Geological Survey Circular 838

Guides to Some Volcanic Terrances in Washington, Idaho, Oregon, and Northern California

HIGH LAVA PLAINS, BROTHERS FAULT ZONE TO HARNEY BASIN, OREGON

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GEOLOGIC SUMMARY

The field trip through the High Lava Plains province of central and south-central Oregon (fig. 1) provides a reconnaissance overview of the Cenozoic volcanic geology along the Brothers fault zone from Bend to Harney Basin, the site of vents for several ash-flow tuffs that cover thousands of square kilometers of southeast Oregon and have volumes of hundreds of cubic kilometers. Most of this area has been mapped only in reconnaissance (Walker and others, 1967; Greene and others, 1972), and many interesting details of the geology have yet to be studied. From east to west the geologic record generally progresses forward in time with some of the youngest volcanic rocks at the west (or Bend) end of the trip and the oldest at the east end in and near Harney Basin. This description is an oversimplification with respect to the basaltic rocks, but the silicic rocks show a well-defined age progression in silicic volcanism described by Walker (1974) and MacLeod, and others, (1976).

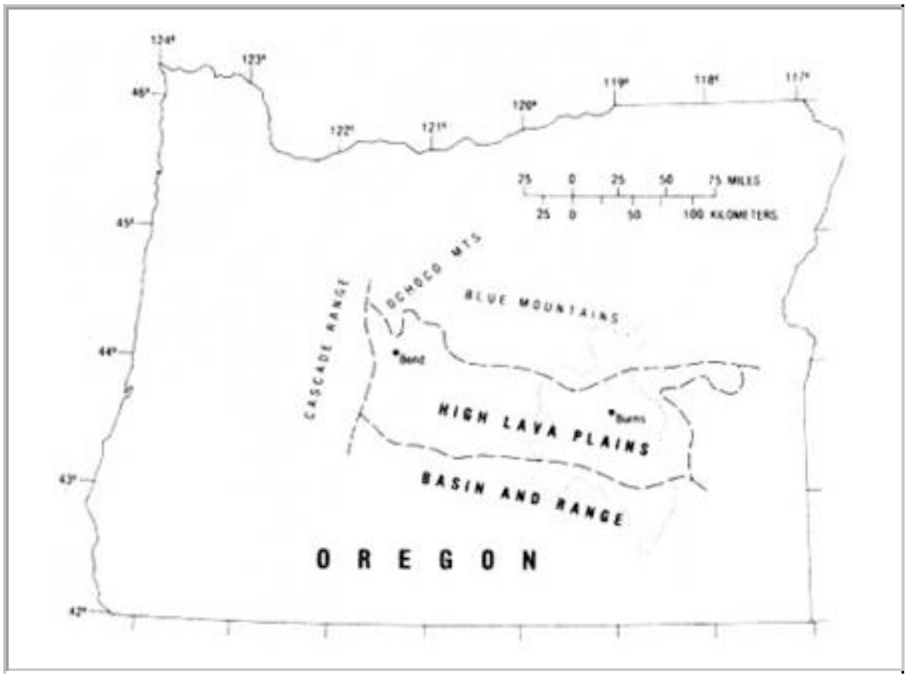

Figure 1.—Index map of Oregon showing the High Lava Plains and approximate outline of Harney Basin (dotted line).

The High Lava Plains province is a middle and late Cenozoic volcanic upland nearly 250 km long and about 80 km wide that extends south-eastward from the Cascade Range to the eastern margin of the Harney Basin. Structurally the province is dominated by a west-northwest-trending zone of en echelon normal faults (fig. 2A, B, and C), generally called the Brothers fault zone, that can be traced for most of the length of the province and perhaps well beyond the province boundaries. Eruptive centers for both basaltic and rhyolitic volcanic rocks are concentrated in the zone of faults and in some nearby subsidiary fault zones. A belt of late Cenozoic basaltic flows, extending from the Bend area nearly to the Idaho border (Walker, 1977), obscures much of the older Cenozoic geology along the Brothers fault zone.

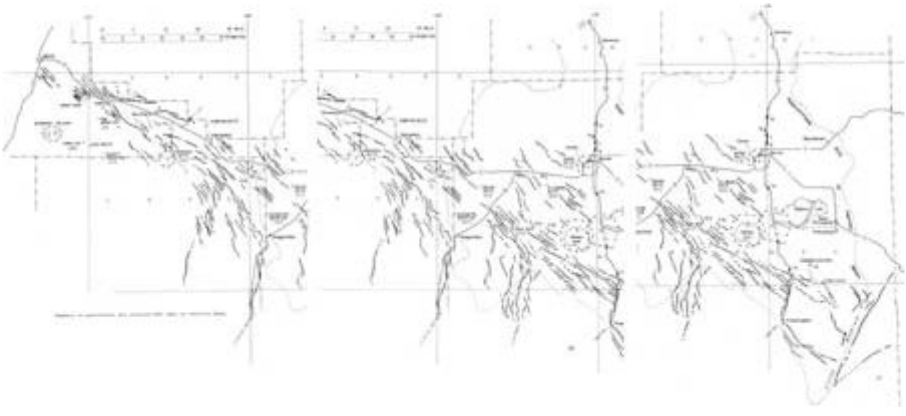

Figure 2A, B, and C.—Map showing route of field trip along U.S. Highway 20 from Bend to Harney Basin, major structural elements, and selected K/Ar ages on rhyolite domes. *(click on links for enlargements in a new window)*

The High Lava Plains are contiguous with and gradational into the Basin and Range province to the south, and many late Cenozoic volcanic rocks and fault structures are common to both provinces. For instance, some of the larger north- and northeast-trending faults, characteristic of the Basin and Range, appear to change to a northwest trend and blend into the Brothers fault zone (Walker, 1977). A comparatively sharp boundary separates the High Lava Plains from the Blue Mountain province to the north, where older Cenozoic and pre-Cenozoic rocks have been brought to the surface in the Blue Mountain-Ochoco Mountains uplift.

Older Cenozoic volcanic and tuffaceous sedimentary rocks are exposed along the northern margin of the High Lava Plains and in kipukas (inliers), such as Pine Mountain, and include parts of the Columbia River Basalt Group and the John Day and Clarno Formations (Walker and others, 1967). In most parts of the province, however, the oldest rocks are aphyric and phyric basalt with plagioclase phenocrysts and minor andesite flows of middle(?) Miocene age. These flows, which have generally been referred to the Steens Basalt, are exposed principally along the southern and southeastern margin of the province; sections hundreds of meters thick are beautifully exposed in the glaciated canyons high on Steens Mountain. An average chemical composition of these middle(?) Miocene flows (table 1, column 1) indicates a moderate to low silica content and a somewhat higher than normal alumina content; this average is not a fair representation, however, of the chemical diversity of these rocks, which vary considerably in their proportions of plagioclase, mostly labradorite, and olivine. Isotopic ages of the basalt flows indicate most were erupted about 15 m.y. ago (Baksi, and others, 1967; Laursen and Hammond, 1974; Walker, and others, 1974).

Table 1.—Average (or typical) chemical compositions ot selected basaltic and rhyolitic rocks.

	1	2	3	4	5
SiO_2	48.4	75.7	75.8	74.7	77.0
Al_2O_3	16.7	13.4	11.6	12.3	12.1
Fe_2O_3	4.3	.26	2.4	2.9	.9
FeO	7.5	.56			
MgO	5.8	.09	.2	.3	.2
CaO	9.0	.90	.4	.5	.6
Na_2O	3.0	3.8	3.8	4.4	3.5
K_2O	1.0	3.7	4.9	4.4	4.9
H_2O	1.5	.43	--	--	--
TiO_2	2.2	.10	.21	.15	.16
P_2O_5	.4	--	.05	--	.15
MnO	.2	.06	.04	--	.17

1. Average composition of 18 middle(?) Miocene flows (from Walker 1969).
2. Typical analysis of early Pliocene (4.9 m.y.), peraluminous rhyolite collected on Glass Buttes (from MacLeod, and others, 1976).
3. Average of 18 analyses of late Miocene Devine Canyon Ash-flow Tuft. All Fe reported as FeO. Calculated water-free (from Walker, in press).
4. Average of 4 analyses ot late Miocene Prater Creek Ash-flow Tuft. All Fe reported as FeO. Calculated water-free (from Walker, in press).
5. Average ot 12 analyses ot late Miocene Rattlesnake Ash-flow Tuft. All Fe reported as FeO. Calculated water-tree (from Walker, in press).

The basalts and andesites were overlapped by tuffaceous sedimentary rocks and subsequently by widespread sheets of ash-flow tuff. The ash-flow tuffs, which are of rhyolitic composition (table 1, columns 3, 4, and 5), spread laterally over tens of thousands of square kilometers of ancestral Harney Basin and form three principal units that are about 9, 8, and 6.5 m.y. old. The total volume of erupted rhyolitic tephra is unknown but is in the range of 1,000-2,000 km. Eruption of this tephra was accompanied by collapse into evacuated magma chambers, which was partly responsible for the development of the large structural depression of Harney Basin. The depression was subsequently filled with younger ash-flow tuffs, tuffaceous sedimentary rocks, and basalt flows and palagonitic sediments—all of middle(?) and late Cenozoic age (Piper, and other, 1939; Baldwin, 1976; Greene and others, 1972). In places, the tuffaceous sedimentary rocks have been diagenetically altered to bentonitic clay minerals, zeolites, and potassium feldspar (Walker and Swanson, 1968), particularly where the sediments were deposited in ancient lakes that filled the lower parts of the structural depression.

In the western part ot the province, the middle and late Miocene basalt flows, ash-flow tuffs, and sedimentary rocks are mostly buried beneath younger late Miocene Pliocene, and Quaternary basalt flows (Williams, 1957; Peterson and Groh, 1965; Walker and others, 1967) that erupted from widely scattered cones, shield volcanoes, and fissures. Charcoal from trees inundated by several young flows on the northwest flanks ot Newberry Volcano have C14 ages of about 6,000 years (Peterson and Groh, 1965, p. 9; Chitwood, and others, 1977). Throughout much ot the western part of the province most ot these younger rocks are mantled with a widespread but discontinuous sheet of unconsolidated coarse to fine pumice fragments, some of which was erupted about 7,000 years ago from Mount Mazama, a large prehistoric Cascade Range volcano that collapsed to form the caldera now occupied by Crater Lake, and some from vents in the caldera ot Newberry Volcano.

A large number of rhyolite domnes and flows occur throughout the length of the High Lava Plains; a typical analysis of one dome (Glass Butte) is given in table 1 (column 2), and several other analyses have been been published elsewhere (MacLeod and others, 1976). The domes and related ash flows, show a well defined age progression that spans the period from late middle Miocene to Holocene time, or about the last 10 or 11 m.y. A large number of radiometric ages (Parker and Armstrong, 1972; McKee and others, 1976) of rhyolite masses along the Brothers fault zone and in areas to the south in the Basin and Range province demonstrate that the progression is not along a single linear belt of age-related domes and intrusive bodies, but rather along a broad front or possibly a series of parallel linear belts. The oldest domes occur at the east and southeast edge of Harney Basin and the youngest at the west near Newberry Volcano.

Age of spatially related late Cenozoic basalt flows and basaltic (palagonitic) sedimentary rocks of greater volume than the rhyolitic rocks within the High Lava Plains province, are too poorly known to demonstrate a parallel age progression. Furthermore, most basalts comparable in age to the rhyolites have yet to be related to vents and their relations to the rhyolites are obscured by the west-northwest-trending band or belt of young basalt flows (Walker, 1977) that extends along the axis of the High Lava Plains and beyond, including basalt accumulations at Jordan (or Morcom) and Cow Lake Craters, east of Steens Mountain, Diamond Craters (Peterson and Groh, 1964) at the northwest base of Steens Mountain, at the west margin of Harney Basin, and in and near Newberry Volcano (fig. 3).

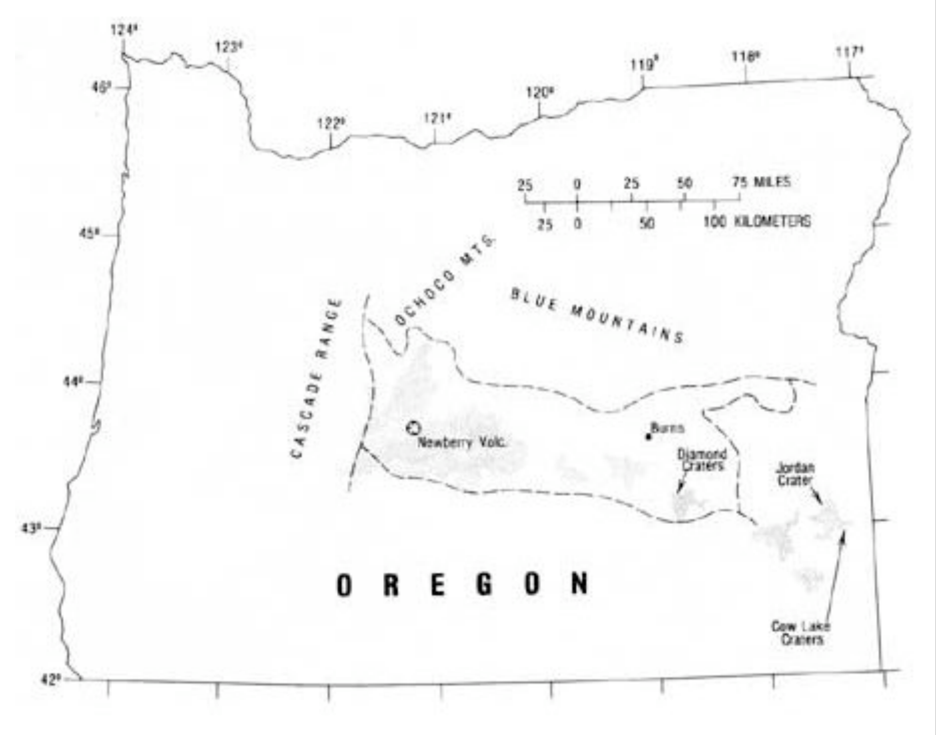

Figure 3.-West-northwest-trending belt of Middle (?) and late Quaternary basalt fields.

The total thickness of Cenozoic volcanic and tuffaceous sedimentary rocks in the province is not known and estimates are based on fragmentary data. These data, however, indicate that the thickness in most areas is more than 1,700 m and that a minimum total volume of basaltic and rhyolitic volcanic material erupted to the surface within the province must be on the order of 30,000 to 40,000 km^3. This widespread and thick sequence of volcanic rocks was erupted from several different kinds of vents of several different ages. Volcanic accumulations older than the age progression and in the form of kipukas consist largely of silicic to intermediate eruptive and intrusive complexes, such as those at Pine Mountain and Wagontire Mountain. Vents related to the age progression, which appear to be concentrated in the Brothers fault zone and subsidiary fractures, consist of domes and dome complexes, such as those at China Hat, East Butte, Quartz Mountain, Glass, Squaw, Palomino, and Burns Buttes, some of which are probably resurgent domes related to very obscure buried calderas in Harney Basin or to ring-fracture zones such as at Frederick Butte.

Most of the older basalt flows in and near Steens Mountain are associated with major dike swarms visible or, glaciated canyon walls on the east face of the high Steens block; younger flows are associated with broad shield volcanoes of low relief, with cinder zones, or with poorly exposed fissure zones. In places the rising basalt magma encountered ground or surface water, which caused fragmentation of the basalt and violent steam explosions. Where the explosions were of moderate intensity basaltic tuft cones, tuft rings, and tuft ridges formed; the best examples of these features are found about 40 to 60 km south of the field trip route at Fort Rock, Table Mountain, St. Patrick Mountain, and Seven Mile Ridge, all either within or marginal to Fort Rock and Christmas Lake Valley, and in the west end of Harney Basin, near Iron Mountain. More violent steam explosions blasted out prominent craters, including such features as Hole-in-the-Ground.

One of the largest and most spectacular eruptive centers is Newberry Volcano near the western end of the province. It is a comparatively young, Pleistocene and Holocene volcano, about 65 km long and 40 km wide surmounted by a caldera that contains East and Paulina Lakes and a variety of volcanic features and products; the walls of the caldera expose several varieties of both basaltic and rhyolitic rocks, including ash-flow tuft, than indicate it has had a more complex volcanic history than most of the other volcanoes in the region (MacLeod, and others, this volume).

The structural pattern of the High Lava Plains province is dominated by the northwest-trending Brothers fault zone which appears to be one of the fundamental structural elements of Oregon. The zone is continuously exposed for nearly 240 km through south-central Oregon. At its west end it appears to swing northward into the Cascade Range at Green Ridge. Southeast of Harney Basin the fault zone may curve southward into Nevada (Stewart, and others, 1975), or it may continue further east to and beyond Jordan Valley near the Idaho border. North of the fault zone, in the Ochoco and Blue Mountains (fig. 1), older Cenozoic and pre-Cenozoic rocks lie at the surface, whereas in and south of the zone a thick mantle of volcanic and volcaniclastic rocks buries the older rocks. Throughout the High Lava Plains, the zone is dominated at the surface by closely spaced (commonly 1/2 - 3 km apart) en echelon normal faults of moderate to small (mostly 10-100 m) displacement that localized many basaltic and rhyolitic vents of late Miocene through Pleistocene age. The abundance and age distribution of vents indicate that recurrent crustal breaking has taken place along the zone for a considerable span of time. Also, many northeast-trending structures in the southern part of the region, including several of the large normal faults that bound major graben, terminate at this discontinuity or, in places, change direction and seem to merge into it. Apparently both the normal faults of the zone and the many volcanic vents along the zone represent only the surface manifestations of deformation on a large deeply buried structure, the exact nature of which is not known. The pattern of normal faults within and near the Brothers fault zone and the relation of many small monoclinal folds to the faults suggests, however, that the zone overlies a deeply buried fault with some lateral displacement; the normal faults denote only adjustment of bristle surface and near surface volcanic and tuffaceous sedimentary rocks. Analysis of surface features permits either right- or left-lateral displacement on this buried structure, although right-lateral movement seems more likely (Walker, and others, 1967; Lawrence, 1976).

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