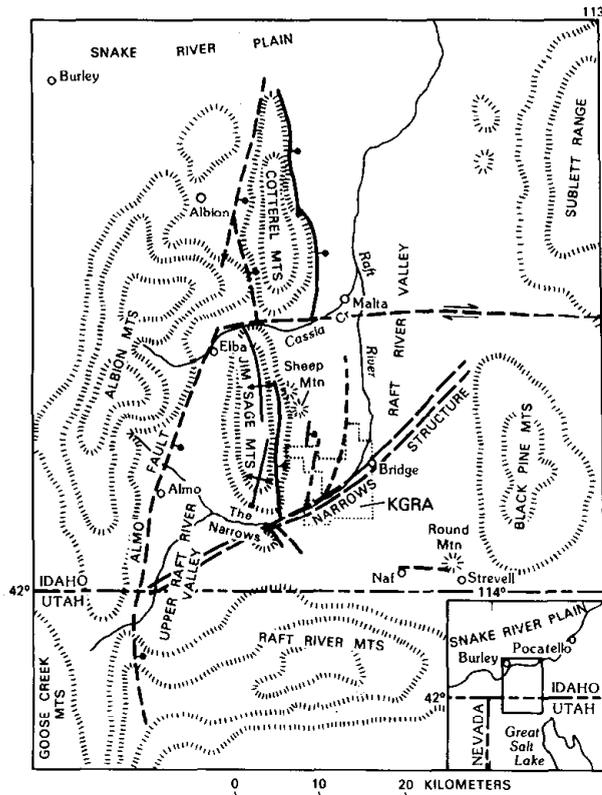


Figure 1. Index map of Raft River region showing Narrows structure (NE-trending dashed lines), structures affecting alluvium (short-dashed lines), transcurrent fault (east-west dashed line), and normal faults (lines with bar and ball showing downthrown side).

resource led to the multidisciplinary investigations in the area, including geologic mapping (Williams and others, 1974; Pierce and others, in press) and exploration with various geophysical techniques. Results of these studies were summarized early by Williams and others (1976) and later by Mabey and others (1978a). Subsurface studies incorporating results from both intermediate (about 300 meters) and deep (about 1,800 meters) boreholes are described by Crosthwaite (1976), Covington (1977a, 1977b, 1977c, 1977d, 1978, 1979a, 1979b), and Oriel and others (1978).

## GEOLOGIC SETTING

The Raft River basin lies in the northeast part of the Basin and Range province just south of the Snake River Plain and about 100 kilometers west of the Overthrust Belt of the northern Rocky Mountains. It is within the Cordilleran thermotectonic anomaly (CTTA) (Eaton and others, 1976, 1978), which is

roughly coextensive with the Basin and Range province and which is interpreted as the product of large-scale crustal extension resulting in high heat flow, much volcanism throughout middle and late Cenozoic time, and basin-range block faulting (Stewart, 1978). In the Raft River region, the base of the magnetized crust (locus of the Curie isotherm for magnetite, about 580°C) is 12 kilometers below sea level, a relatively shallow depth compared with other parts of the Snake River Plain region (Bhattacharyya and Mabey, 1980; Mabey, in press).

The depositional, tectonic, and thermal history of the Raft River region is long and complex, and it has been the subject of considerable study (Armstrong, 1968, 1970, 1976, unpublished data; Compton, 1972, 1975; Compton and others, 1977; Miller and others, 1980; Smith, 1982). In brief, a late Precambrian to Triassic miogeoclinal succession about 12,000 meters thick was deposited on a 2.5 billion-year-old Archean basement of adamellite and subordinate schist, gneiss, and amphibolite. These younger rocks are grouped as the Raft River assemblage (R. L. Armstrong, unpublished data). High-pressure load metamorphism during Jurassic time was followed by thrusting of the Sevier orogeny in Cretaceous time, during which a late Precambrian-Paleozoic allochthon (quartzite assemblage of R. L. Armstrong, unpublished data) was thrust eastward over an autochthon and parautochthon consisting of the Raft River assemblage. Great allochthons of upper Paleozoic rocks were subsequently thrust eastward over older rocks. An early to middle Tertiary rise in thermal activity produced plutons at depth, gneiss domes in the Albion and Raft River Mountains, and considerable attenuation of the sedimentary cover on the domes. Late Tertiary events, largely the result of crustal extension, include folding and faulting to form the present mountains and sediment-filled basins and local rhyolitic volcanism. Eruption of the Jim Sage Volcanic Member coincided with the north-eastward passage of the thermal pulse now under Yellowstone National Park (Christiansen and Lipman, 1972). During the Quaternary, extensional tectonics probably continued, and flood basalts of the Snake River Group were erupted from centers at the north end of the Raft River Valley. Alluvial deposits of main and tributary streams, alluvial fan gravels on piedmont slopes, and successive loess blankets record cyclic climatic changes during Pleistocene time.

## CENOZOIC STRATIGRAPHY

The Raft River basin is filled with Cenozoic rocks

to a maximum known depth, based on deep drilling, of about 1,700 meters. Because of structural complications, however, this thickness only approximates the actual depositional thickness, and rocks penetrated successively in drilling are not necessarily a stratigraphic succession (Covington, 1977a, 1977b, 1977c, 1977d, 1978, 1979a, 1979b, 1980; Pierce and others, in press). Accordingly, the stratigraphic sequence as presently known is fragmentary and is deduced from exposed sections and from well records from which only the existence, but not the magnitude, of folding and faulting can be confidently inferred.

## SALT LAKE FORMATION

The Salt Lake Formation (Hayden, 1869; Mansfield, 1920; Walker and others, 1970) rests on Precambrian rocks in the area of the KGRA and at least as far north as Malta; in the southeastern part of the basin, near Strevell, it rests on Paleozoic rocks. In the KGRA and probably elsewhere, the Tertiary-Precambrian contact is a detachment surface related to growth faulting along listric fault surfaces (Covington, 1980; Pierce and others, in press).

In the Raft River basin, the Salt Lake Formation consists of four principal members: lower and upper tuffaceous members; an intervening wedge of rhyolite flows (Jim Sage Volcanic Member) in the western part of the basin; and the basalt of the northern Cotterel Mountains (Figure 3).

### Lower Tuffaceous Member

The lower tuffaceous member is exposed in the southern half of the Jim Sage Mountains, mainly in the upper valleys of Chokecherry and Cottonwood Creeks and in the bluffs both north and south of the Raft River Narrows. A measured section at the top of the member near Dry Canyon in the southern Jim Sage Mountains consists of 130 meters of gray, poorly consolidated, massive to thin-bedded, tuffaceous sandstone and tuff, and sandy tuffaceous shale and siltstone. Alternating units of sandstone and shale-siltstone are about 10 to 20 meters thick. Thin beds of grit and fine conglomerate are locally present. The lower part of the member, as deduced from well cuttings, consists largely of light brown siltstone alternating with light green tuff and tuffaceous sandstone in units 10 to 30 meters thick (Covington, 1977a, 1977b, 1977c, 1977d, 1978, 1979a, 1979b). Both green and brown rocks are commonly calcareous. Total thickness of the lower tuffaceous member is about 500 meters in the Raft River basin, but it is probably 1,000 meters or more in the upper Raft River basin southwest of the Narrows.

### Jim Sage Volcanic Member

The Jim Sage Volcanic Member is formally proposed herein for a thick wedge of rhyolite flows with subordinate bedded tuff and ash-flow tuff that were deposited on the lower tuffaceous member along the western margin of the Raft River basin. The type locality is in the southern Jim Sage Mountains, approximately 42°05'-42°08' lat, 113°28'-113°32' long. This member forms most of the Jim Sage and Cotterel Mountains and consists of three parts: the lower and upper successions of rhyolite flows and a middle unit of varied lithology. Maximum thickness is about 1,200 meters.

The lower and upper rhyolite flows are similar and consist mostly of dark gray to black, glassy porphyritic rhyolite that weathers dark reddish brown. Individual flows range in thickness from 20 to 200 meters; thicker flows commonly have a light gray devitrified zone in the middle part, enclosed in an envelope of glass. The rock is commonly flow banded, and columnar joints, square in cross section, are well developed. In the field, the lower flows can be distinguished by their reversed magnetic polarity from the upper flows, which have normal polarity.

The middle unit is 30 to 100 meters thick and consists mostly of a distinctive vitrophyre breccia made up of nonsorted particles of angular to subrounded, gray, glassy rhyolite ranging in size from coarse sand to 1-meter blocks set in a matrix of either yellow to orange palagonitic hydrated glass or white to gray tuff. Thin rhyolite flows are locally present. Fine-grained, probably in part lacustrine, sediments above and below the breccia suggest that it formed as an explosion breccia produced by rhyolite flowing into a body of water. In the southern Cotterel Mountains, the breccia is overlain by two thin, vitric rhyolite ash-flow tuffs that were erupted from sources to the east. The tuffs are overlain by a few meters of white to gray tuffaceous sandstone and siltstone.

Volcanic domes that consist of rhyolite similar to the lower and upper rhyolite flows occur in three areas in the Raft River basin (Td, Figure 2): in the southernmost Cotterel Range, just north of Cassia Creek; at Sheep Mountain, east of the central Jim Sage Mountains; and along the southern margin of the basin north of the Raft River Mountains and west of the Black Pine Range. The most prominent of the domes is Sheep Mountain, a roughly conical steep-sided hill about 1 kilometer in diameter that rises 170 meters above the alluvial-fan surface east of the Jim Sage Mountains. The dome consists of an outer shell 10 to 20 meters thick of brown devitrified rhyolite that is generally brecciated and seamed with chalcidony veinlets; the shell encloses a core of gray glassy porphyritic rhyolite. The rhyolite clearly in-

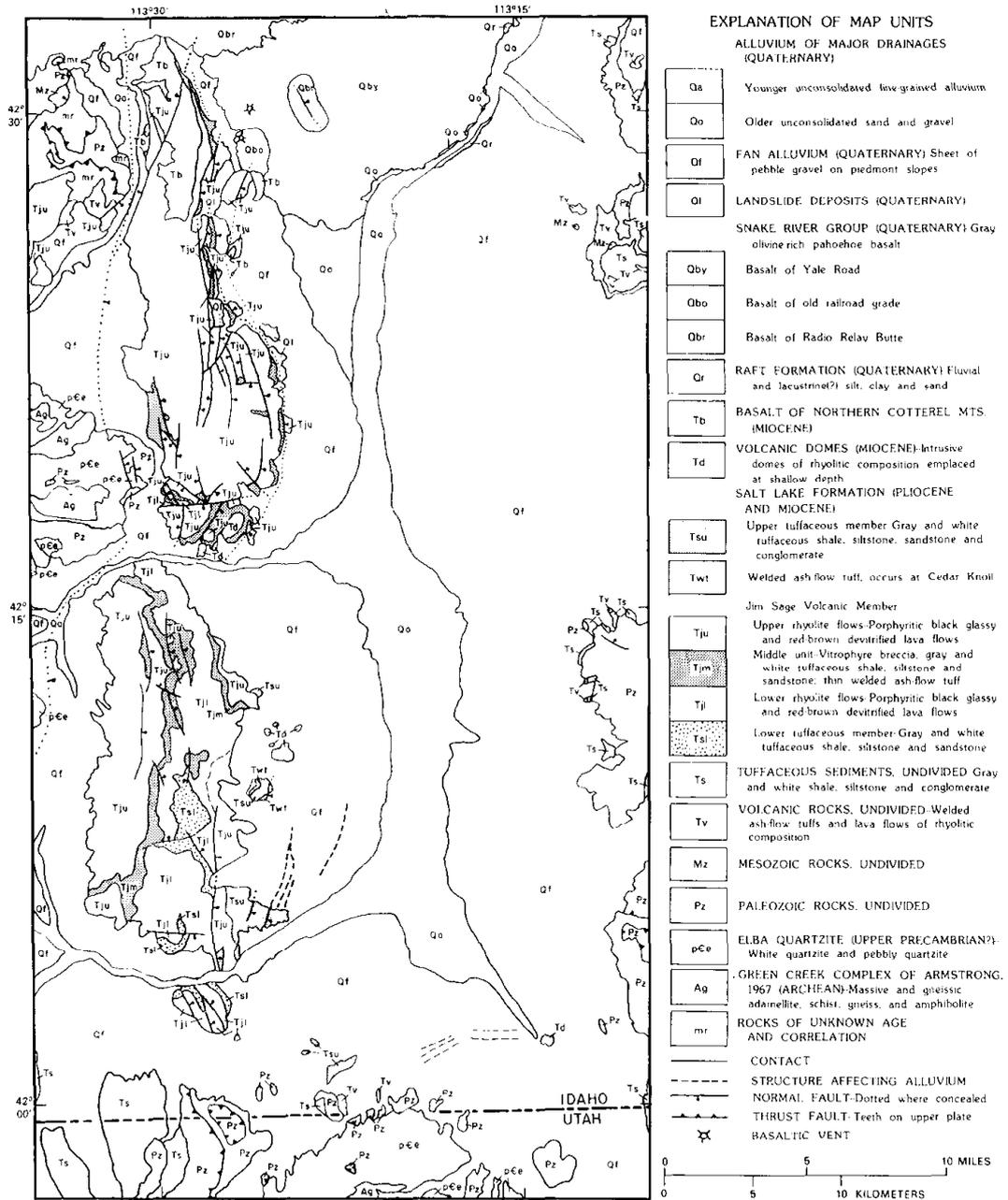


Figure 2. Geologic map of the Raft River basin, Cassia County, Idaho (after Armstrong and others, 1978; Compton, 1972, 1975; Williams and others, 1974; Pierce and others, in press).

trudes the upper tuffaceous member of the Salt Lake Formation; a thin dikelike extension of the shell surrounded by altered tuff is exposed at the east end of Sheep Mountain. At the southern end of the Raft River basin, several small exposures of rhyolite occur along an east-northeast-trending line that appears to be fault controlled.

The rhyolites and volcanic domes of the Jim Sage Member are similar petrographically. Pheno-

cryst content ranges from 6 to 23 percent and averages 15 percent. The phenocrysts are mostly plagioclase (An<sub>29-33</sub>), with 1 to 2 percent augite and pigeonite (hypersthene occurs in the uppermost thick flow), 0.5 percent opaque minerals, 0.5 percent quartz, and rare sanidine. Microscopically, glassy rocks commonly have a well-developed perlitic crack pattern and radial spherulites that indicate incipient devitrification; the groundmass of gray devitrified rock is a

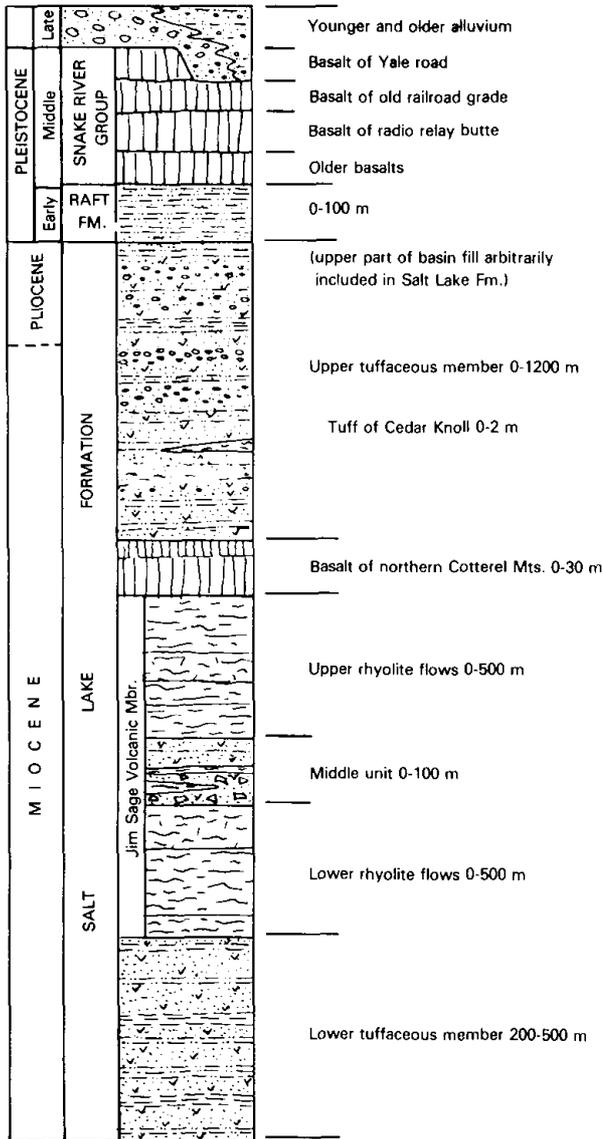


Figure 3. Composite stratigraphic column for Cenozoic rocks, Raft River basin, Idaho.

mosaic of quartz and feldspar. Lithophysal cavities lined with tridymite and feldspar were found only in the thick, uppermost flow of the Jim Sage Volcanic Member and indicate minor vapor-phase activity.

The rhyolites are similar chemically as well as petrographically. Analyses from upper and lower flows and the volcanic dome have the average composition shown in Table 1; deviation from the average is slight. This composition is close to that of the average calc-alkalic rhyolites of Nockolds (1954), but the Jim Sage Volcanic Member contains more total Fe, MgO, CaO, and TiO<sub>2</sub>, and less of the alkalis and silica.

The basalt of the northern Cotterel Mountains

(Figure 3) is the oldest basalt in the Raft River region and consists of two flows that cap the north end of the west-dipping cuesta that forms the range. A single flow is exposed in fault blocks along the northeast and northwest flanks of the range. The rock is gray to light gray with a reddish oxidation tint and contains glomeroporphyritic clots of olivine and plagioclase in a dense groundmass of fine-grained plagioclase, olivine, pyroxene, opaque minerals, and glass (Pierce and others, in press). Radiometric ages (Figure 4) are in agreement with its conformable position overlying the upper flows of the Jim Sage Volcanic Member.

### Upper Tuffaceous Member

The upper tuffaceous member of the Salt Lake Formation is poorly exposed in the Raft River basin and is known mostly from the subsurface. The lower 600 meters or so consists of gray tuffaceous sandstone and tuff and of white, gray, and light green thin-bedded to papery shale and tuffaceous siltstone. Thin rhyolite flows of the upper part of the Jim Sage Volcanic Member are intercalated with the basal beds of the member. East of the central Jim Sage Mountains the tuff of Cedar Knoll, a thin vitric ash-flow tuff, is intercalated with the bedded tuff. A similar but thicker intercalated crystal-poor ash-flow tuff, the tuff of upper Raft River Valley (R. L. Armstrong, unpublished data) forms a low cuesta about 8 kilometers west of the Narrows. The succeeding 500 meters or so of the member is similar in lithology but conglomeratic; most of the pebbles and sand grains were derived from the Jim Sage Volcanic Member. Carbonaceous material is sparsely present in the uppermost part of the member.

In a recent petrographic study, Devine and Bonnicksen (1980) noted that the principal lithologies in the tuffaceous members of the Salt Lake Formation are shale, siltstone, sandstone, and tuff. The sandstones are feldspathic graywackes and lithic wackes (Pettijohn and others, 1972) in which quartz, feldspar, rock fragments, and mica are the principal detrital components. The composition of the texturally immature sediments reflects the mixed provenance from the varied rock types in the mountain ranges surrounding the basin. Devine and Bonnicksen found the volcanic provenance dominant in the upper part of the basin fill, indicated by a high ratio of non-undulatory (volcanic) quartz to undulatory (metamorphic) quartz and by a high ratio of volcanic to metamorphic rock fragments; both ratios strongly decrease downward in the Salt Lake Formation. The downward decrease in average grain size of the basin fill, suggested by the absence of conglomerate in the lower part, is further indicated by a downward decrease in detrital quartz and feldspar and a corres-

ponding increase in nondetrital carbonate (Devine and Bonnicksen, 1980, p. 35).

### Age and Correlation

The Salt Lake Formation is dated as late Miocene, at least in part, by a clustering of radiometric ages between about 7 and 11 million years for the Jim Sage Volcanic Member, by the basalt of the northern Cotterel Mountains, and by the tuff of Cedar Knoll (Figure 4 and Table 2). The oldest dates, from the upper rhyolite flows of the Jim Sage Volcanic Member, indicate an age of 9 to 11 million years; the conformable basalt of the northern Cotterel Mountains is about the same age. The volcanic domes are about 7 to 9 million years old. The ash-flow tuffs in the middle unit of the Jim Sage in the southern Cotterel Mountains were not dated but appear to be the same as tuffs in the Sublett Range, dated by Armstrong at between 8 and 11 million years. A similar date, 8.8 million years, obtained for the tuff of upper Raft River Valley (G. B. Dalrymple, written communication, 1980) suggests that the tuffs in the three areas are the same unit; they are lithologically very similar and preliminary petrographic data confirm the correlation. The youngest radiometric date for the Salt Lake Formation is that of 7 million years for the tuff of Cedar Knoll, although a large analytical uncertainty permits a range of 5 to 9 million years. The tuff of Cedar Knoll is lithologically similar to the Walcott Tuff (Carr and Trimble, 1963; Trimble and Carr, 1976) which has potassium-argon ages of about 6 million years (Marvin and others, 1970; Armstrong and others, 1975).

Preliminary findings from recent paleomagnetic studies by S. L. Bressler (written communication, 1979) permit estimates of the rate of sedimentation in the basin. The magnetic polarity of a suite of samples from intermediate-depth core-hole no. 3, total depth

408 meters (Crosthwaite, 1976), probably indicates a good correlation with the standard late Cenozoic polarity scale and apparently uniform sedimentation rates for the interval represented. Rocks from the bottom of the well apparently date from the top of the Gauss normal epoch, about 2.6 million years, which would indicate an average sedimentation rate of about 160 meters per million years. If sedimentation rates did not vary greatly, the basal sediments in the basin could thus be about 10 to 11 million years old. However, the finer grained rocks in the lower part of the Salt Lake Formation probably accumulated more slowly than the coarser clastics of the upper tuffaceous member, implying a somewhat greater age for the basin, perhaps extending to early Miocene or late Oligocene time.

Although a detailed regional correlation of units is not within the scope of this report, it is worth noting that recently published radiometric dates of volcanic rocks (Marvin and others, 1970; Armstrong and others, 1975, 1980; Williams and others, 1976) have been extremely useful in clarifying the regional stratigraphy. Thus, the Starlight Formation, the Walcott Tuff, and perhaps the Little Creek Formation (Carr and Trimble, 1963; Trimble and Carr, 1976) are probably equivalent to the upper tuffaceous member of the Salt Lake Formation of the Raft River basin. In the Cassia Mountains west of the Albion Mountains, a thick succession of rhyolite ash-flow sheets 8.5 to 11.0 million years old (Armstrong and others, 1975, 1980) overlies the Beaverdam Formation of Axelrod (1964), which contains the Trapper Creek flora of middle Miocene age. The Beaverdam Formation is probably at least partly equivalent to the lower tuffaceous member of the Salt Lake Formation of the Raft River basin.

### YOUNGER CENOZOIC ROCKS

In the northern part of the Raft River Valley, adjacent to the Snake River Plain, basaltic volcanism occurred intermittently from late Miocene to middle Pleistocene time. Quaternary basalts form broad, low shields and thin, tabular flows that blocked and diverted the Raft River eastward. The basalts and surficial deposits, some of which formed as the result of basalt dams, are described by Pierce and others (in press).

#### Raft Formation

The Raft Formation (Carr and Trimble, 1963; Walker and others, 1970), originally the "Raft lake beds" (Stearns and others, 1938), consists of fine-grained lacustrine and fluvial deposits, principally

Table 1. Average chemical composition of the Jim Sage Volcanic Member. (Average of ten rapid-rock analyses by P. Elmore, U. S. Geological Survey, Denver, Colorado.)

	Weight Percent
SiO <sub>2</sub>	72.1
Al <sub>2</sub> O <sub>3</sub>	13.5
Fe <sub>2</sub> O <sub>3</sub>	2.0
FeO	1.8
MgO	0.6
CaO	2.0
Na <sub>2</sub> O	2.7
K <sub>2</sub> O	4.4
TiO <sub>2</sub>	0.6
P <sub>2</sub> O <sub>5</sub>	0.1

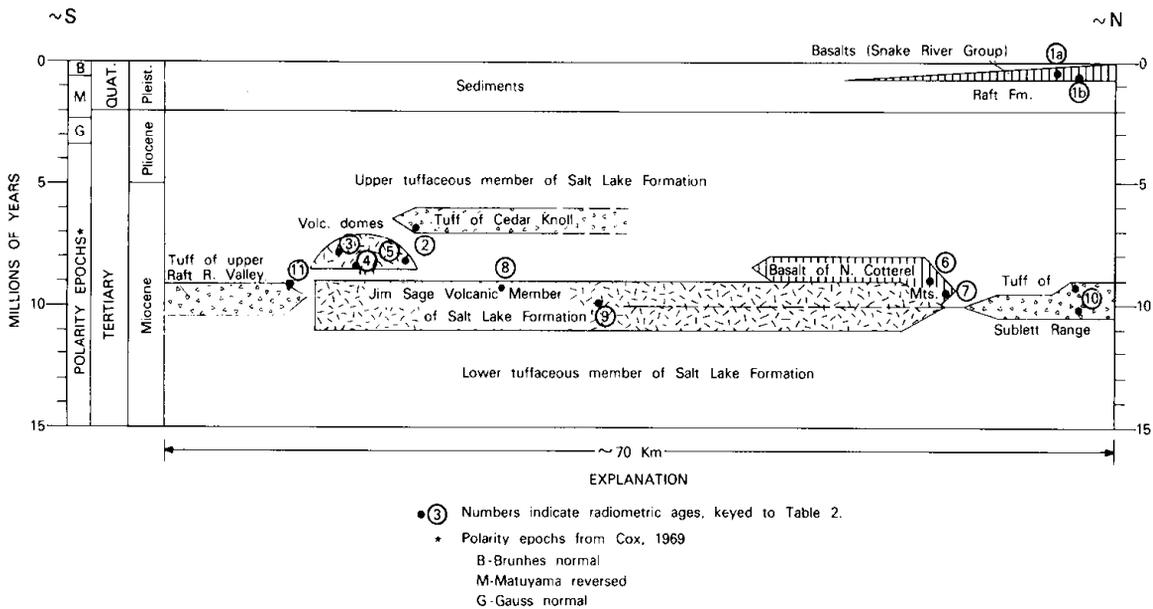


Figure 4. Chronostratigraphic diagram for Cenozoic rocks in the Raft River basin, Idaho.

clay, silt, and sand, that were deposited on the Salt Lake Formation in the Raft River basin. Deposition was probably in response to an elevation of base level by the emplacement of basalt flows of the Snake River Group some distance downstream from the mouth of the Raft River. The formation is not exposed in the southern part of the basin but is believed by Walker and others (1970, p. 24) to be up to 300 meters thick in drill holes. We did not recognize the Raft Formation either in outcrop or in subsurface in the southern part of the basin. In the northern part, the Raft Formation is poorly exposed along the Raft River, but it is well exposed in bluffs

40 meters high along the Snake River a few kilometers east of the mouth of the Raft River. Molluscs collected from localities in that area indicate a Pleistocene age for the Raft Formation (Trimble and Carr, 1976).

**Basalt**

Basalt flows mapped by Pierce and others (in press) as "older basalt" of probable early Pleistocene age occur as kipukas surrounded by younger lava just north of the Cotterel Mountains. The basalt occurs as pahoehoe lava and near-vent pyroclastic deposits and

Table 2. Sources of radiometric age dates (see Figure 4).

No.	Unit	Age, million years	Method	By	Reference
1a	Basalt of Radio Relay Butte	0.446 ± 0.220	K-Ar whole rock	G. B. Dalrymple	Pierce and others, in press
1b	-----do-----	0.709 ± 0.210	-----do-----	-----do-----	Do.
2	Tuff of Cedar Knoll	7.0 ± 2.0	Fission track, zircon	C. W. Naeser	Williams and others, 1976
3	Sheep Mountain volcanic dome	7.8 ± 1.1	-----do-----	-----do-----	Do.
4	-----do-----	8.42 ± 0.2	K-Ar Plagioclase	J. D. Obradovich	Do.
5	Round Mountain volcanic dome	8.3 ± 1.7	Fission track, zircon	C. W. Naeser	Do.
6	Basalt of North Cotterel Mountains	9.2 ± 1.5	K-Ar, whole rock	R. L. Armstrong	Armstrong and others, 1975
7	-----do-----	9.81 ± 1.02	-----do-----	G. B. Dalrymple	Pierce and others, in press
8	Upper rhyolite, Jim Sage Volcanic Member	9.2 ± 0.5	-----do-----	R. L. Armstrong	Armstrong and others, 1975
9	-----do-----	9.4 ± 1.6	Fission track, zircon	C. W. Naeser	Williams and others, 1976
10a	Tuff of Sublett Range	10.4 ± 0.14	K-Ar whole rock	R. L. Armstrong	Armstrong and others, 1975
10b	-----do-----	8.2 ± 0.2	-----do-----	-----do-----	Do.
10c	-----do-----	10.8 ± 0.2	K-Ar, biotite	-----do-----	Do.
10d	-----do-----	9.4 ± 0.14	K-Ar, feldspar	-----do-----	Do.
11	Tuff of upper Raft River Valley	8.8 ± 0.188	K-Ar, plagioclase	G. B. Dalrymple	Written communication, 1980

consists of gray to reddish banded rock with 1 to 2 percent phenocrysts of olivine and plagioclase.

The basalt of Radio Relay Butte (Pierce and others, in press) forms an extensive series of loess-mantled flows across the north end of the basin and makes up Horse Butte, a prominent kipuka (but not a basalt vent) east of the Cotterel Mountains. The basalt is gray, dense, and fine grained, with 1 to 3 percent small phenocrysts of olivine and plagioclase. Two potassium-argon ages (Table 2) of  $0.709 \pm 0.210$  and  $0.446 \pm 0.220$  million years indicate a middle Pleistocene age. Normal magnetic polarity coupled with the potassium-argon dates indicates an age of less than 0.73 million years (Pierce and others, in press).

The basalt of the old railroad grade (Pierce and others, in press) forms pahoehoe flows of limited extent that were erupted from a small vent 3 kilometers southwest of Horse Butte. The basalt is gray, porous, and finely porphyritic, with 2 to 3 percent phenocrysts of olivine 0.2 to 1.0 millimeter in diameter and plagioclase 0.5 to 1.0 millimeter long.

The youngest basalt in the area is the basalt of Yale Road, which was erupted from a vent marked by a fresh-looking crater 2 kilometers west of Horse Butte. Thin but extensive lava flows spread eastward and diverted the course of the Raft River northeastward to its present position. The basalt is light gray, porous, and, like the other basalts, porphyritic with 1.5 to 3.0 percent phenocrysts of olivine, 0.3 to 1.0 millimeter in diameter, and plagioclase laths 0.5 to 1.0 millimeter long. The groundmass consists on the average of about 35 percent plagioclase laths, 15 percent olivine, 25 percent augite, 12 percent opaque minerals, and 12 percent glassy matrix. Groundmass textures range from ophitic to hyalo-ophitic and hyalopilitic.

Both the basalt of the old railroad grade and the basalt of Yale Road have normal magnetic polarity, as measured in the field by a fluxgate magnetometer. They are assigned a middle Pleistocene age (Pierce and others, in press).

## UNCONSOLIDATED SURFICIAL DEPOSITS

Unconsolidated surficial deposits that consist of fan gravels, main stream gravels thickly mantled by fine sediments, and loess cover most of the area shown in Figure 2. Except for the fine sediment along the flood plains of the main drainages, nearly all these surficial materials are Pleistocene in age, deposited under climatic conditions different from the present (Pierce and Scott, 1982 this volume).

## Fan Gravels

Gravelly alluvial fans occur at the base of the ranges bordering the Raft River basin. The clasts in these fans are mostly rhyolite on the west side of the basin, mostly carbonate rocks on the east side, and a mixture of granitic, quartzitic, and carbonate rock on the south side. The fan gravels typically display stone-on-stone fabric, with spaces between stones being either open or filled with a matrix of relatively coarse sand. They are remarkably well washed, containing only 3 to 5 percent silt and clay (Pierce and Scott, 1982 this volume; Pierce and others, in press). Most beds are 0.2 to 0.5 meter thick and more than 5 meters in horizontal extent. Fewer beds are lens shaped, representing channel fills a few meters across and half a meter deep.

Different ages of fan gravels are readily apparent from sequence relations evident on aerial photographs; the age relations between adjacent fan deposits can generally be determined from the drainage pattern on the fans and the relative degree of dissection. For an alluvial-fan sequence deposited by a given stream, as many as eight subdivisions based on age can locally be mapped (Williams and others, 1974; note caret symbols). Age grouping of similar-age fan segments throughout the Raft River basin presents problems because different relative-age criteria need to be used in different parts of the basin. Where thick loess has accumulated on the southern and eastern sides of the valley, the number of loess units on a fan deposit indicates increasing age; on fans where the mantling loess contains one buried soil more strongly developed than the surface soil, the underlying gravels are probably at least 130,000 years old. On the west side of the valley, where loess is thin, ages of fan segments are based on relative-age criteria such as soil development, thickness of carbonate coatings on stones, and degree of fan dissection by drainages that head on the fan (Pierce and others, in press).

Along each drainage that heads well back in a range, a relatively young pie-shaped segment of fan gravel can be identified by (1) little dissection by small drainages on the fan, (2) almost perfect conical form as shown by contours on topographic maps, (3) loess mantle only about 10 to 30 centimeters thick, and (4) weak soil development generally without an argillic B horizon and without any carbonate cementation. The largest such young fan in the Raft River basin is the Meadow Creek-Warm Creek fan which covers about 130 square kilometers north of the Black Pine Range. Based on soil development and relation to Holocene deposits, the gravels forming these relatively young fan segments are estimated to be between 25,000 and 10,000 years old, or late

Wisconsin in age.

Fan deposits of Holocene age are much more limited in extent and were deposited by streams with much less competence than the Pleistocene streams. Fans of poorly bedded silt occur along the margin of Raft River bottoms; they consist mostly of reworked loess. Holocene gravelly deposits occur at the heads of the fans or within the ranges. These deposits are commonly either silty gravel or lobate pods of gravel without a matrix.

### Main Stream Deposits

Deposits of fine-grained Holocene alluvium floor the bottom lands along the perennial streams that cross the basin. This alluvium ranges in thickness from 3 to 5 meters along the Raft River to about 1 meter along Cassia Creek. It is mostly silt, with interbeds of sand, clayey silt, and poorly sorted gravel. Dark gray humic layers are common. Two carbon-14 ages of  $8,370 \pm 250$  and  $7,720 \pm 250$  years were obtained from the lower part of this unit in the southern Raft River basin (Pierce and others, in press). A volcanic ash bed about 2 centimeters thick from about 1.7 meters below the surface of this unit near Bridge has properties similar to the Mazama ash, which is about 6,600 years old (R. E. Wilcox, written communication, 1974). Based on these ages and the stratigraphic relations to upper Pleistocene gravel deposits, this fine-grained alluvium is Holocene in age.

Beneath the fine-grained Holocene alluvium is a well-sorted gravel, which is commonly found in water wells. Exposures of this gravel are scarce, but an excavation 4 meters deep near Bridge did expose 3 meters of fine-grained alluvium overlying well-rounded gravel in an abundant matrix of sand. This gravel represents the main stream equivalent of the younger fan gravel and dates from a time when conditions permitted the entire drainage system to transport gravel. Although the streams transported gravel in late Pleistocene time, in Holocene time they were not competent to carry the entire load of silt through the drainage system, resulting in aggradation of muds along the major drainages (Pierce and Scott, 1982 this volume).

Older deposits of main stream gravel occur in the northern Raft River Valley beneath, adjacent to, and south of the basalt of Yale Road (Figure 2). These gravels are moderately well washed, with only 2 percent silt and clay. Compared with the fan gravels, this main stream gravel generally does not display openwork textures and has about twice as much medium and fine sand (Pierce and others, in press).

The basalt of Yale Road dammed the Raft River and diverted it northeastward to its present course

around the edge of the flow. By studying water-well records the old channel and its gravels can be traced beneath the basalt of Yale Road for about 25 kilometers (Pierce and others, unpublished data). The Raft River aggraded, forming the large, nearly horizontal plain east of Idaho after emplacement of the flow across its path.

### Loess Deposits

Loess consists of wind-blown silt, with smaller amounts of clay and sand. The geologic record in the Snake River Plain region reveals that there has been little wind erosion or loess deposition during Holocene time. However, as a result of disruption of the plant cover by man, modern dust storms erode and redeposit loess at rates that would be geologically quite significant if continued. The surface soil on the upper loess unit is similar in development to soils on Bonneville Flood deposits and younger fan gravels, which suggests both that the upper loess unit has been stable during Holocene time and that the last period of loess deposition occurred in late Pleistocene time. In the Raft River Valley, stratigraphic sequences with multiple loess units, separated from each other by buried soils, show that this two-fold sequence (loess deposition followed by soil development) has been repeated at least four times in about the last half-million years (Pierce and others, in press). Beneath the upper loess unit, a buried soil developed in loess occurs on the basalt of Yale Road, the basalt of the old railroad grade, and the older gravels of the Raft River. This next older loess unit is considered to be about 150,000 years old, for reasons explained in Pierce and others (1982 this volume, Figure 2, column 10).

Loess thickness varies drastically in the Raft River basin and can be related to the following inferred late Pleistocene conditions: (1) westerly winds, (2) large topographic features upwind, (3) local and regional loess sources that at least, in part, were floodplains of Pleistocene rivers, and (4) age of the underlying deposits.

Loess is generally more than 1 to 2 meters thick on most alluvial fans on the eastern side and less than half a meter thick on the western side of the Raft River basin. On the alluvial fans from the Raft River Mountains at the south end of the basin, loess is generally more than 1 meter thick.

An exception to the pattern of thin loess on the west side of the basin occurs along the traces of structures just west of the KGRA. There the loess is commonly more than 1 meter thick and locally as much as 3 meters thick. The lush vegetation associated with springs and shallow ground water along these structures was apparently also present during

loess deposition and aided in trapping loess and preventing it from being eroded.

For the northern Raft River basin, Pierce and others (in press, Figures 1, 2, and 3) show details of the loess distribution as controlled by large- and small-scale topographic features and the prevailing late Pleistocene wind directions. There, at the margin of the Snake River Plain in the lee of the Cotterel Range, a loess drift accumulated in the Horse Butte area. This drift is mostly the upper loess unit and covers about 100 square kilometers to thicknesses of 12 meters.

## STRUCTURAL EVOLUTION OF THE BASIN

The geomorphic setting of the Raft River basin is that of a broad, flat-floored alluvial valley flanked by locally steep-fronted, north-trending mountain ranges. This setting evokes the traditional block-fault structural model for late Cenozoic tectonic activity in the Basin and Range province, long recognized as a region dominated by extensional tectonics. North-trending normal faults are numerous and conspicuous in the Cotterel and Jim Sage Mountains, implying block faulting; but detailed mapping of the area and deep borehole data now require reinterpretation of the structure, with important implications for the geothermal system and regional Cenozoic tectonics.

Mapping has not confirmed the existence of major normal faults along the range fronts (Figure 2). Faults are present on the east flank of the Cotterel Mountains, but offset on any fault is generally less than 200 meters (Pierce and others, in press). Mapping by R. L. Armstrong (unpublished data) in the Sublett Range and by Smith (1982) in the Black Pine Range confirm the absence of range-front faults on the east side of the basin, as first noted by Anderson (1931). Also, frontal faults have not been recognized along the south margin of the basin adjacent to the Raft River Mountains (Compton, 1972, 1975). Normal faults occur at the eastern flank of the southern Jim Sage Mountains and east of the range, but these faults have small displacement; the range is fundamentally a faulted anticline (Figure 2), the east flank of which dips 15 to 35 degrees toward the basin. Steep faults and numerous steep fractures cut the Salt Lake Formation in the KGRA, as observed in core-holes and limited outcrops.

Structures about 3 kilometers east of the Jim Sage Range look like fault scarps (Figure 6), but trenching across these features shows they are not. No offsets with the east side down were observed. The contact

between a unit of unfaulted bouldery alluvium and the underlying indurated greenish tuffaceous sand of the Salt Lake Formation dips 4 degrees for 50 meters across the structurally disturbed zone, whereas the gradient of the alluvial fan surface is about 1.5 degrees. This suggests that tilting across this zone accounts for about two-thirds the relief across it. The bottom of the trench exposed three vertical filled fissures about 10-30 centimeters wide cutting partially indurated sand of the Salt Lake Formation. Vertical offsets of about 10 centimeters occur across two of these fissures, each with the west, or mountain, side down. These filled fissures are locally associated with emergence of ground water in the trench bottom and appear aligned with the zones of higher, lush vegetation (Figure 6). The loess mantling the fan gravels thickens several times in the vicinity of the vertical fractures, apparently as a result of its being trapped by lush vegetation localized along this disturbed zone during times of loess deposition. This thickening of the loess accounts for the remainder of the relief, and for the several escarpments in the structurally disturbed zone (Figure 6). The age of these structures is middle Quaternary or older as indicated both by unaffected units in the upper part of the trench and along strike to the north. These trench exposures favor interpretation of this disturbed zone as formed by extension above a subhorizontal fault rather than a high-angle fault, although a genetic relationship between the zones and the numerous high-angle faults in the basin fill cannot be ruled out.

A steep, northeast-trending tectonic feature, the "Narrows structure" or "Narrows zone" (Figure 1), was postulated from indirect geologic and geophysical evidence (Williams and others, 1976; Mabey and others, 1978a). The nature of this feature is somewhat enigmatic; it lies along a regional northeast-trending magnetic zone interpreted to be a basement feature (Mabey and others, 1978b), and it forms the southern limit of gravity and magnetic highs in the western part of the basin (Mabey and others, 1978a). Yet seismic refraction studies (Ackerman, 1979) and deep seismic reflection profiles do not show this feature (Ackerman, written communication, 1980); its significance in the Cenozoic history is unknown.

Subsurface studies based on deep drilling in the KGRA showed that the basin fill, which is made up of tilted and faulted beds of the Salt Lake Formation, rests on a relatively flat, unbroken surface of early Paleozoic(?) and Precambrian rocks (Covington, 1977a, 1977b, 1977c, 1977d, 1978, 1979a, 1979b). No intervening upper Paleozoic rocks are present even though thick successions of these rocks occur in the surrounding ranges. Subsurface data also indicate many faults and fractures in the fill and the presence

of a tectonic breccia at the Tertiary-Precambrian contact. Mabey and others (1978a) postulated low-angle gravity faulting from the Jim Sage Mountains toward the Raft River basin, with the principal glide plane being the Salt Lake Formation-Precambrian boundary and with displacements of up to 3 kilometers. Seismic reflection profiles across the southern part of the basin in the area of the KGRA confirm the structural complexity of the basin fill, the absence of offset of the basement surface, and the presence of low-angle normal, listric faults. Major low-angle faulting of late Tertiary age is also known to have occurred within the Salt Lake Formation east of the Grouse Creek Mountains (Todd, 1975; Compton and others, 1977; Miller and others, 1980).

Our concept of the late Cenozoic structural history of the Raft River basin that has evolved from surface mapping and geophysical and subsurface studies is one of gravity-induced movement along low-angle glide surfaces (Covington, 1980; Pierce and others, in press). About 20 to 40 million years ago, greatly increased crustal heat flow accompanied emplacement of gneiss domes in a stress field of strong regional extension, culminating in marked uplift of the mountains relative to the surrounding basins. A detachment surface (perhaps the plane of an earlier thrust fault) developed between masses of lower and upper Paleozoic rocks on the domes. The upper Paleozoic rock masses moved eastward on the detachment surface away from the rising gneiss domes, eventually forming the present Sublett and Black Pine Ranges. The ever-widening trough between the gneiss domes and the eastward-moving allochthonous block was filled with tuffaceous sediments; listric glide faults and low-angle faults developed in the fill. Figure 5 is a generalized, partly diagrammatic structure section

extending from the Jim Sage Mountains to the east side of the basin. Two sets of listric faults are portrayed; east-dipping faults are shown as generally younger than west-dipping ones, a relationship suggested by the seismic profiles. These faults and subsidiary fractures, some of which have moved in the Pleistocene and possess the potential for future movement, and the detachment surface at the bottom of the basin fill make up the complex plumbing of the geothermal system. It is unlikely, however, that the faults of this system are the primary conduits that introduce hot water from depth; none extends to depths that, with the ambient geothermal gradient, would supply water at temperatures as high as those observed (Williams and others, 1976).

The Jim Sage Volcanic Member, which forms the Cotterel and Jim Sage Mountains, was erupted only 9 to 10 million years ago into a long, narrow moat east of the Albion Mountains; it was subsequently displaced no more than a kilometer or so eastward along the detachment surface. The relations shown on the geologic map (Figure 2) suggest right-lateral offset, along inferred east-west-trending transcurrent faults, between the Cotterel and Jim Sage Mountains and between the hills of rhyolite north and south of the Raft River Narrows. Possibly the east-west structures mark steep boundaries between major units of the eastward-moving allochthonous mass. A similar suggestion of right-lateral offset between the Sublett and Black Pine Ranges is evident in Figure 1.

Our understanding of the Cenozoic tectonic history is still far from complete and is the subject of our continuing study. Current interpretations differ considerably from earlier concepts of late Cenozoic Basin-Range tectonics, but they are compatible with the structure schemes now evolving from studies in other

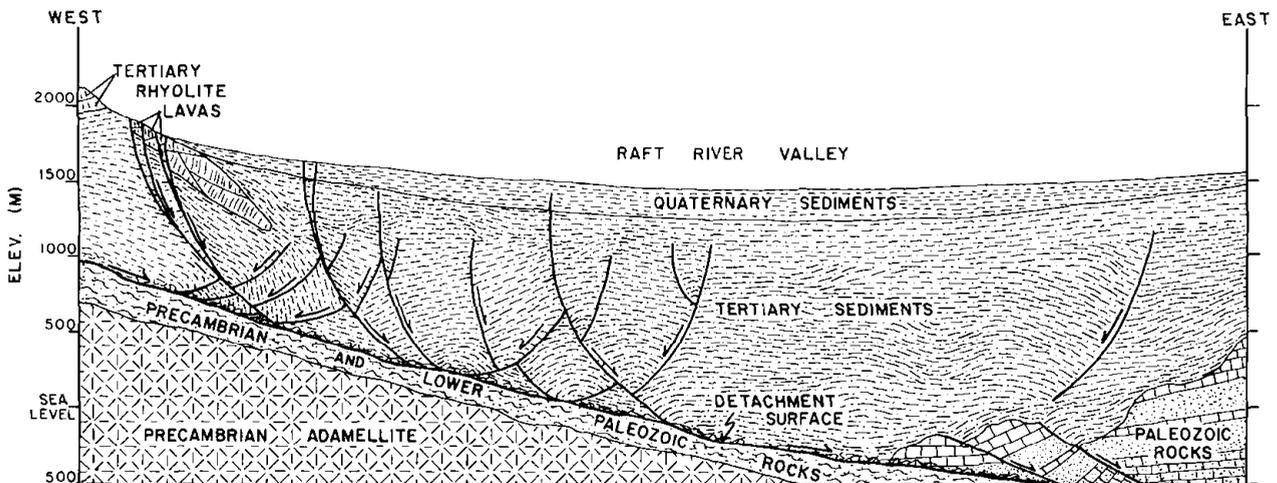


Figure 5. Generalized cross section of the Raft River basin, Idaho, showing interpretative subsurface geology and structure. Arrows show direction of fault movement.

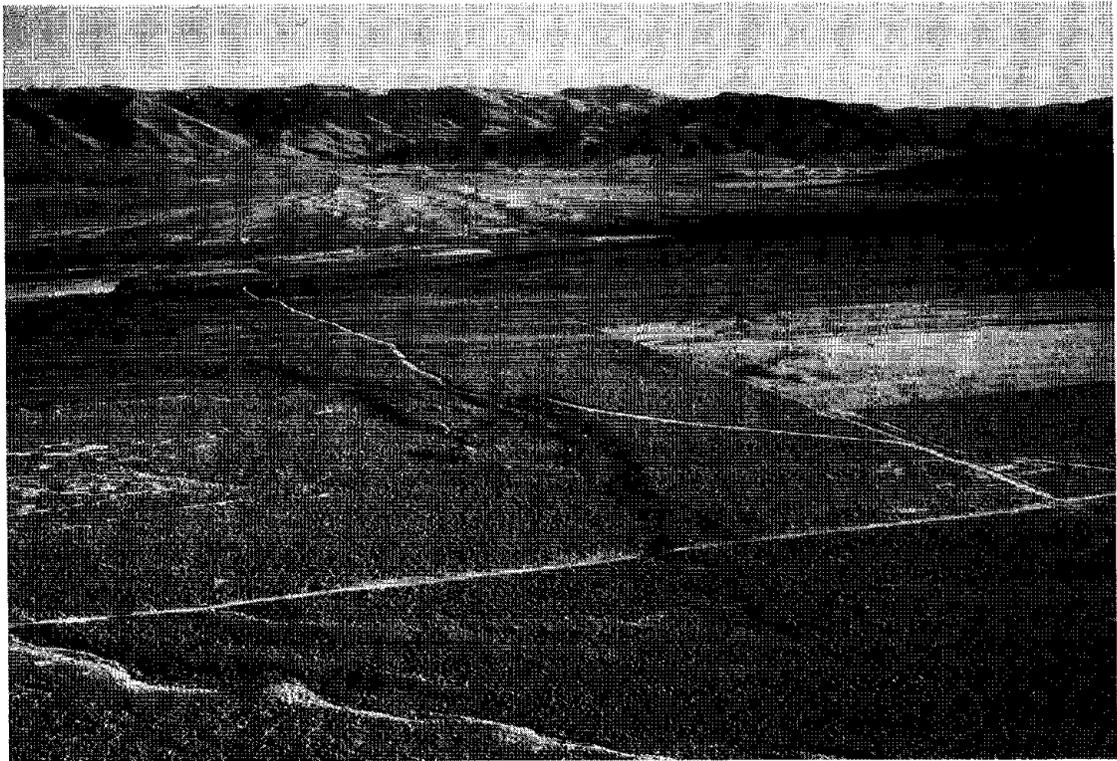


Figure 6. Oblique aerial photograph, looking south, of "Horse well" structurally disturbed zone about 3 kilometers east of the range front of the Jim Sage Mountains. See text for results of trenching studies across this feature. Alluvial-fan gravels on the trend of this structure (lower right) postdate the structural disturbance affecting the alluvium in middle of the photograph.

parts of the Basin and Range province (Davis and Coney, 1979; Crittenden and others, 1978; Howard, 1971; Wright and others, 1974).

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