Prepared in collaboration
with the Mount Rainier
Natural History Association

GEOLOGY OF MOUNT RAINIER
NATIONAL PARK
WASHINGTON

GEOLOGY OF MOUNT RAINIER
NATIONAL PARK

Mount Rainier from the northwest.   Willis Wall and the head of the Carbon Glacier at left.   Mount Adams, 48 miles from Mount Rainier, at upper right.
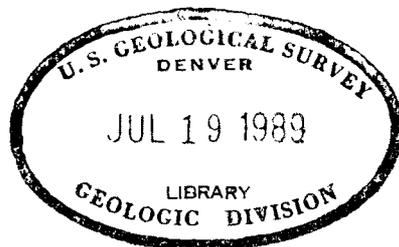Photograph by H. Miller Cowling.

Geology of

Mount Rainier National Park

Washington

By RICHARD S. FISKE, CLIFFORD A. HOPSON, *and* AARON C. WATERS

GEOLOGICAL SURVEY PROFESSIONAL PAPER 444

Prepared in cooperation with the
Mount Rainier Natural History Association

UNITED STATES DEPARTMENT OF THE INTERIOR

STEWART L. UDALL, *Secretary*

GEOLOGICAL SURVEY

Thomas B. Nolan, *Director*

CONTENTS

ILLUSTRATIONS

TABLES

GEOLOGY OF MOUNT RAINIER NATIONAL PARK, WASHINGTON

By Richard S. Fiske, Clifford A. Hopson, and Aaron C. Waters

ABSTRACT

Mount Rainier National Park includes 378 square miles of rugged terrain on the west slope of the Cascade Mountains in central Washington. Its most imposing topographic and geologic feature is glacier-clad Mount Rainier. This volcano, composed chiefly of flows of pyroxene andesite, was built upon an earlier mountainous surface, carved from altered volcanic and sedimentary rocks invaded by plutonic and hypabyssal igneous rocks of great complexity.

The oldest rocks in the park area are those that make up the Ohanapecosh Formation of late Eocene age. This formation is more than 10,000 feet thick, and consists almost entirely of volcanic debris. It includes some lensoid accumulations of lava and coarse mudflows, heaped around volcanic centers, but these are surrounded by vastly greater volumes of volcanic clastic rocks, in which beds of unstratified coarse tuff-breccia, about 30 feet in average thickness, alternate with thin-bedded breccias, sandstones, and siltstones composed entirely of volcanic debris. The coarser tuff-breccias were probably deposited from subaqueous volcanic mudflows generated when eruption clouds were discharged directly into water, or when subaerial ash flows and mudflows entered bodies of water. The less mobile mudflows and viscous lavas built islands surrounded by this sea of thinner bedded water-laid clastics. In compostion the lava flows and coarse lava fragments of the Ohanapecosh Formation are mostly andesite, but they include less abundant dacite, basalt, and rhyolite.

The Ohanapecosh Formation was folded, regionally altered to minerals characteristic of the zeolite facies of metamorphism, uplifted, and deeply eroded before the overlying Stevens Ridge Formation of Oligocene or early Miocene age was deposited upon it. The Stevens Ridge rocks, which are about 3,000 feet in maximum total thickness, consist mainly of massive ash flows. These are now devitrified and altered, but they originally consisted of rhyodacite pumice lapilli and glass shards, which compacted and welded into thick massive units during emplacement and cooling. Subordinate water-laid clastic rocks occur toward the top of the formation, and thin-bedded pyroclastic layers occur between some of the ash flows.

Exposures on Backbone Ridge and on Carbon River below the mouth of Cataract Creek show that in places the thick basal Stevens Ridge ash flows swept with great violence over an old erosion surface developed on rocks of the Ohanapecosh Formation. Masses of mud, tree trunks, and other surface debris were swirled upward into the base of the lowermost ash flow, and lobes and tongues of hot ash were forced downward into the saprolitic mud.

The Stevens Ridge Formation is concordantly overlain by the Fifes Peak Formation of probable early Miocene age, which consists of lava flows, subordinate mudflows, and minor quantities of tuffaceous clastic rocks. The lavas are predominantly olivine basalt and basaltic andesite, but they include a little rhyolite. They are slightly to moderately altered: the ferromagnesian phenocrysts are generally replaced by saponite, chlorite, or carbonate; the glass is devitrified; and the rocks are locally permeated by veinlets of zeolite. Swarms of diabase sills and dikes are probably intrusive equivalents of the Fifes Peak lavas.

The upper part of the Fifes Peak Formation has been mostly eroded from Mount Rainier National Park, but farther north, in the Cedar Lake quadrangle, it attains a thickness of more than 5,000 feet.

The Fifes Peak and earlier formations were gently folded, faulted, uplifted, and eroded before the late Miocene Tatoosh pluton worked its way upward to shallow depths and eventually broke through to the surface. The rise of the pluton was accompanied by the injection of a complicated melange of satellitic stocks, sills, and dikes. A favored horizon for intrusion of sills was along or near the unconformity that separates the Ohanapecosh and Stevens Ridge Formations. Near the major plutonic centers sill is piled on sill along this unconformity to form a complex that resembles a huge cedar-tree laccolith.

At several places the magma broke through to the surface, initiating a series of explosive volcanic eruptions. Only small remnants of the resulting pyroclastic rocks have survived erosion: one of these remnants forms The Palisades, in the northeastern part of the park, and consists of a mass of welded tuff at least 800 feet thick, which grades downward into a plug of rhyodacite. Other plugs and stocks in the area northeast of the park contributed pumiceous debris to the Ellensburg Formation of early Pliocene age in the adjacent Snoqualmie and Mount Aix quadrangles.

Extensive erosion unroofed the Tatoosh pluton and carved canyons as deep as 4,000 feet into the plutonic rocks and sill complexes. This occurred before the Mount Rainier volcano began to take form, probably in early Pleistocene time. The first Mount Rainier eruptions were voluminous lava flows of pyroxene andesite. These flows spilled westward down the canyons of the ancestral Mowich and Puyallup Rivers, astride which the volcano grew. They also obliterated a northeastward-flowing river, remnants of whose canyon now lie buried beneath the lava fill of Grand Park and Old Desolate. With the smoothing of the pre-Rainier topography by these early intra-canyon flows, the volcano began to grow into its present form. Despite rapid contemporaneous erosion by streams and glaciers, the overlapping streams of lava, subordinate mudflows, and thin pyroclastic deposits eventually built a huge cone that was about 1,000 feet higher than the present mountain. Its summit area was then either engulfed by collapse or else deeply hollowed out by erosion. Still later a small cone, which

1

culminates in Columbia Crest, grew along the southeastern edge of the old summit rim. The smoothly rounded sides of this younger cone, almost unscarred by erosion, contrast strikingly with the older intensely eroded cone that forms the main bulk of the mountain.

During the building of the Columbia Crest cone thin falls of pumice and ash spread outward for several miles from the base of the volcano. The youngest extensive pumice sheet, about 500 to 600 years old, probably records the last significant eruption from Mount Rainier.

Rapid erosion, which had so effectively reduced and sharpened the older cone, is still continuing. The huge glaciers, rasping down the mountain's sides, are particularly potent agents of erosion; but avalanches, rockfalls, and a variety of other downslope movements also are rapidly stripping debris from the steeper slopes. Mudflows and destructive slurry floods of melt water, sand, and rock occasionally surge through the lower canyons, eroding great gashes into the unconsolidated glacial and stream deposits that choke the upper ends of the valleys, and spreading huge fans of detritus farther down. The Kautz flood of October 2, 1947, is typical: it moved 15 million cubic yards of debris several miles downstream and redeposited it in a huge fan that temporarily blocked the Nisqually River.

INTRODUCTION

The geology of Mount Rainier and the rugged terrain surrounding it has recieved little attention in the literature. Early exploratory reconnaissances by S. F. Emmons (1879; and *in* King, 1871) in 1870, and by I. C. Russell (1897), G. O. Smith (1897), and others near the end of the last century established that Mount Rainier is a composite volcano built on an elevated platform of altered volcanic and granitic rocks. H. A. Coombs (1936) was the first to publish the results of a fairly comprehensive reconnaissance, accompanied by a map of the park in which the rocks, all of Cenozoic age, are grouped into four major divisions: the Puget Group, the Keechelus Andesitic Series, the Snoqualmie Granodiorite, and the rocks of Mount Rainier volcano. Coombs also gave considerable petrographic detail, which clearly indicated the great heterogeneity and complexity of the rocks within the park, especially of the basement on which the volcano rests. His descriptions and petrographic work emphasized the need of more detailed geologic mapping to determine the number and sequence of formations within the park and to establish their origin and geologic history.

We were therefore pleased when the Mount Rainier Natural History Association invited us to undertake a more detailed study of the geology of the park. A brief preliminary reconnaissance of areas near roads and some of the trails, made by Waters in September of 1957, provided the basis for a plan of fiieldwork. Geologic mapping was started by Fiske, Hopson, and Waters in the summer of 1958, resumed during the summer of 1959, and completed in the summer of 1960. Each of us has spent equal time on the problem; au-

thorship is listed alphabetically. Our results, presented in this report, can scarcely be regarded as more than a detailed reconnaissance. Despite the relatively small area of the park (378 sq mi), generalized groupings of many rock units are necessitated because of the great variety of the rocks, their complex stratigraphic and petrologic relations, and the fantastically intricate map patterns that result from any attempt to show such features as the widespread sill and dike swarms in detail. Not only are many features too small and complex to show on the map reproduction scale of 1 : 62,500, but time was not available during the fieldwork to trace such intricate units in detail.

Physical difficulties also slowed the fieldwork. The altitudes in Mount Rainier National Park range from 1,560 feet where the Ohanapecosh River leaves the park to 14,410 feet at the summit of Mount Rainier.[1] The best sections of the Mount Rainier lavas (those, for example, in Willis Wall and the cliffs at the head of Sunset Amphitheater) are well-nigh inaccessible because these steep slopes are raked by avalanches and bordered by deeply crevassed glaciers. Below altitudes of 5,000 feet much of the park is covered by dense forest and underbrush. The larger streams, swollen in summer by melt water from glaciers, are difficult to cross.

Although these obstacles of intricate geology and difficult terrain necessitated great oversimplifications on the final map, we believe that the major events in the geologic history have been determined, the rocks subdivided into recognizable and distinctive units of formation rank, and the episodes of sedimentation, igneous activity, and hydrothermal metamorphism put in their proper relations within space and time.

ACKNOWLEDGMENTS

---

[1] The altitude of Mount Rainier is given as 14,408 ft on the topographic map, but a recent resurvey by the U.S. Geological Survey has corrected this to 14,410 ft.

Taylor identified a fossil mollusk from near Packwood. Richard E. Fuller supplied photographs and information about the rocks near Ohanapecosh Hot Springs. H. Miller Cowling made available a large collection of airplane photographs, several of which are reproduced in this report. V. R. Bender compiled information about supposed historic eruptions of Mount Rainier, gave other technical help, and assisted in obtaining geologic photographs. D. R. Crandell showed us the stratigraphy of the Quaternary deposits that he had worked on to the west and north of the park (Crandell and Waldron, 1956; Crandell and others, 1958). Thomas L. Wright determined the structural state of the feldspars in the Tatoosh pluton, contributed information about its field relations, and critically read parts of the manuscript. Howard A. Coombs and J. Hoover Mackin read parts of the manuscript and suggested many useful changes.

Laboratory space and equipment was provided by the Department of Geology, The Johns Hopkins University; Waters is also indebted for equipment and space to the University of Oregon and to the National Science Foundation, for his part of the report was written in Eugene, Oreg., during the tenure of a National Science Foundation Senior Postdoctoral Fellowship.

SCOPE OF REPORT

Mount Rainier National Park lies on the west slope of the Cascade Range in Washington (fig. 1). The rocks of the park fall naturally into four major groups: (1) lavas and bedded epiclastic and pyroclastic rocks of early and middle Tertiary age; (2) plutonic masses of Miocene and Pliocene age with associated shallow plugs, sills and dikes, and contemporaneously erupted volcanic rocks; (3) lavas and pyroclastic rocks of Mount Rainier volcano; and (4) unconsolidated deposits, chiefly glacial and fluvioglacial debris, mudflows, ash and pumice sheets of late Pleistocene and Recent age.

These rocks are described in order of age, and the first three groups have received most attention. The main emphasis is on the bedrock geology of the park, especially with reference to volcanism, sedimentation, metamorphism, and plutonic activity. The report deals only incidentally with the exceptionally interesting record of glacial and postglacial events.

BEDDED ROCKS OF EARLY AND MIDDLE TERTIARY AGE

H. A. Coombs (1936), in an early report on the geology of Mount Rainier National Park, assigned all the lavas and volcanic clastic rocks older than Mount
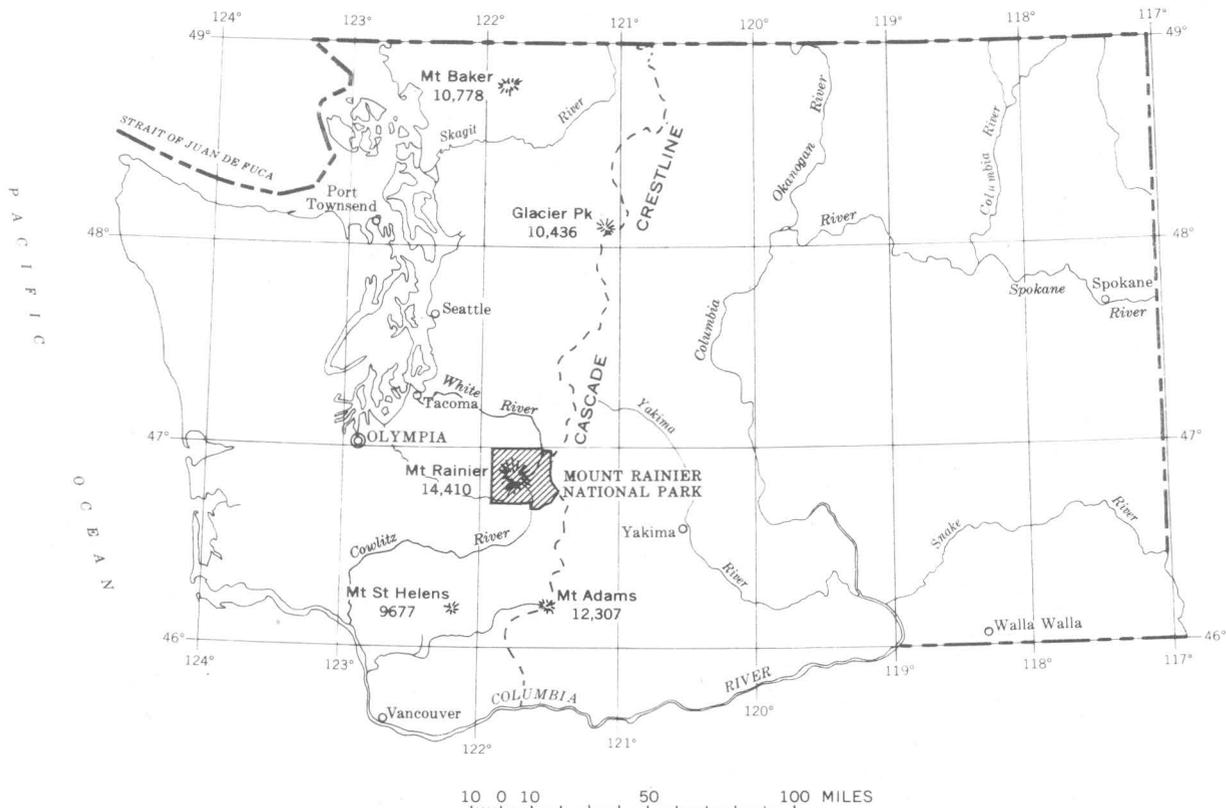
FIGURE 1.—Index map of Washington showing the location of Mount Rainier National Park and positions of large Quaternary volcanoes.

Rainier volcano to the Keechelus Andesitic Series (Smith and Calkins, 1906, p. 8). These rocks, however, are separable into distinct stratigraphic units which differ in lithology, age, and origin. Moreover, Smith and Calkins' original definition and interpretation of the age and stratigraphic position of the Keechelus Andesitic Series has been so modified by later workers that confusion prevails as to the meaning of Keechelus. Accordingly, we have mapped, named, and described the bedded rocks of Mount Rainier National Park without reference to correlation with the Keechelus Andesitic Series. The lower and middle Tertiary rocks are divided into three formations—the Ohanapecosh, Stevens Ridge, and Fifes Peak—whose ages and general character are shown in figure 2. The Keechelus problem has been fully reviewed in a separate publication (Waters, 1961), and one conclusion reached is that Smith and Calkins' lower Keechelus corresponds approximately with the Stevens Ridge and Fifes Peak Formations of this report.

ROCK NOMENCLATURE

There is no universally accepted nomenclature for volcanic clastic rocks. Many terms are used in a different sense by different authors. We have used the following rock names as defined in the papers cited, although in some cases the sorting or grain size parameters have been modified slightly: volcanic breccia (Fisher, 1958, p. 1072); tuff-breccia (Norton, 1917, p. 162); volcanic conglomerate (Wentworth and Williams, 1932, p. 50); volcanic sandstone (Hay, 1952, p. 119); volcanic graywacke (wacke) and volcanic arenite (Williams and others, 1955, p. 292–293). Textural variants of some of the general rock names are indicated by modifiers, such as well sorted and poorly sorted. The classification used is shown in table 1.

TABLE 1.—*Classification of volcanic clastic rocks*

Grade size (mm)	General rock type		Specific rock type	
			Matrix	
			Abundant	Sparse
>2	Clasts angular to sub-rounded	Volcanic breccia	Tuff-breccia	Well-sorted volcanic breccia
	Clasts sub-rounded to well rounded	Volcanic conglomerate	Poorly sorted volcanic conglomerate	Well-sorted volcanic conglomerate
<2>1/16		Volcanic sandstone	Volcanic graywacke	Volcanic arenite
<1/16		Volcanic siltstone		

In this report these terms have been used in a descriptive sense, and are meant to indicate grain size, grain shape, and sorting characteristics; they are applied not only to rocks of known origin, but also to rocks of questionable origin that might be either pyroclastic or epiclastic. On the other hand, names such as mudflow, ash fall, nuée ardente, or volcanic sedimentary rock are used to emphasize the probable origin of the rock.

OHANAPECOSH FORMATION

DEFINITION, DISTRIBUTION, AND THICKNESS

The Ohanapecosh Formation, here first defined, is named from Ohanapecosh Hot Springs, a small park community located on Washington State Highway 5 in the southeast corner of Mount Rainier National Park. Rocks of the Ohanapecosh Formation underlie almost half of the area of the park (pl. 1). The principal areas of outcrop include a broad belt comprising the entire eastern third of the park, the Stevens Peak and Longmire areas extending along the northeast and northwest margins of the Tatoosh Range, and the Mount Wow-Satulick Mountain area in the southwest corner of the park. The Ohanapecosh is also exposed in smaller areas in the valleys of the North Puyallup River and the North and South Forks of the Mowich River, in still smaller areas near the head of the Carbon River, and as septa within the sill complexes of the

Series	Formation	Thickness (feet)	Description
TERTIARY — MIOCENE	Fifes Peak Formation (Top not exposed)	2400+	Basalt, basaltic andesite, and andesite flows; minor amounts of rhyolite flows and tuff-breccia, volcanic sandstone, and volcanic siltstone of epiclastic and pyroclastic origin
TERTIARY — OLIGOCENE	Stevens Ridge Formation	450 to 3000	Rhyodacitic ash flows; subordinate amounts of volcanic breccia, sandstone, and siltstone of epiclastic and pyroclastic origin
	Unconformity		
TERTIARY — EOCENE	Ohanapecosh Formation (Base not exposed)	10,000+	Volcanic breccia, sandstone, and siltstone of epiclastic and pyroclastic origin; local thick accumulations of basaltic andesite lava flows and coarse mudflows; rhyolite

FIGURE 2.—Lower and middle Tertiary formations in Mount Rainier National Park.

Skyscraper Mountain-Lake James area north of Mount Rainier.

Ohanapecosh Hot Springs lies near the center of a thick westward-dipping section of Ohanapecosh rocks. Their thickness in this section, measured across strike from the Cascade crest to Stevens Canyon, is slightly more than 10,000 feet. In the Mount Wow-Satulick Mountain area they exceed 9,000 feet, measured from the park boundary near the Nisqually Entrance Station to the top of Satulick Mountain.

FORMATION BOUNDARIES

Neither the base nor the original uneroded top of the Ohanapecosh Formation is seen in Mount Rainier National Park, nor have they been reached in either of the two measured sections; therefore the thicknesses measured do not represent the total thickness of the formation. The original maximum thickness in the Mount Rainier area probably exceeded 15,000 feet.

West of Mount Rainier National Park rocks of the Ohanapecosh Formation rest upon and intertongue with arkoses and siltstones of the Puget Group. Along the Mowich and Puyallup Rivers Ohanapecosh volcanic clastic rocks are interbedded with and grade into arkoses and carbonaceous shales. This junction is not a time line; in the area just west of the park the arkose intertongues higher and higher into the Ohanapecosh section toward the north. Fisher (1957) reports large-scale intertonguing of volcanic mudflows with arkose in the upper part of the Puget Group in the Elbe-Packwood area southwest of Mount Rainier National Park.

East of Mount Rainier National Park rocks of the Ohanapecosh Formation are in fault contact with highly sheared shales, graywackes, and basic lavas of Mesozoic(?) age. This junction is well shown on Washington State Highway 5, about 7 miles southeast of Ohanapecosh Hot Springs.

Within Mount Rainier National Park, the Ohanapecosh Formation is unconformably overlain by rocks of the Stevens Ridge Formation.

LITHOLOGY

Volcanic clastic rocks and lava flows comprise the bulk of the Ohanapecosh Formation. Lava flows and interbedded coarse volcanic clastic rocks (chiefly mudflow deposits) occur in local thick accumulations that grade laterally into finer volcanic clastic rocks. Ash flows, as defined by R. L. Smith (1960, p. 800), and rhyolite flows form less than 1 percent of the formation. For purposes of description, therefore, the Ohanapecosh Formation is divided into three parts: (1) lava flow-mudflow complexes, (2) adjacent accumulations of volcanic clastic rocks, and (3) ash flows and rhyolites.

LAVA FLOW-MUDFLOW COMPLEXES

Two major complexes of lava flows and associated mudflow deposits are recognized and mapped (pl. 1). The smaller crops out in an irregular north-trending belt east and southeast of Mount Rainier. It is best exposed in the steep cirque walls of the Sarvent Glaciers, about a mile southeast of the Summer Land shelter cabin (fig. 3). The larger lava complex occurs in the Mount Wow-Satulick Mountain area, in the southwestern part of the park, but it also crops out in the valley of the North Puyallup River along the western margin of the park. These lava complexes of the Sarvent Glaciers and Mount Wow areas may once have been parts of one huge volcanic pile; but this cannot be proved, because they are now separated by extensive outcrops of the Tatoosh pluton and by rocks from the Mount Rainier volcano. Lava flows and interbedded mudflows also crop out on Stevens Peak in the southern part of the park. These may be the fringe of still another lava complex lying south of the park.

The lava complexes of the Sarvent Glaciers and Mount Wow areas are both roughly lenslike in cross section as well as in plan. The maximum thickness of the Sarvent complex is about 3,800 feet, and that of the Mount Wow complex is at least 7,000 feet. Lava flows and coarse mudflow deposits form more than 70 percent of these units.

Fresh exposures of Ohanapecosh lava are dark brownish gray, greenish gray, or maroon; most weathered surfaces are brown or maroon. Individual flows are from 10 to 100 feet thick. Primary features, such as columnar jointing, platy partings, and ramp structures, are rarely preserved.

Every flow studied is prophyritic (fig. 4), but the phenocrysts range widely in abundance, forming as little as 5 or as much as 50 percent of the rock. Before being altered all the lavas were very similar in mineral composition: the phenocrysts were plagioclase, augite, and hypersthene; the groundmasses consisted of plagioclase microlites, granular clinopyroxene, magnetite, and glass. The original groundmass textures ranged from intergranular to hyalopilitic, but were most commonly intersertal.

All the Ohanapecosh lavas have been pervasively altered: hypersthene phenocrysts are wholly replaced by chlorite or montmorillonite minerals, or both; augite phenocrysts, more resistant to alteration, are partly or completely replaced by similar products. Most of the plagioclase phenocrysts are replaced by carbonate, epidote, albite, and clay minerals; zeolites are rare. Despite this alteration, complex zoning is visible in most plagioclase phenocrysts. The groundmass min-

erals are highly altered: the glass has been completely devitrified, and most of the clinopyroxene and plagioclase replaced. The groundmass is now a fine-grained aggregate of chlorites, quartz, zeolites, albite, and clay minerals and sparse granules of epidote and sphene. The composition of the plagioclase that survived alteration ranges from about An_{62} to An_{48} in phenocrysts, and from about An_{55} to An_{43} in microlites.[2] These rocks are classified as basaltic andesites because they carry calcic plagioclase typical of basalts but have textures and field associations more characteristic of andesites. More exact classification must await chemical analyses.

---

[2] The anorthite content of plagioclase from lava flows and hypabyssal igneous rocks in the park was found in the following manner: The structural state (disordered—"high-temperature" or ordered—"low-temperature") of unaltered plagioclase crystals from each rock type was determined. The method of Turner (1947) and new unpublished curves constructed by Thomas L. Wright were used to obtain the twin law and structural state. The curves given in Turner are valid for low-temperature crystals; the constructed curves were compiled from published high-temperature data. When a unique solution was not obtained, the graphs given in Tertsch (1942) were consulted to obtain additional An-values and to clarify the structural-state ambiguities. The sampled plagioclases in lava flows and in many dikes and sill rocks were all found to have high-temperature optics; it was therefore assumed that all the plagioclase microlites and small phenocrysts from lavas in the park also have high-temperature optics. The anorthite content of the plagioclase was then obtained by measuring the maximum extinction angle against (010) with a four-axis universal stage and applying this value to the high-temperature determinative curve given by Van der Kaaden (1951, p. 15).

Lava flows and mudflows are associated with flow-top breccias, intercalated well-stratified volcanic clastic rocks, and several kinds of sill rocks. The flow-top breccias are seldom more than 6 feet thick and are of minor importance. The mudflow breccias, however, are locally voluminous. Many individual mudflows are more than 100 feet thick, and in places, for example, at Tamanos Mountain and on the western slope of Mount Ararat, superposed mudflows attain more than 500 feet in aggregate thickness. The mudflow breccias are devoid of stratification and sorting: they contain angular lava fragments, as much as 6 feet in diameter, embedded in fine tuff-breccia (fig. 5). The fragments in the mudflows are similar lithologically to most Ohanapecosh lava flows, and the matrix of the mudflows contains abundant plagioclase and pyroxene crystals of the same composition as those in the lavas. This makes it clear that the lava flows and associated mudflow breccias are from the same source.

The intercalated layers of well-stratified volcanic clastic rocks seldom exceed 100 feet in thickness. Some of these, nevertheless, have been mapped because they give structural control in the otherwise massive and thick-bedded lava complexes. Sill rocks have been mapped in places, but a great many small sills have not been mapped.

FIGURE 3.—Lava complex at the Sarvent Glaciers. Basaltic andesite lava flows and intercalated volcanic clastic rocks of the Ohanapecosh Formation dip gently westward. The prominent planar surface was formed by erosion of softer Ohanapecosh volcanic clastic rocks lying above the thick sequence of lava flows.
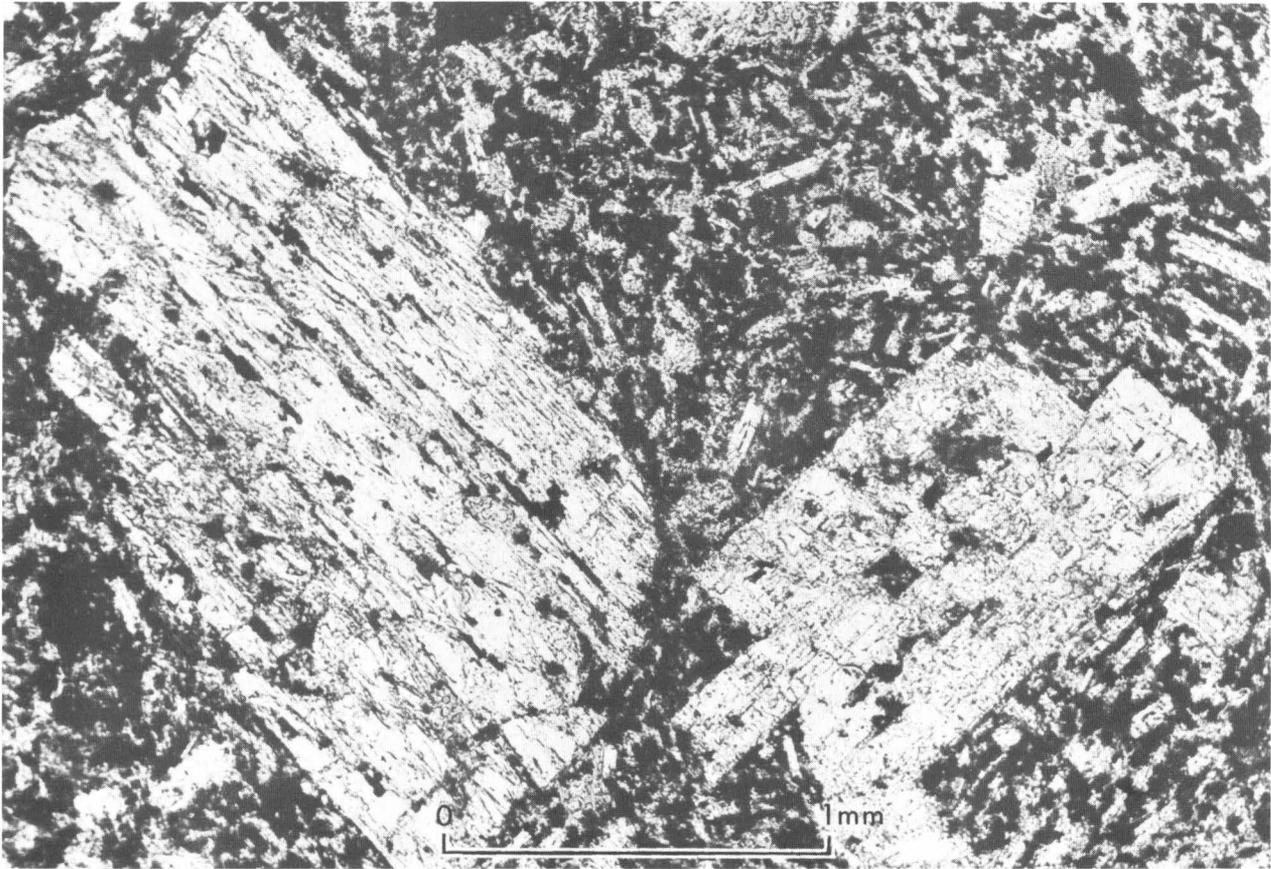
FIGURE 4.—Photomicrograph of altered lava from the Ohanapecosh Formation, one-half mile southeast of Panhandle Gap. The plagio-clase phenocrysts are largely replaced by albite, carbonate, and chlorite; all glass is devitrified; and the ferromagnesian minerals are completely replaced by chlorite and clay minerals.

FIGURE 5.—Coarse mudflow breccia of the Ohanapecosh Formation at Governors Ridge, near the eastern edge of the Sarvent lava complex. Angular to subrounded blocks of lava form a disrupted framework in a matrix of unstratified and unsorted crystal-rich tuff-breccia. Most fragments are andesite, but the light-colored ones are rhyolite.

VOLCANIC CLASTIC ROCKS

The Ohanapecosh Formation consists chiefly of volcanic clastic rocks, which include a wide variety of pyroclastic rocks and minor amounts of epiclastic rocks. This assemblage underlies nearly all the area drained by the Ohanapecosh River in the eastern and southeastern parts of the park. It also occurs in great volume north of Yakima Park, on Crystal Mountain, and in the valleys of the North Mowich and South Mowich Rivers. The volcanic clastic rocks are extremely diverse in both texture and composition. Tuff-breccia (fig. 6) is dominant; associated rocks include well-sorted volcanic breccia (fig. 7), volcanic graywacke (figs. 8, 9), volcanic arenite (figs. 10, 11), and volcanic siltstone (fig. 12).

The character of these rocks can best be illustrated by briefly describing a section that was measured in road cuts along the highway from Stevens Canyon to the Ohanapecosh River and then continued to the east along the White Pass highway (Washington State
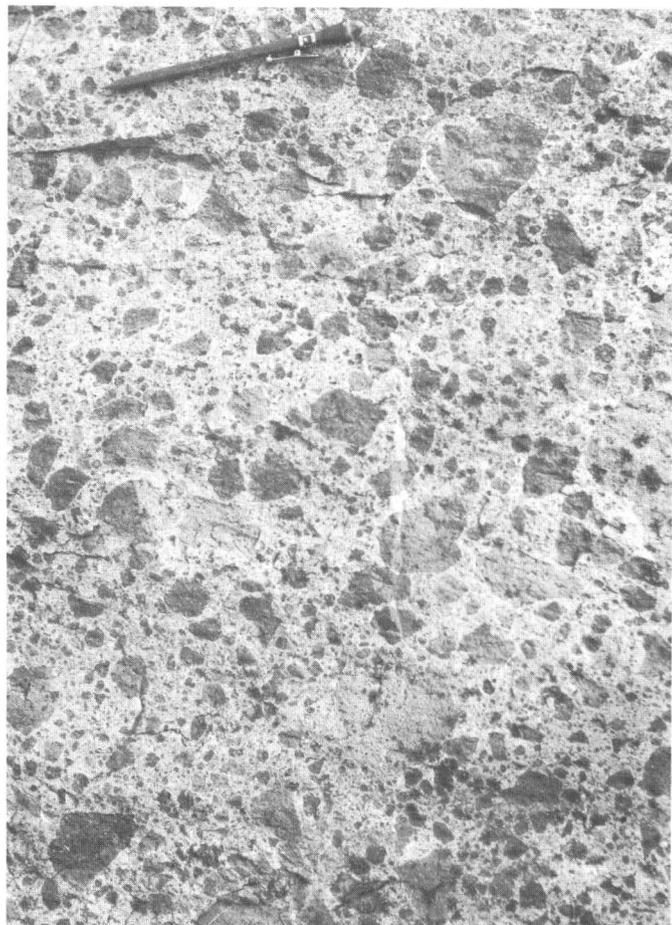
FIGURE 6.—Tuff-breccia from the Ohanapecosh Formation near Double Peak. Angular and subangular andesite fragments form a disrupted framework in an unstratified and unsorted tuffaceous matrix. The fragments are mostly less than 2 inches in diameter.

Highway 5) 4 miles southeast of Ohanapecosh Hot Springs. More than 6,000 feet of volcanic clastic rocks are exposed in this composite section.

The most notable characteristic of the section is the alternation of tuff-breccia beds more than 10 feet thick with thinner bedded volcanic breccias, sandstones, and siltstones (fig. 13). The tuff-breccias, which make up 46 percent of the measured section, form layers about 30 feet in average thickness; but some layers are as much as 350 feet thick. The layers of tuff-breccia are separated by sequences of thin-bedded clastic rocks (fig. 14), in intervals that range from a few inches to 370 feet thick (fig. 15).

The tuff-breccias are massive and thick bedded; they rarely show internal stratification, although in many places they do show an ill-defined vertical grading. Texturally they have a disrupted framework of angular to subrounded rock fragments, rarely more than 2 inches in diameter, lying in a matrix of volcanic graywacke (fig. 6). The fragments include a wide variety of devitrified glassy lavas and pumice lapilli and less abundant holocrystalline lavas and sparse wood fragments. Many of the lava fragments are microvesicular. The graywacke matrix consists of smaller fragments of the same material, plagioclase crystals, and a little quartz and clinopyroxene. The sand-sized grains are dominantly angular, but most layers contain a small proportion of subrounded grains. The outlines of glass shards are generally masked by extensive alteration.

Some of the tuff-breccias consist mainly of one kind of glassy lava or pumice lapilli, but most of them contain a mixture of volcanic fragments.

Even the thickest layers of tuff-breccia in the measured section cannot be traced far along strike because of dense forest cover on either side of highway roadcuts. Elsewhere in the park, however, better exposures make it possible to determine the lateral extent of similar layers. In the area between Cayuse Pass and Tipsoo Lake, a conspicuous tuff-breccia 90 feet thick can be traced with no appreciable variation in thickness for more than a mile along strike, and it undoubtedly continues much farther. Near the crest of the Cascades along the eastern border of the park, many layers of tuff-breccia, from 10 to 50 feet thick, extend with nearly uniform thickness for more than half a mile along strike before being covered by talus and soil. It is apparent that a tremendous volume of fragmental debris is contained in many of the thicker individual tuff-breccia layers.

The thinner bedded and finer grained volcanic clastic rocks interstratified with the thick layers of tuff-breccia consist mainly of well-sorted breccia (fig. 7), volcanic
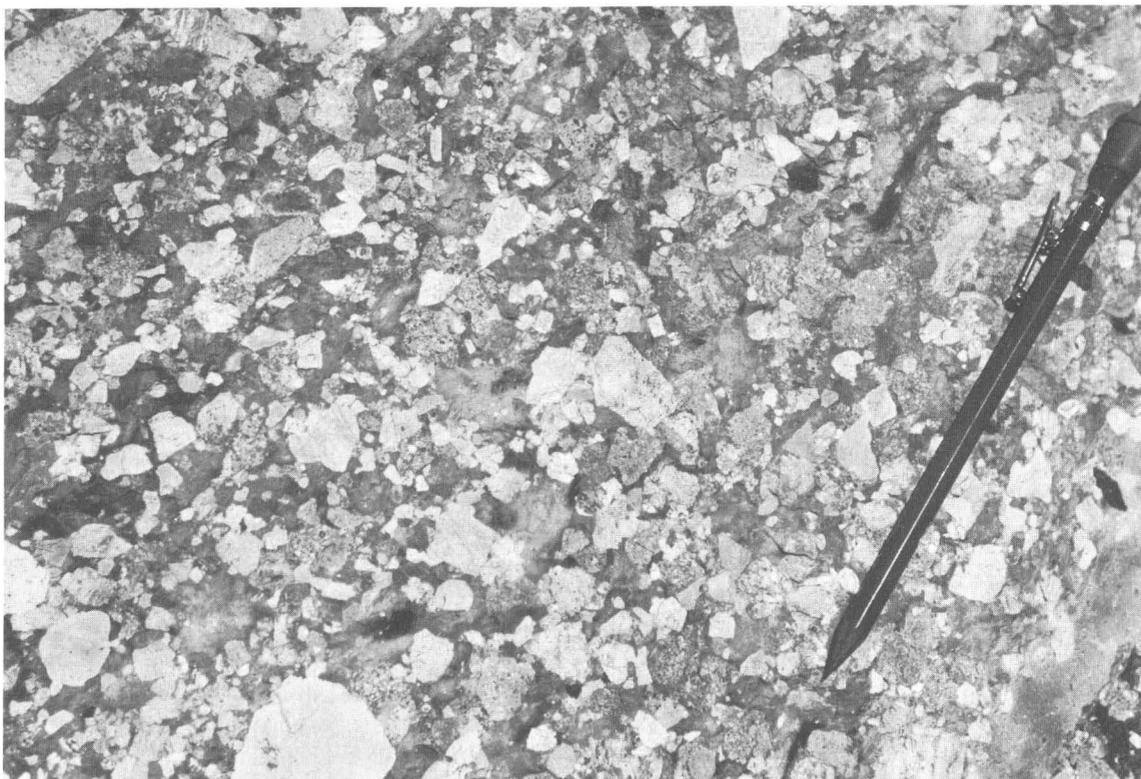
FIGURE 7.—Well-sorted volcanic breccia from a thin-bedded sequence of Ohanapecosh volcanic clastic rocks 1,800 feet west of Cayuse Pass. Angular to subangular lava fragments form an intact framework; sparse tuffaceous material fills the interstices. The pencil parallels bedding.

graywacke (fig. 8), and siltstone; but they also contain some interbedded tuff-breccia. The debris is entirely of volcanic origin and is similar in composition to that in the thick tuff-breccias. Individual beds in these finer grained rocks are from 10 feet to a fraction of an inch thick. The siltstones are generally laminated, and at many localities are delicately crossbedded in laminae rarely more than half an inch thick (fig. 12). The volcanic sandstones are in beds that range in thickness from 1 inch tc 6 feet. Practically all the beds are parallel and many are vertically graded. Some show shallow channeling and poorly developed crossbedding; the thickness of the crossbedded layers averages 4 inches and never exceeds 3 feet. Current ripple marks (fig. 16) are present but are not abundant. The tuff-breccias included in these generally fine-grained sequences are in beds from 1 to 10 feet thick and are similar in composition and texture to the thicker bedded tuff-breccias. The well-sorted breccias are composed of angular to subrounded rock fragments similar to those in the tuff-breccias. Small and large pieces of wood are fairly common throughout the section, but all of this material appears to have been transported to its present position.

FIGURE 8.—Volcanic graywacke containing pockets of tuff-breccia from the Ohanapecosh Formation (roadcut one-half mile east of Nickel Creek). Layers and lenses of coarse and fine material alternate to produce a rude stratification. In both fine and coarse layers abundant fragments of lava, pumice, and broken crystals are embedded in a fine tuffaceous matrix.
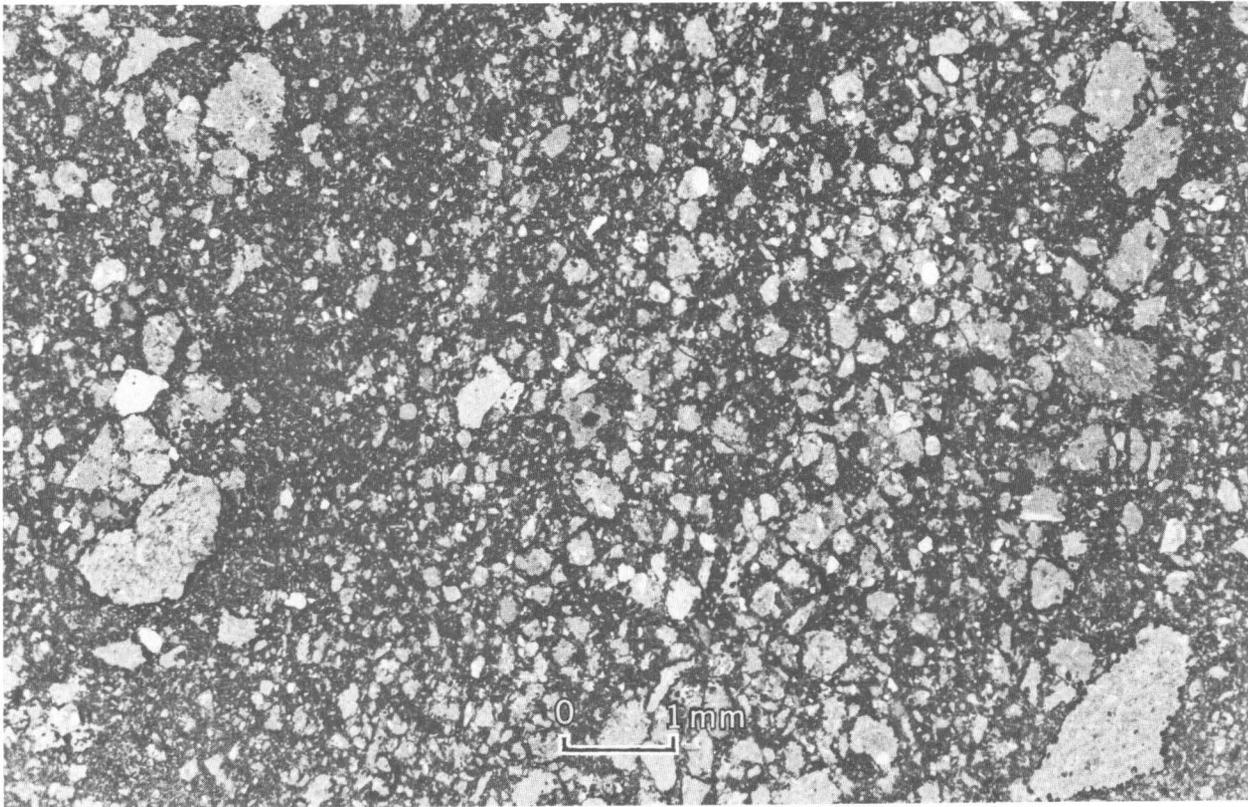
FIGURE 9.—Photomicrograph of volcanic graywacke from the Ohanapecosh Formation in a 30-foot sequence of thin-bedded volcanic clastic rocks on Backbone Ridge. Almost all the larger fragments are microvesicular glassy lava; the fine matrix contains many crystal fragments of fresh clinopyroxene and altered plagioclase.

Contemporaneous deformation of the thin-bedded volcanic clastic rocks is shown by local sharp folding and crumpling at five localities in the measured section. The asymmetry of the folds and the dip of the cross-bedding indicate that the sediments were moving from north to south. Load casts of sandstone into siltstone are present (fig. 16), especially in the thin-bedded layers just beneath thick tuff-breccias.

The metamorphism of the Ohanapecosh volcanic clastic rocks is similar to that of the lava complexes, but it has generally altered the clastic rocks more completely. Plagioclase, both in detrital grains and in the phenocryst of the rock fragments, is largely altered to such minerals as laumontite, wairakite, prehnite, carbonate, epidote, albite, and clay minerals. Hypersthene is completely altered; the scattered grains of clinopyroxene show varying degrees of alteration to chlorite and clay minerals. All the volcanic glass is divitrified to finely divided quartz, zeolites, albite, and celadonite. The fine-grained material forming the matrix of the rock fragments has been almost completely recrystallized; it is now a dense aggregate of zeolites, prehnite, quartz, albite, chlorites, carbonate, magnetite, sphene, leucoxene, and clay minerals.

686-695 O—63——2
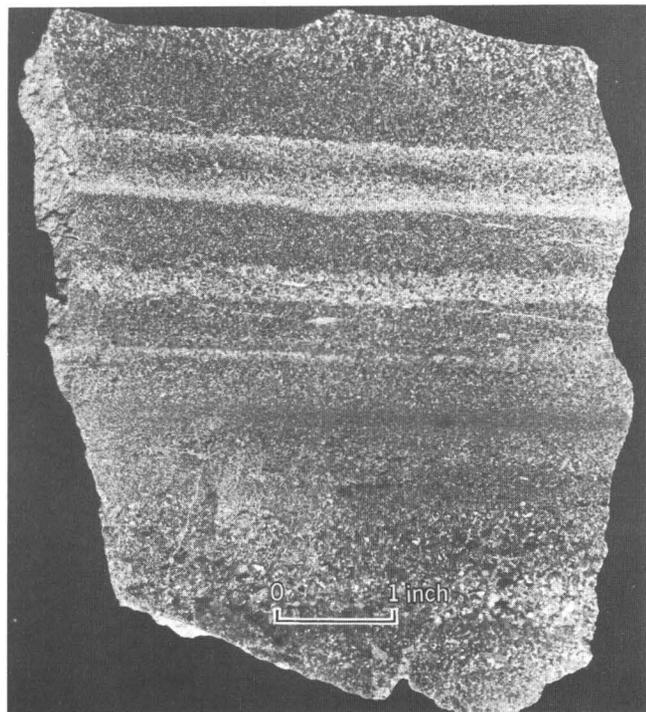
FIGURE 10.—Sawed and polished slab of coarse and fine volcanic arenite from the Ohanapecosh Formation on Backbone Ridge. Fragments are composed of angular to well-rounded glassy lava and pumice.

FIGURE 11.—Photomicrograph of volcanic arenite from the Ohanapecosh Formation on Backbone Ridge. Angular to subrounded fragments of glassy lava and sparse grains of quartz and plagioclase form an intact framework whose voids are filled with metamorphic albite, quartz, and wairakite.
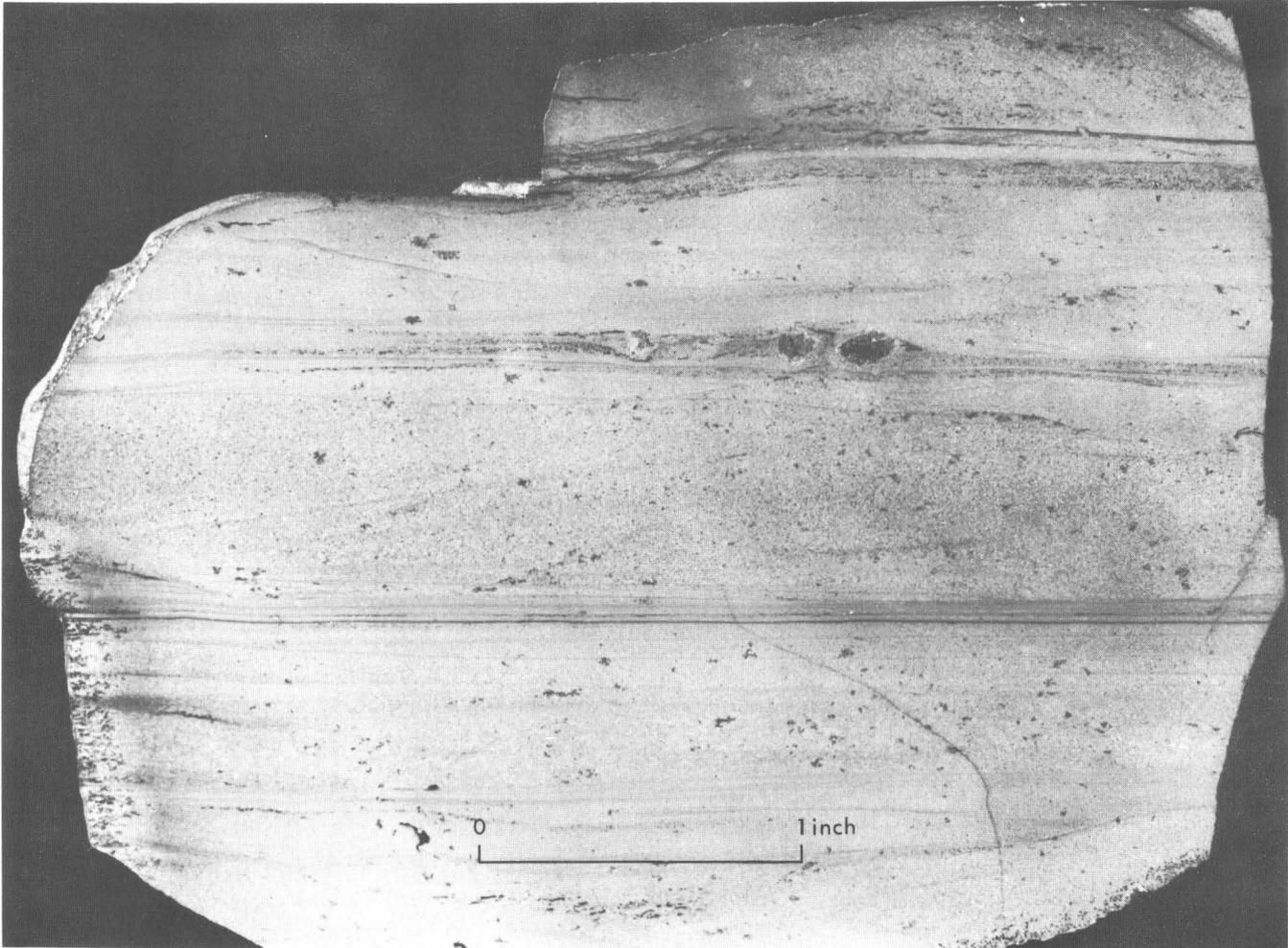
FIGURE 12.—Sawed and polished slab of volcanic siltstone from the Ohanapecosh Formation on the east side of Backbone Ridge. The prominent laminae are parallel, but fainter laminae show small-scale channeling and crossbedding. The large fragments are pumice.

EXPLANATION

Tuff-breccia
Bedding more than 10 feet thick

Tuff-breccia, well-sorted volcanic breccia,
volcanic sandstone, volcanic siltstone
Bedding less than 10 feet thick

Sill rock
Gabbro, diorite, quartz diorite, and granodiorite

Covered intervals
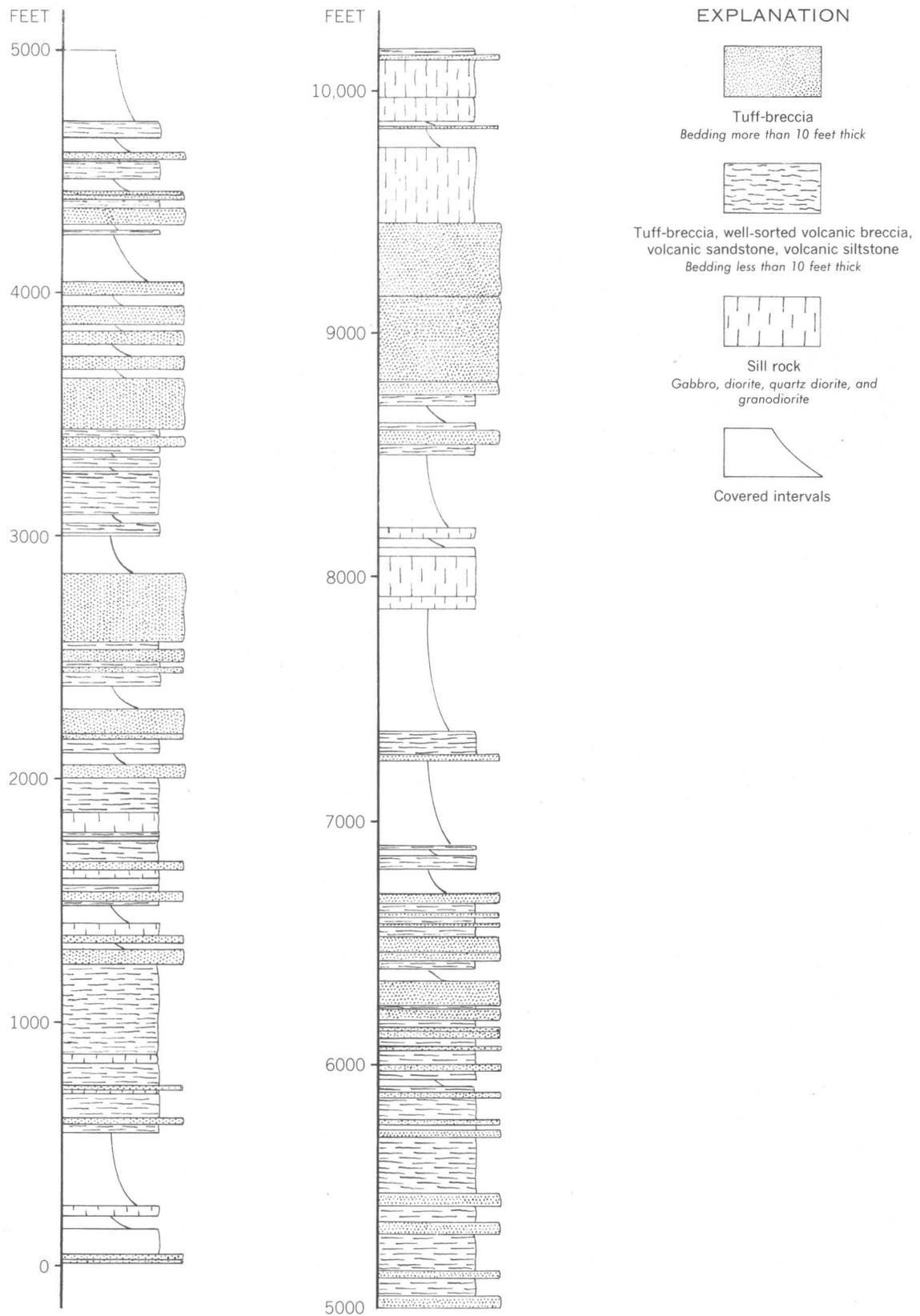
FIGURE 13.—Columnar section of the Ohanapecosh Formation along Stevens Canyon and White Pass highways, Washington.
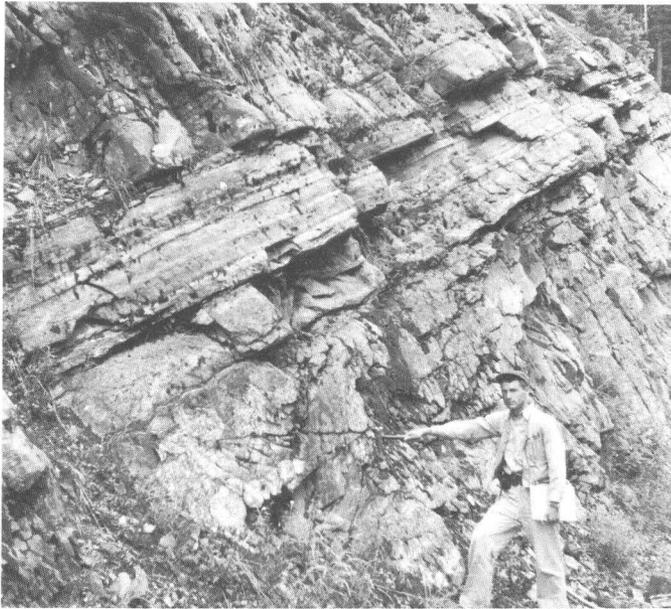
FIGURE 14.—A 20-foot sequence of thin-bedded volcanic sandstones, volcanic siltstones, and tuff-breccias in the Ohanapecosh Formation on the east side of Backbone Ridge.

ASH FLOWS AND RHYOLITES

The only ash flow seen in the measured section of the Ohanapecosh Formation is 40 feet thick and is exposed near the top of Backbone Ridge. It contains abundant and alined pumice lapilli and scattered phenocrysts of quartz, plagioclase, and potassium feldspar. It is similar to the rhyodacitic ash flows that form a large part of the overlying Stevens Ridge Formation. Highly welded ash flows devoid of quartz phenocrysts crop out on the southeast slope of Buell Peak and along the Cascade crest near the head of Panther Creek.

Rhyolite flows are also scarce in the Ohanapecosh Formation within Mount Rainier National Park. Just east of Indian Bar, however, a lens of light-buff rhyolite, 300 feet thick and 2 miles long, crops out in a north-south trending belt. The rhyolite shows contorted flow banding and spherulites; originally it contained phenocrysts of plagioclase in an extremely glassy base, but metamorphism has altered the phenocrysts to albite and transformed the glassy base into an intergrowth of quartz, albite, laumontite, celadonite, saponite, and sphene.
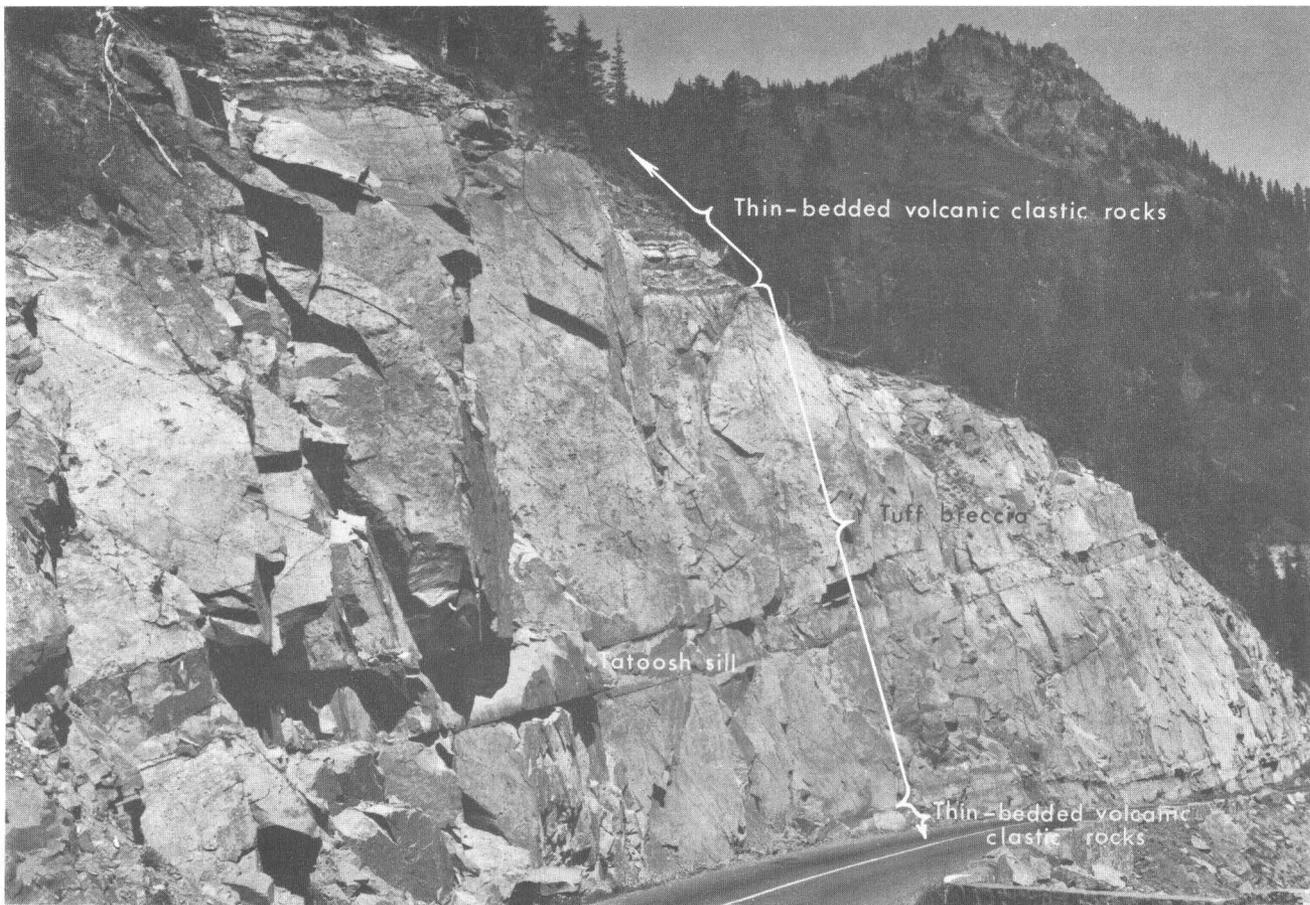
FIGURE 15.—Tuff-breccia, 90 feet thick, from the Ohanapecosh Formation. Roadcut between Cayuse Pass and Chinook Pass. This conspicuous unit lies between thin-bedded volcanic clastic rocks. Repetition of thick tuff-breccias with sequences of thin beds is common throughout much of the Ohanapecosh Formation. A slightly discordant sill, related to the Tatoosh pluton, cuts through the tuff-breccia. Photograph by V. R. Bender.
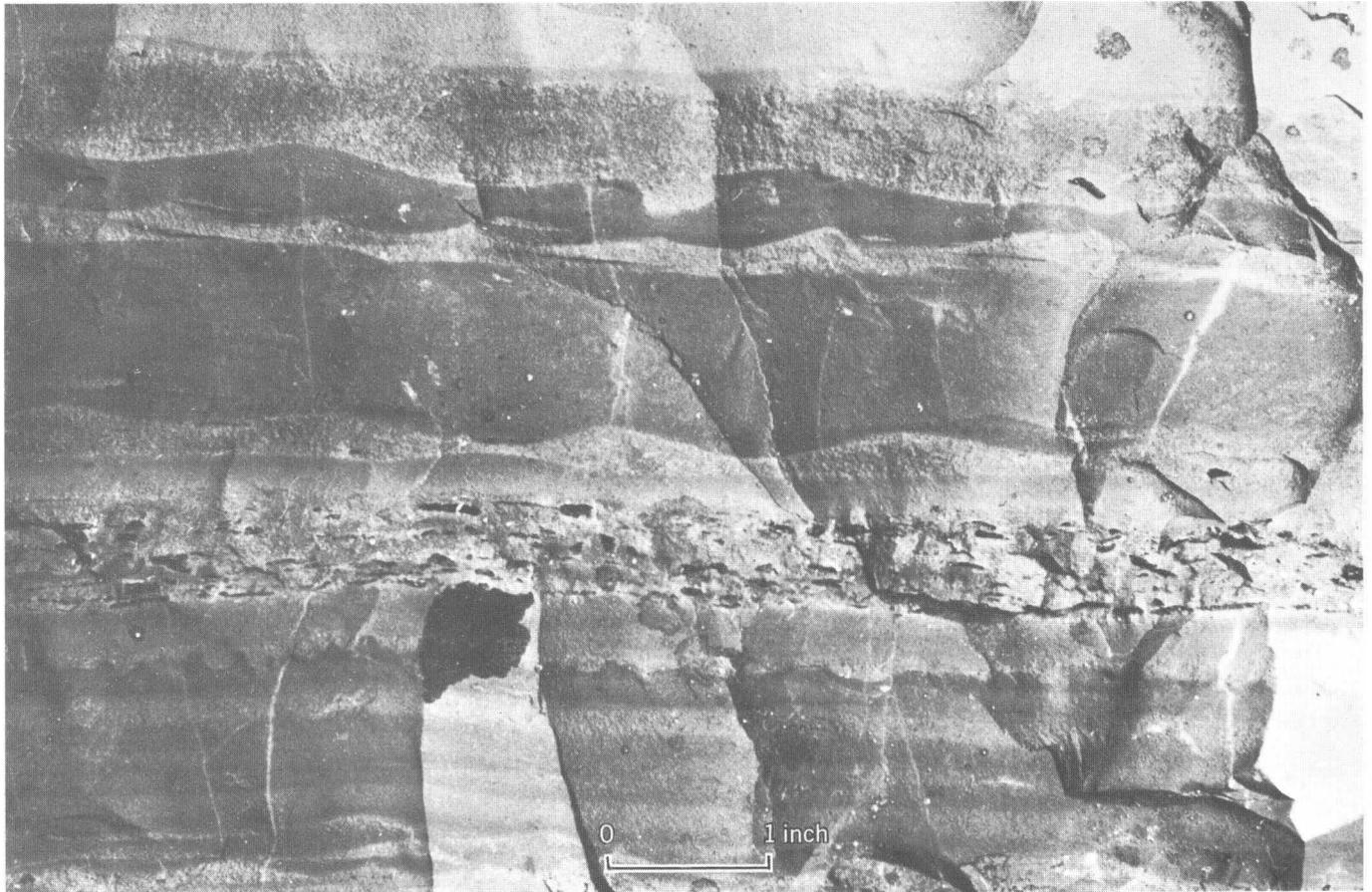
FIGURE 16.—Interbedded volcanic sandstone and siltstone (dark) of the Ohanapecosh Formation from the east side of Backbone Ridge. Note the prominent ripple marks in the upper half of the photograph, the load casts of sandstone bulging downward into siltstone (upper center and at the base of the stratum partly obscured by a dark stain), and abundant flattened pumice lapilli in the sandstone bed extending across the center of the photograph.

ORIGIN

The origin of the Ohanapecosh Formation poses several problems. The most important is the source of the tremendous volume of tuff-breccia and interstratified thin-bedded volcanic clastic rocks. The lava flows and coarse mudflow deposits are relatively minor in volume and are clustered near local centers of volcanism. Did these same volcanic centers supply the more widespread finer grained debris, and if they did, how was it transported from the vents, and in what kind of depositional environment was it eventually accumulated in such quantity?

Most of the volcanic clastic rocks in the Ohanapecosh Formation were probably deposited in water, although broad areas within the basin may have been dry land for brief periods. The lava flows in the Mount Wow and Sarvent lava complexes do not show pillow structures, but some of the flows are splintered and shattered as might be expected if they had been erupted under water. Farther away from the lava complexes deposition in water is indicated by well-developed stratifi-

cation in the thinner bedded parts of the sequence and by numerous graded beds among these thin layers (Fiske and Matsuda, 1964). Shallow channels, crossbeds, ripple marks, and slump structures are found in places, but are rare. Poor development of the current-formed structures indicates deposition by weak and intermittent currents such as might be found in a large lake or a broad sheltered embayment of the sea. The depth of water in submerged parts of the depositional basin undoubtedly varied during Ohanapecosh time. Evidence of weak current action at some stratigraphic levels suggests a shallow-water environment, but most deposits lack these structures and probably formed in deeper water.

The thick tuff-breccias in the Ohanapecosh Formation are interlayered between delicately laminated siltstones and thin graded sandstones, indicating that they also were deposited under water. These extensive sheetlike deposits marked by poor sorting, vertical grading, and rare load casts are subaqueous volcanic mudflows. Like turbidity currents, they appear to have

traveled in response to gravity as dense mixtures of water and volcanic debris. In contrast, however, the subaqueous mudflows carried coarser material, are not as conspicuously graded, and formed layers considerably thicker than those deposited by turbidity currents.

The composition of the Ohanapecosh debris shows clearly that it came from a volcanic source and the abundance of fragile fragments, such as shards and angular shreds of pumice, indicates that most of it was transported and deposited immediately after explosive eruptions. The eruptions that supplied the debris, however, may have occurred either under water or on land. Volcanic clastic rocks of the kind often called epiclastic—those laid down chiefly by mudflows or currents of water—constitute the bulk of the formation. But most of the debris in the Ohanapecosh Formation undoubtedly was supplied by explosive eruptions and not by the subaerial erosion of previously deposited volcanic rocks. Thus, practically all the volcanic clastic rocks in the Ohanapecosh Formation are pyroclastic—some derived from subaqueous eruptions, others from eruptions on land.

It is not always possible to judge whether the debris in a given Ohanapecosh deposit was erupted under water or on land. Some of the mudflows may have formed from eruptions on land (Anderson, 1933) before they entered the submerged basin. Flash floods on the slopes of emergent volcanoes could have eroded both fresh pyroclastic debris and weathered material to form large mudflows that traveled from land far into the submerged basin. Hot nuées ardentes and ash flows erupted on land could have quenched and changed into subaqueous volcanic mudflows on entering water. These processes, however, seem inadequate to supply the huge volume of uniformly stratified debris in the Ohanapecosh Formation. It is more probable that the bulk of the debris was produced by explosive eruptions from underwater volcanoes. Such subaqueous eruptions could have supplied great volumes of quenched pyroclastic debris, which was then carried into deeper water by the subaqueous volcanic mudflows. Additional details that indicate underwater eruptions as the source of both the thin bedded sequences and the thick mudflow layers of the Ohanapecosh Formation are given in Fiske (1963).

Deposition occasionally exceeded basin subsidence, and parts of the basin stood above water. During these periods nuées, ash flows, and normal subaerial mudflows traveled on land. Continued subsidence ultimately submerged these deposits, and deposition of subaqueous mudflows was renewed.

The only Ohanapecosh source areas that have been discovered in Mount Rainier National Park are the Sarvent and Mount Wow lava complexes, and at least a part of the Ohanapecosh fragmental material appears to have been derived from these volcanic centers. The lavas and coarse mudflows that are clustered near the vents represent the less mobile extrusive products that formed local islands in a sea of more mobile volcanic debris.

The most spectacular physiographic feature in the Sarvent lava complex is the South Cowlitz Chimney, a prominent spire soaring 500 feet above its immediate surroundings (fig. 17). This pinnacle is composed of flow-banded rhyolite; the banding is contorted, but generally stands at high angles. As this mass is clearly a plug in one of the Sarvent vents, and consists of rock very similar to many of the felsite fragments in the Ohanapecosh clastic rocks, it may well mark the source of part of the fragmental debris.

Five other vents have been mapped in and near the Sarvent area. Two of these vents (Double Peak and the knob 1,000 ft west of Barrier Peak) contain highly brecciated basaltic andesite lithologically similar to the Ohanapecosh lava. The remaining three are small plugs of basaltic andesite, with minor satellitic dikes, which lie on a north-south line just east of the Cowlitz Divide. Other centers of volcanism, some of them outside the park and perhaps others not exposed at the present erosion surface, have doubtless contributed debris to the Ohanapecosh Formation. Volcanic vents are probably present in the Mount Wow lava complex, though none have been positively indentified.

AGE

No identifiable fossils were found in the Ohanapecosh Formation within Mount Rainier National Park. About 2,000 feet west of the park boundary on the north side of the North Puyallup River (NW¼NE¼ sec. 21, T. 16 N., R. 7 E.), the following plant fossils were collected from volcanic clastic rocks of the Ohanapecosh Formation and were identified by Roland W. Brown:

Sabal sp.
Carya sp.
Castanea sp.
Chaetoptelea sp.
Cercidiphyllum sp.
Hydrangea sp.
Platanus sp.
Cinnamomum sp.
Laurus sp.
Sassafras sp.
Liquidambar sp.

According to Dr. Brown (written communication, 1960), this floral assemblage indicates a late Eocene age. The southeastward projection of these fossiliferous strata into the park passes near Round Pass and through the saddle just north of Tumtum Peak (fig. 18).

Fisher (1957) describes additional fossil localities in areas south and southwest of the park. Fisher's Copper Creek locality (NE¼ sec. 19, T. 15 N., R. 7 E.) occurs in arkosic sandstones of the Puget group which intertongue with the Ohanapecosh Formation. The fossil locality lies stratigraphically below the oldest Ohanapecosh rocks exposed in the southwest corner of the park. Roland W. Brown (cited in Fisher, 1957) assigned an Eocene, "perhaps late," age to these rocks. Fisher's Packwood locality (SW¼ sec. 8, T. 13 N., R. 9 E.) appears to lie at about the same stratigraphic position as the North Puyallup River locality of this report. Brown dated the Packwood floral assemblage as late Eocene.

Ellingson[3] cites a fossil collection taken by Wolfe from a tongue of coal-bearing arkose in the Ohanapecosh Formation on Summit Creek, about 2 miles southeast of Mount Rainier National Park. The arkose lies about 1,500 feet stratigraphically below the oldest Ohanapecosh rocks exposed in the southeast corner of the park. Wolfe assigned a probable early Eocene age to the flora, but Roland W. Brown (written communication, 1960) states that the list of species, as quoted by Ellingson, "would appear to indicate a middle Eocene or perhaps a slightly younger age." Since Dr. Brown identified the fossil plants from all the other localities in the Mount Rainier area, as well as from many other Tertiary localities in the Pacific Northwest, his estimate of the age of the Summit Creek flora is accepted for this report.

---

[3] Ellingson, J. A., 1959, General geology of the Cowlitz Pass area, central Cascade Mountains, Washington: Washington Univ. [Seattle], unpublished M.S. thesis.

FIGURE 17.—South Cowlitz Chimney, a plug of flow-banded rhyolite in the Sarvent lava complex. The south (right) side of the plug flares outward, truncating and overriding nearly horizontal bedded rocks of the Ohanapecosh Formation. The plug's sheer rock face is about 500 feet high.

EXPLANATION

Rocks of Mount Rainier volcano
Glacial ice not shown

Tatoosh pluton
Only main body in southern
part of the park is shown

Stevens Ridge and Fifes Peak Formations
and possible correlatives

Ohanapecosh Formation and interbedded
arkose of the Puget Group

30
Strike and dip of Ohanapecosh strata

Index to localities

1. North Puyallup River
2. Copper Creek (Fisher, 1957)
3. Packwood (Fisher, 1957)
4. Summit Creek (Ellingson, 1959)

Sources of Regional Geology

Fiske, Hopson, and Waters
(this report)

Geology unknown

Modified from
Fisher (1957)

N

FIGURE 18.—Geologic sketch map showing fossil localities in the Ohanapecosh Formation and interbedded arkose of the Puget Group, near Mount Rainier National Park.

The Ohanapecosh rocks cropping out in the south-west corner of the park are no doubt mainly of late Eocene age, but it is possible that a part of the 3,000 feet of unfossiliferous Ohanapecosh rocks overlying the dated strata in the Round Pass-Tumtum Peak area are early Oligocene, and that rocks stratigraphically below the oldest Ohanapecosh rocks in the southeast corner of the park are of middle Eocene age.

The rocks of the Ohanapecosh Formation in Mount Rainier National Park may therefore range from middle Eocene to early Oligocene in age, but available data suggest that they are mostly of late Eocene age.

STEVENS RIDGE FORMATION

DEFINITION, DISTRIBUTION, AND THICKNESS

The Stevens Ridge Formation is named for Stevens Ridge, a prominent spur just north of Stevens Canyon, in the south-central part of Mount Rainier National Park (fig. 19).

The Stevens Ridge rocks are not present in large volume within the park, but they crop out at many scattered localities. The principal areas are on Back-bone Ridge, on Stevens Ridge, along the crest of the Tatoosh Range, on the northeastern part of Indian Henrys Hunting Ground, and in numerous small areas in the northern and northwestern part of the park.

The thickness of the Stevens Ridge Formation ranges from 450 feet on the northeast side of Unicorn Peak to about 3,000 feet in the Northern Crags, near the source of the Carbon River.

FORMATION BOUNDARIES

The Stevens Ridge Formation lies unconformably upon rocks of the Ohanapecosh Formation. The overlying Fifes Peak Formation is concordant, and presumably conformable, with the Stevens Ridge rocks.

The angular discordance between the Stevens Ridge Formation and the underlying Ohanapecosh rocks is highly variable within Mount Rainier National Park. On Backbone Ridge, in the southeast corner of the park, it ranges from 0° to 20°. The two formations are nearly concordant along the Muddy Fork of Cowlitz River and at Indian Bar. In the northern part of the park the discordance increases. On McNeeley Peak in the Sourdough Range, it is about 15°; near the head of the Carbon River it is 60°, and at some places in the canyon of the North Mowich River it is almost 90°.

The contact between the Stevens Ridge and the overlying Fifes Peak Formation is poorly exposed in the park. At Unicorn Peak, Fifes Peak lava flows concordantly overlie Stevens Ridge volcanic clastic rocks. In the Mother Mountain-Ipsut Creek area, the contact
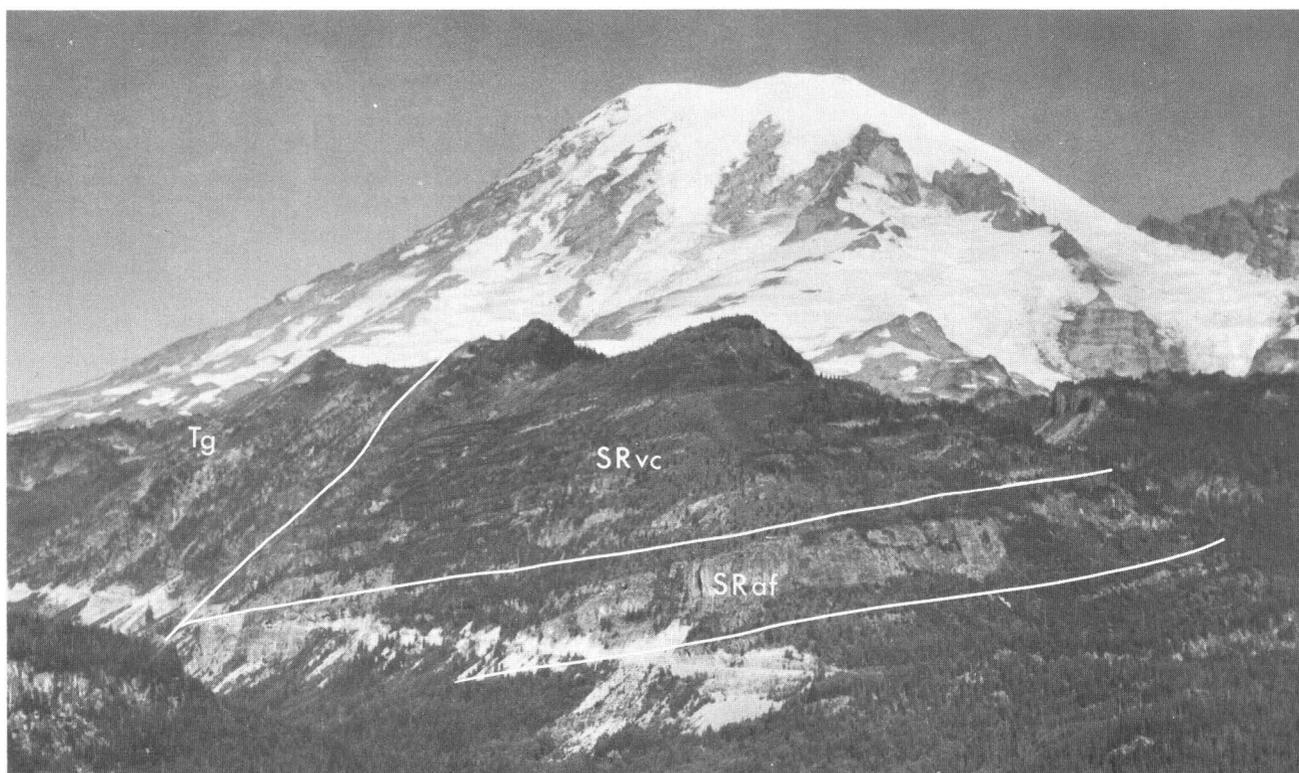
FIGURE 19.—Stevens Ridge Formation at its type locality. Ash flows and intercalated volcanic clastic rocks (SRaf) at the base are concordantly overlain by well-bedded volcanic clastic rocks (SRvc). The top of the formation is not preserved at this locality. Rocks of the Tatoosh pluton (Tg) sharply crosscut the bedded rocks of the Stevens Ridge Formation.

between the 2 formations is nearly planar over an area of 12 square miles. The Stevens Ridge and Fifes Peak Formations are therefore concordant and probably conformable.

LITHOLOGY

The Stevens Ridge Formation consists of ash flows and volcanic clastic rocks. Most of these rocks contain quartz, and they are generally light colored; both characteristics help to distinguish them from the darker colored generally quartz-free rocks of the Ohanapecosh and Fifes Peak Formations.

The generalized columnar sections (fig. 20) show the nature of the Stevens Ridge Formation at six localities within the park. Ash flows occur mainly in the lower part of the formation. Volcanic clastic rocks are most abundant in the upper part, but they also occur as intercalations between ash flows and as thin accumulations near the base of the formation.

ASH FLOWS

Distinctive features of the Stevens Ridge ash flows are their generally light color, the abundance of flattened and alined pumice fragments, and large round phenocrysts of quartz (fig. 21). Their colors on fresh fractures are white to dark gray; weathered surfaces are light buff to gray. The fresh rock is extremely firm; the weathered rock is softer and may be fissile. In some ash flows this firmness is due to sintering and welding at the time of emplacement, but devitrification and other secondary changes have also indurated the rocks. Individual ash flows are from 15 to 350 feet thick. Still thicker ash flows—or cooling units as defined by R. L. Smith (1960)—may be present in the Mother Mountain-Northern Crags area, but detailed study was impossible in that area where poor exposures alternate with nearly vertical cliffs. The tops of individual ash flows are nearly planar. Rude columnar jointing is developed in a few of the thicker flows.

Phenocrysts form 20 to 65 percent of the ash flow rocks. Most of them consist of subhedral and euhedral plagioclase, which shows faint oscillatory and progressive zoning. The average composition of individual crystals ranges from about An_{44} to An_{32}; the composition of most crystals is about An_{36}. Anhedral quartz phenocrysts are rarely absent and often form 15 percent of the rock. Some are as much as 8 mm in diameter, and generally show rounded and deeply embayed outlines. Potassium feldspar phenocrysts rarely compose more than 5 percent of an ash-flow deposit, and appear to be absent in many. Ferromagnesian minerals are rare in most of the ash flows; but some of them contain sparse grains of biotite, hornblende, augite, and hypersthene.

Flattened pumice fragments are common (fig. 21), and give a vague banding to some rocks. They are about 1 inch in average length, though some are as much as 12 inches long. Foreign rock fragments are generally scarce; but they are fairly common in the darker colored ash flows, and in a few they constitute 50 percent or more. They rarely exceed 4 inches in diameter and generally consist of porphyritic hornblende andesite.

Glass shards and small pumice fragments, which formed the original matrix of the ash flows, have been devitrified and altered. Relict textures show partial outlines of the thin-walled vesicles in bits of pumice and rare remnants of the delicate shards in half of the 30 thin sections studied; devitrification and alteration have obliterated most of the primary textures in the remainder.

Alteration and devitrification has greatly modified the mineral composition. The plagioclase is partly or completely replaced by albite, zeolites, carbonate, epidote, and clay minerals, and the ferromagnesian minerals by chlorites, epidote, and carbonate. Pumice lapilli and the glassy matrix of these rocks have been almost completely devitrified. Quartz, albite, and potassium feldspar are the most common devitrification products; carbonate, chlorites, celadonite, and other clay minerals are present in varying amounts. Finely granular celadonite is often concentrated in pseudomorphs after pumice fragments.

Petrologic classification of the Stevens Ridge ash flows is difficult without chemical analyses. The plagioclase-potassium feldspar ratio of phenocrysts averages 20 to 1, but sodium cobaltinitrite staining of devitrified matrix material shows that moderate to large amounts of potassium feldspar are present. The ash flows are therefore tentatively classified as rhyodacitic.

VOLCANIC CLASTIC ROCKS

The Stevens Ridge volcanic clastic rocks are composed of fragments ranging from several feet in diameter to sand- and silt-sized particles. The finer grained material predominates. It is well bedded, and commonly shows festoon cross-stratification and cut-and-fill structures. The beds range in thickness from a fraction of an inch to more than 30 feet and average about 1 foot. Layers containing crossbeds are generally 4 to 6 inches thick. Coarse-grained material occurs mainly in thick-bedded tuff-breccias, but it forms only a small part of the Stevens Ridge Formation.

Pumice lapilli, quartz and plagioclase crystals, and fragments of glassy lava compose most of the recognizable clasts in the thinly stratified rocks. Devitrified ash forms the bulk of the matrix material.
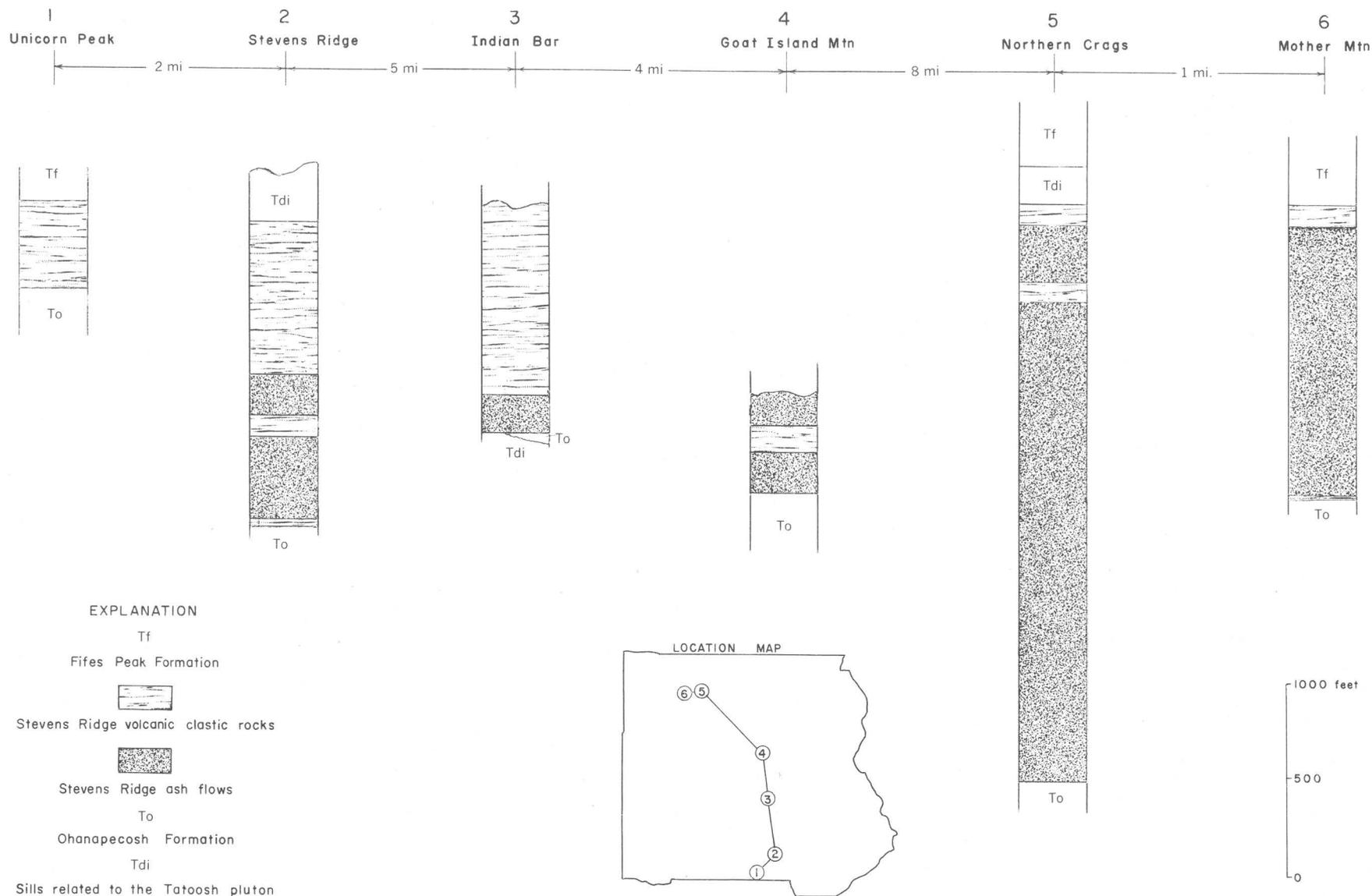
FIGURE 20.—Generalized columnar sections of the Stevens Ridge Formation.

The volcanic clastic rocks in the Stevens Ridge Formation are probably of both epiclastic and pyroclastic origin. The current structures indicate that much of the material was water laid, but the abundance of crystals and pumice lapilli suggests the ultimate pyroclastic origin for much of the debris. Many parallel-bedded strata showing no evidence of current action could be ash-fall deposits.

The volcanic clastic rocks in this formation are even more altered than the ash-flow deposits. Most of the feldspar crystals are partly or completely replaced by zeolites, carbonate, epidote, and clay minerals, and all the ferromagnesian minerals are completely replaced by chlorites and carbonate. The matrix of these rocks is now a finely granular aggregate of quartz, feldspar, the zeolites (wairakite and laumontite), chlorite, and clay minerals.

CONTACT BETWEEN THE STEVENS RIDGE AND OHANAPECOSH FORMATIONS

At many places the unconformity between the Stevens Ridge Formation and the underlying Ohanapecosh Formation is obscured by sills from the Tatoosh pluton which have invaded this zone of structural weakness. The contact is well exposed, however, at three localities: on the west side of the Carbon River just north of Cataract Creek, in roadcuts of the Stevens Canyon highway about a mile west of the Box Canyon, and in highway roadcuts along the eastern crest of Backbone Ridge.

The oldest rocks exposed at the Carbon River locality are steeply dipping dark-green and maroon tuff-breccias of the Ohanapecosh Formation. These rocks are overlain unconformably by about 15 feet of dark-gray and black volcanic clastic rocks of the Stevens
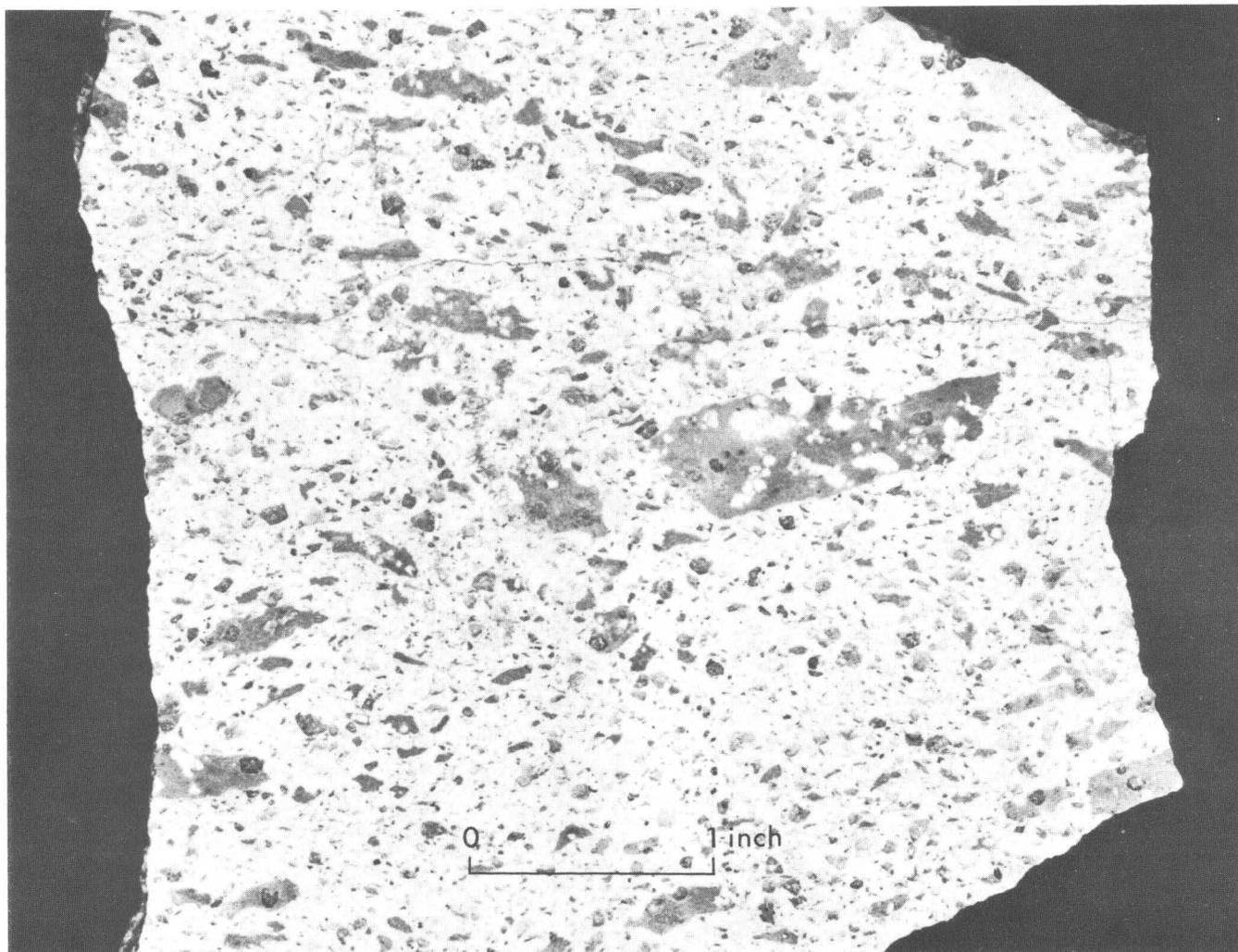
FIGURE 21.—Sawed and polished slab of rhyodacite welded tuff from the thick basal ash flow of the Stevens Ridge Formation at its type locality. Flattened lapilli of porphyritic pumice (dark gray because of alteration and devitrification) are rudely alined parallel to the base of the ash flow. Large phenocrysts of quartz (also dark gray) and plagioclase (white) lie in the devitrified shard-rich matrix.

Ridge Formation, which are, in turn, overlain by 1,500 feet of light-colored ash-flow deposits. Violent emplacement of the lowermost ash flow is indicated by crumpling and brecciation of the underlying rocks. Fragments of the clastic rocks are incorporated in the lower 6 feet of the ash flow; crystals of quartz and plagioclase and bits of altered glassy material from the ash flow have been forcibly injected along bedding surfaces and into fractures crossing the basal clastics.

At the Stevens Canyon locality, a small part of the rugged pre-Stevens Ridge erosion surface is well exposed in highway roadcuts. A steep-sided gully partly filled with boulders and cobbles had been incised into Ohanapecosh volcanic clastic rocks, and a small hill of these older rocks protruded upward; both are now buried under the basal Stevens Ridge ash flow. As at the Carbon River locality, the first Stevens Ridge deposits were thin-bedded pumice and ash-rich clastics. These deposits covered the gully previously cut into the Ohanapecosh rocks, but were not thick enough to bury the hill on the unconformity surface. An ash-flow sheet 350 feet thick then blanketed the area, tore up and incorporated fragments of the basal Stevens Ridge clastic rocks, and completely overwhelmed the small hill of Ohanapecosh rocks.

The contact between the Stevens Ridge and Ohanapecosh Formations on Backbone Ridge is one of the most spectacular outcrops in Mount Rainier National Park. Here, the basal ash flow of the Stevens Ridge Formation lies directly upon slightly truncated Ohanapecosh volcanic sandstones, shales, and tuff-breccias. The contact relations are well exposed for nearly a third of a mile along the highway in steep-sided roadcuts. At this locality, the surface of unconformity was a broad flat area. It was mantled with unconsolidated mudflow and stream deposits and with a conspicuous brick-red saprolite derived from the Ohanapecosh rocks. Scattered trees grew on this lowland.

The 300-foot ash flow at the base of the Stevens Ridge Formation swept across this landscape with catastrophic violence. The lower part of the swirling hot avalanche of pumice and glass dust churned up the loose debris on the surface, and lobelike tongues of hot ash burrowed into the soft water-saturated saprolite. Fragments of bedrock, saprolite, stream-rounded pebbles, and bits of macerated wood were swirled upward as much as 50 feet into the base of the ash flow (fig. 22). The existing vegetation was destroyed; some of the larger tree trunks and limbs are still recognizable, but they were engulfed in the base of the ash flow and thoroughly charred. Crystals of quartz and feldspar from the ash flow are embedded deep within the charred wood fragments; they were

also forced downward into the underlying bedrock and into local pockets of surficial debris that were incorporated in the flow (fig. 23).[4] The thickness of the chaotic mixture of ash flow and underlying material ranges from 0 to 50 feet and averages 20 feet. Above this zone, the ash flow is homogeneous, massive, and free from foreign fragments; below it, the bedded rocks of the Ohanapecosh Formation are undisturbed.

The basal Stevens Ridge ash flow on Backbone Ridge was probably slightly welded immediately after emplacement, and subsequently it was thoroughly indurated into a firm rock by devitrification and secondary alteration (fig. 21). This prominent cliff-forming member is traceable from the Backbone Ridge locality to Stevens Canyon, a distance of 4 miles, and its lithologic character remains practically unchanged within this interval.

The conspicuous brick-red saprolite formed by subaerial weathering of Ohanapecosh rocks before the deposition of the Stevens Ridge Formation is exposed at two other localities. Relict textures of Ohanapecosh tuff-breccias and volcanic sandstone are well preserved in soft clayey saprolite exposed in highway roadcuts on the west slope of Backbone Ridge (SE¼SW¼ sec. 31, T. 14 N., R. 10 E.). A similar, but less well exposed, saprolite crops out beneath the basal Stevens Ridge ash flow, about half a mile west of the Indian Bar shelter cabin.

ORIGIN

The thin accumulations of pumice and ash-rich volcanic clastic rocks at the base of the Stevens Ridge Formation at Stevens Canyon and at Mother Mountain probably represent ash falls that resulted from the earliest volcanic activity in Stevens Ridge time. Some of this pyroclastic debris was transported to low areas and stratified by running water.

The ash flows forming the main part of the Stevens Ridge Formation contain at least 25 cubic kilometers[5] of rhyodacitic pyroclastic material, and probably much more. The wide area covered by some of the individual ash flows is indicated by their great extent as cliff

---

[4] Goodspeed, Fuller, and Coombs (1941) interpreted these relations as due to metasomatic replacement of a coaly sediment to the point where it resembled a dacite breccia. We concluded that although most plagioclase crystals have been partly altered to laumontite, and some secondary quartz and albite might have been introduced from the devitrifying ash flow, most of the quartz and feldspar crystals did not grow by replacement, but were physically injected and kneaded into the soft red saprolitic mud and underlying Ohanapecosh rocks during emplacement of the ash flow.

[5] The volume estimate of 25 km³ assumes an average thickness of 400 feet over an area of 80 square miles. This estimate of area is the sum of the outcrop areas within the park, probable underground extensions between major areas of exposure, and small areas in which the Stevens Ridge rocks were engulfed by the Tatoosh pluton. This figure, however, is much too low, because it fails to take account of eroded extensions of the ash flows or known accumulations in areas north and east of the park.

formers. The source vents for these ash flows have not yet been identified.

The volcanic clastic rocks intercalated with and overlying the ash flows were probably derived in part from ash falls and in part from the attrition of nearby ash-flow deposits. Running water sorted and stratified much of this material and produced well-developed crossbedding and cut-and-fill structures.

The Stevens Ridge Formation was deposited upon a surface of considerable relief. From the generalized columnar sections of the formation (fig 20) it has been inferred that this relief was at least 1,500 feet. This inference is based on the assumption that the Stevens Ridge and Fifes Peak Formations are conformable and that the upper surfaces of the ash flows are nearly planar; these two conditions are everywhere observed to exist in the field. The disapperance of the ash flows between Stevens Ridge and Unicorn Peak and the 1,500-foot thinning of the ash flows between Northern Crags and Mother Mountain are therefore probably due to the ruggedness of the topography at the beginning of Stevens Ridge time.

AGE AND CORRELATION

No diagnostic fossils were found in the Stevens Ridge Formation in Mount Rainier National Park. Conifer needles collected from Stevens Ridge clastic rocks at several localities do not indicate the age of the formation (Roland W. Brown, written communication, 1960).

The age can be roughly estimated, however, from tentative correlations with dated fossiliferous rocks in two areas near the park. The first locality lies 1.5 miles north of Tieton Reservoir dam in the Mount Aix quadrangle and 15 miles southeast of the boundary of the park; the second is three-quarters of a mile north of Packwood Wash., and 8 miles south of the park.

At the Tieton Reservoir, Grant (1941) discovered a well-preserved lower jaw of the middle Oligocene to lower Miocene oreodont, *Eporeodon*, in light-colored tuffaceous sandstones and siltstones that lie stratigraphically below Fifes Peak lava flows and coarse mudflow deposits. During a reconnaissance of the Tieton Reservoir locality we found that these tuffaceous sedimentary rocks lie concordantly beneath, and are in places intercalated with, the basal Fifes Peak lava flows

FIGURE 22.—Lower part of the basal ash flow, Stevens Ridge Formation, on Backbone Ridge. The light-colored ash flow churned through soft saprolite on the pre-Stevens Ridge land surface and included fragments of the saprolite and less weathered Ohanapecosh rocks in the lower 8 feet of the deposit.

and coarse mudflow deposits. Similar stratigraphic relations also appear in the Fifes Peaks area to the northwest, where Warren (1941) first defined the Fifes Peak Andesite. Here light-colored tuffaceous sedimentary rocks, locally interbedded with ash flows and rhyolite, lie beneath the lavas and mudflows that make up the Fifes Peak Andesite.

Abbott (1953) included all these varied tuffaceous sedimentary rocks and acidic lavas in his Fifes Peak Andesite. He recognized a definite angular discordance between them and a still older series of fragmental volcanic rocks. This older series, called lower Keechelus by Warren (1941) and Abbott (1953), is in part equivalent to the Ohanapecosh Formation of this report, but the original lower Keechelus of Smith and Calkins (1906, p. 8) consists mainly of equivalents of the Fifes Peak and Stevens Ridge Formations (Waters, 1961).

The Stevens Ridge Formation in Mount Rainier National Park is also a light-colored rhyodacitic unit lying conformably beneath the lavas of the Fifes Peak Formation and unconformably above the Ohanapecosh Formation. It therefore appears probable that the basal tuffaceous part of the Fifes Peak Andesite as described by Abbott in the Mount Aix quadrangle and the rocks containing the oreodont remains near Tieton Reservoir were deposited during the same episode of acidic volcanism that formed the Stevens Ridge Formation in Mount Rainier National Park. The validity of this correlation, however, cannot be accepted without reservation until the complex stratigraphic relations between Mount Rainier National Park, Fifes Peaks, and Tieton Reservoir have been studied in detail.

In the Packwood area, gently dipping tuffaceous siltstone and sandstone unconformably overlie more steeply

FIGURE 23.—Lower part of the basal ash flow, Stevens Ridge Formation, on Backbone Ridge. Closeup showing chaotic intermixture of ash flow (light colored) and fragments of saprolite from the pre-Stevens Ridge land surface. Crystals of quartz and feldspar from the ash flow are deeply imbedded in many of the included fragments. Photograph by Richard E. Fuller.

dipping Ohanapecosh strata. An extensive tabular complex of fine-grained quartz diorite (identical with many Tatoosh sills in the park) has been intruded along this unconformity and has engulfed parts of the overlying rocks. In the park, also, the Stevens Ridge Formation unconformably overlies the Ohanapecosh Formation, and the contact is obscured by many sills. These similarities suggest that the rocks near Packwood may be equivalent to a part of the Stevens Ridge Formation, but the area between Packwood and the park must be mapped before we can fully judge the value of this evidence. The Packwood locality [6] has yielded a fresh-water snail identified by D. W. Taylor (written communication, 1960) as *Viviparus*. According to Taylor,

Within the late Eocene to mid-Miocene age range suggested [letter from R. S. Fiske, Jan. 12, 1960], a younger age is more likely than an older, about late Oligocene or Miocene. So far as the material alone goes, the age could be Oligocene to early Pliocene.

The age range suggested by this one fossil overlaps that indicated by the *Eporeodon* found north of Tieton Reservoir dam. Thus at the Packwood locality, as at the Tieton locality, some indication of age is afforded by a single fossil. The Stevens Ridge Formation is therefore tentatively regarded as having been deposited sometime in the period extending from middle Oligocene through early Miocene.

FIFES PEAK FORMATION

DEFINITION

Warren (1941) assigned the name Fifes Peak Andesite to a heterogeneous assemblage of lava flows, mudflows, and volcanic clastic rocks that crop out in the eastern and northern parts of the Mount Aix quadrangle, which borders the Mount Rainier 30-minute quadrangle on the east. According to Warren (1941, p. 800), "The proportion of flows to agglomerates and tuffs varies greatly from place to place; on the whole, the pyroclastic materials undoubtedly have the greater volume." Warren thought the Fifes Peak Andesite to be correlative with the upper and less altered part of the Keechelus Andesitic Series, defined by Smith and Calkins (1906) in the Snoqualmie quadrangle (just north of the Mount Aix quadrangle), but later work (Waters, 1961) indicates that the Fifes Peak and the Stevens Ridge Formations are parts of Smith and Calkins' lower Keechelus). Warren (1941, p. 799, 800) further assumed that a still older and more altered group of rocks, which lie unconformably beneath the

[6] USGS Cenozoic loc. 22021, Lewis County, Wash. Mount Rainier quadrangle, 1 : 125,000. NE¼ SW¼ sec. 15, T. 13 N., R. 9 E. Small quarry on west side of Washington State Highway 5, about three-quarters of a mile north of Packwood. R. S. Fiske, 1959.

Fifes Peak Andesite in the Mount Aix quadrangle, was "the lower part of the Keechelus" of Smith and Calkins. These so-called lower Keechelus rocks of Warren are at least in part correlative with the Ohanapecosh Formation.

Abbott (1953) extended the mapping of the Fifes Peak Andesite to the northwest corner of the Mount Aix quadrangle. He included in it not only the lavas and volcanic clastic rocks of andesitic and basaltic composition, but also acidic tuffaceous rocks that underlie them. Abbott predicted that extensions of the formation would be found in the Cedar Lake quadrangle, directly north of Mount Rainier National Park. The present study substantiates Abbott's prediction and demonstrates that both the andesitic lavas and the underlying acidic tuffaceous rocks also extend southwestward into Mount Rainier National Park. Here, however, we have subdivided Abbott's Fifes Peak Andesite: we mapped and defined the acidic tuffaceous rocks as the Stevens Ridge Formation and mapped the overlying andesitic and basaltic lavas, together with minor amounts of volcanic clastic rocks interbedded with the lavas, as a separate formation. For these andesitic and basaltic rocks we propose the name Fifes Peak Formation. The Fifes Peak Formation includes nearly all the Fifes Peak Andesite as originally defined by Warren; but it excludes the acidic materials beneath it that were mapped as part of the Fifes Peak Andesite by Abbott.

Separation of the Fifes Peak Andesite into the Stevens Ridge and Fifes Peak Formations is useful because the two formations are lithologically distinct and can readily be separated in mapping.

FORMATION BOUNDARIES

The Fifes Peak Formation lies concordantly upon rocks of the Stevens Ridge Formation, and this basal contact is assumed to be conformable. The top of the Fifes Peak Formation is not exposed in Mount Rainier National Park, but the Yakima Basalt of late Miocene and early Pliocene age overlies these rocks along the eastern margin of the Mount Aix quadrangle, about 16 miles east of the park.

DISTRIBUTION AND THICKNESS

The Fifes Peak Formation has been removed by erosion from most of Mount Rainier National Park. Remnants are confined to high altitudes along the trough of the broad Unicorn Peak syncline (pl. 1). The largest remnants of the formation are in the northwestern part of the park near Mowich Lake and at the top of Chenuis and Scarface Mountains. The only

remnant in the southern part of the park is a capping, 1,000 feet thick, at the top of the Unicorn Peak (fig. 24).

The Fifes Peak Formation is 2,400 feet thick in the Castle Peak area three-quarters of a mile east of Mowich Lake. The thickness of other erosional remnants does not exceed 1,200 feet. Reconnaissance in the Cedar Lake quadrangle north of the park, however, indicates that the total thickness may exceed 5,000 feet.

LITHOLOGY

Lava flows form at least 80 percent of the Fifes Peak Formation; volcanic clastic rocks are next in abundance, and ash flows are rare. The lava flows are chiefly in the range from olivine basalt to glassy hypersthene-augite andesite, but a little rhyolite—constituting less than 1 percent of the formation—is interbedded with the more mafic lavas in the area northwest of Mowich Lake. Most Fifes Peak flows are from 50 to 150 feet thick, but one widespread basal flow, 500 feet thick, forms a prominent scarp on the southeast face of Mother Mountain and extends southeastward, across Cataract Creek, until it disappears beneath the Mount Rainier lavas. In relatively small exposures the thinner Fifes Peak lavas appear to be sheetlike in form, but more continuous exposures in the Cedar Lake quadrangle

north of the park show that many flows are lensoid in cross section and seldom more than a mile wide. The bulk of the formation is evidently a plexus of narrow lava streams and mudflows radiating outward from volcanic centers, instead of a sequence of sheetlike flows.

Primary structural features, such as columnar jointing, platy partings, ramp structures, and flow-top breccias, are common in the Fifes Peak lavas. The preservation of these structures and the less thorough alteration of the lavas aid in distinguishing Fifes Peak lavas from the somewhat similar lava complexes within the Ohanapecosh Formation.

Fresh exposures of the Fifes Peak basalts and andesites are generally dark brown, dark gray brown, or greasy black; weathered surfaces are buff brown and reddish brown. The dark colors in fresh outcrop and the reddish tints on weathered rocks are due chiefly to the presence of abundant iron-rich saponite which replaces olivine and other ferromagnesian minerals. These mafic lavas are invariably conspicuously porphyritic (fig. 25). Before alteration, phenocrysts of plagioclase, olivine, augite, and hypersthene were present in nearly every flow—although a few flows contained only olivine and augite and others only hypersthene and au-

FIGURE 24.—Remnants of the Fifes Peak Formation (Tf), 1,000 feet thick, on Unicorn Peak. Lava flows, forming the cliffs, alternate with less resistant volcanic sedimentary rocks. These nearly flat strata lie in the broad trough of the Unicorn Peak syncline. Rocks of the Stevens Ridge Formation (Ts) lie in the foreground.

gite as the ferromagnesian phenocrysts. The euhedral plagioclase phenocrysts generally show faint oscillatory zoning and rather strong progressive zoning; most zones in single crystals are in the range labradorite to calcic andesine, but, in some crystals, zones in the interior part may be as calcic as An_{85}, or as sodic as An_{35}. The range in composition among the microlites is much narrower: in 75 percent of the thin sections studied the microlites range only from about An_{52} to An_{47}, and the extreme ranges in composition of crystals from different flows is only about An_{55} to An_{40}.

Without chemical analyses it is difficult to give precise names to these rocks. Using the composition of the plagioclase as the sole basis of classification, many of them straddle the border between basalts and andesites. The rocks we classify as basalt are rich in olivine and chlorophaeite, show intergranular or intersertal textures, contain abundant clinopyroxene and magnetite in their groundmasses, and generally contain labradorite microlites. The hypersthene-augite lavas with glassy groundmass rich in plagioclase microlites are called andesites; the term "basaltic andesite" is appropriate for numerous flows that contain a little olivine and abundant plagioclase near An_{50}.

The mafic lavas of the Fifes Peak Formation are only slightly to moderately altered. Most of the plagioclase is fresh, but in some flows it is partly altered to clay minerals, carbonate, and very rarely epidote. The augite shows minor alteration to saponite and carbonate; the hypersthene is more extensively replaced by saponite, chlorite, and carbonate; the olivine is completely replaced by saponite or, more rarely, by carbonate. Volcanic glass has wholly devitrified; chlorophaeite is completely decomposed.

The few rhyolite flows in the Fifes Peak Formation are poorly exposed at the head of Falls Creek and in bluffs at the base of Tolmie Peak just west of Eunice Lake. Before alteration these light-buff generally flow-banded rocks contained sparse phenocrysts of plagioclase in a glassy base, but the glass has been completely devitrified to produce abundant fine-grained quartz and feldspar which in places shows incipient spherulitic structures. The plagioclase phenocrysts and parts of the groundmasses of these rocks are replaced by carbonate and clay minerals.

Tuff-breccia and volcanic sandstone form less than 10 percent of the Fifes Peak Formation in Mount Rainier National Park. These rocks occur in pockets
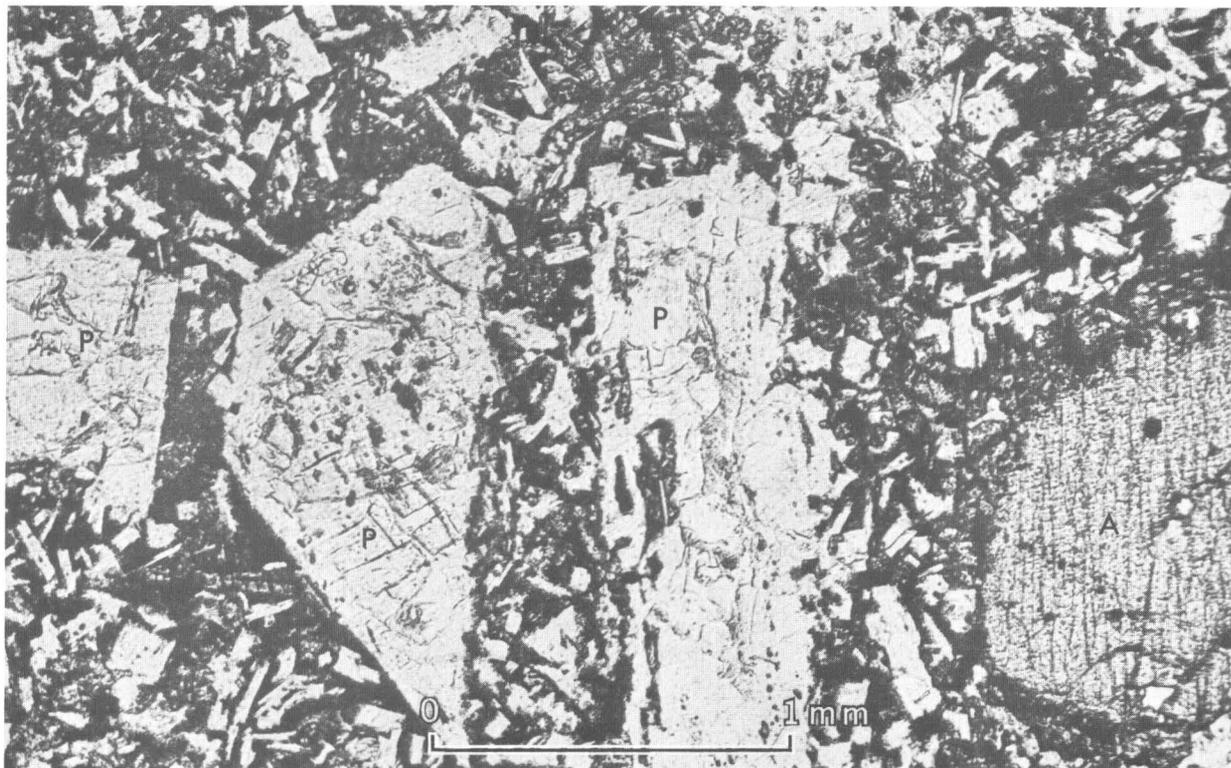
FIGURE 25.—Photomicrograph of porphyritic, intergranular basalt of the Fifes Peak Formation. The phenocrysts are plagioclase (P), augite (A), and hypersthene (not shown), and the microlites are plagioclase and clinopyroxene. Glass, chlorophaeite, and hypersthene are partly altered to clay minerals. The rock comes from the ridge crest 3 miles north of Slide Mountain and about 1 mile north of the park boundary.

and in thin intercalations between the lava flows. They are best exposed on the southeast side of Arthur Peak (2 miles north of Mowich Lake) and near the top of Unicorn Peak. The thickness of these detrital intercalations seldom exceeds 50 feet.

Lithic fragments predominate in both the breccias and the sandstones; they include all varieties of Fifes Peak lavas plus abundant pumiceous material. Broken plagioclase crystals of the kind found as phenocrysts in Fifes Peak lavas are scattered throughout. These fragmental rocks are more severely altered than the lava flows; the alteration products include chlorite, epidote, zoisite, carbonate, zeolites, and clay minerals.

A single ash flow in the Fifes Peak Formation was found along the Mowich Lake road about a half a mile west of Mowich Lake. It is a light-gray rock that contains abundant flattened and alined pumice lapilli as much as 6 inches in length and numerous light-colored fragments of foreign lava, some as much as 3 inches in diameter, in a matrix that originally consisted mainly of glass shards. The rock has been completely devitrified; the alteration products include clay minerals, carbonate, and celadonite.

ORIGIN

The Fifes Peak Formation contains an enormous volume of andesitic and basaltic lava flows. Individual flows were not extensive, but the lava field they formed covered much of this part of the Cascade Range. These lavas must have spread from many centers of volcanism, yet no vents which directly connect with individual piles of Fifes Peak lava have been positively identified in Mount Rainier National Park. A central accumulation of Fifes Peak lava is well exposed, however, on Castle Mountain, 3 miles east of the park in the northwest corner of the Mount Aix quadrangle, where a small dike swarm built an elongate shield volcano of Fifes Peak basalt. On the southwest-facing scarp of Castle Mountain (fig. 26) individual flows can be traced for about 2 miles from the top of the dissected shield.

A prominent dike swarm that cuts the Ohanapecosh and Stevens Ridge Formations on Backbone Ridge is composed of porphyritic hypersthene-augite basalt and diabase, very similar in mineral composition and texture to many of the Fifes Peak lavas. These dikes probably fed a part of the Fifes Peak lava field. Similar dike swarms crop out also on the east slope of Stevens Peak, on Mount Wow in the southwest corner of the park, and at the head of the North Puyallup River. Except for the small remnant on Unicorn Peak, erosion has swept away the lavas that probably were erupted from these dikes.

Some of the thin accumulations of volcanic sediment that alternate with the lavas were probably reworked from nearly contemporary pyroclastic deposits; others probably formed by the subaerial weathering and erosion of Fifes Peak lava flows. Streams redistributed and concentrated this material into thin sheets and pockets.

AGE

No diagnostic fossils were found in the Fifes Peak Formation, but its approximate age can be estimated from its stratigraphic position. The Stevens Ridge Formation, which is of Oligocene or early Miocene age, concordantly underlies the Fifes Peak Formation. Abbott (1953, p. 111, 116) says that the Yakima Basalt overlies the Fifes Peak Formation unconformably in the northern part of the Mount Aix quadrangle; the superposition of the Yakima Basalt upon Fifes Peak Formation in the eastern part of the Mount Aix and Snoqualmie quadrangles has also been recorded by Warren (1941) and Waters (1961). The Yakima Basalt is of late Miocene and early Pliocene age. The age of the Fifes Peak Formation is therefore tentatively assigned to the interval ranging from middle Oligocene to late Miocene; the regional relations to other formations suggest that it is probably early Miocene.

INTRUSIVE DIABASE AND BASALT

Sills and dikes of diabase and basalt invade the Ohanapecosh and Stevens Ridge Formations at many localities in the southern and western parts of the park. Some of these intrusive bodies have been mapped (pl. 1), but many others have not.

The thickest and most extensive plexus of diabase sills crops out in the canyon of the Muddy Fork of Cowlitz River; the diabase is especially well exposed on the walls of Box Canyon, a narrow slot incised by the glacier-fed waters of that stream. These sills form a tabular intrusive complex, at least 5 miles long and about 1,000 feet in maximum thickness, composed mostly of single and multiple sills but containing septa of Ohanapecosh volcanic clastic rocks. The thicker septa are shown on the geologic map, but many smaller ones are unmapped. Basalt and diabase sills also crop out at Longmire and in the western end of the Tatoosh Range east and southeast of Longmire. Those sills whose contacts can be seen are from 15 to 120 feet thick; still thicker sills may be concealed in the poorly exposed areas near the junction of Stevens Creek with the Muddy Fork of the Cowlitz River, and along the park boundary southeast of Longmire. Thick sills are remarkably homogeneous, and they show few joints.

The diabase contains glomeroporphyritic clots of ferromagnesian minerals that form conspicuous dark-

green blotches. The microscopic texture ranges from hypidiomorphic to diabasic and porphyritic; ophimottling is visible in several sills near Longmire.

The chief primary minerals are plagioclase (55 to 75 percent), augite (8 to 35 percent), quartz (0 to 15 percent); the common accessories are titaniferous magnetite and apatite. The plagioclase, which is in the disordered, "high-temperature" state, ranges in composition from about An_{60} to An_{52}; most crystals show normal progressive zoning, but little oscillatory zoning. The augite (in part titaniferous) shows a $2V$ in the range between 52° and 58°. Quartz occurs in irregular interstitial patches. The diabase is not much altered; but albite, epidote, uralitic amphibole, and chlorite are present in small amounts. In some sills, tabular concentrations of uralitic amphibole appear to be pseudomorphs after orthorhombic pyroxene.

Swarms of basalt dikes of the same composition as the sills crop out in the southern and western parts of the park. Four of these dike swarms have been somewhat schematically mapped (pl. 1); but it was impossible to map more than a few of the dikes, most of which are only from 6 inches to 15 feet in thickness. Some multiple dikes, however, are more than 20 feet thick. The dikes are similar in mineralogy and degree of alteration to the sills; a few of the dikes are vesicular.

The sills and related dikes of diabase and basalt are probably a hypabyssal phase of the Fifes Peak lavas. This correlation is based on the mineralogic similarity between these rocks and the Fifes Peak lavas and on the fact that the sills and dikes appear to be of the same age as the Fifes Peak rocks—they cut the Stevens Ridge Formation but are cut by offshoots from the Tatoosh pluton.

FIGURE 26.—Castle Mountain, a deeply dissected Fifes Peak shield volcano in the northwest corner of the Mount Aix quadrangle, just outside the east boundary of Mount Rainier National Park. The lava flows were fed from a plexus of nearly vertical dikes. Three dikes are shown by arrows.

ZEOLITIC METAMORPHISM OF THE OHANAPECOSH, STEVENS RIDGE, AND FIFES PEAK FORMATIONS

The rocks of the Ohanapecosh Formation, and to a lesser degree those of the Stevens Ridge and Fifes Peak Formations, are characterized by greenish-gray and drab-green colors caused by the presence of fine-grained alteration minerals. Most of the rocks, also, are extremely well indurated: fragmental rocks break across clasts and matrix, well-stratified volcanic clastic rocks seldom part readily along bedding planes, and columnar or platy joints in lavas have generally been healed by new minerals. These features are due to a widespread pervasive alteration which has caused extensive recrystallization of the glass and more susceptible minerals. Waters (1955b) noted a similar alteration in the lower Tertiary rocks throughout much of the Cascade Range in southern Washington and northern Oregon; thus, this alteration is of regional extent.

ZEOLITE FACIES

P. Eskola (1939, p. 345) proposed a zeolite facies of metamorphism, but he later discarded the concept because there is a lack of equilibrium in most zeolitized rocks, and because the composition of the secondary mineral assemblage probably depends in part on the composition and concentration of introduced solutions (Turner, *in* Fyfe and others, 1958, p. 216).

D. S. Coombs (1954) described extensive zeolitization of a 28,000-foot section of Triassic volcanic graywackes in southern New Zealand. This area is free of igneous intrusions and has undergone little structural deformation. The main factors promoting the mineralogic transformations were increase of temperature and pressure accompanying deep burial in a geosyncline.

The sequence of mineralogic changes from top to bottom of the New Zealand section is listed by Coombs and others (1959, p. 60) as follows:

1. Alteration of glass in tuffs to heulandite or analcime.
2a. Replacement of the assemblage analcime-quartz by albite-quartz.
2b. Replacement of heulandite by laumontite and of detrital calciferous plagioclase by albite and laumontite.
3. Substitution of pumpellyite and prehnite for laumonite.
4. Post-tectonic precipitation of stilbite in joints.

Despite the simplicity of the geologic setting Coombs emphasizes the "gross overlapping of zones." Traces of heulandite occur deep in the section, and small amounts of epidote and laumontite are found near its top.

Turner (in Fyfe and others, 1958) set up a "zeolitic facies" of metamorphism (which we call the zeolite facies following D. S. Coombs and others, 1959), largely based on the regional zeolitization in New Zealand as described by D. S. Coombs (1954). This facies em-

braces "only regionally developed zeolitic assemblages that largely replace the pre-existing rocks and conform to the mineralogical and chemical requirements of a metamorphic facies * * *" (Turner, *in* Fyfe and others, 1958, p. 216). The zeolite facies is characterized by extensive albitization of plagioclase and by the presence of laumontite in place of heulandite and of quartz and albite in place of analcime. In Coombs' simplified sequence of mineralogic changes listed above, the transition between diagenesis and zeolite metamorphism lies between (1) and (2a).

ZEOLITE FACIES METAMORPHISM IN MOUNT RAINIER NATIONAL PARK

The Ohanapecosh Formation is generally more altered than the overlying Stevens Ridge and Fifes Peak Formations. Fragmental rocks of all three formations are more recrystallized than the massive lava flows and ash flows; yet, even in the most highly altered rocks, the original clastic or igneous textures are well preserved.

Many Ohanapecosh rocks contain either laumontite or wairakite,[7] and varying amounts of albite, quartz, prehnite, chlorite, epidote, and clay minerals. Laumontite or wairakite is commonly concentrated with albite in pseudomorphs after detrital plagioclase; these zeolites plus quartz and albite often replace both detrital matrix materials and glass in rock fragments. Joint fillings of chabazite and stilbite are common in the Ohanapecosh rocks on the east side of Backbone Ridge (W. S. Wise, oral communication, 1959).

Zeolitization can be recognized in the field from several conspicuous features. Detrital plagioclase crystals partly or completely replaced by zeolites and albite are characteristically milky white. Light-gray and buff spots and bands in volcanic siltstones and fine-grained volcanic sandstones indicate concentrations of laumontite or prehnite (fig. 27). Some tuff-breccias, originally rich in pumiceous debris, contain spherical replacement masses of laumontite as much as 5 inches in diameter (fig. 28). Most rocks containing wairakite or prehnite are extremely well indurated, but rocks rich in laumontite are often soft and friable, probably because the laumontite has been softened by hydration during the present weathering cycle.

Common mineral assemblages in Ohanapecosh volcanic clastic rocks include laumontite-albite-quartz (fig. 29), wairakite-albite-quartz (fig. 30), and prehnite-albite-quartz. Common accessories in all these assemblages are epidote, chlorite, carbonate, sphene,

---

[7] W. S. Wise (1959) first identified wairakite as a rock-forming mineral in the rocks of Mount Rainier National Park from samples supplied by us.

leucoxene, celadonite, saponite, and other montmorillonoids. Comparison of the Ohanapecosh volcanic clastic rocks in the park with the thick volcanic graywacke sequence in New Zealand (D. S. Coombs, 1954) shows similarities and differences. Both sequences of rocks originally contained abundant volcanic glass, and both now contain metamorphic mineral assemblages that include laumontite, prehnite, albite, and quartz. The distribution of mineral assemblages, however, is far more complicated in the Ohanapecosh Formation than it is in the rocks described by Coombs. Wairakite and laumontite occur sporadically throughout the upper 8,000 feet of the 10,000-foot measured section (fig. 13), whereas laumontite is the only zeolite observed in the lower 2,000 feet. Prehnite is locally abundant in an interval from 6,000 to 8,000 feet below the top of the section, but it is scarce elsewhere. A few rocks from nearly all stratigraphic levels contain little or no zeolite, but others nearby contain more than 20 percent.

Mineralogic equilibrium is seldom attained in the zeolitized Ohanapecosh rocks. Even though all bits of pumice and originally glassy material have been completely replaced, relict detrital quartz and augite are nearly always partly preserved, and at least some remnants of plagioclase generally survive. Nevertheless, because of the widespread and usually complete replacement of glass and easily susceptible minerals by albite, laumontite, prehnite, and other new minerals, it is evident that the Ohanapecosh rocks belong in the zeolite facies of metamorphism as defined by Turner. The closely associated wairakite-bearing rocks are here included in this facies also.

Owing to the preliminary nature of the present investigation, we offer only a few tentative conclusions regarding the causes of this complex alteration. The zeolite facies minerals in the Ohanapecosh rocks were probably formed by a combination of two main processes: deep burial with accompanying rise in temper-
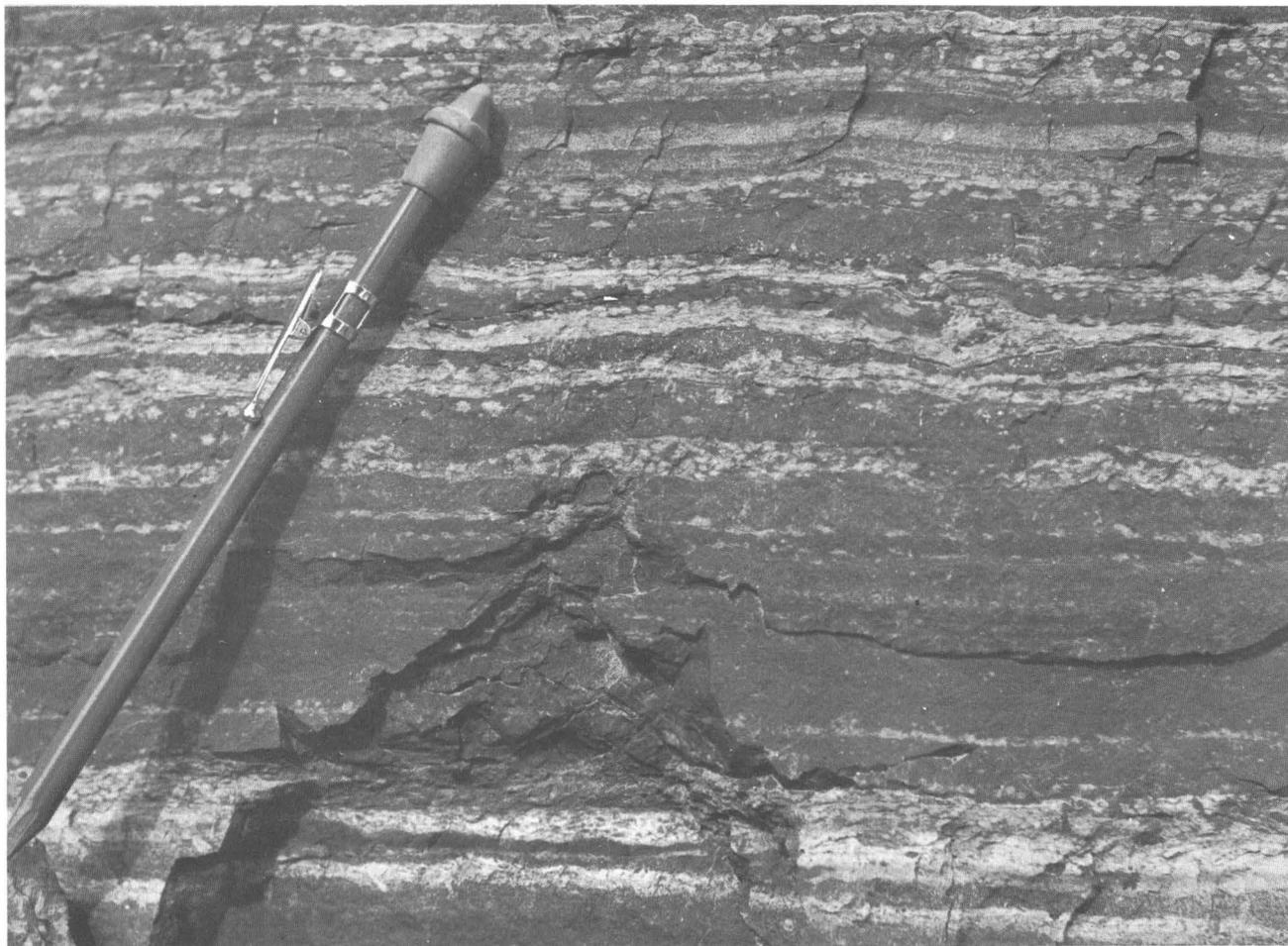
FIGURE 27.—Megascopic features of Ohanapecosh metamorphism. Highly indurated volcanic siltstones on the east side of Backbone Ridge. Concentrations of finely granular prehnite lie parallel to the bedding, forming globular spots which have coalesced into white bands along some bedding planes.

ature and pressure (as in the New Zealand rocks described by Coombs) and far-reaching dispersal of heat from igneous intrusives. The effects of these two processes are superimposed and are difficult or even impossible to distinguish.

The effect of deep burial is indicated by the abundance of zeolites in some areas distant from known or suspected intrusive bodies. Most of this loading and recrystallization probably took place before the deposition of the Stevens Ridge and Fifes Peak Formations, for in most areas the Ohanapecosh rocks are more altered than the overlying rocks regardless of the stratigraphic level of the Ohanapecosh Formation exposed at the unconformity. Loading effects may have been complicated, however, by renewed zeolitization of the Ohanapecosh accompanying or following deposition of the Stevens Ridge and Fifes Peak Formations.

Igneous intrusive masses also had an important part in the metamorphism of the Ohanapecosh Formation. Ohanapecosh rocks were first invaded in middle Oligocene to early Miocene time by diabase sills and dikes related to the Fifes Peak lavas. They were then repeatedly injected, probably in late Miocene or early Pliocene time, by voluminous sills, dikes, and stocks from the Tatoosh pluton. The wide distribution of

these intrusive rocks throughout the park, together with their suspected presence at shallow depths in many other areas, indicates that the regional geothermal gradient in the Ohanapecosh Formation was highly distorted on at least two different occasions in middle and late Tertiary time.

Heat from these intrusive bodies formed contact aureoles extending as much as 300 feet from the intrusives, with the growth of biotite, hypersthene, amphibole, and cordierite. In addition, this heating probably caused the growth of zeolites and associated minerals over much wider areas. Circulation of heated connate water could have dispersed large quantities of heat far beyond the limits of the hornfels aureoles, and it might have had considerable effect on the zeolite facies minerals which had previously been formed as a result of deep burial.

The complex distribution of zeolite-facies minerals in the Ohanapecosh Formation thus probably records two periods of lithostatic loading and at least two periods of dispersal of heat from igneous intrusions. It is impossible at present (1960) to estimate the relative importance of each of these, but we believe that the two most important were the load metamorphism accompanying and immediately following the deposition of
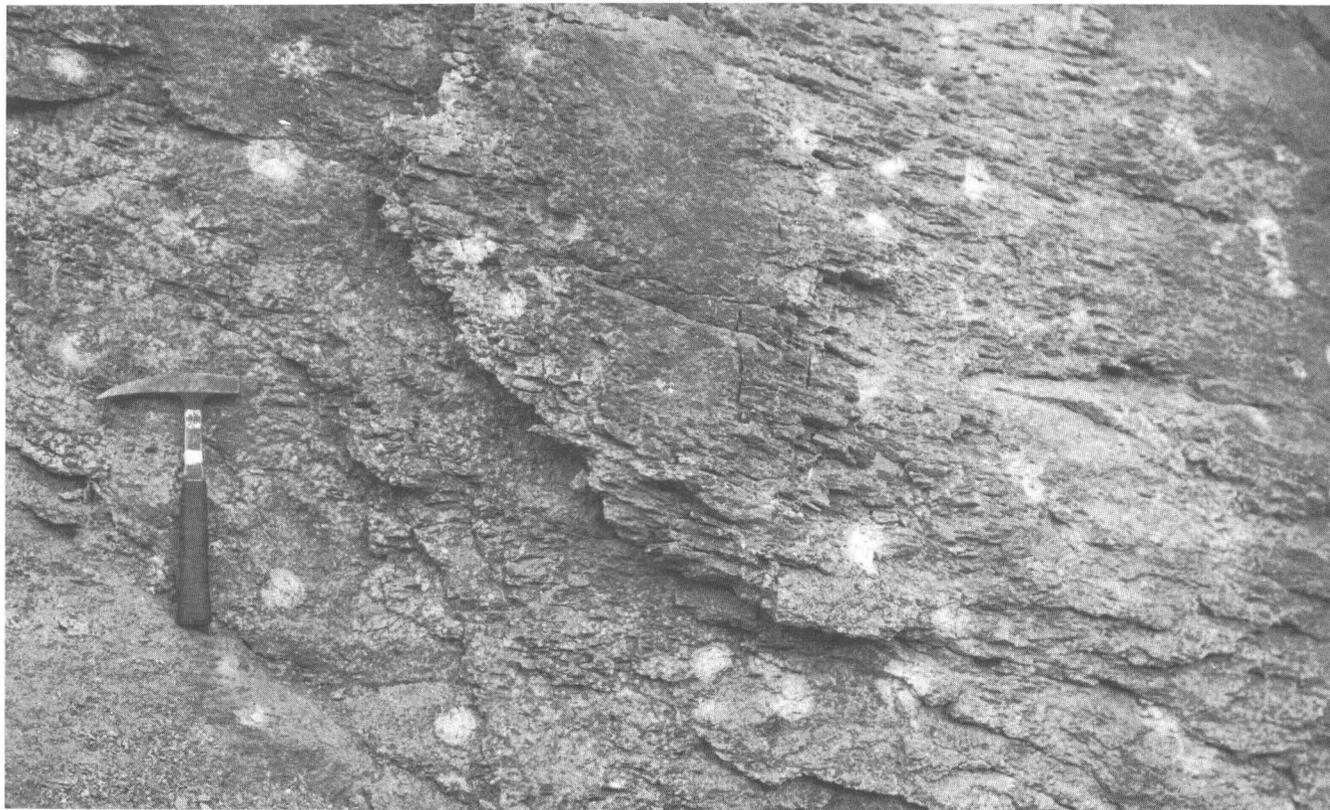
FIGURE 28.—Megascopic features of Ohanapecosh metamorphism. Spherical concentrations of laumontite in pumiceous tuff-breccia on east side of Backbone Ridge. Note the friability of the rock.

the Ohanapecosh Formation and the dispersal of heat (perhaps through water-saturated rocks) that accompanied the later intrusion of the Tatoosh pluton.

Zeolite metamorphism is much less extensive in the Stevens Ridge and Fifes Peak Formations than in the Ohanapecosh Formation. The glass in the Stevens Ridge ash flows has been completely devitrified to quartz and feldspars; zeolites are rare in these rocks, but chlorite, celadonite, and epidote are locally abundant. In Fifes Peak lavas, too, all glass has devitrified, but the rocks are considerably less altered than most of the Ohanapecosh lavas. Fifes Peak lavas generally contain fresh plagioclase, though their ferromagnesian minerals show varying degrees of alteration to carbonate, saponite, and other montmorillonoids. In some places, however, the Stevens Ridge and Fifes Peak volcanic clastic rocks are as thoroughly zeolitized as Ohanapecosh rocks of similar lithology, and they contain abundant chlorite, epidote, laumontite, wairakite, and albite.

The most zeolitized Stevens Ridge and Fifes Peak rocks are found near large intrusive bodies (for example, at Unicorn Peak, Eagle Peak, Tolmie Peak, and Sluiskin Mountain, all of which lie near the Tatoosh pluton or its complexes of satellitic sills). In areas several miles from large exposed intrusives secondary minerals other than the normal devitrification products are generally restricted to celadonite, saponite, nontronite, and other montmorillonoids.

These space relations indicate that the zeolitic alterations of the Stevens Ridge and Fifes Peak Formations is related to the intrusive bodies. Extensive circulation of heated underground water was probably the most important means of heat dispersal and was the main cause for the local zeolitization. This process was probably important in the metamorphism of the Ohanapecosh Formation as well, but its effects in that formation cannot be clearly separated from those of regional load metamorphism. The role of igneous intrusion is more apparent in the Stevens Ridge and
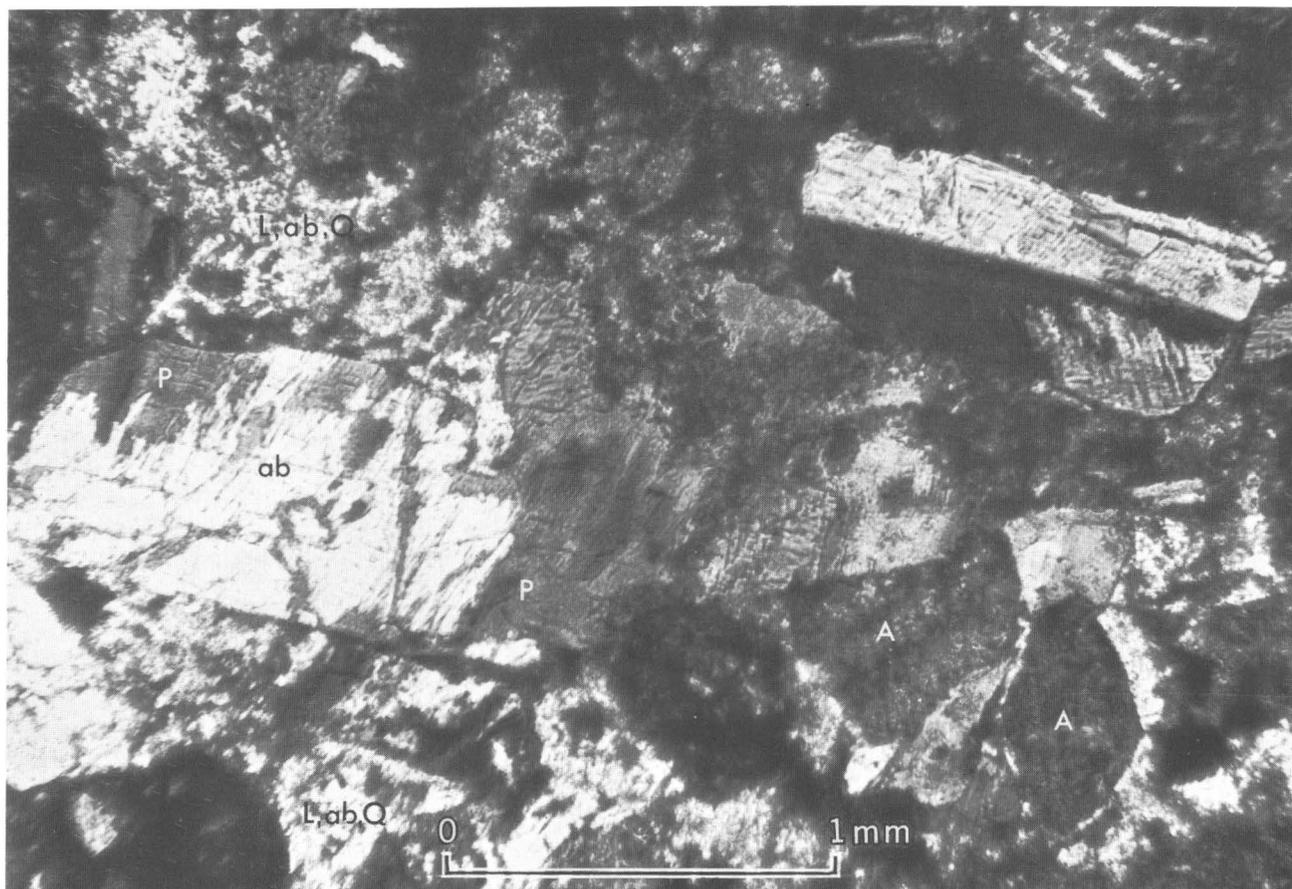
FIGURE 29.—Microscopic features of Ohanapecosh metamorphism. Photomicrograph of laumontite-bearing tuff-breccia from 5 miles southeast of Ohanapecosh Hot Springs. Original crystals of plagioclase are largely replaced by albite (ab), but remnants (P) survive. Finely granular aggregates of laumontite, albite, and quartz (L, ab, Q) replace much of the matrix. Clastic augite (A) remains relatively unaltered.

Fifes Peak Formations, in which the effects of load metamorphism appear to have been minor.

In summary, many rocks of the Ohanapecosh Formation are in the zeolite facies of metamorphism. This alteration was promoted largely by increase of temperature and pressure accompanying deep burial in the Ohanapecosh depositional basin in Eocene and perhaps early Oligocene time. Since then, the distribution of Ohanapecosh mineral assemblages has been greatly complicated by thermal adjustments caused by igneous intrusive activity. Diagenetic mineralogic alteration is characteristic of the Stevens Ridge and Fifes Peak Formations throughout much of the park, but in these formations a zeolite facies metamorphism was also developed locally near large instrusive bodies.

STRUCTURES OF THE BEDDED ROCKS

The early and middle Tertiary rocks have been deformed by regional folding on axes that trend north-northwest. Folds were developed in the Ohanapecosh Formation before the deposition of the Stevens Ridge and Fifes Peak Formations. Renewed folding along the same axes then deformed the Stevens Ridge and Fifes Peak rocks and accentuated the folds in older strata. Normal faults of northwesterly trend block out the east end of the Tatoosh Range and cross Indian Henrys Hunting Ground. Still later, folding and warping accompanied the intrusion of the Tatoosh pluton. Regional uplift of the Cascade Range, probably accompanied by some accentuation of the former folds, also affected the rocks in Pliocene and Quaternary time.

FOLDS

The most important structural element in Mount Rainier National Park is the Unicorn Peak syncline, named for a rugged spire on the crest of the Tatoosh Range (fig. 24). From Unicorn Peak, near the south boundary of the park, the axis of this broad fold extends north-northwest, directly beneath Mount Rainier, and then leaves the park about 3 miles east of its northwest corner. Reconnaissance shows that the fold extends at least 5 miles farther to the northwest, and an
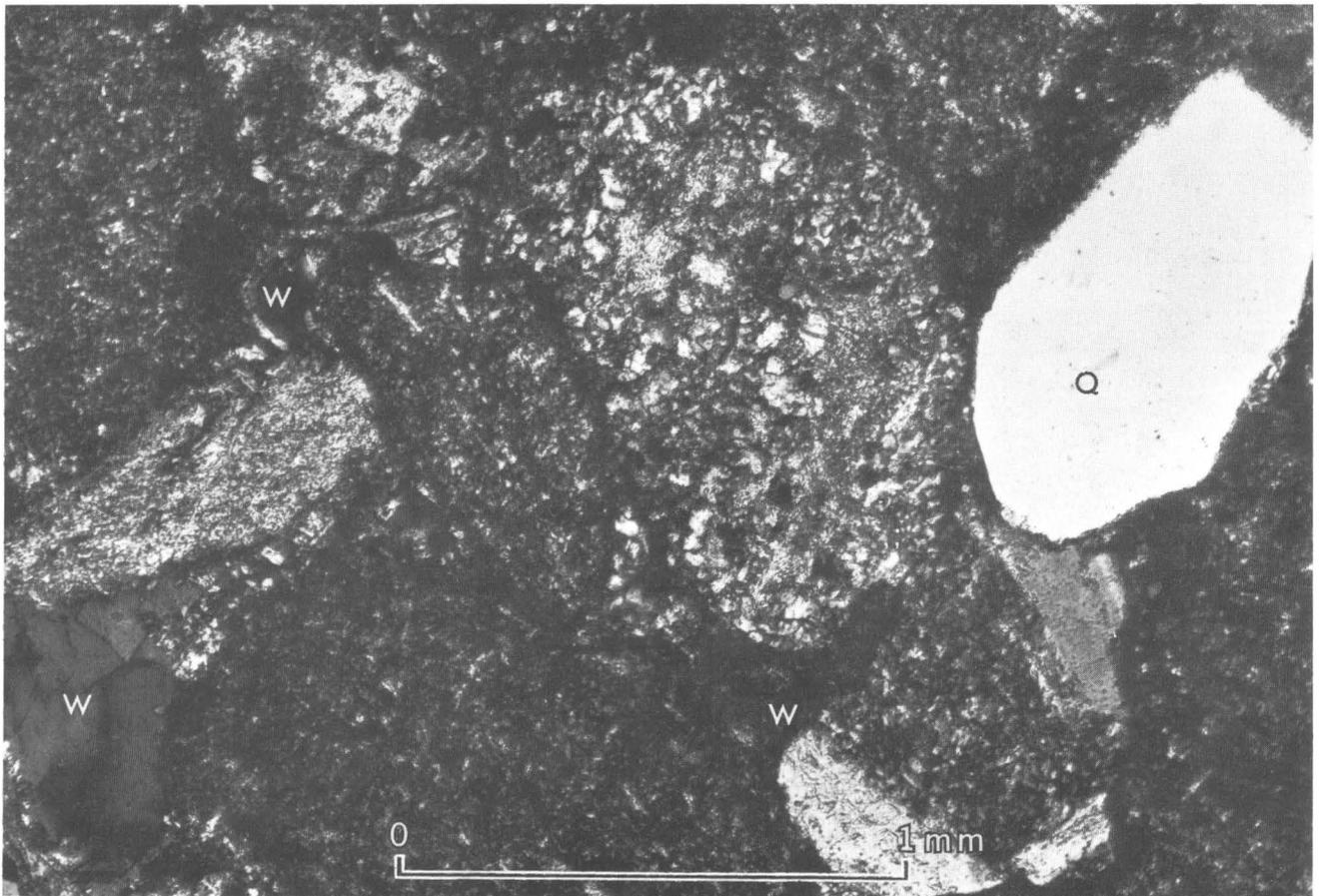
FIGURE 30.—Microscopic features of Ohanapecosh metamorphism. Photomicrograph of wairakite-rich volcanic arenite from east side of Backbone Ridge. Glassy lava fragments and very minor amounts of quartz (Q) form an intact framework. All volcanic glass has devitrified; plagioclase microlites in lava fragments are mostly albitized; wairakite (W) fills nearly all voids.

even greater distance beyond the boundaries of the park on the southeast—a known extent of at least 30 miles. The Unicorn Peak syncline is flanked by two north-northwest trending anticlines (fig. 31): the Chinook Pass anticline, whose axis passes through the northeast corner of the park; and the Skate Creek anticline (described by Fisher, 1957), whose axis misses the southwest corner of the park by about 2 miles. Each anticline is about 15 miles wide; the Unicorn Peak syncline is more than 20 miles across. These three folds probably belong to the system of large tranverse folds that extend across the central part of the Cascade Range and far beyond into the Columbia River plateau (Waters, 1955a).

In general form the Unicorn Peak syncline is a broad simple structure, but in detail it is somewhat complicated because folding has occurred in more than one episode, because minor folds appear on its limbs, and because the strata have been irregularly deformed by intrusive bodies. Where it crosses the central part of the park, moreover, the trough of the syncline has been partly obliterated by the crosscutting Tatoosh pluton, and it is concealed over wide areas by the lavas from Mount Rainier. Nevertheless, this broad syncline is clearly outlined by the restriction of the Fifes Peak and Stevens Ridge Formations entirely to the shallow trough of the structure and by the restriction of the older Ohanapecosh Formation to the anticlinal areas on either side (fig. 31).

The syncline is nearly symmetrical where is crosses the southern part of the park; here the maximum dips in the Ohanapecosh rocks on either limb rarely exceed 35°—the few steeper dips and aberrant strikes probably record local deformation caused by underlying intrusive igneous masses. In the central part of the park the southwest limb of the fold—largely obscured by the lavas from Mount Rainier—is complicated by a small syncline that extends from Indian Henrys Hunting Ground to the North Puyallup River. Still farther northwest the west limb of the Unicorn Peak syncline steepens and the fold in the Ohanapecosh rocks becomes strongly asymmetrical—west and south of Mowich Lake the Ohanapecosh rocks dip 65° to 88° eastward; corresponding westward dips on the northeast limb of the fold are mostly less than 30°.

This asymmetry of the Unicorn Peak syncline in the northeast corner of the park is not reflected in the overlying Stevens Ridge and Fifes Peak Formations; in these rocks the dips seldom exceed 20°, and are generally less than 10° over broad areas near the trough of the syncline. In nearly all parts of the park rocks of the Ohanapecosh Formation dip more steeply than rocks of the overlying Stevens Ridge and Fifes Peak Formations: in the Backbone Ridge area the Ohanapecosh rocks have an average westward dip of 20°, whereas the overlying Stevens Ridge Formation dips only 15°; on Goat Island Mountain the corresponding average dips are 21° and 8°; at McNeeley Peak, 20° and 5°. The greatest discordance, however, is in the northwestern part of the park where Ohanapecosh strata near the junction of Spray Creek and the Mowich River dip 78° to 88° east and are overlain by Stevens Ridge and Fifes Peak rocks which dip less than 10° to the northwest.

These relations show that flexing of the Unicorn Peak syncline began before the Stevens Ridge and Fifes Peak Formations were deposited; later folding along the same axis has deepend the syncline in the Ohanapecosh rocks and has formed a broad shallow trough in the Stevens Ridge and Fifes Peak Formations.

A minor syncline that extends from Indian Henrys Hunting Ground to the North Puyallup River has already been mentioned. Several other minor folds are found in the park. A northwest-plunging anticline underlies Huckleberry Park, but its axis can be traced for only about 3 miles. At the head of the North Puyallup River a poorly defined anticline and syncline trend north-south, but they are exposed for less than a mile before disappearing beneath the lavas and glaciers of Mount Rainier.

The Chinook Pass anticline, where it crosses the northeast corner of the park, is not well defined because the Ohanapecosh strata have been tilted out of position by sills. The strata are also irregularly deformed, probably by large underground cupolas on the top of the Tatoosh pluton. Abrupt steepening of the dip, and marked contortions in the regional strike of the strata, can be seen at several places near the crest of the anticline where its axis lies west of White River and north of Yakima Park. Farther south, the Ohanapecosh rocks about a mile southeast of Seymore Peak are contorted into a small northward-trending syncline which lies near the axis of the major Chinook Pass anticline; this folding was probably caused by the emplacement of nearby small intrusive bodies. Poorly defined flexures at Cayuse Pass and at Buell, Seymour, and Shriner Peaks also probably were caused, at least in part, by forceful intrusion of large unexposed masses of the Tatoosh pluton. .Similar features occur along the edges of the Tatoosh pluton where it cuts the Unicorn Peak syncline. An example of a flexure definitely related to upward rise of the core of the pluton is easily seen on the

EXPLANATION

Rocks of Mount Rainier volcano

Tatoosh pluton
Only the large coarse-grained masses are shown

Stevens Ridge and Fifes Peak Formations

Chiefly Ohanapecosh Formation, but also includes arkosic
sedimentary rocks of the Puget Group west and southeast
of the Park

Contact

Fault
Dashed where inferred.   U, upthrown side; D, downthrown side

Anticline, showing trace of axial plane

Syncline, showing trace of axial plane
Dashed where concealed by Mount Rainier lava

Strike and dip of beds

Horizontal beds

Boundary of Mount Rainier National Park

0   3   6   9 MILES

FIGURE 31.—Simplified geologic map of Mount Rainier National Park and bordering areas showing the major folds and faults. The north-northwest-trending belt of Stevens Ridge
and Fifes Peak rocks, which passes beneath Mount Rainier, marks the trough of the Unicorn Peak syncline.

south side of Stevens Ridge, where volcanic clastic rocks of the Stevens Ridge Formation are bent 20° upward from their near-horizontal position by the crosscutting pluton.

FAULTS

The Stevens Peak fault trends southeast across the east slope of Stevens Peak and continues southward along the east base of the Tatoosh Range. Reconnaissance shows that it extends at least 8 miles south of the park boundary in the Cowlitz River valley. The fault is nearly vertical, for its trace crosses irregular topography with little change in course. At Stevens Peak it has displaced the contact between the Stevens Ridge and Ohanapecosh Formations about 3,000 feet vertically, with downthrow on the northeast. This fault is clearly older than the Tatoosh pluton: the Tatoosh contact is not offset by the fault, no shear zone crosses the granodiorite on strike with the fault, and granophyric dikes related to early phases of the Tatoosh pluton invade the fault zone in the valley of Maple Creek.

The Devils Dream fault crosses Indian Henrys Hunting Ground and roughly parallels Devils Dream Creek in the southwestern part of the park. It displaces the base of the Stevens Ridge Formation at least 1,200 feet vertically; the downthrow is on the northeast, as it is on the Stevens Ridge fault. Other faults, unmapped, probably occur in the park, but the heavy forest cover and the scarcity of marker beds in the Ohanapecosh Formation make it difficult to detect and trace them.

AGE RELATIONS OF STRUCTURAL FEATURES

Ages of folding and faulting in Mount Rainier National Park can be bracketed within fairly narrow limits. Folding of the Unicorn Peak syncline began after the deposition of the Ohanapecosh Formation of late Eocene age. The Stevens Ridge and Fifes Peak Formations were deposited upon folded and deeply eroded Ohanapecosh rocks, probably in the time interval from middle Oligocene through early Miocene. All the rocks were then folded and faulted along the preexisting structural lines before the intrusion of the Tatoosh pluton, probably in late Miocene time. This pluton was injected so forcefully as to cause additional local folding.

Regional upwarping accompanying the Pliocene and Quaternary rise of the Cascade Range probably accentuated the earlier structures, but it is difficult to judge the amount and character of post-Miocene deformation because of the scarcity of bedded rocks younger than the Fifes Peak Formation. The early intracanyon lava flows from Mount Rainier are not noticeably disturbed by either folding or large-scale faulting.

TATOOSH PLUTON AND RELATED HYPABYSSAL AND VOLCANIC ROCKS

According to S. F. Emmons (1879, p. 54), the presence of granitic rocks on the slopes of Mount Rainier was first noted by Lieutenant Kautz in 1857, but Kautz' report was mistakenly interpreted to mean that the granite occurred as a dike which had invaded the lavas of the volcano. The geologic relations were correctly interpreted by Emmons during his exploration and ascent of the mountain in 1870. He reported (in King, 1871, p. 163) that the Nisqually Glacier had "cut through the more yielding strata of volcanic rocks, and come upon an underlying and unconformable mass of syenite."

I. C. Russell (1897, p. 361–379) and G. O. Smith (1897, p. 422–423) during their reconnaissance of Mount Rainier in July 1896 discovered a granitic mass along the Carbon River near the present park boundary and found a related but finer grained intrusive body at nearby Chenuis Falls. On reaching the front of the Carbon Glacier they were surprised by the abundance of granitic boulders in ice descending from a volcanic mountain. Farther up the glacier they noted outcrops of granite in Goat Island Rock, a nunatak protruding through the glacier. Still farther up they reached the granitic mass of Moraine Park and Mineral Mountain and observed lavas from Mount Rainier resting upon it (pl. 1). On the south side of the mountain they confirmed Emmons' observation at the Nisqually Glacier and concluded: "Granite is exposed on the slopes of Rainier where erosion has cut through the overlying lava, and it is plain that the volcanic cone rests upon an elevated platform of older rock." (G. O. Smith, 1897, p. 423).

Smith and Mendenhall (1900) were the first to establish that the granitic rocks in the core of the Cascade Range are of two ages, and that the younger rock, widely exposed in a batholith at Snoqualmie Pass 45 miles north-northeast of Mount Rainier, is of late Tertiary age. They also suggested (p. 229) that the granitic rock underlying Mount Rainier might correlate with the "granite" at Snoqualmie Pass. A few years later Smith and Calkins (1906) mapped the Snoqualmie quadrangle (which includes Snoqualmie Pass), named the batholith the Snoqualmie Granodiorite, firmly established the intrusive relation of the granodiorite to the Keechelus Andesitic Series and other Teritiary formations, and assigned it to the late Miocene. In the Mount Rainier area, however, little additional geologic work was done until H. A. Coombs (1936) identified the granitic rock as granodiorite, delimited some of its outcrops, established its intrusive relations to the Keechelus Andesitic Series, and correlated it with the Snoqualmie Granodiorite.

VARIABILITY AND COMPLEXITY OF THE INTRUSIVE BODIES

The intrusive rocks of the Mount Rainier area are considerably more complex than the early reconnaissance workers realized. They range from hypersthene-augite diorites through hypersthene-hornblende quartz diorites, and hornblende-biotite granodiorites and quartz monzonites to granophyres, porphyries, rhyodacites, and vitrophyres. They occur in amazingly intricate sill complexes, dike swarms, and irregular composite stocks, as well as in a mass of batholithic dimensions (pl. 1). Many sills and dikes were emplaced beneath such a thin cover that the magma vesiculated. Furthermore, the pluton broke through to the surface at The Palisades and elsewhere with explosive violence, shattering the partly congealed granitic rocks intruded earlier, erupting pyroclastic material on the surface, and solidifying within the vents as inclusion-filled welded tuff, vitrophyre, and rhyodacite. These rocks were then reinvaded by granodiorite porphyry and aplite from below. Near the vent areas, parts of the earlier crystallized margin of the pluton were explosively brecciated, riddled with vertical zones of miaroltic cavities and vesicles, and altered by streaming gases. Thus, the intrusive rocks record a complex history of magmatic rise, in part under truly plutonic conditions, but also under subvolcanic conditions of rapid chilling and dehydration because of close approach to the surface. The resultant variations in cooling history and endomorphic alteration have produced an extremely varied and complex suite of rocks.

Despite this great variability, the plutonic rocks of Mount Rainier National Park are not unique. They are strikingly similar to those of the nearby Snoqualmie batholith (Smith and Calkins, 1906; Fuller, 1925), whose position with respect to the Tatoosh mass is shown on figure 32. They also show similarities to other middle Tertiary intrusive masses in the Cascade Mountains; among these are the one north of Spirit Lake, near Mount St. Helens (Verhoogen, 1937), the Star Peak stock (Felts, 1939), the Cloudy Pass intrusion (Cater, 1960), and the Laurel Hill intrusion at the southwest base of Mount Hood. All these possibly connect underground into a continuous plutonic body that underlies the entire northern Cascade Range at depth, but since there is no direct connection between them at the surface, and since the presence of a continuous plutonic mass beneath the Cascades is hard to prove, it seems best to designate the pluton of Mount Rainier National Park by a separate name instead of consider-ing it part of the Snoqualmie mass as H. A. Coombs (1936, p. 167) did. It is accordingly named the Tatoosh pluton from the Tatoosh Range, whose precipitous cirques and cliffs give magnificent exposures of the intrusive rocks and their contact effects.

ROCKS OF THE MAIN TATOOSH PLUTON AND ASSOCIATED ROCKS

The medium- to coarse-grained granitic rocks forming the core of the Tatoosh pluton and the larger stocks are shown separately on plate 1 from the closely related sill and dike complexes that cluster along the borders of these larger masses. In some places the rocks of the core and those of its hypabyssal sheath of sills and dikes intergrade, so that the boundaries between them are arbitrary, but in many places the contacts are sharp and easily located. In a few places the two kinds of rocks are separated by thin masses of plutonic breccia.

DISTRIBUTION

The largest body of coarse-grained rocks in Mount Rainier National Park—the core of the Tatoosh pluton—is a roughly oval mass extending underneath Mount Rainier from the central part of the Tatoosh Range to Vernal Park, but it also has a domelike offshoot to the northeast along White River. About 50 square miles of coarse-grained rock are exposed at the surface; but if the lavas and glaciers of Mount Rainier were removed, the area of exposure would be more than doubled (pl. 1 and fig. 32).

The contact of this mass curves across the Tatoosh Range in a large rounded lobe that cuts discordantly through the gently folded and faulted rocks of the Ohanapecosh and Stevens Ridge Formations. Contacts with the Ohanapecosh Formation are steep, clearly defined, and marked by a border of hornfels; the pluton, however, throws out extensive sill and dike complexes along and near the unconformity at the base of the Stevens Ridge Formation, and in such areas contact relations are extremely complex.

North of the Tatoosh Range the granitic rocks disappear beneath the lavas of Mount Rainier, but they reappear in volume along the north base of the mountain, and also in scattered outcrops along its east and west flanks. Sill and dike complexes border the pluton throughout most of its periphery, but they are especially prominent on Skyscraper Mountain, Sluiskin Mountain, and in the Lake James area, north of Mount Rainier.

EXPLANATION

Mount Rainier lava

Plugs and small stocks of pyroxene quartz diorite, dacite, rhyodacite, vitrophyre, and welded tuff
Sills and dike swarms omitted

Plutons and large stocks of granodiorite, quartz diorite, and quartz monzonite

- - - - - - -
Crestline of Cascade Range
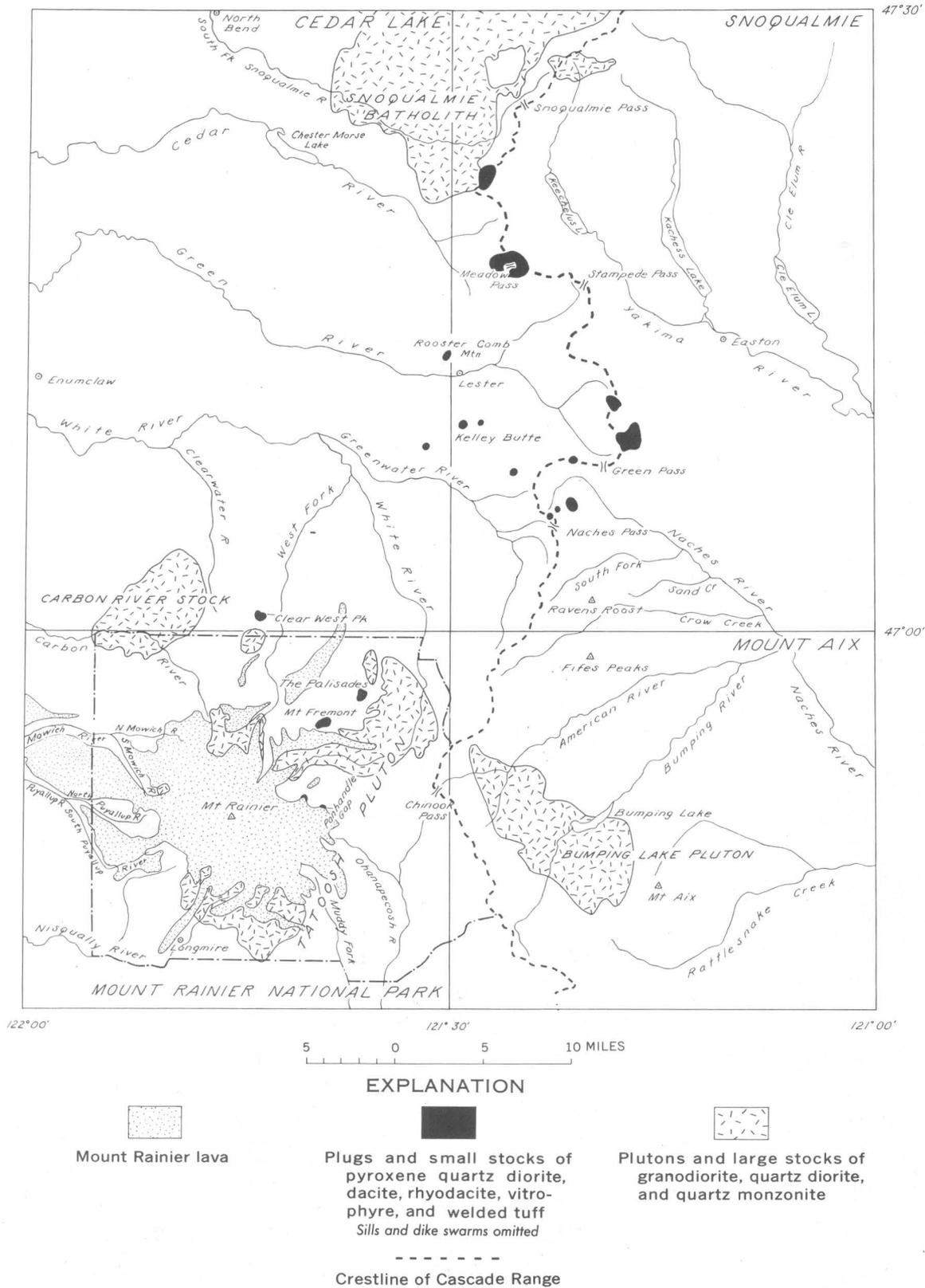
FIGURE 32.—Reconnaissance map of a part of the Cascade crest, central Washington, showing known areas of Miocene or early Pliocene plutons, stocks, and plugs. Outlines of the Snoqualmie batholith and larger stocks in the Snoqualmie and Cedar Lake quadrangles from Smith and Calkins (1906) and Fuller (1925); outlines of the Bumping Lake pluton in the Mount Aix quadrangle modified from Abbott (1953).

Erosion of a few thousand feet of roof rock would doubtless reveal large underground extensions of the pluton beneath these voluminous sill complexes. An outlying stock of coarse granitic rock north of Adelaide Lake does cut the sills. A much larger stock with its own retinue of sills occupies the valley of Carbon River in the northwest corner of the park (fig. 32). It is nearly joined to the heavily silled area on Sluiskin Mountain by a group of sills and dikes that cluster on Chenuis Mountain and near the front of the Carbon Glacier (pl. 1). A stock of fine-grained rock, similar to many of the sill rocks, lies on the park boundary west of Mowich Lake.

The Bumping Lake pluton (Abbott, 1953), comparable in size to the Tatoosh pluton, lies in the Mount Aix quadrangle about 5 miles east of the park (fig. 32). The complex of sills and dikes that crops out along the Cascade divide in the eastern part of the park suggests that the two plutons may be connected underground.

PETROGRAPHIC AND FIELD VARIATIONS

From a casual examination of some outcrops of the coarse granitic rocks, such as those exposed in roadcuts at the entrance to White River campground and in the bluffs around the foot of the Nisqually Glacier, one might suppose that the Tatoosh pluton consisted of even-grained uniform fresh granodiorite. This is far from true; the rocks forming the core of the pluton vary widely in mineral composition and texture, and have undergone a surprising amount of alteration for so young an igneous mass.

Hornblende-biotite granodiorite (fig. 33) is the most abundant rock in the central part of the pluton and in the larger stocks. Quartz monzonite, generally of somewhat finer grain (fig. 34), is also abundant, and it predominates near the roof. Other rocks, less abundant and more irregularly distributed, include several varieties of pyroxene-bearing quartz diorite. Granophyre occurs in minor quantity, and a granophyric mesotasis forms the groundmass of some of the porphyritic quartz monzonites and porphyritic quartz diorites.

All these rocks are characterized by marked variations in texture and composition. The granodiorite and quartz monzonite show a general but irregular decrease in grain size toward the roof of the pluton, but even the superficially massive-appearing granodiorite from the central and deeper parts of the mass shows marked changes in texture and mineral composition over short distances. The chief minerals of the gran-
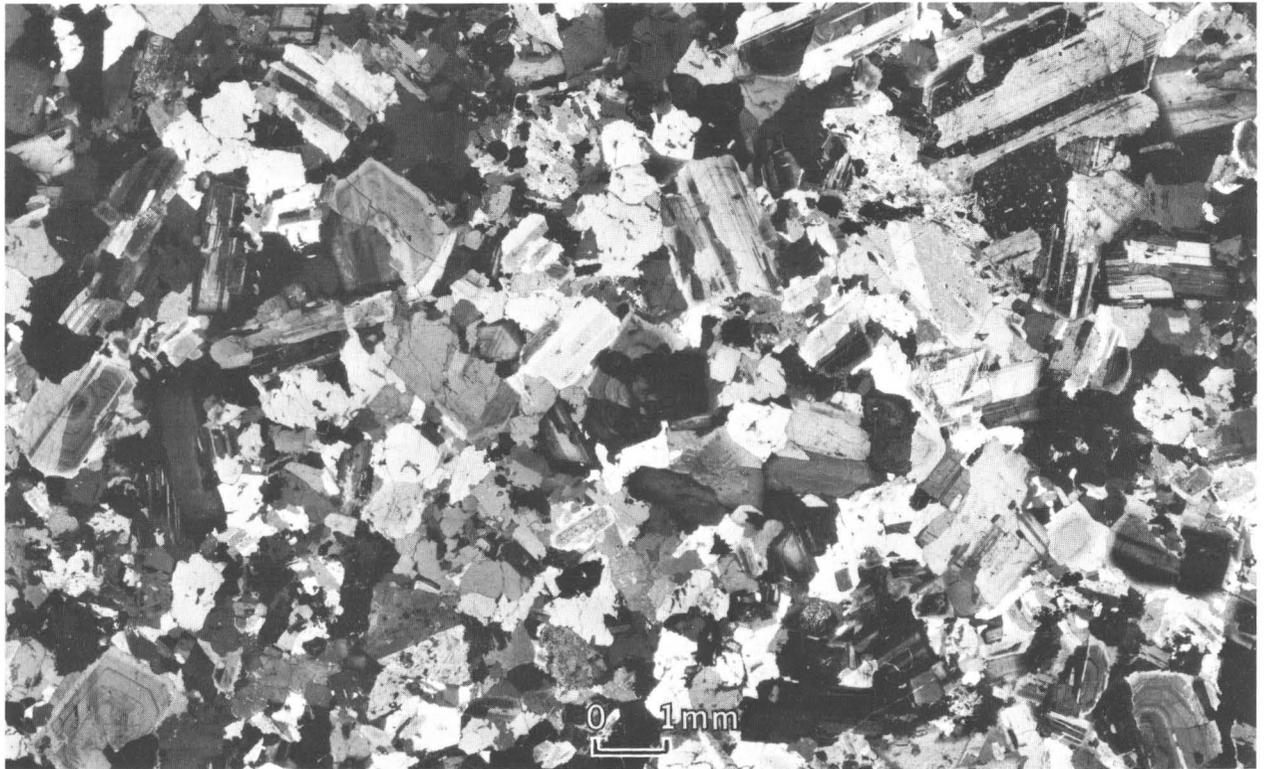
FIGURE 33.—Photomicrograph of hornblende-biotite granodiorite from a nunatak near the foot of the Nisqually Glacier. The texture is typical of many of the coarse-grained rocks from the core of the Tatoosh pluton. Note the strikingly euhedral to subhedral outlines and the strong zoning of plagioclase; also the ragged outlines of the interstitial quartz and potassium feldspar. Photograph by Thomas L. Wright.

odiorite are plagioclase, quartz, orthoclase, hornblende, and biotite, but the proportions of these minerals vary widely and erratically from place to place. The variations are on every scale—from those seen in almost every large outcrop to others noticeable within the space of a single thin section. Marked variations also occur in the composition and degree of alteration of the individual minerals.

Plagioclase, characterized by striking oscillatory and progressive zoning, occurs in all the rocks of the main pluton. The average composition of the plagioclase from the granodiorite and quartz monzonite is andesine; but most of the crystals are strongly though irregularly zoned, and the range of composition in individual crystals may be as great as from An_{65} to An_{10}. The kind and amount of zoning varies considerably, even among individual crystals from the same rock. In many thin sections, however, unzoned cores of calcic plagioclase, dotted with rounded inclusions of fresh pyroxene, are surrounded by a broad sheath of oscillatory zoned plagioclase, which, in turn, is bordered by a narrow outer rim showing strong progressive zoning. The unzoned cores probably record an early period of slow intratelluric crystallization before the rise and rapid congealing of the Tatoosh magma at higher

levels in the crust; the outer zones record a much more complex history of late crystallization.

Additional evidence of this complex history is attested by the structural character of the plagioclase. Thomas L. Wright (written communication, 1960) has made a special study of the structural state of the feldspars; he finds that the unzoned intratelluric cores consist of fully ordered low-temperature plagioclase, whereas the zoned outer parts are partially ordered, with "transitional" optics. Evidently the plagioclase cores became fully ordered during slow intratelluric crystallization, but the outer zones which crystallized rapidly and probably metastably, reached only a partly ordered configuration. In granodiorites and quartz monzonites from the southern part of the pluton Wright finds that the composition of the unzoned cores is generally in the range from An_{55} to An_{50}, the sheath of oscillatory zones is near An_{40}—the range of oscillations is seldom more than about 5 percent anorthite—and the outer rims show an abrupt decrease in anorthite content to about An_{20-15} at the edge of the crystals. We find similar ranges in composition of the feldspars from the northern part of the pluton, except that the unzoned cores of pyroxene-bearing plagioclase from some porphyritic quartz monzonites are more calcic, generally

FIGURE 34.—Photomicrograph of fine-grained porphyritic quartz monzonite from the Tatoosh pluton, one-half mile northeast of Paradise Inn. Phenocrysts and clots of coarse plagioclase are enclosed in a fine matrix composed mostly of quartz and alkali feldspar.

in the range from An_{60} to An_{55}. Plagioclase in quartz diorite from the Carbon River stock also shows these strongly calcic cores.

The potassium feldspar is orthoclase; in most rocks it shows no observable inversion toward microcline, the stable potassium feldspar in most granitic rocks. Orthoclase microperthite or cryptoperthite is common, although in the finer grained rocks the orthoclase is nearly homogeneous. The amount of exsolution differs widely among crystals within a section, and even within a single crystal. The orthoclase is commonly anhedral and interstitial. Quartz is also interstitial, and in some rocks the two minerals occur in irregular granophyric intergrowths (fig. 35). The content of potassium feldspar and quartz in both the granodiorite and the quartz monzonite is variable; either mineral may change several percent in the space of a few feet. The ferromagnesian minerals also vary in kind, amount, and alteration over short distances. Hornblende and biotite are most common in the granodiorite and quartz monzonite, but a little uralitized pyroxene is generally present and is abundant in some rocks. Pyroxenes include both hypersthene and augite, but in most rocks hypersthene is extensively uralitized. Large masses of coarse-grained pyroxene-rich granodiorite and quartz monzonite are exposed near the southern border of the pluton on Rampart Ridge and Eagle Peak. Fine-grained pyroxene-quartz diorite is common as flat sheets high in the Carbon River stock.

In the roof zones quartz monzonite and various kinds of more mafic rocks ranging from pyroxene-quartz diorite to granodiorite are closely associated. In some places they occur together, intercalated in flat sheets with vague to relatively sharp contacts. At other localities pyroxene-quartz diorite is invaded and infiltrated by underlying quartz monzonite. In still other places sheetlike bodies of plutonic breccia, composed of slightly rounded fragments of fine-grained quartz diorite embedded in a coarser grained and more felsic matrix, overlie or are enclosed within larger masses of quartz monzonite. Less commonly the quartz monzonite forms the matrix of the plutonic breccia. These roughly tabular masses of plutonic breccia generally crop out near, but not always directly at, the roof of the pluton. They may have crystallized early against the roof, or even as sills just above the roof, and then have been pried off and partly stoped away by continued rise of the magma (fig. 38). Some, however, are strung

FIGURE 35.—Photomicrograph of granophyric quartz monzonite from the Tatoosh pluton, southeast slope of Plummer Peak. Note the patchy but interstitial distribution of the granophyre. Such rocks are common in the parts of the Tatoosh pluton that are inferred to lie in the root zone beneath volcanic vents. Photograph by THOMAS L. WRIGHT.

out along nearly vertical contacts—an example is the plutonic breccia at the southeast wall of the pluton where crossed by the Stevens Canyon road.

In many places the roof is underlain by a thick porphyritic "chill zone" composed of quartz diorite containing a small amount of granophyric matrix, or it is underlain by porphyritic quartz monzonite containing phenocrysts of both plagioclase and orthoclase cryptoperthite set in a more abundant granophyric groundmass (fig. 35). These fine-grained potassium-rich rocks generally contain both hypersthene and augite in addition to the more abundant hornblende and biotite. A significant textural and mineralogic variation found in parts of the chill zone is the appearance of vugs and vertical swarms of large vesicles filled with quartz, amphibole, epidote, chlorite, tourmaline, and other hydrous minerals (fig. 36).

Diorites and quartz diorites, confined chiefly to areas near the roof of the pluton and to its satellitic stocks and sills, are highly variable. Some pyroxene-quartz diorites are similar to those that form stocks in the Snoqualmie quadrangle (Smith and Calkins, 1906; Fuller, 1925); they contain altered hypersthene, less altered plagioclase and augite, and a little hornblende and quartz. Equally abundant are fine- to medium-grained hypersthene-hornblende-quartz diorites with little augite. The small amount of quartz and potassium feldspar—present in many quartz diorites as granophyre—fills spaces between the plagioclase laths and embays their edges; as quartz and orthoclase increase in amount, these rocks grade through granodiorite into quartz monzonite.

The variations in both composition and texture within the pluton are gradational: the mass does not consist of sharply bounded intrusions of different ages, but it is
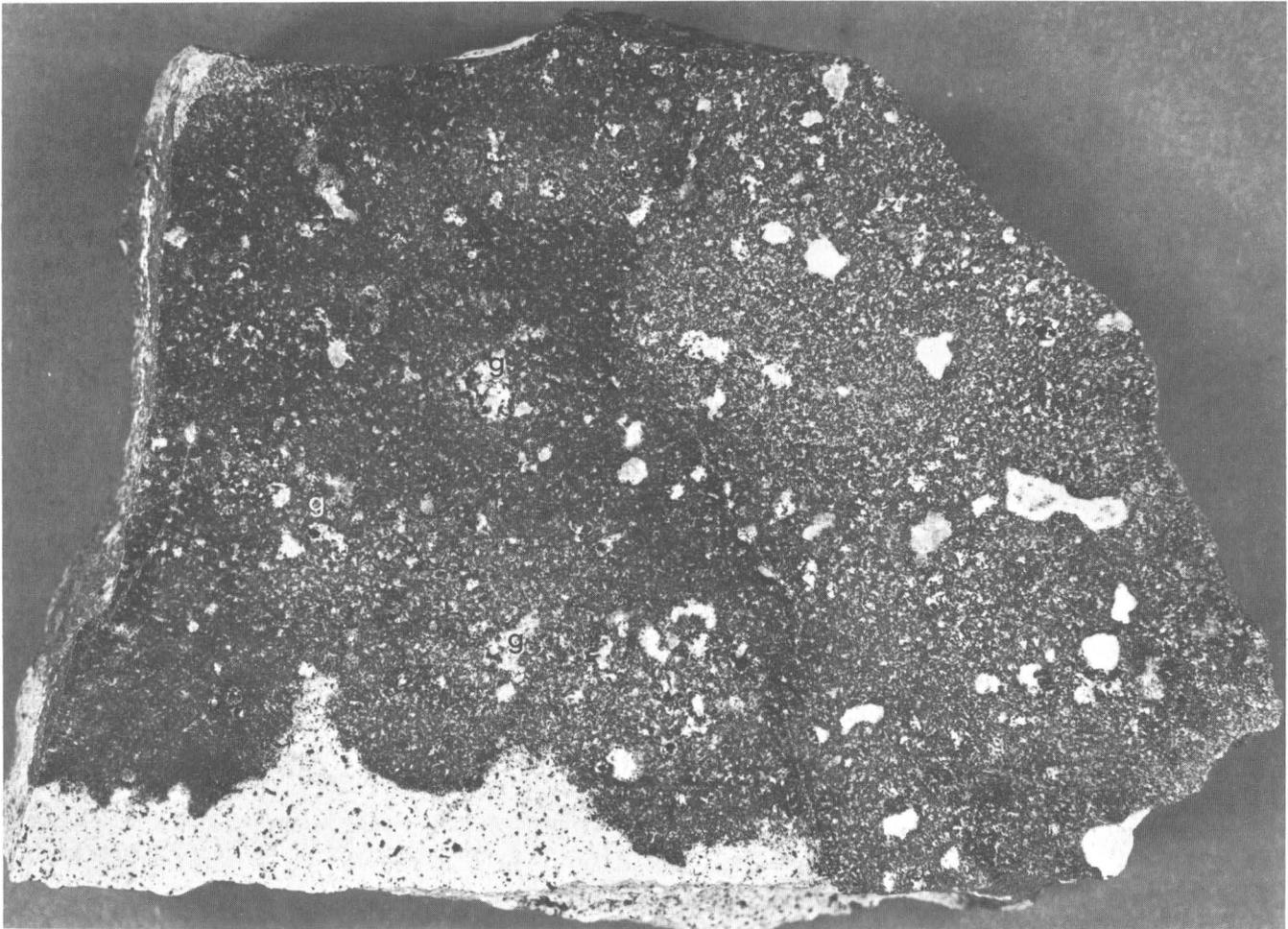
FIGURE 36.—Vesiculated fine-grained quartz diorite at the roof of the Tatoosh pluton, White River Park, one-fourth mile south of Hidden Lake. The round to amoebaform vesicles are filled chiefly by quartz. Small inclusions of grandodiorite (labeled g), were hydrothermally altered before being caught up in the quartz diorite. Before the quartz diorite had completely congealed it was invaded (lower edge of photograph) and in places transformed into a contact breccia (not shown in photograph) by a later pulse of porphyritic quartz monzonite.

characterized by subtle and irregular transitions over short distances. Crosscutting relations are found, as where quartz monzonite invades quartz diorite in the plutonic breccias and border zones (fig. 36), but intergradation is much more common. In the field one can often see unequivocal crosscutting relations between the fine-grained sills and dikes of the bordering hypabyssal complexes, but even there the evidence indicates no great lapse of time between intrusions.

Several characteristics of the rocks from the main pluton suggest that the magma crystallized prematurely under subvolcanic conditions as the result of dehydration under a cover so thin that volatiles streamed upward through the roof. In places, moreover, the magma broke through to the surface with explosive violence. Among features noted in the field that support this interpretation are the occurrence in the pluton near its roof of (1) pneumatolytically filled miarolitic vugs and vertical strings of large vesicles, (2) zones of explosively brecciated rocks, (3) local masses of aphanite or even vitrophyre, and (4) transitions between the roof rocks of the pluton and volcanic plugs. The remarkably widespread occurrence of a fine-grained porphyritic zone as much as 300 feet or more thick adjacent to the roof of the pluton in some areas but not in others is not easily explained as an effect of chilling by simple heat conduction into the wallrocks. But it could well have resulted from rapid transfer of heat and volatiles to the surface during contemporaneous volcanic action. The sudden and explosive vesiculation and the streaming upward of volatiles, especially near the roots of the volcanic vents, would have caused an abrupt drop in vapor pressure, hastening solidification of the rest magma and enabling the thick porphyritic zone to form.

Such upward streaming of volatiles has evidently happened at the north end of Mazama Ridge, where an early crystallized plutonic breccia was shattered into a mass of sharply angular fragments (fig. 37) which occupy a cylindrical zone—possibly the root of a former vent. The openings between the fragments are filled with amphibole, quartz, scapolite, magnetite, and apatite deposited from volatiles streaming through them. As much as 2,000 feet below this breccia the quartz monzonite is pervaded by a fine-grained granophyric mesostasis similar to that shown on figure 35. This granophyre apparently is the product of a sudden final crystallization that took place after most of the magma had solidified. Its occurrence below the breccia suggests that the violent explosion which produced the breccia liberated volatiles from this part of the congealing pluton, and the sudden transfer of heat and drop in vapor pressure then caused the (remaining) magma, of granophyric composition, to crystallize immediately. This interpretation is strengthened by field relations at other localities. Similar angular breccias, whose voids are also filled with pneumatolytic minerals, occur in the western Sourdough Mountains associated with aphanitic and even vitrophyric rocks that crystallized beneath the roof of the pluton. Records of vesiculation and rapid crystallization can also be seen near The Palisades in the northeastern part of the park (fig. 36). Evidence is given below to show that at both The Palisades and in the Sourdough Mountains the pluton broke through to the surface and caused contemporaneous volcanic activity.

The almost complete absence of marked flow banding and mineral lineation in the rocks throughout the pluton gives additional evidence that the magma crystallized rapidly; there was no prolonged interval during which the material moved as a viscous crystal mush.

Thus, the field relations indicate rapid crystallization of the Tatoosh magma under subvolcanic conditions, and the erratic variations in mineral composition and the complex zoning of individual minerals also suggest a highly variable physicochemical environment of crystallization.

ALTERATION

In contrast to the relatively fresh granodiorites and quartz monzonites of Mesozoic Age in the Cascade Range, the much younger rocks of the Tatoosh pluton are greatly, though unevenly, altered. Their feldspars are clouded; in the more altered rocks, calcic cores of plagioclase are completely saussuritized and the rims partly sericitized. The biotite is partly altered to chlorite, and the augite is partly uralitized. In some rocks these changes have gone farther: flamboyant crystals of epidote and quartz replace much of the feldspar, hypersthene and augite are completely replaced by amphibole, and hornblende is changed to mixtures of chlorite and epidote. Zeolitic replacement of feldspars, and of the fine-grained groundmass in some rocks, occurs in a few places. The margin of the pluton is locally impregnated with pyrite. Some brecciated zones in the granitic rocks are filled with such minerals as secondary amphibole, scapolite, sphene, apatite, tourmaline, chabazite, and stilbite.

HORNFELS BORDERS

Contacts of the pluton with the Ohanapecosh Formation are discordant, generally steep, relatively straight, and without marked admixture of intrusive rock and wall rock. In a few places—for example, along the west border of the small stock on Lost Creek—the contact is irregularly crenulated, as if the pluton had pushed into the more yielding layers of the Ohanapecosh clastics in rounded bulblike masses.

FIGURE 37.—Explosion breccia inferred to have been formed in the root zone beneath a Tatoosh volcanic vent.   North end of Mazama Ridge. The original rock before the explosive shattering was an early crystallized roof phase of the Tatoosh pluton.   After the explosion, gases streamed through the shattered rock and filled its voids with amphibole, magnetite, scapolite, apatite, and quartz.   These minerals also replace the edges of the fragments and the walls of cracks that penetrate them.

These steeply dipping contacts are rimmed by black or dark-gray hornfels in a zone varying from a few tens of feet to 300 feet in thickness. The hornfels resists weathering and erosion: a high ridge of hornfels rims the northeastern border of the pluton along White River, and a ragged wall of hornfels encircles the stock on Lost Creek. The original rock has thoroughly recrystallized and the hornfels looks homogeneous and sugary; but weathered surfaces reveal original outlines of clastic grains, pyroclastic fragments, and relict phenocrysts. The black color of the hornfels is imparted by metamorphic hornblende, hypersthene, and biotite; cordierite and tourmaline are abundant in some rocks. Original zeolites and clay minerals have been altered to quartz and feldspar. Some of the feldspar in the hornfels appears to be original—it pseudomorphs the outlines of feldspar phenocrysts and microlites in the original rock—but it is limpid and clear, unlike the dusty and altered feldspars characteristic of the original zeolitized lavas. In a few places much material was evidently added to the hornfels by gases streaming from the pluton. In the Sourdough Mountains, for example, near areas where the pluton is inferred to have broken through to the surface, tuff-breccias have been converted to rocks rich in pale-green amphibole, tourmaline, scapolite, zeolites, and other minerals.

In general, simple hornfelsized borders occur only where the pluton is in contact with the lower part of the Ohanapecosh Formation; where it has risen to stratigraphic levels as high as the Stevens Ridge Formation, its border is generally an unbelievably complex assemblage of hypabyssal rocks.

ROCKS OF THE HYPABYSSAL COMPLEXES
SILL AND DIKE SWARMS

The Tatoosh pluton is bordered by complex swarms of sills and dikes, which are similar in composition to the rocks of the core but are even more varied in texture. These sills and dikes are so numerous and their interrelations so complex, that for most areas no attempt was made to show them individually on plate 1. Where they are widely spaced, as along the ridge crests near the eastern boundary of the park, the larger sills and dikes were mapped; but great numbers have been omitted, especially in forest-covered valleys where exposures are poor.

One of the areas in which the sills are so abundant as to constitute the dominant rock extends from Skyscraper Mountain to Lake James. Here sill is piled on sill until 50 to 90 percent of the rock is intrusive. Septa of wallrock between sills are thin or have been so shredded and mangled by crosscutting dikes and sills that in places an indescribable jumble of interpenetrating intrusive rocks replaces the orderly succession of superposed sills.

Septa of wallrocks are much thicker and more abundant in the great swarm of sills and dikes that extends along Crystal Mountain between Chinook Pass and Silver Creek (pl. 1). Even here, however, all the larger sills are composite; two to five or more intrusions share a single chamber. Many of the younger sills cluster near the centers, or intrude along the edges of older sills; but some wander irregularly, breaking from one side to another, subdividing, or even turning abruptly upward and leaving the sill chamber as dikes. Excellent examples of dikes branching from the tops of two prominent sills can be seen just south of the Deadwood Lakes. Several of these are large enough to be shown on the map (pl. 1). Small dikes branching upward from a sill are well exposed in roadcuts just east of the westernmost tunnel between Box Canyon and the bridge over Stevens Creek. Composite and multiple sills can be seen in roadcuts between Cayuse Pass and Chinook Pass and along the ridge between Klickitat Creek and Deadwood Creek.

The intrusive rocks of the sill complexes are chiefly aphanitic to medium-grained porphyries, ranging in composition from diorite to quartz monzonite. Somewhat coarser sheetlike bodies of quartz diorite, quartz monzonite, and granodiorite occur within the tops of some of the larger stocks and show features that suggest huge "cedar-tree laccolith" complexes (fig. 38). These complexes appear to represent a transition between the main massive body of the stock and the surrounding complex of individual sills separated by septa of wallrock. Such a transition is especially indicated by relations in the roof zone along the southern margin of the Carbon River stock. The following observations were made chiefly in the area between June and Falls Creeks, where the stock consists mainly of tabular masses of medium-grained granodiorite and quartz diorite. In these rocks the plagioclase and hornblende form a semidiabasic framework which encloses a much finer grained but erratically distributed groundmass containing nearly all the potassium feldspar, quartz, and biotite. The rocks with abundant quartz and orthoclase generally have rudely granophyric groundmasses.

Especially significant is the mode of occurrence of these contrasting rock types in the field. They form tabular sheets, 50 to 200 feet thick and of great lateral extent, which dip 5° to 20° and conform in a general way to the southward dipping roof of the stock. Sheets near the top of the pile tend to be finer grained than those below. At the actual contact with the Stevens Ridge Formation is a thick mass of almost aphanitic quartz diorite porphyry, similar to that in the sills which riddle the Stevens Ridge Formation along the border of the Carbon River stock on Chenuis Mountain.

It seems probable that the granitic rock exposed in the southern part of the Carbon River stock was nearly all emplaced in the form of sill-like sheets even though septa of wallrocks are generally lacking or are represented only by a few fragments enclosed in the edges of the tabular intrusive bodies. The sills apparently were injected so rapidly and in such great volume that the top of the pluton center swelled into a huge cedar-tree laccolith with only a few intercalated septa of sediments. Evidence that the sills were intruded separately is still preserved in the finer grain size along the original sill contacts and in the character of the jointing; massive jointing in the central part of each sill contrasts with fine fracturing along its margins. The closely jointed parts weather rapidly, so that the steep south wall of Carbon River canyon is a series of cliffs and benches; each cliff marking the central massive part of a sill. Separate intrusion is further indicated by minor differences in composition and texture between adjacent sheets.

Except for the slightly coarser grain and the general absence of wallrock septa between them, these intrusive rocks do not differ notably from the sills that form the great sill complex of the Sluiskin Mountain area. On the precipitous east wall of the West Fork canyon, just east of Sluiskin Mountain, only 1 sedimentary septum, 100 feet thick, was seen in a 1,000-foot thickness of multiple sills.

PREFERRED ZONES OF SILL INTRUSION

The sills tend to cluster at or near the unconformity between the Ohanapecosh and Stevens Ridge Formations (fig. 38). The great pile of sills extending northward from Skyscraper and Sluiskin Mountains along the West Fork of White River follows this unconformity and invades the Stevens Ridge Formation just above it. Most of the sills in the intrusive complex near the front of the Carbon glacier, and also those of the Tatoosh Range, cluster near this unconformity.

As noted previously sills of gabbro and diabase, perhaps of the same magmatic cycle as the Fifes Peak, also follow this unconformity, and were later invaded by Tatoosh sills. An example of a Tatoosh sill cutting a Fifes Peak sill can be seen at the top of the west portal of the highway tunnel about 300 feet west of the Box Canyon Bridge.

The development of these interesting hypabyssal rocks in such profusion along the unconformity testifies to the relative ease with which the Tatoosh magma spread laterally beneath the light pumiceous rocks of the Stevens Ridge Formation after it had penetrated through the denser and less permeable Ohanapecosh strata below. At the unconformity the magma could lift and burrow beneath the thin cap of light rocks; furthermore, the near approach to the surface permitted vesiculation and the streaming of volatiles into the thin permeable roof. This hastened solidification and facilitated the upward breakthrough of magma from the center of the sill through its newly chilled roof to feed intrusions at slightly higher levels. The sill complexes of the Tatoosh pluton, however, are not all confined to this one stratigraphic level. Those at Crystal Mountain and on Shriner Peak invade the Ohanapecosh Formation well below the unconformity; those along the crest of Chenuis Mountain are mainly near the top of the Stevens Ridge Formation. The Fifes Peak Formation is the host of Tatoosh sills at Tolmie Peak, near Knapsack Pass, and along the border of the Carbon River stock.

CONTEMPORANEITY OF SILL COMPLEXES AND PLUTON CORE

The picture just outlined, however, is oversimplified. The interrelations between pluton core and hypabyssal sheath are complicated in other ways here described.

The pluton core only rarely grades upward and outward into sills; in most places it cuts them abruptly, or is separated from them by an intervening plutonic breccia composed of angular or rounded fragments of sill rock invaded by a coarser grained matrix. For example, the large composite sill on the ridge between Klickitat and Deadwood Creeks is cut off, invaded, and locally reduced to fragments enclosed in a plutonic breccia at its junction with the core of the pluton in White River valley. The extension of the pluton between Mineral Mountain and Vernal Park also slices across the sill complex exposed in Skyscraper and Sluiskin Mountains, as does the stock north of Adelaide Lake.

Symbols: Tdi, sills and dikes; Tg, granodiorite and related rocks; Tv, pyroclastic rocks; To, Ohanapecosh Formation; Ts, Stevens Ridge Formation; Tf, Fifes Peak Formation.

Tf

Tf

Ts

Ts

Ts

Ts

Tdi

Tdi

To

To

Tg
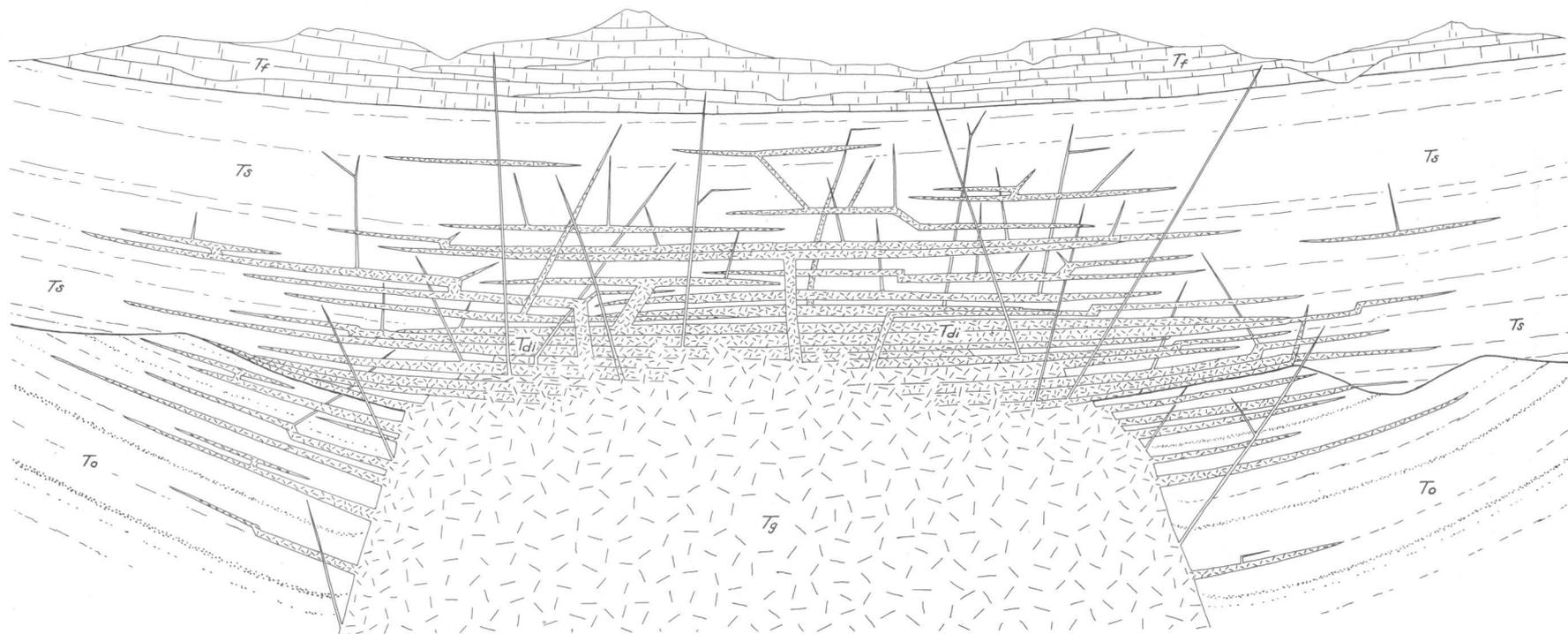
A
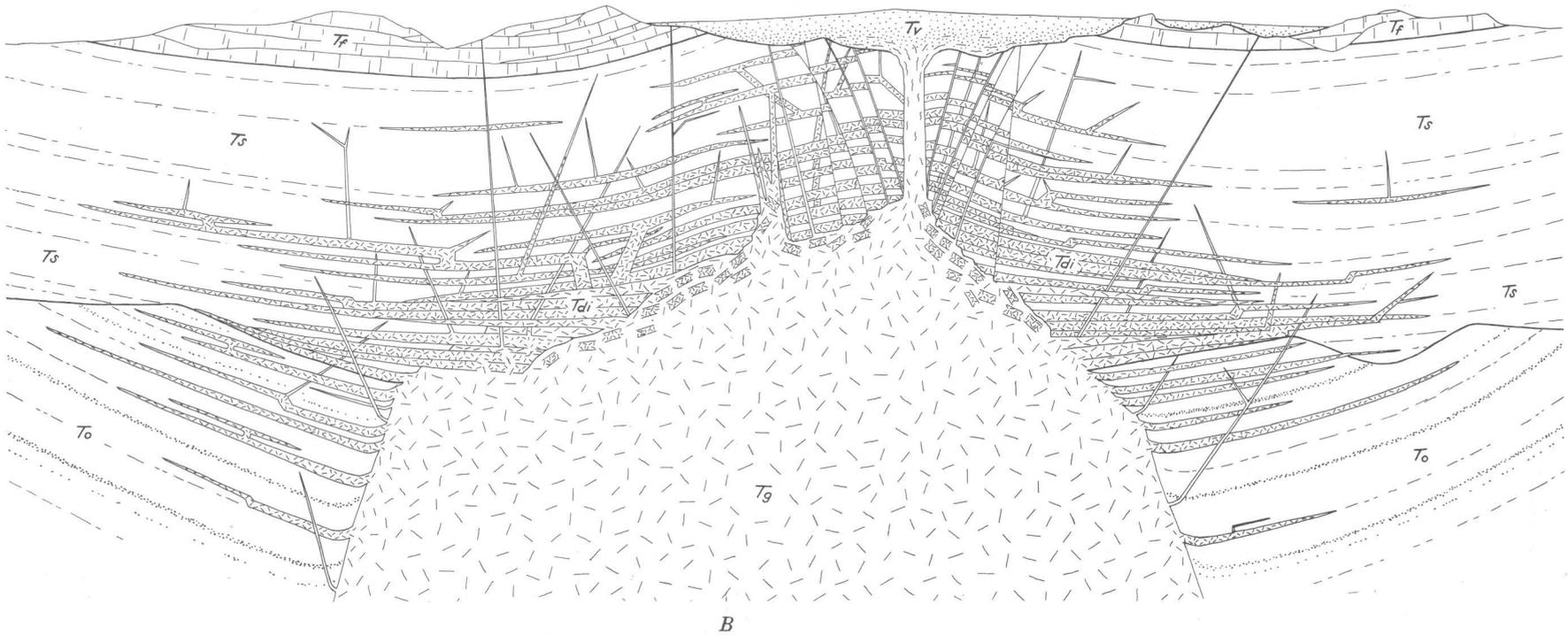
FIGURE 38.—Schematic cross sections illustrating the inferred relations between pluton core, sill swarms, and contemporaneous pyroclastic rocks. *A.* An early stage in the rise of a small pluton showing the development of a "cedar-tree laccolith" complex of sills and dikes in the roof zone at and near the unconformity between the Ohanapecosh and Stevens Ridge Formations. *B.* A later stage showing continued rise of the pluton into its cortege of early sills and local breakthrough of the magma to the surface forming contemporaneous pyroclastic rocks. A still later stage involving continued rise of the pluton and invasion of its own volcanic products by late dikes and sills, as at The Palisades and in the Cedar Lake quadrangle (Fuller, 1925), can also be visualized.

These contact relations seem to establish that the sills are at least slightly older than the main mass of the pluton—they do not invade the core of the pluton, and in many places the pluton invades them. But on the other hand, there is equally good evidence that the sills were intruded after parts of the core had already solidified; the most striking evidence is afforded by inclusions that are thickly sprinkled through many sills and dikes. In general, the sill rocks have the same range in composition as the rocks from the main pluton and its roof zone, but they are finer grained and most of them are markedly porphyritic. Many of the fine-grained to aphanitic sills of granodiorite porphyry and granophyric quartz monzonite porphyry are crowded with broken particles of minerals and small chips of coarse-grained rock obviously torn from the underlying pluton. Many of these fragments came from hydrothermally altered parts of the pluton, for flow banding in the groundmasses of some felsitic sills cuts sharply across the edges of previously broken crystals of chloritized biotite and saussuritized plagioclase and swirls into cracks in inclusions of altered granitic rock. In many sills broken sheaves of epidote replacing plagioclase, pieces of chloritized biotite, clots of uralitized pyroxene, and small chunks of granodiorite are thickly sprinkled in a fine-grained matrix consisting of unaltered plagioclase, potassium feldspar, quartz, and hornblende. Broken large crystals of uralitized augite and hypersthene occur in sills of fine-grained quartz monzonite porphyry whose indigenous mafics are hornblende or biotite. So closely do these inclusions and xenocrysts resemble the altered rocks from the core and roof zone of the Tatoosh pluton that there is little doubt that they were derived from it.

It may be significant that the sills and dikes containing abundant inclusions of minerals and rocks derived from the core of the pluton are also the ones that most commonly show records of vesiculation. Perhaps these sills were formed by the sudden and explosive boiling of parts of the half-solidified core of the pluton, causing rapid expansion and simultaneous intrusion into the walls. Such features seem to be most abundant near areas in which the pluton broke through to the surface and initiated a volcanic phase.

ROCKS OF THE VOLCANIC PHASE

Vesiculating magma in the higher cupolas of the Tatoosh pluton broke through to the surface at several points, causing large-scale volcanic activity. Visible manifestations of this contemporaneous volcanic phase include plugs whose connection to the deeper parts of the pluton can be traced, as well as the normal surface products of volcanic eruptions, such as welded tuffs, pumice lapilli tuffs, and lava flows. In Mount Rainier National Park the easily removable pyroclastic deposits have mostly been swept away by erosion, but this deep erosion has fully laid bare the connection between volcanic plugs and parent pluton. A small remnant of welded tuff survives at The Palisades, in the northeastern part of the park, and remnants of a hypersthene-oxyhornblende andesite flow possibly connected with this phase of volcanism are found at Windy Gap and Bee Flat in the Carbon River drainage.

WELDED TUFF AND PLUG AT THE PALISADES

At the northwestern corner of White River Park a great cliff of columnar jointed black rock, known as The Palisades, rises abruptly from the headwaters of Lost Creek. The rock in the cliff is a rhyodacite welded tuff, as much as 800 feet thick, which rests upon and grades into an underlying plug of rhyodacite. The welded tuff extends south from The Palisades, forming an irregular-topped plateau about a mile long and half a mile wide, rimmed by striking cliffs, 50 to 500 feet high, carved from joint columns more than 100 feet long. West of Hidden Lake this plateaulike remnant pinches into a narrow neck, only to expand again to the south and form the cap of Marcus Peak (pl. 1).

The welded tuff is obviously a remnant of a formerly much more extensive sheet—the edge of the tuff is today a high cliff that is constantly losing volume by erosion. It is unlikely, however, that the welded tuff maintained its full thickness for more than a mile or two in any direction from the plug that fed it, for this remnant lies within, and appears to have formerly overflowed, a downsagging structural bowl in the underlying Ohanapecosh rocks. In and near this bowl the Ohanapecosh strata are cut by sills of granodiorite porphyry and rhyodacite. Over a broad zone about 2½ miles in diameter the sedimentary rocks and associated sills dip inward from all sides (pl. 1). The bowl-like structure in the Ohanapecosh rocks is entirely independent of the regional structure—the strata appear to have sagged into an underlying magma chamber, perhaps immediately after it had been partly eviscerated by explosions to the surface.

The contacts of the welded tuff with the underlying Ohanapecosh rocks are complex, but they leave no doubt that at least the central part of the welded tuff rests upon, and grades downward into, a large plug of flow-banded and spherulitic rhyodacite which fills the center of the sag in the Ohanapecosh rocks. The rhyodacite plug (and in places the welded tuff above it)

cuts sharply across Ohanapecosh strata. The contact dips steeply toward the center of the plug on its northeast, northwest, west, and southeast edges. But beneath Marcus Peak the welded tuff rests directly on Ohanapecosh rocks and interleaved sills; here the base of the welded tuff dips gently to the north, conformable to the sag of the underlying Ohanapecosh strata (fig. 39).

At most places the contact between the welded tuff and the underlying plug is hidden by talus and rock glaciers, but an exceptionally interesting section of it is well exposed on the cliffs and steep slopes northwest of Hidden Lake. Here one can observe a transition of the rhyodacite plug upward into the welded tuff (fig. 40). Just above the talus slope streaming down to Hidden Lake are low cliffs of a porphyritic microcrystalline rhyodacite which forms part of the plug. Faint to distinct flow banding in the rhyodacite conforms to the walls of the plug; it stands nearly vertical or dips 65° or more toward the center of the plug. Above this steeply inclined rhyodacite are two rudely horizontal zones; both of these zones vary in thickness, but the lower averages about 50 feet and the upper 35 feet. The lower zone has been formed by the sudden flaring out of the vertically banded rhyodacite in the lower part of the plug to an almost horizontal position. This abrupt bending outward of the flow bands is accompanied by a tremendous increase in the number of spherulites, by the formation of much thinner, more distinct, and more contorted flow bands, and by the appearance of much glass (now devitrified) in the groundmass of the rhyodacite. Indeed, the upper part of this zone is a vitrophyre thickly crowded with bands of spherulites, and containing collapsed vesiculated patches infiltrated with coarse-grained devitrification products and pneumatolytic minerals. Some vesiculated layers grade into thin zones of breccia composed of altered froth fragments enclosed between layers of spherulitic vitrophyre.

This lower zone of horizontally flow banded spherulitic vitrophyre is cut by vertical breccia dikes from a few inches to more than 15 feet in thickness. In places the dikes subdivide, or pinch and swell abruptly, but most have clean-cut walls against the flow-banded vitrophyre. They are filled with angular fragments, which consist chiefly of altered pumice and pumiceous vitrophyre but also include many pieces of the adjacent dense spherulitic vitrophyre and typical sill rocks from the subjacent Tatoosh pluton. When traced upward, some of these dikes merge into the breccia that forms the upper horizontal layer.

This upper horizontal zone, which consists of a breccia about 35 feet thick, grades rather abruptly downward into the banded spherulitic vitrophyre which composes the lower zone. The breccia of the upper zone contains abundant fragments, originally chunks of frothy spherulitic pumice, that have collapsed into flat pancakes. These pumice fragments are completely devitrified and impregnated with amphibole, quartz, zeolites, and other secondary minerals. Spaces between them are completely sealed by the products of devitrification and pneumatolytic activity, giving the rock a deceptive massive appearance which is belied on close examination of surfaces etched by weathering. The altered rock weathers quickly and forms slopes of loose granules.

At its top this upper zone of altered breccia grades abruptly within a few feet into the great sheet of columnar welded tuff that rises in a high cliff above Hidden Lake and continues southward until it joins The Palisades. The welded tuff differs from the underlying zone of compacted breccia; its matrix consists of tightly welded glass shards and small bits of collapsed pumice instead of the compacted, but partly open, framework of coarse fragments that characterizes the underlying breccia. The tuff also is considerably less altered than the breccia.

A close genetic relation between the breccia and the overlying welded tuff is indicated by several features, two of which are easily visible in the field: (1) The striking columnar joints of the welded tuff pass downward without interruption into the top of the underlying breccia. Here, however, they are not conspicuous in outcrops because the breccia is so susceptible to
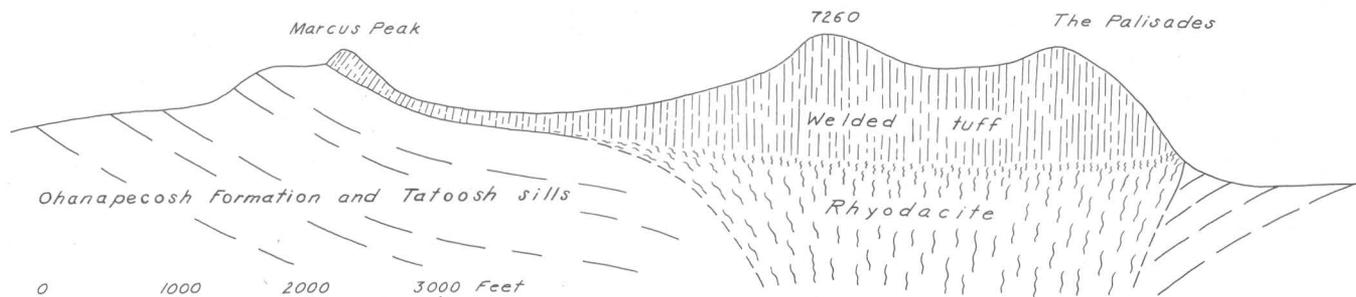

FIGURE 39.—Generalized cross section of The Palisades showing a remnant of welded tuff, more than 800 feet thick, grading downward into a rhyodacite plug. The welded tuff fills a downsagging area in the underlying Ohanapecosh rocks.

Molten magma must still have been available in the underlying Tatoosh chamber after solidification of the welded tuff, for a dike of coarse granodiorite porphyry injects the tuff south of the saddle between Marcus Peak and The Palisades.

The microscopic character of the welded tuff also shows its genetic relation both to the underlying rhyodacite plug, and to the subjacent Tatoosh pluton. The welded tuff appears glassy to aphanitic on fresh fractures, closely similar to vitrophyres from the western part of the Sourdough Mountains, but thin sections reveal its typical vitroclastic texture. It consists mainly of glass shards and tiny fragments of pumice, so thoroughly compacted and welded that before devitrification the rock must have resembled obsidian or perlite. The shards are draped around abundant rock fragments and broken minerals (figs. 41, 42). Most of the fragments of rock are like those in the sills that form the sheath of the Tatoosh pluton. Many euhedral crystals of plagioclase enclosed in the tuff were probably phenocrysts in the magma that frothed to form the glass shards, but other mineral grains clearly were derived from plutonic rocks that had been hydrothermally or deuterically altered before they were shattered and incorporated in the explosion cloud that deposited the welded tuff.

The inclusions, other than euhedral phenocrysts, that were noted in three large (3-in. by 2-in.) thin sections cut from the welded tuff are of several kinds:

1. Abundant broken crystals of slightly altered plagioclase. Some are tabular, others blocky and with marked oscillatory and progressive zoning; most of these were torn from medium- to fine-grained plutonic rocks, for many of them are attached to other altered minerals of the same grain size—chiefly chloritized biotite, hornblende, quartz, and uralitized pyroxene. The cores of some of the zoned plagioclase xenocrysts contain the characteristic tiny inclusions of rounded pyroxene found in zoned plagioclase from many parts of the Tatoosh pluton.

2. Numerous ragged grains of altered hornblende and biotite; also grains and clots of pyroxene in all stages of alteration to uralitic amphibole (fig. 42). These minerals were partly uralitized or chloritized before being enclosed in the tuff, for the shards are compacted against broken surfaces that cut both altered and unaltered parts.

3. Sparse to abundant rock fragments. The most numerous of these are chips of medium- to fine-grained rocks like those found in the sill and dike complexes along the border of the Tatoosh pluton (fig. 41). Some are encased in a lacy edging of plagioclase pheno-

crysts which apparently stuck to their margins after the inclusions were engulfed in the magma but before the eruption. Much less abundant are fragments of granodiorite like that found in the core of the pluton, and bits of hornfels derived from tuffaceous sedimentary rocks; still less abundant are chips of andesite (fig. 42) probably derived from lavas or mudflows in the Ohanapecosh Formation.

The rhyodacite plug rocks from beneath the welded tuff range in texture from microcrystalline to glassy. Even the microcrystalline specimens consist of unusually small andesine microlites, hornblende needles, pyroxene granules, and interstitial glass, tightly packed in a typical flow-banded pilotaxitic pattern. Phenocrysts of zoned plagioclase (about An_{50} to An_{30}) are abundant, hypersthene phenocrysts are less numerous, and there are a few resorbed phenocrysts of monoclinic pyroxene. The rhyodacite contains sparse inclusions of the same kinds found in the welded tuff.

The glassy rocks from the plug are highly spherulitic and mostly devitrified. The glass contains forked microlites of plagioclase, grains of pyroxene, and tiny scopulites of hornblende. Devitrification of glass released much quartz and potassium-rich feldspar. Quartz and potassium feldspar were also detected in the microcrystalline parts of the plug, chiefly in vugs and as concentrations along the better defined flow bands.

Such rocks are called rhyodacite in this report, but they raise problems in nomenclature. Chemically the glassy rocks are similar to granodiorite; but if they had not been devitrified, the quartz and potassium feldspar would have remained occult in the glass, and such rocks would have been called andesite. Because the spherulitic vitrophyre in the plug at The Palisades has been largely devitrified, and because almost complete magmatic crystallization has taken place in the microcrystalline parts of the plug, it can be seen from petrographic studies, even without chemical analysis, that the magma was highly siliceous and moderately potassic. These rocks are accordingly called rhyodacite, a name that closely expresses both their chemical character and their genetic relationship with the rock that makes up most of the Tatoosh pluton.

ROOT ZONE OF A PLUG IN THE WESTERN SOURDOUGH MOUNTAINS

The western part of the Sourdough Mountains is a rugged ridge that extends in a curving line from just north of Sunrise Lodge to Fremont Lookout. This ridge is carved from a complicated melange of intrusive rocks which include small amounts of coarse-

grained granitic rocks and medium- to fine-grained sill rocks, but it consists mainly of a porphyritic black aphanite with a waxy to dull luster.

Texturally the aphanite ranges from vitrophyric to hypocrystalline or even microcrystalline. Some parts are crowded with spherulites and most of it is flow banded, but the bands are 1/4 to 3 inches thick—much thicker than the finely laminated flow bands characteristic of acidic lavas. Some of this coarse banding is due to alternations of spherulitic and nonspherulitic layers, but nonspherulitic parts of the mass are also banded.

These rocks are nearly identical in mineralogy to the rhyodacite from the plug at The Palisades. The more crystalline varieties consist of a nearly irresolvable mat of andesine and hornblende needles, variable amounts of pyroxene, sparse quartz, rare biotite and sanidine, and considerable devitrified glass. Strongly zoned plagioclase and minor amounts of hypersthene and augite occur as phenocrysts.

The relations of this aphanite to the Tatoosh pluton, which crops out to the south in the canyon of White River, are not easily interpreted. Unfortunately the contact between aphanite and pluton has been almost completely buried beneath the Yakima Park intracan-

yon flow from Mount Rainier. This contact is exposed, however, although rather poorly, on the ridge half a mile northeast of Sunrise Lodge, and here the granodiorite invades vitrophyre. A contact between granodiorite and vitrophyre is also exposed on the floor of the cirque just northeast of Fremont Lookout, at the head of a tributary of Huckleberry Creek. Here the granodiorite grades into a coarse breccia composed of fragments of fine- to medium-grained granitic rocks immersed in a faintly flow-banded matrix of vitrophyre, showing that here the vitrophyre is the younger. Toward the south these fragments decrease in abundance, and the vitrophyre changes to a microcrystalline aphanite. Similar breccias, but less well exposed, were noted in the cirque west of Forest Lake. At several places the aphanite cuts porphyry sills considered to be offshoots of the Tatoosh pluton, but at a few places the vitrophyre is cut by dikes of similar porphyry. Evidently the aphanite and vitrophyre, younger than the granodiorite in some places but older than it in others, formed during an intermediate stage of the complicated cooling history of the Tatoosh pluton.

The contacts of the aphanite with the Ohanapecosh Formation, although clearly intrusive, are complex in form and attitude. The vitrophyre and microcrystal-
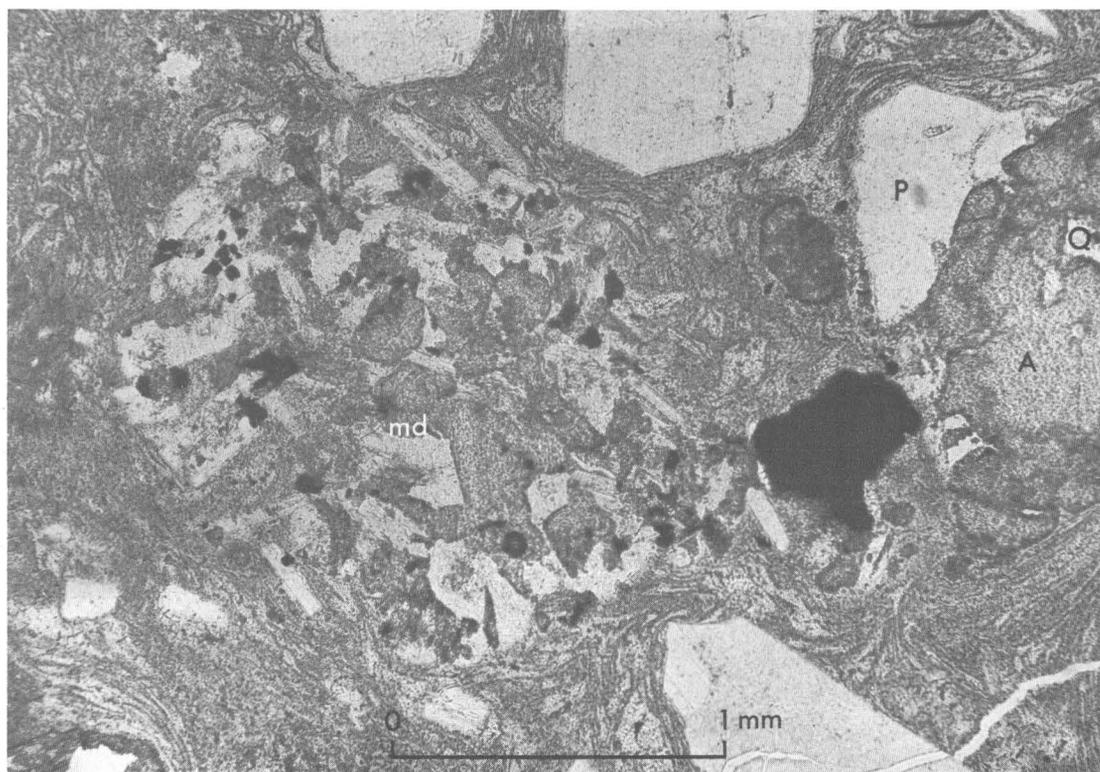
FIGURE 41.—Photomicrograph of welded tuff from The Palisades. Devitrified welded shards are compacted around a fragment of microdiorite (md) probably torn from a Tatoosh sill, and around a fragment of coarse-grained plutonic rock composed of plagioclase (P), chloritized augite (A), and quartz (Q).

line rhyodacite form sills in the Ohanapecosh Formation at and near McNeeley Peak, but in most of the area between McNeeley Peak and Cold Basin the main body of the rhyodacite cuts across the clastic rocks nearly at right angles to their strike. At the few places in this area where the contact could be seen it dips 45° to 65° to the south; here the Ohanapecosh rocks project beneath the igneous mass as an inclined, but discordant, floor. This southward-dipping contact is also reflected in the internal structure of the overlying aphanite. The north side of Mount Fremont, just south of the contact, is composed of successive wedges of flow-banded aphanite that also dip south. Each successively higher wedge dips more steeply than the one directly beneath it, and its base also truncates the flow banding of the underlying wedge. These relations recall the complicated fanlike flow bands and shear structures commonly observed in the throats of large volcanic necks and domes of rhyolite and dacite.

The inference that these crosscutting wedges of aphanite on Mount Fremont may mark the throat of an old volcano, through which the congealing Tatoosh pluton broke through to the surface, is supported by the fact that the vitrophyre in the tops of the higher wedges is highly vesicular. The jagged spires that form the southern part of the crest of Mount Fremont consist for the most part of highly amygdaloidal vitrophyre in which vertical swarms of vesicles are filled with a pale amphibole accompanied by minor amounts of quartz, epidote, chlorite, and scapolite (fig. 43). Yet the pluton must have remained molten at shallow depths below, for the pneumatolytically altered amygdaloid is invaded by a coarse aplite dike just beneath the crest of the south peak of Mount Fremont.

Still other features suggest that these aphanites of the western Sourdoughs may be merely the deeper parts of a larger and more complicated plug of The Palisades type. Vitrophyre from near the margin of the supposed plug (that is, from the floor of the cirque west of Mount Fremont and from ledges about half a mile east-northeast of Fremont Lookout) show thin layers of collapsed and tightly welded pumice between wider layers of vitrophyre. Parts of the amygdular rock at the summit of Mount Fremont are
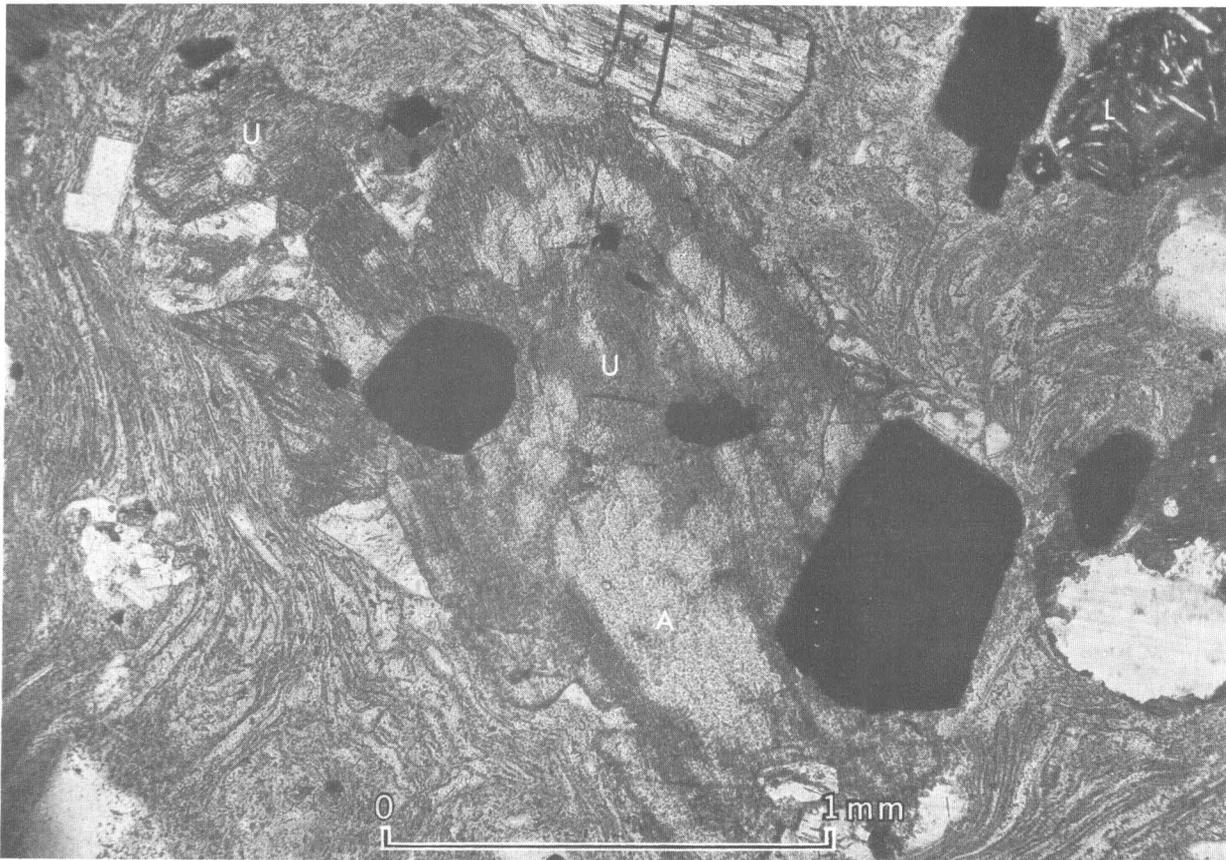
FIGURE 42.—Photomicrograph of welded tuff from The Palisades. Inclusions set in a matrix of devitrified welded shards. Augite (A) from a coarse phase of the Tatoosh pluton had been partly uralitized (U) before its incorporation in the tuff. At upper right is a fragment of lava (L) probably derived from the underlying tuff-breccias of the Ohanapecosh Formation. Note stretching and compaction of shards between fragments.

shattered into a coarse breccia of angular fragments. A similar explosion breccia has burst through Ohana-pecosh rocks as well as vitrophyre in an irregular area along the summit of the Sourdough Mountains half a mile southeast of Mount Fremont. This breccia consists of sharply angular fragments with voids partly to completely filled with pneumatolytic minerals, chiefly pale-green amphibole, quartz, epidote, tourmaline, scapolite, pyrite, and magnetite. In the aphanite adjacent to these shatter zones the work of streaming gases is also recorded by vertical swarms of coarse amygdules filled with amphibole. These may grade into small vertically alined bodies of breccia or into a rock crisscrossed with joints that have been filled with veinlets of amphibole, quartz, and zeolites. These features probably resulted from long-continued accumulation of volatiles, followed by their explosive release, during the rise of the main pluton into its own previously solidified volcanic cap.

It thus appears likely—although erosion has destroyed much of the evidence—that in the western Sourdough Mountains, as well as at The Palisades the Tatoosh magma broke through to the surface with explosive violence. Breakthroughs probably occurred in still other parts of the Park. The brecciated and pneumatolytically altered rocks of Mazama Ridge, underlain by a great thickness of fine-grained quartz monzonite with a granophyric groundmass, was cited earlier as the probable root zone of a Tatoosh volcano. The occurrence of strongly amygdaloidal rock in sills and in the fine-grained roof zone of the pluton, and the presence in these rocks of inclusions like those in the welded tuff of The Palisades and in the aphanites of the Sourdough Mountains, afforded evidence of violent underground vesiculation (fig. 36), clearly indicating that the Tatoosh magma was emplaced under little cover over wide areas. Under such conditions local breakthroughs to the surface would have been inevitable. Many plugs that probably connect with the Tatoosh pluton at depth have been located outside the boundaries of the park. Two of them will be briefly described here. A cluster of plugs and small stocks near Naches Pass is mentioned on page 61.

One huge plug of flow-banded vitrophyre and contemporaneous welded tuff forms Clear West Peak, which lies only a mile north of the Park boundary and just west of the West Fork of White River. As seen in the cliffs overlooking White River, the lower part of this plug consists of an outer carapace of strongly flow-banded vitrophyre with well-developed columnar joints. The columns lie nearly horizontal against the cooling surface of older rocks that form the wall of the plug, but toward its center they curve upward to a

nearly vertical position. The plug is composite; its upper part is composed of separate roughly vertical fillings of gray welded tuff and orange to brick-colored oxidized welded tuff. These are a little coarser grained and contain larger pumice lapilli, but otherwise are petrographically identical with welded tuff from The Palisades. One of the most interesting parts of the Clear West plug, however, is a large mass of completely undevitrified obsidian and perlite—much the freshest rock found in any Tatoosh plug. In thin
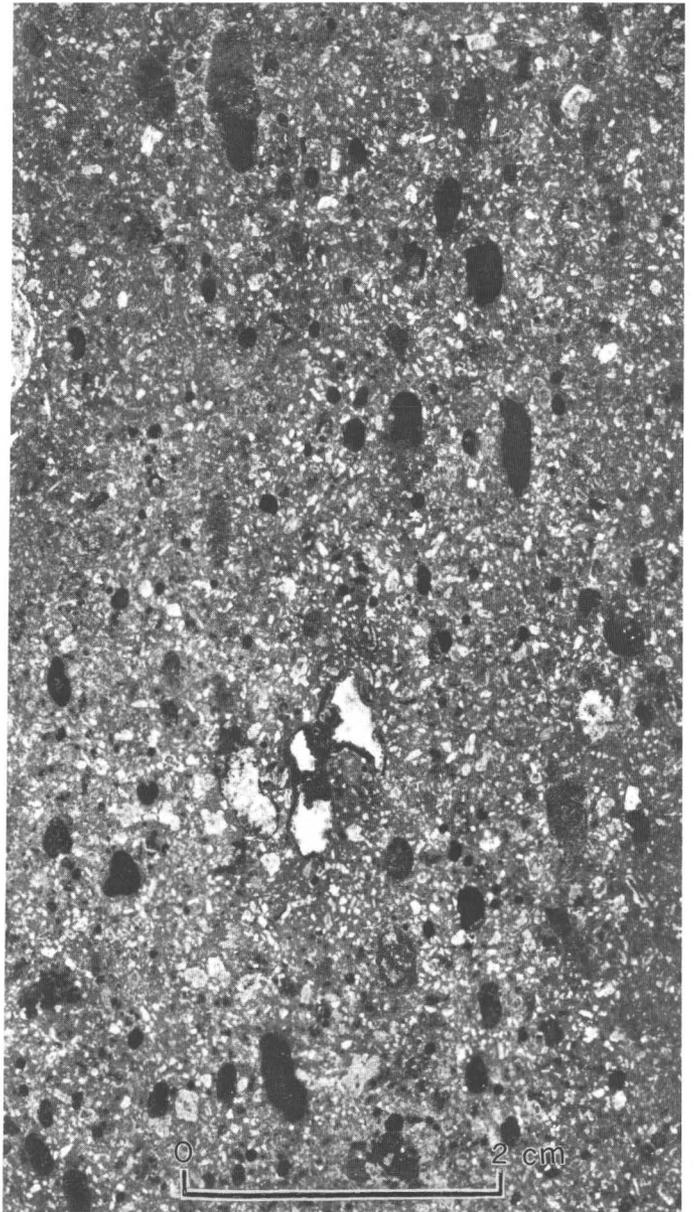
FIGURE 43.—Sawed and polished slab of vesiculated intrusive vitrophyre from the roof zone of the Tatoosh pluton, crest of Mount Fremont, Sourdough Mountains. The rock contains swarms of vertically alined amygdules filled with amphibole, magnetite, biotite, and chlorite; three large vesicles near the center of the photograph show a later central filling of quartz. This vitrophyre marks the root zone of a vent through which the Tatoosh magma broke through to the surface.

section it shows sparse phenocrysts of andesine and hypersthene, enclosed in a brown glass dusted with feathery scopulites of hornblende that resemble those from the famous pitchstones of Arran. A thick dike-like mass of vitrophyre, probably an offshoot from the Clear West plug, is exposed on Scarface Creek, about a mile to the southeast.

Another large plug, composed chiefly of altered hypersthene-hornblende rhyodacite, cuts Ohanapecosh rocks on the ridge between Skate and Willame Creeks south of Mount Rainier National Park. It shows highly contorted banding and vesiculated zones near its top. A large group of similar plugs described on page 61, lies north of Naches Pass, which is located in the Snoqualmie quadrangle about 7 miles from the northeast corner of Mount Rainier National Park.

RELATION OF THE TATOOSH PLUTON TO NEARBY AREAS OF TERTIARY INTRUSIVE ACTIVITY AND VOLCANISM

SNOQUALMIE PASS AREA

The idea that the Tatoosh pluton broke through to the surface and initiated a volcanic cycle is not a new concept for the middle-Tertiary granitic masses of the Cascade Mountains. More than 35 years ago Richard E. Fuller (1925), showed that the much larger Snoqualmie batholith had deroofed itself explosively. Fuller called attention to Smith and Calkins' (1906, p. 9) description of four stocks of pyroxene diorite, two of which (the Meadows Pass stock and a stock near Green Pass) they described as grading laterally and upward into flows of the Keechelus andesitic series. Smith and Calkins considered these to be "the roots of the volcanoes from which the Keechelus volcanics were erupted." Fuller's contribution was to show that these stocks are cupolas on the subjacent Snoqualmie batholith and to demonstrate that not only at Meadows Pass but also at localities within the adjacent Cedar Lake quadrangle there is a gradation between the pyroxene diorite and the Snoqualmie batholith. The pyroxene diorite forms a "cap" or border facies in the higher cupolas of the Snoqualmie mass.

Fuller concluded, on the basis of work in the Cedar Lake quadrangle and review of the occurrences described by Smith and Calkins, that the Snoqualmie mass deroofed itself explosively in many places. He suggested that most of the fragmental material so conspicuous in the upper part of the Keechelus andesitic Series (as contrasted with the deformed and more altered lower part of the Keechelus) was produced by this explosive deroofing. With waning of explosive activity the vents finally filled with a "cap

of andesite," which grades into pyroxene diorite deeper in the vents. In Fuller's words (p. 58, 91, 93) :

the vast bulk of the fragmental material was formed by the deroofing of the Snoqualmie batholith during its final great advance * * *. The great loss of volatiles resulting from these eruptions caused a premature solidification of the deep-seated magma * * *.

When the [magma] reached the surface it would lose its gas and quickly solidify, blocking the vent. While the batholith still contained a high content of volatile matter, the pressure of the gases imprisoned by this constantly forming crust would be sufficient to form a series of explosions. This would account for the great thickness of fragmental material in the Keechelus series. The final gradational passage from andesite to pyroxene diorite would have been formed when the gas pressure had at last diminished sufficiently to permit the survival of the final cap. At no great depth below this dense andesitic cap the slow heat conductivity of the rock would permit a fairly normal solidification, and the formation of a medium-grained rock * * *.

The few remaining pent in gases liberated by the crystallization of the granodiorite are considered sufficient to account for the higher degree of alteration of [this cap of] pyroxene diorite [than of the core of the batholith].

There is abundant evidence in the area to prove the chilling of the batholith by dehydration. The texture of these rocks is invariably closer to a porphyry than to a true plutonic type. Even specimens well within the batholith appear in thin section like dike rocks.

The history of the Snoqualmie batholith thus appears to have paralleled that of the Tatoosh pluton. But in the Snoqualmie-Cedar Lake district, areas of contemporary pyroclastics produced by deroofing have survived erosion, whereas near Mount Rainier—aside from the welded tuff at The Palisades—only a few tuffaceous and vitrophyric vent fillings remain. In Mount Rainier National Park the evidence of contemporaneous volcanic activity must rest largely on the abnormal characteristics of the pluton, its cortege of vesiculated and inclusion-filled sills, and the root zones of a few volcanic necks. For evidence based on the relation of extensive lava flows and pyroclastics to rhyodacite plugs and quartz-bearing pyroxene diorite stocks of Tatoosh affinities, we must cite the Naches Pass area, just northeast of Mount Rainier National Park.

NACHES PASS AND SAND CREEK AREAS

Naches Pass lies in the Snoqualmie quadrangle, about 7 miles northeast of Mount Rainier National Park (fig. 32). It is the locality where Smith and Calkins (1906, p. 8) described a clear-cut separation of the Keechelus Andesitic Series into an upper Keechelus composed of fresh lavas and a lower Keechelus composed of more altered and deformed flows and pyroclastics. In the limited time at their disposal

they were unable to trace this separation throughout the quadrangle, because (p. 8):

In areas farther south, the separation was found to be impracticable. In the lower series some andesite quite similar to that of the upper flows was seen, plainly overlain by the Ellensburg formation. The effort to draw a boundary between this rock and the supposed post-Miocene lava was unsuccessful since the older rock is less altered than usual, probably because it consists of massive lava.

The Sand Creek area lies 5 miles southeast of Naches Pass; here Smith and Calkins found the Keechelus Andesitic Series to rest upon the Yakima Basalt and to be overlain by pumiceous and conglomeratic sedimentary rocks of the Ellensburg Formation.

Field review of these areas in the light of the stratigraphic succession of Mount Rainier National Park indicates some significant interrelations in stratigraphy and geologic history between these parts of the Snoqualmie quadrangle, adjacent parts of the Mount Aix quadrangle on the south (Warren, 1941), and Mount Rainier National Park on the west.

Warren (1941, p. 799, 801, 809) assumed that the Fifes Peak Andesite, which he split off from the Keechelus Andesitic Series along the common border of the Mount Aix and Snoqualmie quadrangles, was the equivalent of Smith and Calkins' upper Keechelus at Naches Pass. But Warren (written communication, 1961) had not visited the Naches Pass area; he found an unconformity in the Mount Aix quadrangle between a deformed series of volcanic clastic rocks and his overlying Fifes Peak Andesite and associated tuffaceous sedimentary rocks, and he reasoned that this was the unconformity Smith and Calkins had described as separating the upper and lower parts of the Keechelus at Naches Pass. The relations at Naches Pass, however, show clearly that the Fifes Peak Formation corresponds with the lower part of the Keechelus of Smith and Calkins, and that the upper part of the Keechelus thus lies unconformably upon the Fifes Peak Formation (Waters, 1961). The unconformity that Warren recognized in the Mount Aix quadrangle is apparently the one that separates the Ohanapecosh Formation from the overlying Stevens Ridge Formation in Mount Rainier National Park.

At Naches Pass a large remnant of upper Keechelus lava lies in a topographic sag eroded in the yellow tuffs and black lava flows of the Fifes Peak Formation. The most abundant upper Keechelus rock is a highly glassy dull-black pyroxene andesite with thin wobbly columnar joints. As noted by Smith and Calkins, it crops out in a huge cliff overlooking the headwaters of the Middle Fork of Naches River. This cliff has been eroded from a flow that filled a former valley to a

depth of more than 300 feet. This topographic relation was noted by Smith and Calkins (1906, p. 8), who wrote: "The lava forming this ridge appears to have poured out after the extensive erosion which the Cascade Mountains suffered in Pliocene time."

The valley-filling flow is joined on the north by a different flow of highly porphyritic hypersthene andesite (probably a rhyodacite chemically), characterized by a glassy groundmass which shows, under the microscope, an unusual pattern of flow-alined crystallites. The glass is thickly sprinkled with peculiar needle-shaped microlites of pyroxene, forked crystallites of plagioclase, and feathery scopulites of hornblende.

Near the headwaters of Sand Creek, 4 miles southeast of the tableland underlain by these glassy lava flows, Smith and Calkins mapped a remnant of the Ellensburg Formation which caps the crest of the ridge between Crow Creek and the South Fork of Naches River. Just west of this remnant of the Ellensburg the ridge culminates in a peak that reaches an altitude of 6,221 feet as marked in the Snoqualmie Folio. This peak, now called Ravens Roost (fig. 32), is underlain by andesite of the Fifes Peak Formation, but on the east slope of the peak volcanic conglomerates and thin-bedded pumice deposits at the base of the Ellensburg Formation rest unconformably upon, and bank against, eroded edges of the Fifes Peak lava. Farther down this ridge, in sec. 19, T. 18 N., R. 13 E., andesite flows, one of which has the same peculiar groundmass texture that was noted in the porphyritic lava at Naches Pass, are interbedded with the pumiceous sedimentary rocks of the Ellensburg Formation. At this locality, however, these sedimentary rocks and flows rest not on the Fifes Peak Formation, but upon Yakima Basalt, which crops out in the canyon of Sand Creek and in a tributary of the South Fork of Naches River. Farther south, on the walls of the canyon of Crow Creek, the Yakima Basalt covers and laps against the accumulation of lavas and mudflows clustered around the Fifes Peaks. Warren (1941) demonstrated that the Yakima Basalt overlies the Fifes Peak Formation in the Mount Aix quadrangle. He noted that the "Fifes Peak andesite formed a rugged shore for the Yakima basalt flows—a shore against which the basalt piled up to a height of at least 1,500 feet."

Comparison of the stratigraphic relations at Naches Pass and Sand Creek strongly indicates that the upper Keechelus of Smith and Calkins is contemporaneous with the Ellensburg Formation and associated flows at the Sand Creek locality. We infer, also, from the character of the lava that occurs in these flows and as detrital fragments in the Ellensburg Formation, that

the flows and pyroclastics were extruded during the volcanic phase that accompanied emplacement of the Tatoosh pluton. Three miles east of the Sand Creek locality Smith and Calkins (1906, p. 8) collected a specimen from a flow of "hypersthene andesite," which they describe as having a hyalopilitic base with abundant brown glass. They note that this rock "is perfectly fresh, and therefore probably belongs to one of the later post-Miocene flows"—a deduction that is confirmed by the appearance of Yakima Basalt in the bed of the Ncahes River just downstream from this locality. The specimen was chemically analyzed in the laboratories of the U.S. Geological Survey, with results that Smith and Calkins (1906, p. 8) comment on as follows:

The rock contained an unexpectedly high percentage of silica and the analysis as a whole bears a strong resemblance to that of the Tertiary granodiorite, to be described later, but the andesite is more siliceous. [It contains 62.77 percent of SiO_2; the granodiorite contains 60.49 percent]. According to the new quantitative classification, it falls in the same division, tonalose. It is evident that the glass is rich in silica and potash, and that, if it were completely crystallized, it would contain much quartz and considerable orthoclase, both of which are essential constituents of the granodiorite.

Chemically the rock is a rhyodacite, not an andesite; both it and the lavas near Naches Pass are similar in petrographic character to the vitrophyric rhyodacites from The Palisades and from Clear West Peak.

Plugs and stocks, probably the source of the flows in the upper part of the Keechelus Andesitic Series and of the pyroclastic debris in the Ellensburg Formation, occur nearby. Several plugs cluster in the mountainous area that stretches from the headwaters of Naches River to beyond Lester in the Green River valley. These are exposed in Pyramid Peak, Kelley Butte, Rooster Comb, and several unnamed buttes (fig. 32). Dikes and sills also invade the Fifes Peak Formation and the underlying Stevens Ridge Formation in this area. A few of these plugs consist of basic material, and perhaps fed basalts and andesites of the Fifes Peak Formation; but most of them are filled with the characteristic pyroxene andesites and hypersthene-hornblende andesites (probably rhyodacites on chemical analysis) similar to the Tatoosh-related plugs, dikes, and sills in Mount Rainier National Park.

In this same general area Smith and Calkins mapped three stocks of pyroxene diorite of the kind that Fuller (1925) showed to be cupolas on the top of the Snoqualmie pluton from which large amounts of pyroclastic debris had been erupted.

East of the region that contains Mount Rainier, Naches Pass, and Snoqualmie Pass, with their stocks of pyroxene diorite and associated rhyodacite plugs, lies the main area of outcrop of the Ellensburg Formation (G. O. Smith, 1903). The middle part of the Ellensburg Formation is composed chiefly of huge fans of reworked pyroclastic debris. Most of this formation consists of thin-bedded but strongly cross-stratified and channeled layers of ash and pumice, thicker bedded deposits dropped from water-pumice slurries, and volcanic conglomerates with a pumiceous matrix (Waters, 1955a, p. 673–675). These materials rapidly coarsen to the west. Northwest of Nile (which is in the Naches valley, 18 airline miles east of The Palisades plug and about 25 miles southeast of the cluster of plugs and pyroxene diorite stocks at the head of the Naches River basin) the Ellensburg Formation consists of coarse mudflows and volcanic conglomerates interspersed with pumiceous tuffs which are in part ash falls and in part water laid. As noted above, lava flows are intercalated with the Ellensburg farther up Naches valley and on the ridge crest bordering its tributary Sand Creek. Fuller (1925) regarded the great accumulations of pyroclastics in the upper part of the Keechelus andesitic series of the Cedar Lake quadrangle as having been formed by the explosive deroofing of the Snoqualmie batholith.

Most of the lava fragments in the conglomerates and mudflows of the middle part of the Ellensburg Formation are of "hypersthene-augite andesites" and "hypersthene-hornblende andesites" similar to those found in the lavas in the upper part of the Keechelus Andesitic Series at Naches Pass and in the porphyritic plugs of the Naches Pass-Lester region. Especially abundant are fragments of gray to black highly porphyritic lava and of a pale-pink to orange-red oxidized variant of the same lava. Both of these contain abundant strongly zoned phenocrysts of plagioclase, phenocrysts of hypersthene, and phenocrysts of augite or hornblende or both. Much more abundant than the lava fragments are fragments of porphyritic pumice. Significantly, these lava and pumice fragments are nearly everywhere accompanied by a few fragments of rocks characteristic of the Tatoosh sills and, more rarely, by fragments of granodiorite like that which forms the bulk of the Tatoosh and Snoqualmie plutons. Small chips of these intrusive rocks enclosed in fragments of pumice have been seen in many outcrops. It is reasonable to assume, therefore, that the great fans of pumiceous debris, volcanic conglomerates, and mudflows that characterize the middle part of the Ellensburg Formation are largely the water-laid explosive products from the deroofed Snoqualmie and Tatoosh plutons. The pyroclastic material was carried eastward from the vent areas by streams and

spread out on the newly formed depositional plain of Yakima Basalt and nonvolcanic sediments of the lower part of the Ellensburg Formation.[8]

It should be emphasized that the Ellensburg Formation does not consist entirely of volcanic materials derived from the Snoqualmie and Tatoosh plutons. The lower part of the Ellensburg Formation in the Yakima region contains much nonvolcanic debris deposited by an ancestral Columbia River. Some pumice layers are interstratified with this material, but volcanic sediments become dominant only in the middle part of the formation. The middle part of the Ellensburg, moreover, is overlain, in places with distinct unconformity, by sheets of coarse gravel composed of debris eroded from a wide variety of rocks from the Cascade Mountains and from flows of the Yakima Basalt. In the past these gravels have been included as part of the Ellensburg Formation, but they probably should be separated out as a distinct formation. Furthermore, even the prevailing andesitic (rhyodacitic) sediments of the middle part of the Ellensburg are interstratified in places with debris eroded from nearby uplifts of Yakima Basalt and other rocks (Waters, 1955a, p. 674).

EVOLUTION OF THE TATOOSH PLUTON: A SUMMARY

It is evident that the rise of the Tatoosh pluton involved an extremely complex series of events whose record is still not completely deciphered. The field and petrographic work already done, however, serves to outline some of the processes involved in the origin of this complex assemblage of igneous rocks and also permits some interpretations as to their timing.

The earliest magma to crystallize at exposed levels in the pluton chamber appears to have been a quartz diorite containing both hornblende and pyroxenes. Slightly later, but overlapping in time, are the grano-

[8] In a recently published report Leonard M. Gard, Jr. (1960) states: "It is here suggested that the [Snoqualmie] magma reached the surface and that the volatile constituents were given off during explosive volcanism." This volcanism, he claims, might have furnished the "hornblende andesite debris" found in the Pliocene Ellensburg formation and in Miocene deposits on the west flank of the Cascade Range.

Although Gard cites Fuller's master of science thesis, written in 1925, he does not give Fuller credit for the idea of explosive deroofing of the Snoqualmie mass, mentioning only Fuller's conclusion that the lack of ore deposits and late differentiates was due to loss of "volatile constituents." Yet the central theme of Fuller's thesis is that explosive deroofing produced the large quantities of pyroclastic debris that characterizes the "upper Keechelus" as defined by Smith and Calkins (1906, p. 8). See, for example, the quotations from Fuller (1925) on p. 59 of this report.

Our contribution to this particular aspect of the geology of the Cascade Range has been to correlate the upper Keechelus of the Naches Pass region with the Ellensburg formation and to show that in Mount Rainier Park a volcanic phase accompanied the intrusion of the Tatoosh pluton. More than 35 years ago Fuller contended—rightly we believe—that the same was true of the Snoqualmie mass.

diorites, quartz monzonites, porphyries, granophyres, and vitrophyres now found in the pluton core and in its hypabyssal sheath, and also the rhyodacite, and voluminous pumiceous pyroclastics that characterize the contemporaneous volcanic phase. Rise of the main pluton and of its satellitic stocks to the unconformity at the base of the Stevens Ridge Formation permitted easy penetration of sills and dikes into the thin light cover of ash flows and sediments. Large stocks and cupolas apparently did not invade the rocks above the unconformity, however, until the wallrocks were so riddled by sills that they formed a structure resembling a cedar-tree laccolith above each major center of intrusive activity (fig. 38). Arching and stretching of the sills by continued upwelling of the pluton beneath them may explain why clusters of dikes break upward from the tops of many sills. Late rise of the core of the pluton also accounts for the present strongly crosscutting character of the core of the pluton against dikes and sills intruded low in the stratigraphic section.

If these deductions are correct, the Shriner Peak complex of dikes and sills, near the east border of Mount Rainier National Park, probably lies above a stocklike mass not yet revealed by erosion, and the amygdaloidal intrusives near the top of Shriner Peak may once have connected with a volcanic vent. The great complex of sills in the Skyscraper-Sluiskin-West Fork area probably lies above a large underground extension of the pluton core to the north; the extensive sills near the front of the Carbon glacier probably hides an underground connection between the Tatoosh pluton and Carbon River stock. The southern part of the Carbon River stock displays the crest of a deeply eroded plutonic center which, nevertheless, still preserves subtle records of its early stage of growth in a cedar-tree roof of coarsely crystalline sills separated only rarely by septa of wallrock. The main core of the Tatoosh pluton reveals a still more deeply eroded area wherein the continued rise of the pluton has broken most of the junctions between core and early members of its sheath of sills, but in which some parts of the roof still survive and give evidence that the hypabyssal complexes and the pluton core developed contemporaneously.

Intermittent volcanic activity would be expected during the rise and emplacement of such shallow masses, for the arching of the roof rocks above the magmatic chamber and the displacements attending the intrusion of the many sills could not have failed to open fissures to the surface. Many fissures were probably sealed, as fast as they were formed, by con-

gealing of sticky magma in narrow dikes; but some large plugs certainly reached the surface and poured forth great volumes of pyroclastic material. In the Mount Rainier area most records of this surface volcanism have been destroyed by erosion, but The Palisades plug, with its cap of welded tuff, is a remnant proving that at least locally the Tatoosh pluton breached the surface with explosive violence. Root zones of other plugs, partly engulfed by continued rise of the magma, are recognized with less certainty at other localities. Several definite plugs, exposed at shallower levels of erosion, have been located outside the park, along with pyroclastics and lavas that they fed.

AGE OF THE TATOOSH PLUTON

The time of emplacement of the Tatoosh pluton has been bracketed by direct and indirect methods within fairly narrow limits. The youngest rocks cut by the pluton in Mount Rainier National Park are lavas of the Fifes Peak Formation, whose age lies in the interval from middle Oligocene through early Miocene. In the park these rocks were gently folded and removed by erosion from large areas before the pluton broke through to the surface at The Palisades. It is therefore probable that emplacement of the pluton did not begin before Miocene time.

The upper time limit of intrusion is fixed by the earliest lavas from Mount Rainier; they rest unconformably upon the coarse-grained rocks of the pluton core. A fairly long period of erosion must have intervened, however, between the congealing of the pluton and the birth of Mount Rainier volcano, probably in Pleistocene time. The age of the pluton thus appears to be bracketed, by these direct methods, between early Miocene and late Pliocene time.

An indirect method of dating can be based on a study of the pyroclastic rocks that are contemporaneous with the Tatoosh granitic rocks and the closely related Snoqualmie Granodiorite. Fuller has given strong evidence that deroofing eruptions from the Snoqualmie mass fed the large volume of fragmental material in the upper part of the Keechelus Andesitic Series. The products of explosive eruptions from both the Snoqualmie and Tatoosh areas were doubtless transported both eastward and westward by streams, and they ought to be recognizable in the sedimentary record of the adjacent areas. The middle part of the Ellensburg Formation, to the east of the Snoqualmie and Tatoosh areas, consists of such products, apparently deposited by streams whose headwaters repeatedly were choked by coarse pyrolastic material from Snoqualmie and Tatoosh eruptions.

From these relations the age of the pluton can be bracketed more closely than from the evidence gathered in the Park. The Ellensburg Formation was orignally assigned to the Miocene (G. O. Smith, 1903; Smith and Calkins, 1906), but it is now generally agreed that the middle part of the formation is of early Pliocene age (Axelrod, 1950; Chaney and Axelrod, 1959), although the question of whether the lower part of the Ellensburg—which also contains some pyroclastic debris similar to that found in the middle part of the Ellensburg—is of late Miocene or early Pliocene age has never been fully resolved. The lower part of the Ellensburg Formation lies conformably upon Yakima Basalt, and is interbedded in places with its uppermost flows. The Yakima Basalt ranges in age from late Miocene to early Pliocene (Waters, 1955a; Lohman, cited in Sheppard[9]). Fragments of Tatoosh and Snoqualmie in the Ellensburg Formation are thus no younger than early Pliocene. Snavely and others (1958, p. 66) have described a series of tuffaceous sedimentary rocks of early Pliocene age from near La Grande, about 30 miles southwest of Mount Rainier National Park. These tuffaceous rocks contain aboundant hypersthene, hornblende, and augite, possibly derived from Tatoosh pyroclastics.

Based on these correlations, the emplacement of the Tatoosh and Snoqualmie plutons thus appears to have occurred in the interval between the end of Fifes Peak deposition (early Miocene) and the building of the huge fans of pyroclastic debris that make up the middle part of the Ellensburg Formation (early Pliocene). This inference is supported by an absolute age determination on rock from the Snoqualmie batholith, made by Lipson and others (1961, p. 460). From brown biotite in biotite-hornblende granodiorite exposed in a roadcut on U.S. 10, southwest of Snoqualmie Pass (47°23′ N., 121°28′ W.), they obtained a potassium-argon date of 17 million years.[10]

[9] Sheppard, R. A., 1960, Petrology of the Simcoe Mountains volcanic area, Washington: Unpublished Ph. D. dissertation, The Johns Hopkins University, 153 p.

[10] Since this report was written Prof. R. E. Folinsbee (written communications, Nov. 29, 1962, and Jan. 17, 1963) has reported the A/K ages of biotite from two samples of Tatoosh granodiorite collected by us :
1. Biotite, Tatoosh granodiorite, roadcut 100 yd south of east abutment, Nisqually River bridge, Mount Rainier National Park. 46°47′ N, 12°46′ W. University of Alberta sample AK No. 398. Calculated age 14.7 ±1 million years.
2. Biotite, Tatoosh granodiorite, small quarry, White River valley at foot of Goat Island Mountain, Mount Rainier National Park. 46°53′ N, 121°31′ W. University of Alberta sample AK No. 429. Calculated age 13 ±1 million years.
These results strongly support the inference from stratigraphic and paleontologic data that the emplacement of the Tatoosh pluton occurred in late Miocene and early Pliocene time.

LAVA FLOWS AND MUDFLOWS OF UNKNOWN AGE

In the northwestern part of the park, on the southeast wall of Chenuis Mountain north of Spukwush Creek, there is a high bench known as Bee Flat. This bench is the remnant of a flow of unaltered hypersthene-oxyhornblende andesite (probably a dacite or rhyodacite chemically), which partly filled a former canyon cut in the Stevens Ridge Formation. Another remnant of the same lava lies at the head of Spukwush Creek, just northwest of Windy Gap. Marked changes in topography have occurred since this flow solidified. The Carbon River has cut down nearly 2,000 feet beneath the floor of the canyon that was filled by this lava, and the flow is crossed almost at right angles by an early intracanyon lava flow from Mount Rainier volcano. The Mount Rainier flow spilled north from Old Desolate through a gap between Crescent and Sluiskin Mountains, crossed the hypersthene-oxyhornblende andesite flow at Windy Gap, and extended from there to a point beyond Marjorie Lake (pl. 1). Talus has obscured the actual junction between the two flows at Windy Gap, but the topographic relations indicate that a stream canyon at least 300 feet deep was cut across the Bee Flat lava before arrival of the Mount Rainier flow.

The lava at Bee Flat may have been erupted during the period of Tatoosh volcanism. Shallow Tatoosh intrusive bodies west of Summer Land shelter cabin, at Panhandle Gap, near Indian Bar, and in other places in the park consist of hypersthene-oxyhornblende dacite very similar in appearance and mineralogy to this flow. These bodies, or the plug that forms Clear West Peak, may have been the feeders for this lava flow.

Fisher (1957) has also mapped unaltered andesites and felsites, tentatively assigned to the Pliocene, near Elbe, a few miles west of the southwest corner of Mount Rainier National Park.

A thick sequence of mudflows and coarse stream deposits, containing abundant cobbles of hypersthene-hornblende andesite very similar to those in the Ellensburg Formation of Naches valley, lies in the headwaters of Voight Creek—a tributary of Carbon River—and in the Mowich and Puyallup drainage areas just west of Mount Rainier National Park. Apparently only small remnants of these deposits occur within the park, and they have not been distinguished on the geologic map. Early pyroxene andesite flows from Mount Rainier rest upon these mudflows at two localities: (1) Just west of the park boundary, a Mount Rainier flow occupies a shallow valley eroded in mudflows near the head of Voight Creek; this lava flow, or one

very like it, also caps mudflow deposits a little farther south, along the Voight Creek-Mowich River divide. (2) Stream gravels and slurry-flood deposits containing hypersthene-hornblende andesite pebbles underlie Rainier lava near the head of the South Puyallup River, inside the park. According to D. R. Crandell (1963), the mudflow and stream deposits unite into a great compound fan which filled the ancestral Mowich and Puyallup valleys in the Cascade foothills and which formerly extended westward into the Puget Sound lowland. Crandell named these deposits the Lily Creek Formation.

The mudflows and volcanic clastic deposits of the Lily Creek Formation may have had their source in pyroclastic eruptions from Tatoosh volcanoes. Tatoosh vents are known at The Palisades, Mount Fremont, and Clear West Peak, and possibly others were located near Panhandle Gap, Shriner Peak, and Mazama Ridge. Mount Rainier volcano grew astride the former courses of the ancestral Mowich and Puyallup Rivers—before its growth these streams probably drained the area in which the Tatoosh vents were located. The fan of pyroclastic outwash of the Lily Creek Formation thus may have been built from debris flushed westward down the Mowich and Puyallup Rivers from Tatoosh volcanoes on the west slope of the Cascades, just as the huge fan of mudflows and pumiceous clastics of the Ellensburg Formation in Naches valley spread eastward from similar volcanoes around Naches Pass.

A different explanation of the origin of the Lily Creek Formation, however, is given by Crandell (1963). He mapped the glacial and interglacial deposits northwest of Mount Rainier National Park in the Puget Sound lowland, and although he was not able to establish the stratigraphic relations directly, he thinks it probable that the Lily Creek Formation is of Pleistocene Age and younger than the earliest of four stages of glaciation that affected the Puget Sound lowland. He shows, moreover, that the mudflows and stream deposits of the Lily Creek Formation filled the ancestral Mowich and Puyallup valleys, but not the adjacent valleys to the north and south. Having thus traced the Lily Creek deposits up the Mowich and Puyallup nearly to Mount Rainier, he concluded that voluminous eruptions during the first stages of volcanic activity on Mount Rainier must have supplied the hypersthene-hornblende debris in the Lily Creek Formation.

Crandell's explanation is tenable, but in our opinion it does not accord with the composition of the volcanic fragments found in the Lily Creek Formation. The

glassy hypersthene-hornblende lava in the cobbles of the Lily Creek deposits is similar to many volcanic and hypabyssal rocks of Tatoosh affinities, whereas the Mount Rainier lavas, in both the main cone and the early intracanyon flows that extend far outward from the base of the volcano, are pyroxene andesites virtually without hornblende. Remnants of hornblende crystals, thickly rimmed or completely replaced by magnetite or hematite, do occur in a few flows, but hornblende is nowhere a major constituent. Olivine is sparsely present in more than half of the Rainier flows and is abundant in some, whereas olivine is not a constituent of the hypersthene-hornblende andesites (probably dacites and rhyodacites) that characterize the Lily Creek and Ellensburg Formations. Finally, the eruptions from the volcanoes that supplied such vast quantities of fragmental debris to the Lily Creek and Ellensburg Formations must have been mainly explosive, whereas Mount Rainier erupted chiefly lava.

It might be argued that early eruptions from Mount Rainier built a pyroclastic cone of hypersthene-hornblende andesite, which is now completely covered by a later lava cone of pyroxene andesite and olivine andesite. But it seems very unlikely that, even after long-continued erosion, all remnants of a volcano that contributed such a hugh volume of pyroclastic material as is now found in the Lily Creek deposits could be completely hidden beneath the present cone of Mount Rainier. The huge cirques of Willis Wall and Sunset Amphitheater bite deeply into the present cone, but the boulders carried by the glaciers flowing from them consist of pyroxene andesite and olivine-pyroxene andesite.

It is not yet possible, however, to make a final choice between these two hypotheses, and furthermore, these are not the only possible interpretations of the observed facts. Tatoosh volcanism might not have ended in the late Miocene or early Pliocene, but it could have persisted in a few small areas into the early Pleistocene. The Lily Creek Formation, moreover, may consist of reworked materials from earlier Tatoosh volcanic deposits. Unconsolidated pumice-rich pyroclastics and stream-deposited materials laid down on the crest and western slope of the Cascades would be subject to continual erosion and redeposition by streams; such deposits in the Ellensburg Formation of Yakima valley were reworked all through Pliocene and Pleistocene time, and they are being reworked today (Waters, 1955a). The Lily Creek deposits, taken as a whole, are not so uniform in composition as the Ellensburg deposits of the upper Naches valley: they

contain abundant material from the older rocks of the central Cascades (Crandell, 1963) that differs from the dominant cobbles of hornblende-bearing andesite. They also contain more clay and much less pumice than the Ellensburg Formation does in upper Naches valley. In this regard the Lily Creek deposits resemble the widespread sheets of coarse gravel, composed partly of reworked Ellensburg debris, that rest unconformably upon the middle part of the Ellensburg Formation in the Yakima valley between Ellensburg and Cle Elum.

The final answer to these questions regarding the source and age of the Lily Creek deposits awaits further fieldwork. It will come from a careful analysis of the Quaternary and older deposits that are found in terraces and flood plains in all the valleys that radiate from Mount Rainier. Some of these valley fills are thick and stratigraphically complex; they record an exceedingly eventful history of erosion, volcanism, sedimentation, and glaciation. Interstratified falls of ash and pumice provide opportunities to correlate the deposits from valley to valley. The older valley fills are seldom preserved except beneath early intracanyon flows from Mount Rainier; elsewhere, they have been stripped away by erosion and replaced by later lava flows, mudflows, glacial deposits, ash falls, and stream deposits. Our work on the bedrock geology did not afford time to work out the history of these deposits in detail, but such a study would do much to clarify the source and age of the Lily Creek Formation.

MOUNT RAINIER VOLCANO

Southward, 9,000 feet above you, so near you must throw your head back to see its summit, is grand Mount Tacoma; its graceful northern peak piercing the sky, it soars single and alone. Whether touched by the glow of early morning or gleaming in bright noonday, whether rosy with sunset light or glimmering, ghost-like, in the full moon, whether standing out clear and cloudless or veiled among the mists it weaves from the warm south winds, it is always majestic and inspiring, always attractive and lovely. It is the symbol of an awful power clad in beauty. (Bailey, Willis, 1883).

Capping the Cascade Range for 600 miles, from Mount Baker, near the Canadian border to Lassen Peak in northern California, stands a row of snow-clad volcanoes. Eleven of these exceed 10,000 feet in height and between are scores of lesser size. But highest of all is Mount Rainier—Ta-co-man in the language of the Indians from Puget Sound—whose broad bulk towers 8,000 feet above the peaks at its foot and 14,410 feet above sea level (fig. 44).

FIGURE 44.—Mount Rainier and the Cascade crest from the north. Mount Rainier, altitude 14,410 feet, occupies the foreground. In the distance three other large Quaternary volcanoes rise above the deeply dissected platform of older lavas and volcanic clastic rocks: Mount Adams (alt 12,307 ft, 48 miles from Mount Rainier) at the left; Mount Hood, Oregon (alt 11,243 ft, 103 miles) near center, just to the left of the top of Mount Rainier; and Mount St. Helens (alt 9,671 ft, 50 miles) at extreme right. Photograph by H. Miller Cowling.

Mount Rainier is a composite volcano, built chiefly of andesitic lava flows, but also containing mudflows, breccia, and ash. It rose to its greatest height before the close of Pleistocene time, for now, unlike Mount St. Helens and some of her sisters to the south, it is deeply scarred by erosion. It was aptly described by Matthes (1914, p. 7), who said that it "is not a simple cone tapering to a slender pointed summit like Fuji Yama, * * * but rather, a broadly truncated mass resembling an enormous tree stump with spreading base and broken top."

Early reports give information on Mount Rainier's structure, the petrography of its lavas, and its large system of glaciers. Although the earliest explorers recognized the mountain's volcanic origin (Dr. W. F. Tolmie, 1833; Capt. J. C. Fremont, 1845), the first important observations on its geology were those of S. F. Emmons (in King, 1871, p. 161–165; 1879). Reconnaissances by I. C. Russell (1897), G. O. Smith (1897, 1900), and H. Landes (1905) added to knowledge of the volcano and its physiographic setting. The main petrographic features of its lavas were described many years ago by Hague and Iddings (1883), Oebbeke (1885), G. O. Smith (1897), and more recently by H. A. Coombs (1936) who also mapped the extent of Mount Rainier's lavas. The mountain's heavy mantle of glaciers drew early attention (Tolmie, 1833: King, 1871; Kautz, 1875) and has been vividly described by Russell (1897) and Matthes (1914).

TOPOGRAPHY COVERED BY THE EARLY ERUPTIONS

Mount Rainier volcano was built upon a rugged surface carved mainly in rocks of the Tatoosh pluton and the Stevens Ridge Formation (fig. 45). Stubs of the pre-Rainier surface more than 6,500 feet above sea level project though the lavas around the base of the cone between the upper Carbon and Russell Glaciers (7,000 ft), above Mineral Mountain (6,900 ft), near St. Elmo Pass (7,200 ft), below the terminus of the Fryingpan Glacier (7,200 ft), near Cowlitz Rocks (6,600 ft), at Glacier Island (7,650 ft), near Tokaloo Rock (7,500 ft), and near the terminus of the Edmunds Glacier (about 6,500 ft). The volcano (14,410 ft) thus rises 7,000 to 8,000 feet above the ridges and peaks on the surface that it buried. Thick intracanyon flows, the earliest Mount Rainier lavas exposed, provide striking evidence of the ruggedness of that surface.

Some flows are banked against ancient canyon walls as precipitous as any of those one can see in the park today. Others filled canyons to depths of 2,000 feet, but the original walls of these canyons have long since been removed by erosion. One immense flow, making up the bulk of Ptarmigan Ridge, is at least 1,200 feet thick. Such thick flows could not have spread upon a surface of low relief; they were confined in canyons.

Although the region was mountainous when the first eruptions began, the maximum relief at that time cannot be closely estimated. Because canyon cutting continued throughout the growth of the cone, the early relief cannot be measured by merely finding the difference in altitude between the bases of the lowest canyon-filling Rainier lava flows and the highest summits carved from older rocks, for if the flow is young, the canyon it occupies could have been cut during late Mount Rainier time. A better estimate can be derived from the difference in altitude between the bases of the earliest lava flows and the highest remnants of older rocks; this estimate, however, gives only a minimum figure, because erosion has reduced the height of all exposed pre-Rainier summits.

FIGURE 45.—Unconformity between Mount Rainier lava and Stevens Ridge Formation at Margaret Falls. The 500-foot cliff over which the cascade tumbles was carved from an intracanyon flow of Mount Rainier andesite (Qra). This flow rests on tilted tuffaceous sedimentary rocks of the Stevens Ridge Formation (Ts) in the foreground. The cliff cuts diagonally across the intracanyon flow, revealing the curving contact of the flow against the wall of the old canyon that the lava filled. Columnar jointing in the lava fans out to meet the canyon wall, staying perpendicular to the cooling surface. Recent ablation moraine from Cowlitz Glacier mantles bedrock in the foreground.

No criteria were found by which the earliest flows can be surely recognized. Lava flows in valleys that do not conform to Mount Rainier's present radial drainage pattern are presumably old. The amount of canyon cutting that followed the filling of a valley by a lava flow indicates, in a general way, the flow's relative age. Continuous flows that rest on, or close to, present-day canyon floors are presumably younger than those whose dismembered remnants are left clinging high on canyon walls. Lava flows are surely old whose remnants now form peaks or ridges, with no vestige of former canyon walls remaining, but whose exceptional thickness (greater than 500 ft) shows them to have once been canyon-filling flows.

Judging by these criteria, the oldest flows are in the northern and western parts of the park. One of these, a thick intracanyon flow whose remnants form the mountain called Old Desolate and the flat surface of Grand Park, followed an ancient valley running northeastward from Moraine Park to the lower part of Huckleberry Creek, oblique to Mount Rainier's present radial drainage pattern. The deep canyon of the West Fork of the White River is entrenched across this flow, separating it into two segments. Near Vernal Park the altitude at the base of the Grand Park flow is 4,500 feet, but Sluiskin and Skyscraper Mountains, carved from pre-Rainier rocks, rise nearby to altitudes of 7,015 and 7,065 feet, respectively, showing that the minimum relief was more than 2,500 feet when the flow was erupted. Seven miles downstream, where the flow ends, its base is at 3,000 feet, indicating a minimum relief of 4,000 feet between the end of the flow and the top of Skyscraper Mountain. Along the North Puyallup River, pre-Rainier rocks near Tokaloo Rock also rise 4,000 feet above the base of early Rainier flows.

The early drainage can be partly reconstructed from remnants of the older intracanyon flows, as shown for the course of the vanished northeast-trending canyon of the ancient Grand Park River. Many flows that filled former canyons now stand up as ridges because the original streams, when displaced by lava, eroded new channels along the margins of the flow. Such reversals of former topography are common, but they are nowhere better displayed than at Sunset Park. Here the broad divide between the Mowich and Puyallup Rivers is carved from Mount Rainier lavas that filled the canyon of an ancient westward-flowing stream to a depth of at least 2,000 feet. The steep north wall of this former canyon, where the lavas are banked against older rock, is exposed along the divide between the North and South Mowich rivers. The south wall of the former canyon stood where the pres-

ent canyon of the North Puyallup River is now located. The old canyon, cut by the ancestral Mowich River, probably headed somewhere to the east of Ptarmigan Ridge.

The canyon of the ancestral Puyallup River, filled by at least 2,000 feet of lava, lay a few miles to the south. It may have headed beneath Mount Rainier's present summit or even much farther east. Remnants of its intracanyon lavas form Klapatche Ridge and St. Andrews Park. They abut against older rocks in the north wall of the old canyon at Klapatche Park and near Tokaloo Rock and in its south wall along the crest of Emerald Ridge.

The ancestral White River, heading on Mount Rainier's northeast flank, was also displaced by lava. A single intracanyon flow, 900 feet thick at Burroughs Mountain, clings high on the north wall of the present canyon, forming the flat surface at Yakima Park (fig. 46). The end of this flow is marked by subhorizontal columns of glassy lava exposed along the Yakima Park road below Sunrise Point.

Remnants of early intracanyon flows mark the courses of several other ancestral streams, but none received such great quantities of lava as the ancestral Mowich, Puyallup, and Grand Park canyons (fig. 47). The volcano evidently grew astride the headwaters of these streams, flooding their canyons with its early lavas and detrital material shed from its growing slopes. The other major stream valleys probably did not originally head within the area now covered by Mount Rainier, but they have extended their headwaters to the cone as it grew larger and began to develop large radial streams on its newly formed slopes.

NATURE OF THE EARLY ERUPTIONS

During its early growth the eruptions from Mount Rainier consisted chiefly of the lava that formed the early intracanyon flows. Not only were these flows large, but many were erupted in rapid succession; at Klapatche Ridge the flows that filled the canyon of the ancestral Puyallup River are stacked one above the other, each covering the undissected surface of its predecessor. It thus appears that in its youth the volcano grew rapidly, built up by large and frequent eruptions of andesite lava. Later, as the cone grew larger, intracanyon flows became smaller and less frequent; and the latest intracanyon flows, instead of covering one another, banked against new canyon walls carved in their predecessors, forming terraces with the latest flow at the bottom.

A few unconsolidated mudflow and glacial deposits are intercalated with the intracanyon lavas. Examples are found west of Spray Park and near Klapatche Park, where bouldery deposits of mudflow or of glacial origin lie both beneath and between the lava flows. Few of these deposits are thick or continuous, and all the larger ones lie beneath basal lava flows, as at Burroughs Mountain. Where the original topographic form of the unstratified bouldery deposits is no longer preserved, it is difficult to decide whether they are tills or mudflows; both contain poorly sorted boulders, cobbles, and pebbles of Mount Rainier lava in a silt- to clay-sized matrix. Striated or faceted boulders cannot here be taken as distinctive of till, for glacial debris on the volcano's slopes may have been incorporated in the mudflows.

Little pyroclastic material is associated with the early canyon-filling lavas; great volumes of lava poured forth, but explosive eruptions were small and infrequent.

BUILDING OF THE MAIN CONE

The main cone of Mount Rainier rises above the early intracanyon flows that form the base of the volcano. It was built up by hundreds of thin lava streams interbedded with breccias and subordinate layers of ash and pumice. These rocks are magnificently exposed in continuous 2,000 to 3,000-foot sections, at Willis Wall (fig. 48), on Little Tahoma Peak, and in the headwall of the South Tahoma Glacier (fig. 49). Sections that are less impressive, but more accessible, are to be seen at Gibraltar Rock and in many other wedges and cleavers around the mountain.

The flows are thin high on the mountain, where they drained rapidly down the steep slopes, but they thicken toward the broad lava apron at its base. This apron includes some flows more than 200 feet thick, such as the one at Comet Falls, but few of the flows above 9,000 feet are more than 50 feet thick.

FIGURE 46.—Remnant of an early intracanyon lava flow from Mount Rainier. The flow extends from the base of Mount Rainier (upper left) through Burrough Mountains (BM) and Yakima Park (YP). Its distal end is marked by a fan of semihorizontal columns, well exposed in the roadcut on the sharp bend below and to the left of Sunrise Point (SP). The flow, about 800 feet thick at Burroughs Mountain, partly filled the former canyon of White River. This river (WR), now reestablished along the south margin of the flow, is entrenched 1,300 feet (opposite Yakima Park) below its former level. A small remnant of Rainier lava also caps the west end of Goat Island Mountain (GM).

FIGURE 47.—Map showing area underlain by lava from Mount Rainier volcano.   Between 8,000 and 14,000 feet, contour lines at thousand-foot
intervals are indicated by short dashlines.

FIGURE 48.—Cirques at the head of the Carbon Glacier bite deeply into Mount Rainier volcano. Scores of thin flows and interlayered breccias are exposed in Willis Wall (left) 3,600 feet high and on Liberty Ridge, the sharp arete separating the two cirques. The stratigraphic section of lava flows and breccias exposed in Willis Wall is about 1,600 feet thick. Note the oversteepened ice walls, 200 to 500 feet high at the heads of both cirques. Avalanches and rockfalls constantly tumble from these cliffs, loading the heads of the glaciers.

FIGURE 49.—Upper part of Mount Rainier, looking east. The mountain is built of thin outward-dipping lava flows and layers of breccia, well displayed in the headwall of the South Tahoma Glacier (right foreground), and in Sunset Amphitheater (left). A thick layer of light-colored pumice (P) is conspicuous in the headwall of Sunset Amphitheater. On the east (right) margin of Sunset Amphitheater, just to the right of the pumice layer, a light-colored mass of unstratified rock (PL) can be seen. This is solfatarized andesite, probably the west edge of the volcano's central plug. Note that the lava flows in both Sunset Amphitheater and the South Tacoma headwall slant upward, probably toward a vent that lay 500 to 1,000 feet above the mountain's present summit. The upper ends of these lava flows are truncated by the rim of a large summit basin, outlined by Point Success (PS) and the ridge extending from Liberty Cap (LC) to the top of Russell Cliff (RC). This summit basin, breached by the Tahoma Glacier (TG), was probably formed either by collapse and engulfment or by erosional hollowing out of the top of the cone. The Recent summit cone, forming Columbia Crest (CC), surmounts the east rim and fills the east half of the summit basin. The remainder of the basin is filled with snow and ice. Photograph by H. Miller Cowling.

The apron of young lavas fingers out radially into intracanyon flows younger than those previously described. Some—those near Nickel Creek and at Ricksecker Point, for example—must still be fairly old, for streams have cut far below their base. Others, such as the small flow below the Winthrop Glacier, are clearly much younger, for they rest directly on the present canyon floors.

A spectacular example of a young intracanyon flow, easily viewed from the Stevens Canyon highway, is the sharp ridge of strikingly columnar lava separating Stevens and lower Maple Creeks. It advanced down Stevens Canyon to Maple Falls, diverting Maple Creek and shifting its junction with Stevens Creek nearly a mile downstream (fig. 50). This flow is found only on the south side of lower Stevens Creek, and fans of radiating horizontal columns at its distal end suggest that the flow was no wider at the time it solidified than it is now, though it fills only about a third of the width of Stevens Canyon. The entire north side of the flow, moreover, is a chaotic jumble of curved columns which flare out to a horizontal position against a vanished former cooling surface near the edge of the present lava cliff. These relations suggest that the flow was erupted when the lower part of Stevens Canyon was occupied by a glacier, and that below Sun-

beam Falls the flow followed a melt-water channel along the south edge of the glacier (fig. 50).

Remnants of a much older intracanyon flow, its base more than 500 feet above the present stream level, cling to the south wall of Stevens Canyon, forming the topographic feature called The Bench. Stevens Canyon was therefore invaded by at least two flows, separated by a period of canyon deepening. Throughout the building of Mount Rainier many streams were forced to cut new canyons through lava flows that recurrently flooded the upper parts of their valleys.

Throughout most of its history Mount Rainier erupted magma of uniform andesitic composition, but this magma solidified into rocks whose appearance varies greatly according to their manner of cooling. The thick slowly cooled flows consist mainly of porphyritic light-gray lava, but their bases, and also their sides where the flow has banked against a canyon wall, consist of black glassy lava. The basal part of a thick flow is generally columnar, and a few flows also have an upper tier of columns; but in most flows the upper part is platy, slabby, or hackly. Thin lava flows on the mountain's upper slopes, partly oxidized by escaping gases, are of various shades of red, pink, brown, and gray; some are compact, but most are vesicular. Many of these thin lavas are flow banded; black glassy streaks and layers alternate with reddish or pale-gray frothy layers. High on the mountain both massive and intensely shattered and agglutinated lava is common, but columnar structure is rare.

Interstratified with these lavas high on the mountain are breccias formed by explosive shattering of the lava flows. They contain fragments of porous or compact glassy lava, ranging from small chips to blocks as much as 10 feet across, in a matrix of sand- and silt-sized glass particles and crystals. The blocks in these breccias are not a varied assemblage of different kinds of lava, such as might be torn by explosions from the walls of a conduit, but are glassy and scoriaceous equivalents of the adjacent partly crystalline lava into which they grade. The blocks have rough slaggy surfaces, many of which are reddened by oxi-

FIGURE 50.—Remnants of two intracanyon flows from Mount Rainier in Stevens Canyon. Remnants of the earlier flow (Qra₁) form Mazama Ridge (MR) and The Bench (B). The present Stevens Creek (SC) is entrenched 500 feet below the base of this flow at The Bench. A much younger flow (Qra₂) rests on the present canyon floor, but it is found only on the south (left) side of Stevens Canyon. This flow, which descended through upper Stevens Canyon (USC) beyond the edge of the photograph on the right, may have followed a marginal melt-water channel along the south wall of a glacier that filled Stevens Canyon. This interpretation is also suggested by chaotic fans of curved columnar joints, perhaps formed by chilling of the lava against ice. These joints are developed all along the north side of the present remnant of the flow. The younger flow dammed Maple Creek (MC) and shifted its junction with Stevens Creek to the point (J), about a mile downstream.

dation, or they are covered by agglutinatelike crusts. The breccia commonly grades up through pods and lenses of slaggy, brecciated, and agglutinated lava into less shattered and reddened lava, and finally into massive lava forming a definite flow. Other breccias are on the tops of flows, and grade downward into them. Still other breccias do not visibly grade into flows, but they enclose small to large roughly concordant streaks, pods, and tongues of slaggy lava. (fig. 51).

The breccias that grade upward into lava flows were formed by steam explosions at the base of lava as it moved down slopes mantled with mud and melting snow and ice. Lava that burrowed beneath melting ice or mud, on the other hand, has breccia on its top. Where thin lava streams slid downhill mixing with large amounts of slushy snow and melt water, the entire flow was disrupted by steam explosions, and the resulting mixture of mud and blocky debris continued

FIGURE 51.—Mudflow breccias enclosing streaks, lenses, and tongues of partly brecciated lava. Such mixed accumulations of lava and breccia probably formed from thin lava streams that were shattered by steam explosions as they glided down steep slopes and mixed with mud, slushy snow, and melt water. The dips shown in the photograph are primary and are as much as 30°. The cliff is about 100 feet high. South wall of Little Tahoma Peak, about 400 feet below its summit.

down the slope in mudflows. In thicker flows irregular tongues and streaks of lava, still unshattered, have survived in the chaotic mixture (fig. 51). All gradations can thus be found on Mount Rainier between autobrecciated lava, hydrovolcanic breccia (Fisher, 1960, p. 997), and laharic breccia.

Still other breccias were probably formed from pyroclastic material blasted from vents during vulcanian explosions. Some of these pyroclastic materials accumulated where they fell, others avalanched into place as mudflows. It seems unlikely, however, that there are many breccias of this kind, for the lava flows on the lower slopes are rarely interlayered with vulcanian-type breccias, and spindle-shaped bombs are seldom found on the mountain. A few beds of pumice lapilli are interstratified with the lava flows, as at Gibralter Rock (fig. 52). Most of these pumice and ash layers are less than 5 feet thick, but high in the cliffs of Sunset Amphitheater is a conspicuous white pumiceous layer at least 75 feet thick (fig. 49). Pumice falls and other pyroclastics make up only an insignificant part of the cone.

Early writers, nevertheless, disagreed as to whether Mount Rainier is largely a pyroclastic cone or was built up chiefly by lava flows. I. C. Russell (1897, p. 360) recognized the volcano's composite character but overemphasized the abundance of ejectamenta, concluding that "The mountain * * * was built largely by the material thrown out by explosions from a summit crater." Matthes (1914, p. 7), also says that "Rainier has built up its cone * * * with cinders and bombs, and with occasional flows of liquid lava * * *" G. O. Smith (1897, p. 423), on the other hand, held that "the breccias, agglomerates, and tuffs, although of striking appearance, are, perhaps, less important elements in the construction of the composite cone." H. A. Coombs (1936, p. 174) "concurs with Smith in estimating the bulk of the mountain to be composed of flows." Our observations agree with Smith and Coombs as to the relative abundance of lavas and breccias, but we also hold that nearly all the breccias are shattered lava flows, not pyroclastics.

CENTRAL PLUG AND RADIAL DIKES

High on the east wall of Sunset Amphitheater stands a cliff of massive yellowish rock flanked on either side by thin-bedded flows and breccias—it appears to be part of the plug that filled the volcano's central conduit (fig. 49). The yellowish cliff itself, overhung by a 200-foot ice wall and constantly raked by avalanches, is inaccessible, but a short distance below it the Tahoma and Puyallup Glaciers carry yellowish float that could only have come from this source. This rock

was originally a porphyritic andesite; however, it is now thoroughly opalized and contains some kaolinite, tridymite, cristobalite, and also a little pyrite mostly altered to limonite. The strongly altered and mineralized condition of the yellowish rock suggest that it came from a plug; this suggestion is strengthened by its massive appearance, standing in contrast to the thin-layered flows and breccias on either side. It strikingly resembles the solfatarized vent rocks of the Brokeoff volcano, near Mount Lassen (Williams, 1932, p. 244–247, 263–265).

Radial dikes on Mount Rainier were first recognized by G. O. Smith (1897, p. 417), who described a prominent vertical dike cutting flows and breccias in the north face of Little Tahoma Peak. Thick dikes at St. Elmo Pass and along the northwest face of The Wedge were noted by H. A. Coombs (1936, p. 173). On the west side of the mountain, a huge dike forms Tokaloo Rock and the spine of Puyallup Cleaver (fig. 53); other dikes are visible in the high cliffs at Sunset Amphitheater and on the cleavers beside the Tahoma and Puyallup Glaciers. All these dikes are vertical, and radiate from the direction of Mount Rainier's summit. The dikes consist of pale-gray porphyritic andesite closely resembling that of the lava flows; probably they fed lava flows, but no direct connection has been found.

ERUPTIONS AT ECHO ROCK AND OBSERVATION ROCK

Echo Rock and Observation Rock, at the head of Spray Park, are dissected remnants of two satellite cones which erupted floods of olivine andesite after Mount Rainier was almost fully grown (fig. 54). Streams of lava from these vents flowed down Mount Rainier's northwestern flank, burying the low divide between the Carbon and North Mowich Rivers and flooding their headwaters. At Spray and Seattle Parks the flows rest on older Mount Rainier lavas, but they are readily distinguished by their large olivine phenocrysts.

When the eruptions from these vents began, the North Mowich River forked at Tillicum Point, and Ptarmigan Ridge—which is composed of much older Mount Rainier lava—formed a sharp divide between the forks. The northern valley, draining the present Spray Park area, was subsequently filled by thin flows of olivine andesite from the Observation Rock vent; three of these flows are banked against Ptarmigan Ridge at Tillicum Point. Spray Park is a glacially dissected structural surface built from these flows. Some of the lava flowed down the North Mowich canyon at least to its junction with the South Mowich, for a rem-

FIGURE 52.—Interlayered lava, breccia, tuff-breccia, and pumice near the top of Gibraltar Rock. In the lower part of the photograph a thick mass of andesite breccia, composed of shattered lava blocks and smaller fragments of slaggy and agglutinated lava in a matrix of crystals and glass particles, lies at the base (foreground). Above, on the right, is a wedge of dark tuff-breccia capped by a continuous layer of white pumice 3 feet thick. A lava flow, 20 feet thick, and with conspicuous platy jointing, lies above the pumice. The highest unit, composed of brecciated lava nearly 80 feet thick, slants upward away from the observer to the top of the cliff.

nant of olivine andesite clings to the north wall of North Mowich canyon opposite this junction. Olivine andesite flowing north and northeastward from the vents flooded the upper Carbon River basin and ponded against Old Desolate. Thick remnants of this ponded lava form the two flat benches projecting from the west side of Old Desolate (fig. 54); erosion has severed their connection with the flows surrounding Echo Rock.

Mount Rainier had grown to nearly its full height before the eruptions at Echo and Observation Rocks, yet the olivine andesite flows date back at least as far as the last major glaciation. Since their eruption the following events have occurred: (1) The Spray Park constructional surface has been deeply scoured by glaciers; (2) a large cirque has been cut in the once-continuous bench of ponded lava at Old Desolate; and (3) the upper Carbon and North Mowich Rivers have cut through the flows and deeply into the underlying rocks.

The Carbon River, before the olivine andesite eruptions, did not head on the slopes of Mount Rainier as it does today. Had it done so, the highly fluid olivine andesite could not have ponded to a depth of 500 feet against Old Desolate but would have drained down the Carbon River canyon (fig. 54). This canyon, however contains no remnants of these flows, nor, below Goat Island Rock, of the older pyroxene andesite flows from Mount Rainier. There must have been a barrier at Northern Crags, even before the early pyroxene andesite was erupted.

Sluiskin and Crescent Mountains may be remnants of a continuous ridge that formerly extended from Northern Crags to Mount Pleasant or Mother Mountain, separating the ancestral Carbon River on the north from the ancestral Mowich and Grand Park Rivers on the south. The voluminous early floods of pyroxene andesite flowing northward from Mount Rainier banked against this ridge, and then continued westward into the canyon of the Mowich and eastward into the canyon of the former Grand Park River.

Lava flowing north and northwestward may eventually have blocked the drainage to the Mowich River on the west, and thus have enabled new streams, flowing across the surface of these flows, to spill over the divide at Northern Crags and into the valley of Carbon River. Once the divide was breached, the Carbon River captured the entire drainage from Moraine and Seattle Parks, and from part of Mount Rain-

FIGURE 53.—Radial dike of Mount Rainier pyroxene andesite cutting volcanic breccia on Puyallup Cleaver. The dike is about 20 feet wide, and shows poorly developed columnar jointing perpendicular to its walls. The headwall of Sunset Amphitheater, with its conspicuous white pumice band, is visible in the background.

FIGURE 54.—Mount Rainier, satellitic volcanoes, and the Carbon Glacier. Scores of thin lava flows are exposed in Willis Wall (WW), the steep
headwall of the Carbon Glacier. Observation Rock (OR) and Echo Rock (ER) are dissected satellitic volcanoes, which erupted olivine
andesite upon Mount Rainier's northwestern flank late in its history. Olivine andesite (Qroa) from these vents covered earlier Mount
Rainier lava (Qra) at Seattle Park (SP) and ponded against a ridge of early Mount Rainier intracanyon lava at Old Desolate (OD). The
two flat benches projecting from the west (right) side of Old Desolate are remnants of this ponded lava. Ponding occurred because the
Carbon River had not yet opened the deep slot through the foreground ridge (Crescent Mountain and Northern Crags); neither the
olivine andesite nor earlier Mount Rainier lava drained down Carbon River through this gap. Before the olivine andesite eruptions,
streams from Mount Rainier's northern slopes flowed northwestward (to the right) past the present site of Seattle and Spray Parks to
the Mowich River. Photograph by the National Park Service.

ier's northern slopes. The point of capture is marked by a narrow gorge (fig. 54) and an abrupt steepening of the gradient of the Carbon River canyon between Goat Island Rock and the mouth of Cataract Creek.

DESTRUCTION OF THE SUMMIT

It has long been recognized that Mount Rainier was once considerably higher than it is now. The essential facts were given by I. C. Russell (1897, p. 360–361), who also proposed that this loss of height was probably due to an explosion:

The profiles of the mountain and the character of its summit show that at the time of its greatest perfection and beauty it rose as a tapering cone, with gently concave sides, to a height about 2,000 feet greater than its present elevation. At a later date it was truncated, probably by an explosion, which removed the upper 2,000 feet and left a summit crater 2 to 3 miles in diameter. Remnants of the rim of this immense crater now form Peak [Point] Success and Liberty Cap. Subsequently explosive eruptions partly filled the great crater and formed two smaller craters within it. The rims of the smaller craters are still clearly traceable, although at present the depressions they encircle are nearly filled with snow. A moderately prominent point between the two youngest craters, known as Crater Peak [Columbia Crest], is now the actual summit of the mountain.

Matthes (1914, p. 7) held the same opinion:

At one time the mountain attained an altitude of not less than 16,000 feet, if one may judge by the inclination of the lava and cinder layers visible in its flanks. Then a great explosion followed that destroyed the top part of the mountain, and reduced its height by some 2,000 feet. The volcano was left beheaded, and with a capacious hollow crater, surrounded by a jagged rim.

The outline of the supposed former crater is most clearly seen when viewed from the air (fig. 49). Point Success and the ridge extending from Liberty Cap to the top of Russell Cliff are remnants of its rim. It is deeply breached on the west by the Tahoma Glacier and on the northeast by the Winthrop Glacier. The young cone forming Columbia Crest later grew near its east rim, and now nearly fills the former depression, masking the breach in its northeastern wall. The youth of this cone is attested by its smooth unscarred slopes.

Russell and Matthes probably overestimated the mountain's former height. A projection, on carefully drawn profiles, of the average dip of the lava flows (25°) from the tops of Point Success, Liberty Cap, Russell Cliff, and Gibraltar Rock, indicates a maximum former height of 15,500 feet. If we accept H. A. Coombs' reasonable suggestion (1936, p. 203) that the volcano probably rose to a blunt summit with a crater, rather than to a sharp peak, its maximum height would not have been much greater than 15,000 feet.

Russell and Matthes surmised that a catastrophic explosion destroyed Mount Rainier's former summit, leaving a huge crater at 14,000 feet. But the reconnaissance nature of their work left little opportunity to search for supporting evidence, and neither of them gave any. Coombs (1936, p. 203) regarded their concept as oversimplified:

The writer hesitates to attribute the present configuration of the summit area to one huge explosion removing the upper 2,000 feet. The distribution of the pyroclastics and their alternation with lava flows on the upper reaches of the mountain suggest intermittent explosive activity, probably breaching first one side of the crater and then the other.

The volume of the rocks that formed the missing summit is at least 800 million cubic yards. If this great amount of material had been blasted outward by a single huge explosion, one would expect to find remnants of course blocky debris still preserved on protected upland surfaces around the base of the mountain, but none has been identified.

The possibility remains that the summit was fragmented by smaller explosions, as Coombs has suggested. We can speculate that the breach in the western wall of the summit depression, through which the Tahoma Glacier escapes, the gap in the eastern rim now covered by the cone at Columbia Crest, and perhaps even the huge hollow of Sunset Amphitheater, were blasted out by lateral explosions following the rise of plug domes into the crater, as described by Williams (1934, p. 236) for the destruction of Shastina's west wall. But accumulations of blocky debris beneath these supposed explosion sites have not been recognized, and the breaches in the volcano's summit can just as well be explained by glacial erosion aided by slides and rockfalls.

The two most likely alternatives to decapitation by explosions are the engulfment of the upper part of the cone by subsidence into underlying magma, or the rapid eroding out of an extensive solfatarized area at the summit. We have been unable to obtain proof of either of these alternatives, but evidence in favor of each can be briefly summarized.

Engulfment of the summit might have occurred as a result of rapid outpouring of magma from the vents of Echo and Observation Rocks. No definite time connection between the opening of these vents and the collapse of the summit has been established; both appear to have occurred before the end of the last Pleistocene glaciation. The young summit cone (Columbia Crest), astride the supposed ancient crater's east rim (fig. 49), may be likened to a rim volcano associated with cauldron subsidence (Richey, 1932; Anderson, 1941, p. 358–363; Williams, 1941, p. 295–296). Also, inward step faulting which offsets the

pumice band in Sunset Amphitheater might be evidence of collapse, through visible displacement is small.

Calderalike depressions in the volcanoes on Tahiti (Williams, 1933) and on Banks Peninsula, New Zealand (Speight, 1917), have been attributed to hollowing out of the summit areas by erosion. Williams (1941, p. 307–308) discusses erosion calderas in general:

> Immense, amphitheaterlike depressions may be developed in the cores of volcanic cones by ordinary processes of erosion. These processes are greatly favored where the original crater walls are composed of massive, strongly jointed lavas interbedded with layers of unconsolidated ash, for under such conditions the headward cutting of streams into beds of ash, together with copious subterranean drainage through them, rapidly undermines the walls and causes them to collapse. In this manner the valley heads assume a cirque-like form.

At Lassen Volcanic National Park, the collapse caldera of the Brokeoff volcano (Williams, 1932, p. 242–252) has been widened and deepened by the erosion of its soft solfatarized core.

If Mount Rainier's central plug was strongly altered, as suggested by the float from Sunset Amphitheater, it would be less resistant to erosion than the sheath of hard lavas. Once the protective sheath was breached, hollowing at the rotten core would rapidly follow. Headward cutting by glaciers might create a deep cirquelike depression. Or the core material, saturated with melt water and perhaps also softened by stream from below, might slump out in mudflows. The spaces that look like breaches in a crater rim may have been opened by slides and avalanches, followed by glaciers whose cirques ate rapidly headward into the rotton core.

RECENT ACTIVITY

Volcanic activity at Mount Rainier declined greatly after destruction of the summit. Recent activity consisted of (1) the building of the present summit cone and (2), perhaps simultaneously, the emission of explosive eruption clouds of sufficient size to blanket the surrounding area under thin falls of ash and pumice.

SUMMIT CONE

The young cone that forms the present summit of the mountain (fig. 49) grew at the eastern rim of the old summit depression. The top of the cone, once thought to be the highest point in the United States, was called Columbia Crest (Matthes, 1914, p. 8). The Columbia Crest cone rises 800 feet above the rim of the summit hollow. Its smooth slopes, almost unscarred by glacial erosion, stand in sharp contrast to the ragged remnants of the earlier main cone (fig. 49) and clearly show that it was built since the last Pleistocene glaciation. The cone is composed of blocky black glassy lava; some of this lava flowed westward, flooding the floor of the old summit depression, and some of it poured northeastward down the mountain's flank.

The top of the young cone is dimpled by two small craters (fig. 55), first described by Russell (1897, p. 361). The older crater is partly overlapped on its eastern side by the younger, which has a perfectly circular rim, 1,300 feet in diameter, canted slightly to the east. Both craters are filled with ice and snow, but steam jets issuing from the eastern crater have melted out caverns beneath the edges of the ice. During most of the year the rim of this crater is swept bare of snow by the howling gales that constantly batter the summit, but the lower slopes of the cone are deeply buried under snow and ice.

BLANKETS OF RECENT PUMICE AND ASH

The plateaus and ridge tops extending out from the base of Mount Rainier are mantled with thin sheets of Recent pumice and ash. Few ash falls are more than a foot thick, and most of them are measured in inches; but thicker deposits several feet in depth consist of pumice washed into stream channels, or of ash and volcanic dust blown by the wind into sheltered pockets on leeward slopes. Extensive reworking of the original pumice and ash falls by erosion makes it difficult to ascertain how many distinct ash falls there are or to correlate them from one ridge top to the next. It is possible, moreover, that at least some of the thinner sheets of fine ash were not erupted from Mount Rainier, but have drifted into this area from eruptions of Mount St. Helens and Glacier Peak, both of which are surrounded by extensive and voluminous ash blankets (Carithers, 1946; Rigg and Gould, 1957).

We have recognized two main ash-fall sequences which coarsen toward Mount Rainier, and therefore must have been erupted from it. Each sequence consists of two or more separate ash falls, erupted at closely spaced intervals. The two sequences, however, are separated by a considerable interval of time during which extensive erosion of the older ash occurred. Excellent exposures of the older sequence are found in Mist Park near the headwaters of Cataract Creek and in the headwaters of Marmot Creek. In Mist Park this ash blanket mantles the rounded surfaces of the moraines and thickens in the hollows. The ash is easily recognized by two distinctive features: (1) It is mostly altered to a greasy yellow clay containing abundant montmorillonite and (2) at its base there is a conspicuous brown pavement of iron oxides which stains the moraine (or other underlying material) and in many places cements loose materials at the basal contact into a hard crust from a fraction of an inch
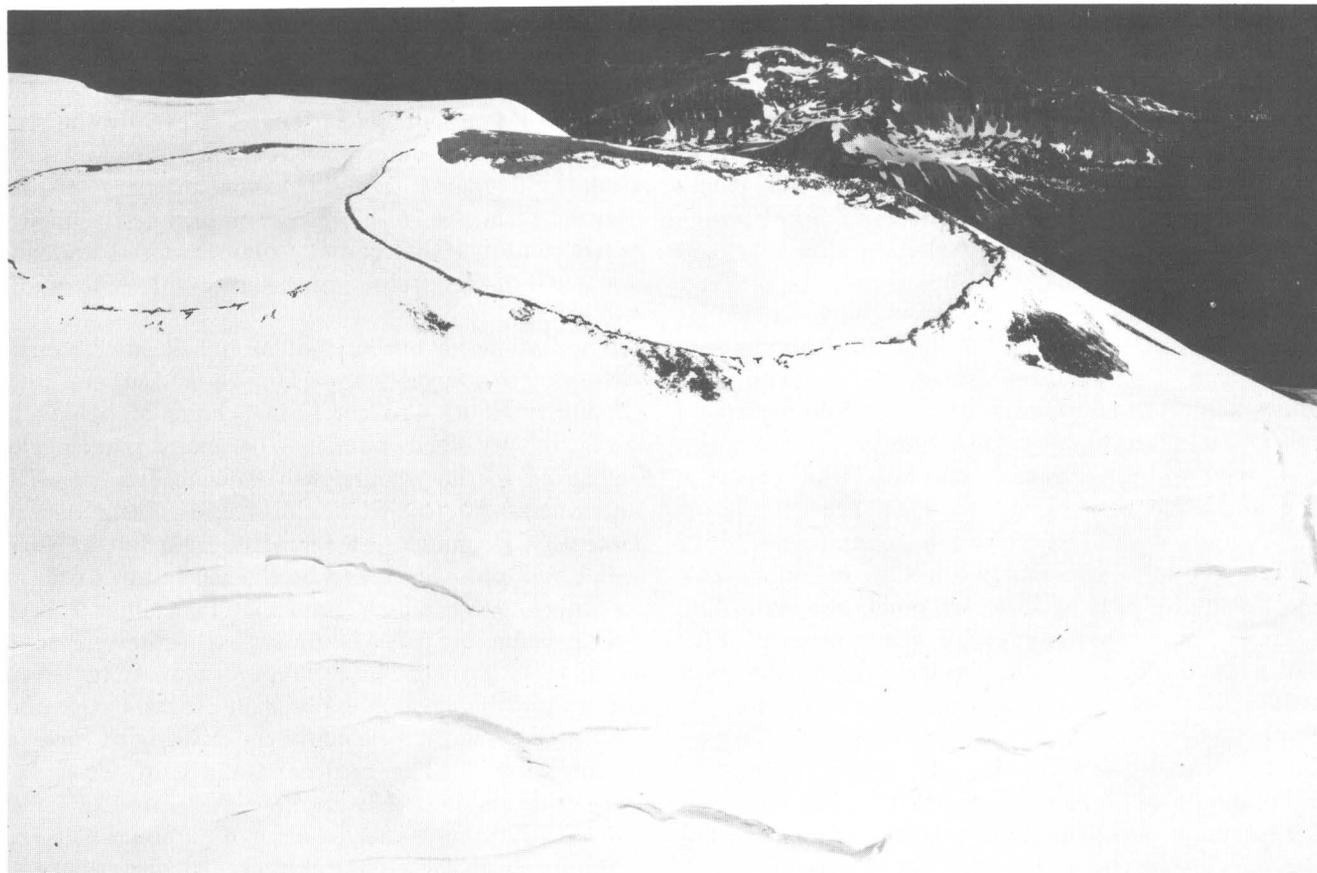
FIGURE 55.—Columbia Crest: twin craters indent the Recent cone at the summit of Mount Rainier. The east crater (right), 1,300 feet in diameter, has partly obliterated the older west crater. Both are perennially filled with snow and ice, but raging winds have cleared their rims, exposing slaggy black hypersthene andesite lava and cinders. Steam jets issue near the crater rims. Register Rock, the actual summit, is at the west (left) side of the east crater rim. Photograph by H. Miller Cowling.

to a little more than an inch in thickness. The ash must originally have contained iron sulfides and sulfur, which contributed to the strong leaching and oxidation.

Similar ash is also widely distributed around the northeast base of the mountain, where it has been described by Crandell and Waldron (1956, p. 359–360; pl. 1). Here, too, it is underlain by a well-developed pavement of brown iron oxide. Along the ridgetop due west of Glacier Basin, where the ash rests on a porous andesite flow from Mount Rainier, iron oxides that leached downward from it have cemented the upper surface of the andesite into a dense erosion-resistant brown crust. These iron-oxide crusts and the altered clay-rich character of the ash indicate long-continued decomposition of glass since the ash was deposited. Crandell and Waldron (1965, p. 360), on the other hand, believe that this material was altered to montmorillonite before it was ejected from the volcano.

On the uplands a few miles from Mount Rainier this older blanket of altered ash is less than 2 feet thick, but it thickens toward the volcano. In most outcrops

its structure suggests deposition during a single ash fall; but it is hard to be certain of this, for much of the material has been reworked, mixed with talus and surface wash, and otherwise so modified by erosion, deposition, and creep that its original structure has been almost obliterated. At a few localities, however, it is clearly underlain by another ash fall, composed of coarser pumice lapilli. Pumice and ash from the older sequence of ash falls has also been reworked into the valley fills. Terraces trenched by the headwaters of White River contain thick stream deposits and mudflows, some of which are composed mostly of pumice and ash. The glass in these deposits is now altered to a sticky clay.

Most of this older blanket of pumice and ash was swept away by erosion before the younger pumice sequence was laid down. The younger sequence is much the more widespread; it covers nearly all flat surfaces for 10 miles or more beyond the base of the volcano, and sediments reworked from it occur in all the valleys. This younger sequence, distinguished by its fresh glassy pumice, consists of three, and possibly more, succes-

sive ash falls, the upper two of which are widespread and give this ash sequence its distinctive character.

1. The older of the two ash falls is fresh pale-yellow to gray mostly sand-sized pumice, accumulated in a single sheet from 3 inches to as much as 2 feet in thickness. At most places it is covered by a few inches to a foot or more of forest duff, colluvial deposits, or slope wash; but on gently sloping meadows and other nonforested areas it is directly at the surface. A few rapidly aggrading streams have buried it deeply; for example, the 1947 flood of Kautz Creek excavated a deep channel whose walls reveal this pumice sheet buried beneath 18 to 55 feet of stream deposits and mudflows. This pumice is the most widespread of the different ash falls that comprise the young sequence.

2. The sheet of fresh yellow pumice described above is overlain, apparently without an appreciable time break, by a thin sheet of much coarser white to brown pumice which crunches underfoot. This pumice is perfectly fresh; broken fragments show hairlike filaments of completely unaltered glass. It occurs chiefly on the northeast side of Mount Rainier, but it has also been recognized on the south side. On the slopes of Crystal Mountain 12 miles northeast of the summit cone, the pumice lapilli are mostly less than one-half inch in diameter; at Yakima Park 7 miles from the summit they are as much as 2 inches; at the tops of Burroughs and Goat Island Mountains —3 to 4 miles from the summit—the deposit contains blocks of crunchy pumice as much as 6 inches in length. This pumice probably represents the last significant eruption from Mount Rainier.

Still younger than any of the pumice and ash layers described are thin sheets to mere films of powdery white ash composed entirely of tiny fresh glass shards. This ash is thickest where the wind has drifted it into sheltered hollows. Such fresh powdery ash is not confined to the vicinity of Mount Rainier; it is widespread throughout central and eastern Washington. It probably came from Mount St. Helens (Carithers, 1946).

AGE OF MOUNT RAINIER

We do not know when the growth of Mount Rainier began. Its first eruptions are intracanyon flows poured out upon a mountainous landscape carved mainly in the late Miocene or early Pliocene rocks of the Tatoosh pluton. This pluton had been unroofed over wide areas before Rainier volcanism began. On the other hand, so much erosion has taken place since the first intracanyon

flows from Mount Rainier were erupted that it does not seem likely that they are younger than early Pleistocene.

Mount Rainier reached its greatest height no later than the last major glaciation, as shown by the stripping of its thick outer shell, of which Little Tahoma Peak is a remnant, and by the presence of huge cirques. But the volcano has remained active until recently, building the cone at Columbia Crest and dusting much of the surrounding countryside with pumice and ash.

The last major outburst of Mount Rainier occurred about 500 to 600 years ago (Hopson and others, 1962). On upper Kautz Creek a 10-inch layer of pale-yellow sand- to pea-sized pumice—the main pumice sheet belonging to the younger ash sequence—is buried beneath about 55 feet of mudflow and stream deposits. Beneath the pumice are remnants of a buried forest, which was probably killed by the ash fall or by a pumice-slurry flood which immediately followed it. A stump belonging to this forest, still standing in the position of growth with its roots spreading beneath the pumice blanket, is exposed in Kautz Creek about three-quarters of a mile above the crossing of the Wonderland Trail. The radiocarbon age of wood from this stump is 350±250 years; [11] therefore the pumice above the stump cannot be older than about 600 years.

Sandwiched between mudflows 30 feet above the pumice are remnants of a second buried forest; increment borings by Nelson [12] in stumps from this horizon show that trees in this forest lived for at least 300 years before they, in turn, were overwhelmed by additional mudflows. Nelson has also shown that the age of the modern forest, now growing on the surface of the mudflows that buried the second forest, is about 250 years. The pumice sheet, buried beneath stream deposits that underlie both these forests, cannot be less than about 550 years old.

The field evidence thus indicates that the last major eruptions from Mount Rainier occurred about 550 to 600 years ago. Yet old newspaper accounts tell of eruption clouds rising from Mount Rainier on at least 14 different occasions between 1820 and 1894. No new ash or lava was observed at any of these times, however, and it seems likely that dust rising from large rockfalls and avalanches was mistaken for eruptions (Hopson and others, 1962). Whether the volcano is now dead or only dormant is a matter of speculation.

---

[11] Laboratory No. W–925, Field No. W60–6. Measurements made in the laboratory of the U.S. Geological Survey, under the direction of Meyer Rubin, Dec. 7, 1960.

[12] Nelson, J. W., 1958. Soil properties related to vegetation and time on the Kautz Creek flood area, Mount Rainier, Washington: Washington Univ. [Seattle], unpublished M.S. thesis, 56 p.

DISSECTION OF MOUNT RAINIER

The main body of Mount Rainier has been deeply dissected, and none of its original constructional surface remains. The young cone that culminates in Columbia Crest was formed only after nearly all this dissection had been accomplished.

GLACIAL EROSION

Evidence of vigorous glacial erosion has been recognized by all who have worked on the mountain, but none have portrayed it so vividly as Matthes (1914, p. 10):

so intense and so long-continued has been the eroding action of the ice that the cone is now deeply ice-scarred and furrowed. Most of its outer layers, in fact, appear already to have been stripped away. Here and there portions of them remain standing on the mountain's flanks in the form of sharp-crested crags and ridges, and from these one may roughly surmise the original dimensions of the cone. Mere details in the volcano's sculpture, these radial masses are, some of them, so tall that, were they standing among ordinary mountains, they would be reckoned as great peaks. Particularly noteworthy is Little Tahoma, a sharp, triangular tooth on the east flank, that rises to an elevation of 11,117 feet. In its steep, ice-carved walls, one may trace ascending volcanic strata aggregating 2,000 feet in thickness that point upward to their place of origin, the former summit of the mountain, which rose almost half a mile higher than the present top.

Impressive too, are the great cirques, whose steeply slanting headwalls truncate scores of flows. Willis Wall (fig. 48), Sunset Amphitheater, and the headwall above the South Tahoma Glacier (fig. 49) are the best examples. Profiles through the mountain show its present slopes to slant more steeply than the flows from which it is built; these flows now appear as outward sloping shingles. Because Mount Rainier is so deeply dissected, it is now far more rugged and precipitous than it was at any time during the past.

Previous workers, without exception, have ascribed the denudation of Mount Rainier to glacial action. Both Russell and Matthes described the mountain's present system of glaciers and cited impressive evidence of their eroding power. Russell (1897, p. 379–385) used Mount Rainier as a model to illustrate the progressive transformation, through glaciation, of an ideally smooth volcanic cone to a rugged "matterhorn." H. A. Coombs (1936, p. 204–210) described the glacial landforms on the mountain and in adjacent parts of the park. Today glacial action is rapidly cutting back and deepening the furrows in the volcano's flanks, and during the Pleistocene glacial stages it must have been even more effective.

ROCKFALLS, MUDFLOWS, AND SLURRY FLOODS

Other processes, such as avalanches, rockfalls, and landslides, have also contributed to the erosion of the mountain. Any approach to Willis Wall, Russell Cliff, Sunset Amphitheater, or the headwall of South Tahoma glacier is extremely hazardous because of the constant tumbling of avalanches and rockfalls from these oversteepened slopes. Brown dust clouds from cascading masses of ice and rock can be seen billowing up the face of the mountain on almost any warm sunny day. The climbing route along the Gibraltar ledge is changed or temporarily abandoned from time to time because of the falling away of parts of the ledge. Furthermore, every valley that heads on Mount Rainier contains complex fills composed of mudflows, coarse bouldery stream deposits, and debris from talus and larger rockfalls, glacial till, and outwash.

It is thus evident that although glaciers were the main factor in the denudation of the mountain, they have been aided by a wide variety of downslope movements. Some of these, including most of the rockfalls and avalanches, assist glacial erosion by replenishing the heads of the glaciers, speeding their flow, and providing them with cutting tools. But other tumbling and sliding masses of rock, snow and ice start huge mudflows, which rush down the sides of the mountain and follow stream valleys for miles beyond its base. Intermediate between mudflows and heavily loaded streams are the slurry floods of turbulent water carrying mud and sand, pebbles and boulders, chunks of ice, and uprooted trees that are generated where the crevassed and moraine-covered ends of glaciers and the frozen ground ahead of them break loose under the hydraulic pressure of impounded melt water or heavy rains and race down the valleys in devasting surges.

Just such a slurry flood broke loose from the end of the Kautz Glacier on October 2, 1947, and devastated the entire length of Kautz Creek valley. It has been vividly described by Grater (1948, p. 278):

The area affected by the flood stretched from the Kautz glacier, high on the slopes of Mount Rainier, for approximately 6½ miles down the Kautz valley to its junction with the Nisqually River * * * at least 50,000,000 cubic yards of solid material were excavated by the raging waters and were either spread out through the forests in the lower Kautz valley or poured into the Nisqually River. There the flood forced the river out of its channel, causing thousands of dollars worth of damage to houses and property farther downstream * * *.

The material being moved along by the flood was certainly not water; it was more the consistency of good cement * * *. Several feet in depth, it flowed sluggishly along, bringing with it large boulders, some of which measured approximately thirteen feet in diameter! Small wonder that few things were able to withstand its full force * * *. Here and there, large trees, hun-

dreds of years old and measuring up to 5 feet in diameter, were ground and hammered by tons of sand and huge boulders until they could resist no longer. Throughout their entire length they would begin to shake and weave crazily; then they would fall with a roar that was almost lost in the noise of the flood. Those that were able to withstand the onslaught were literally chewed and ground almost in half by the liquid sandpaper flowing past them * * *.

In the vast amphitheater once occupied by the glacier there was now only a deep gorge, approximately 300 feet deep and 1,000 feet wide at the widest point. The glacier itself had been destroyed for a distance of at least one mile and swept away by the floodwaters. Here and there remnants of the old ice field still perch precariously on the steep walls of the newly cut canyon of the Kautz.

The Kautz flood left a deposit which varies from stratified to unstratified, and which in places shows a crude upward grading in size of transported boulders. The first maximum surge of the flood also left a marginal rampart of boulders, in form like a narrow lateral moraine, that winds through the forest on either side of the devastated area for the first 4 miles from the source of the flood.

Every large valley heading on Mount Rainier shows abundant evidence that it has been swept by similar slurry floods and mudflows. Active volcanoes with thick accumulations of loose ash and blocky debris are especially favorable sites for mudflows. In such settings mudflows are generally of two kinds: (1) hot volcanic mudflows generated where lava flows, ash flows, or dense eruption clouds enter streams or slither down steep slopes mantled with snow, melt water, and mud; (2) cold mudflows developed by gravity sliding of water-soaked debris. Combinations are possible, as when a volcanic explosion triggers a slide or mudflow in water-saturated materials below, or when a minor explosion at the base of a plug or spine causes the rock to collapse and cascade down the mountainside.

Graphic accounts of hot mudflows caused by volcanic action have been published by Curtis (1903) and by Van Bemmelen (1949). As already noted, many volcanic mudflows, recorded in typical laharic breccias, were formed during the active building stage of Mount Rainier; they occur between and beneath the lava flows.

These hot mudflows, however, must be distinguished from the cold mudflows and slurry-flood deposits found in the alluvial fills of the larger valleys draining from Mount Rainier. Cold mudflows break loose from the slopes of a volcano in several different ways. Van Bemmelen's observations (1949, p. 191) of Indonesian volcanoes are pertinent:

The normal lahars (cold mudflows) are not specifically volcanic. They originate by heavy rainfall on slopes covered with loose material, or by earthquakes. Such secondary

rain-lahars of the Merapi may attain a length of 25–30 km, and one wet lahar of the Ruang was 40 km long.

Mount Rainier's slopes are not thickly mantled with pyroclastic debris like those of the Indonesian volcanoes, but their thinner coatings of loose material, chiefly disintegrating breccia and glacial debris, may have given rise to mudflows of this type. Others may have occurred where the weight of overlying lava caused slipping on layers of water-soaked ash or breccia between steeply dipping flows.

The importance of huge avalanches, rockfalls and landslides in carrying material from the mountain's upper slopes, and perhaps in starting mudflows and slurry floods, should not be discounted. According to Plummer,[13] a rockfall from Liberty Cap (the northwestern peak of Mount Rainier) during the earthquake of 1870 removed about 80 acres from along its southern edge. Van Padang (1939, p. 36) reports that at Raung volcano, Java, a large segment of cone slid away, leaving a narrow wedge-shaped scar bounded by steep escarpments, as if a gigantic slice of pie had been removed. The slide started a cold lahar that traveled 60 kilometers. Such slides started by avalanches, could have happened repeatedly on Mount Rainier, but their scars, enlarged and deepened by glaciers, would be difficult to recognize.

Finally, the hydrologic conditions on glacier-clad volcanoes, such as Mount Rainier, are especially conducive to the formation of mudflows and slurry floods. Valleys radiating from them are invariably choked with moraines and outwash, stream deposits, and mudflows, all with fairly steep initial slopes. At the valley heads these fills grade upward into huge aprons of talus, rockslides, and morainal debris. The valley deposits, being coarse and containing very little clay, are excellent aquifers. Large springs emerge from every valley fill that heads on Mount Rainier. The volcanic cone itself, composed as it is of highly jointed lavas and cavernous breccias, is also an ideal storehouse for underground water.

In a cold climate the surface layers of the valley-filling sediments are sealed during at least the winter months by impermeable frozen ground. Under the fluctuating climatic conditions that attended the Pleistocene epochs of glaciation, permafrost must have thickened and waned in the valley fills. But the fragile dams of frozen ground thus formed were unstable, and at times may not have been capable of withstanding the hydraulic pressure built up within the unconsolidated

[13] Plummer, F. G., 1893, Recent volcanic activity in Washington; a paper read before the Tacoma Academy of Science, Mar. 3, 1893: newspaper clipping in C. B. Bagley Scrapbook, Washington Univ. [Seattle] Library, v. 2, p. 163–165.

valley sediments upstream. Heavy rains or periods of excessive melting could also increase the head of water to the point where a weakened dam suddenly collapsed, enabling the hydraulically dilated sedimentary fill beneath its surface to surge out in a huge mudflow or slurry flood.

In recently glaciated areas of northwestern Europe and eastern Canada masses of unconsolidated glacial sediments, on valley floors with much lower initial slopes than those radiating from Mount Rainier, have suddenly broken loose during the spring thaw or after autumn rains. A spectacular example is the 1893 mudflow in the Vaerdalen, Norway, in which 111 people lost their lives and others survived only after a terrifying jostling ride in their staunch wooden houses (Reusch, 1901; Holmsen, 1952–53). The Vaerdalen mudflow consisted mostly of glacial sands that glided downstream on a base of marine clay whose upper layers were rendered mobile, apparently from displacement of the stabilizing salt water by fresh water (Holmsen, 1952–53; Rosenquist, 1952–53). At Mount Rainier slippery montmorillonite-bearing volcanic sediments interstratified in the thick valley fills might play a similar role.

OSCEOLA MUDFLOW

Crandell and Waldron (1956) believe that 4,800 years ago a large eruption from Mount Rainier initiated an enormous mudflow—they named it the Osceola mudflow—which ran 45 miles down the White River valley and spread out upon the Puget Sound lowland as a lobe 20 miles long and 3 to 10 miles wide. They believe that the deposit's clayey component, containing abundant montmorillonite, "was derived from hydrothermally altered older rocks prior to eruption rather than from subaerial alteration of volcanic ejecta during and subsequent to the eruption" (1956, p. 360). They suggest that the eruption involved "lateral ejections of clayey material from the northeast side of Mount Rainier" (p. 349), and estimated the minimum volume of the deposit, which they thought to represent a single short-lived mudflow, as 1.5 billion cubic yards (p. 357).

Several difficulties arise, however, in attributing such a huge mudflow to volcanic ejection of already montmorillonitized rock. Mud eruptions have been observed at other volcanoes (for example, Van Bemmelen, 1949, p. 198–199), but their volume is insignificant compared to that of the Osceola mudflow. They were produced by phreatic explosions, where magma erupted beneath sediment-filled lakes or large swamps. But Crandell and Waldron did not visualize the "Osceola eruption" as phreatic. If the deposit's clay component (which they estimated at 25 percent of the minimum

total volume) was explosively ejected from a vent area of montmorillonitized rock, it would empty a "clayey pocket" half a mile wide and a third of a mile deep, or of comparable volume. No such crater-like hole has been identified on Mount Rainier; specimens, moreover, of the hydrothermally altered core rock from Sunset Amphitheater do not have a large clay fraction, and the clay does not contain detectable montmorillonite. When the clay-rich fraction was X-rayed on a Phillips diffractometer, no hint of the basal montmorillonite peak (normally a strong reflection) was found, and this reflection could not have been masked by the abundant cristobalite in the sample, as other important montmorillonite peaks might have been.

We consider it more likely that most of the montmorillonite in the Osceola mudflow came from altered water-laid pumiceous sediments and pumice-slurry flood deposits interstratified in the valley fills along the headwaters of White River. Clay-rich remnants of stratified pumiceous deposits more than 100 feet thick still cling to the walls of Inter Fork valley. Pumiceous debris must have choked the headwaters of these streams at the time when pyroclastic eruptions formed the older of the two widespread ash-fall sequences described on page 80. The pumiceous sediments as noted above show clayey alteration comparable to that in the older ash blanket. Remnants of the air-deposited part of the ash blanket, with its characteristic iron-stained basal crust, are plentiful, moreover, on the high ground near the head of the Osceola mudflow. After the ash was altered, creep of this slippery clayey material downslope would have contributed additional montmorillonite to the valley fills.

Crandell and Waldron (1956, p. 355, 360; fig. 1) recognized the blanket of montmorillonite-rich material on flat summits high above the head of the Osceola mudflow, but they refer to it as the "air-laid facies" of the Osceola mudflow and believe that it was formed by the fall of particles of already montmorillonitized rock during the eruption that supposedly generated the Osceola mudflow in the valley below.

Brown iron-stained crusts, presumably formed by thorough weathering, underlie the "air-laid facies," and fragments of iron oxide crust have been incorporated in the mudflow. These fragments afford evidence of a period of long-continued oxidation and leaching of the ash following its deposition but before the mudflow occurred.

The composition of the boulder and pebble fraction of the Osceola mudflow, moreover, is not in keeping with its suggested origin from a volcaic eruption. Volcanic mudflows tend to be made up largely, if not

entirely, of material derived from the volcano itself. Counts of Osceola pebbles show about 35 percent of Mount Rainier lava and 65 percent of older central-Cascade rocks (Crandell and Waldron, 1956, p. 352; and one count made by us). This pebble ratio accords much better with a mudflow coming from a mass of stream and glacial deposits, derived mainly from terrain surrounding Mount Rainier, than with one that came almost entirely from the volcano.

We suggest, instead, that the Osceola mudflow was formed by the collapse and flowing out of a thick fill of water-saturated sediments in the upper part of the White River valley, in much the same way as the Kautz flood of 1947 at Mount Rainier or the mudflow that swept Vaerdalen, Norway, in 1893. This interpretation is supported not only by the composition of the materials composing the mudflow, but also by the presence of remnants of a thick stratified fill that forms terraces in the upper part of the White River drainage area. Crandell and Waldron (1956, p. 357, pls. 2, 3) refer to a part of this fill—a deposit 250 feet thick that forms a terrace on the south wall of the Inter Fork valley—as a remnant of the Osceola mudflow. But this deposit is not an Osceola remnant; its lower part consists chiefly of well-stratified sands and gravels, and its upper part of water-laid pumiceous sediments and pumiceous mudflows now partly altered to a sticky clay. These stratified materials are exposed in a high cut bank whose face is almost completely plastered over with large boulders and slope wash that conceal the stratified materials, except in deep gullies and on exceptionally steep slopes (Crandell and Waldron, 1956, pl. 3, p. 358).

Nearly all the terrace fill, moreover, consists of pre-Osceola stratified deposits, which must formerly have extended completely across the valley and for many miles downstream. The large volume of the Osceola mudflow, and its high percentage of non-Rainier rocks, could be accounted for by supposing that this old valley fill collapsed and spurted downstream, just as the fill on upper Kautz Creek did in 1947. This collapse could have been triggered by a volcanic eruption, or it might have been entirely independent of volcanic action. Volcanic action might also have caused the collapse indirectly, for the intermittent growth of the late Columbia Crest cone must have caused the size and regimen of the glaciers on and near the summit of Mount Rainier to vary greatly from time to time. Rapid melting of the summit glaciers by volcanic heat could have sent avalanches and large volumes of melt water cascading into the headwaters of White River.

The presence of montmorillonite-rich pumiceous sediments in the pre-Osceola fill raises other possibilities as to the cause of collapse: these slippery sediments are unstable, and if the upper end of the valley fill containing them received additional load—as, for example, by an advance of Emmons Glacier, a landslide or a rockfall, or the cascading debris from a summit eruption—they might give way. Once sliding started, failure of the thin dams of clay-rich tuff might suddenly release water-logged sediments under strong hydraulic pressure from beneath these dams, rapidly spreading the area of collapse and increasing the mobility of the mudflow.

Crandell and Waldron have made a significant contribution in recognizing the Osceola deposit as a mudflow—it had earlier been described as a till—and in pointing out distinctive features by which mudflows can be recognized. We disagree only with their view that this particular mudflow is the direct result of a volcanic eruption of already montmorillonitized rock from Mount Rainier.

SUMMARY

In summary, the rapid erosion of Mount Rainier is now being accomplished largely by glaciers, but the work of large and small rockfalls, avalanches, slurry floods, and various kinds of mudflows should not be underestimated. By restoring Mount Rainier's former constructional surface on profiles drawn to conform to the original slopes of its lavas, it is calculated that the volume of the cone was formerly at least half again as great as it is now. The glacial deposits and mudflows lining the upper valleys leading from Mount Rainier, voluminous as they are, do not appear fully commensurate with the huge amount of debris that must have been shed from the cone. Obviously, much of the material removed by glaciers and downslope movements has been carried away by streams; the flood plains along the lower courses of the Cowlitz, Puyallup, White, and other rivers that head on Mount Rainier contain some of this material, but other parts have been carried on out to sea.

PETROGRAPHY AND MINERALOGY

Mount Rainier erupted lavas of remarkably uniform composition throughout most of its history—a fact first recognized by Hague and Iddings (1883). Nearly all the lavas are pyroxene andesite; olivine andesite was erupted from satellitic volcanoes at Echo Rock and Observation Rock. None of the andesites contain visible alkali feldspar or quartz, but they do contain much glass from which these minerals might have crystallized with slower cooling. Chemically (table 2) the pyroxene andesites are nearly as silicic as Nockold's average dacite (1954), and are even slightly more potassic; the olivine andesite is probably slightly more

mafic than the older lavas from Mount Rainier (no chemical analyses of it are available).

TABLE 2.—*Chemical analyses and molecular norms of Mount Rainier lavas and rocks of similar composition*

	Mount Rainier andesites					Dacite	Anatectic quartz diorite	
	1	2	3	4	5	6	7	8
SiO₂	60.53	63.50	62.08	63.57	61.50	63.58	61.8	57.5
Al₂O₃	17.01	16.91	17.25	17.06	17.14	16.67	17.0	18.0
Fe₂O₃	2.12	1.12	1.80	1.17	1.55	2.24	2.1	2.7
FeO	3.58	3.61	3.32	3.32	3.85	3.00	3.5	3.9
MgO	3.47	2.28	2.91	2.78	2.96	2.12	2.4	3.6
CaO	5.79	5.14	5.43	5.11	5.80	5.53	5.2	6.9
Na₂O	4.01	4.14	4.18	4.21	4.11	3.98	3.8	3.8
K₂O	1.58	1.84	1.84	1.72	1.65	1.40	1.7	1.0
H₂O⁺	.25	.09	.16	.09	.20	.56	} 1.2	1.0
H₂O⁻	.12	.05	.09	.09	.04			1.0
TiO₂	.95	.81	.20	.18	.91	.64	.71	.98
P₂O₅	.26	.24	.85	.73	.25	.17	.33	.32
MnO							.11	.12
MnO₂	.10	.09	.08	.08	.10			
CO₂							.05	.07
	99.77	99.82	100.19	100.11	100.06	100.00	99.9	99.9

Molecular norms

	1	2	3	4	5	6	7	8
Q	13.1	17.5	15.3	17.2	14.3	20.0	17.4	10.9
Or	9.5	9.8	11.1	10.0	9.6	8.4	10.0	6.0
Ab	34.2	34.5	35.1	35.8	34.8	33.6	34.5	34.5
An	23.7	23.2	21.4	21.2	23.8	23.5	23.0	29.5
C			.5	.5			.8	
Σ salic	80.5	85.0	83.4	84.7	82.5	85.5	85.7	80.9
Wo								0.8
Di	3.1	0.9			2.7	2.7		
En	7.9	5.4	7.3	6.9	6.6	4.6	6.8	10.2
Fs	3.0	3.5	4.2	4.9	4.0	2.4	3.2	2.8
Mt	3.0	3.0	2.6	1.6	2.3	3.3	2.3	2.9
Il	1.8	1.5	.5	.3	1.7	1.2	1.0	1.4
Ap	.7	.7	2.0	1.6	.2	.3	.8	.8
Cc							.2	.2
Σ femic	19.5	15.0	16.6	15.3	17.5	14.5	14.3	19.1
Normative plagioclase	An₄₁	An₄₀	An₃₈	An₃₇	An₄₁	An₄₁	An₄₀	An₄₇

1. Andesite from Burroughs Mountain. (Coombs, H. A., 1939, p. 1506).
2. Andesite from the fissure [radial dike] at St. Elmo Pass. (Coombs, H. A., 1939, p. 1506.)
3. Andesite from a locality near the terminus of the Nisqually Glacier. (Coombs, H. A., 1939, p. 1506.)
4. Andesite from Paradise. (Coombs, H. A., 1939, p. 1506.)
5. Andesite from the summit of Mount Rainier, at Register Rock. (Coombs, H. A. 1939, p. 1506.)
6. Nockolds' average dacite+dacite obsidian; average of 50 analyses (Nockolds, 1954, p. 1015.)
7. Hornblende-biotite quartz diorite, east side of Rock Creek, Holden quadrangle, Washington. Composite sample of 8 specimens. (Crowder, 1959, p. 866-869.)
8. Biotite-hornblende quartz diorite, south side of Hart Lake to Mount Maude, Holden quadrangle, Washington. Composite of 7 samples. (Crowder, 1959, p. 866-869.)

The Mount Rainier pyroxene andesites are strongly porphyritic lavas (figs. 56, 57) containing conspicuous plagioclase phenocrysts and smaller pyroxene phenocrysts which tend to occur in glomeroporphyritic clots. The light-gray lavas from the thicker flows at the base of the mountain are pilotaxitic (fig. 56), whereas the dark lavas from thin flows and breccias high on the mountain and the jet-black lavas from the basal parts of the intracanyon flows are chiefly hyalopilitic (fig. 57). The glass from these lavas is mostly fresh, but in some, especially the badly shattered and reddened lavas of the upper cone, it is partly devitrified.

The mineralogy of these andesites is so uniform that a general description will apply to all. In the following paragraphs the principal minerals are described, in their usual order of decreasing abundance.

As the plagioclase in the Mount Rainier lavas has been described in detail by H. A. Coombs (1936, p. 175–180), only its most salient features are noted here. The phenocrysts, many of which are strongly zoned and complexly twinned, range from less than 0.1 to 3.0 mm in diameter; their average diameter in a given flow may be anywhere from 0.2 to 2.0 mm. Many of the larger phenocrysts are glomerocrysts, formed where several smaller crystals clotted together in rudely parallel orientation and continued to grow as one. Most of the phenocrysts are progressively zoned, ranging from An₄₅₋₆₀ (rarely as much as An₇₀) at the core to An₃₀₋₄₅ at the rim. Narrow oscillatory zones are generally superimposed on the broad progressive zones. The microlites are andesine, generally homogeneous and of about the same composition as the rims of the phenocrysts in the same rock; on the average about An₃₅₋₄₀.

A few flows contain crystals with marked reverse zoning at the margin; in these crystals the anorthite content in a narrow outer rim exceeds that in the main body of the crystal by about 5 to 10 percent.

There is often a striking diversity in the character of the plagioclase phenocrysts from the same rock. For instance, within a single thin section, some of the phenocrysts may be strongly zoned but others virtually unzoned; the zoning may be progressive in some phenocrysts but reversed in a few others; some may contain resorbed inner zones whereas others do not; some may have abundant inclusions of glass or magnetite whereas others are clear. The histories of different crystals from the same rock has evidently differed greatly, which may reflect considerable turbulence and metastable crystallization in the magma chamber.

Thomas L. Wright has determined the structural state of a number of plagioclases from several different Mount Rainier flows, using the optical methods described on page 6. All the crystals studied, including the cores of the larger phenocrysts, have "high temperature" disordered structures.

Hypersthene is present in all Mount Rainier lavas examined, and is the dominant dark mineral in most. It forms prismatic phenocrysts as much as 2 mm long (average 0.2 to 1 mm), and small granules in the groundmass. Glomeroporphyritic clots of hypersthene alone, or hypersthene and augite, are common. In some flows the hypersthene was unstable during the later stages of crystallization: some of the hypersthene

FIGURE 56.—Photomicrograph of porphyritic hypersthene-augite andesite from Mount Rainier. Pheno-crysts of plagioclase (P) and hypersthene (H) occur in a pilotaxitic groundmass that consists chiefly of plagioclase and clinopyroxene. The large hypersthene phenocryst (H) is enclosed by a narrow rim of augite (A). Basal intracanyon flow, near the head of South Mowich River.

phenocrysts are corroded, and others are partly replaced by augite; in such rocks hypersthene is absent from the groundmass. More commonly, the hypersthene has sharply defined rims of augite, but is not replaced by it (fig. 56). Some hypersthene andesites from Mount Adams and Mount Hood show thin rims of pigeonite enclosing the hypersthene phenocrysts, but this was not observed in the Mount Rainier lavas. In many of the bright-red highly oxidized rocks, hypersthene phenocrysts are thinly coated with exsolved magnetite.

The composition of hyperstenes from eight different flows, determined by measuring their $2V$ (Kuno, 1954, p. 40), is mainly in the range from about En_{60} to En_{75}. The phenocrysts in most of the flows are progressively zoned, the ratio Mg : Fe being 5 to 10 percent greater in their cores that at their rims or in the tiny hyperstenes from the groundmass. But this is not always true: in a specimen from the exceptionally thick flow at Tillicum Point, for example, the phenocrysts are about En_{65-70} and the small groundmass crystals En_{84}.

Clinohypersthene, first noted in Mount Rainier lava by H. A. Coombs (1936, p. 184), is abundant in nearly every thin section. $Z \wedge c$ ranges from $3°$ to $16°$. The extinction angle of some phenocrysts increases markedly, but always gradually, from core to rim.

Augite is slightly subordinate in abundance to hypersthene in most of the flows. It forms stout phenocrysts as much as 2 mm long (average 1 mm), and is generally also present in the groundmass. Augite forms rims around hypersthene, but the reverse was not observed.

Most of the augite is a normal Ca-rich variety, with $2V = 54°$ to $56°$, but some flows also contain a Ca-poor augite, in which the $2V$ may be as low as $35°$. Here the normal augite crystallized last, forming rims around the Ca-poor augites and occurring as small grains in the groundmass. No pigeonite was observed.

More than half the andesite flows contain olivine, but so sparsely is it distributed that one must sometimes examine several thin sections from the same rock before seeing a grain of it. Most of the grains of olivine are partly resorbed, or they have reaction
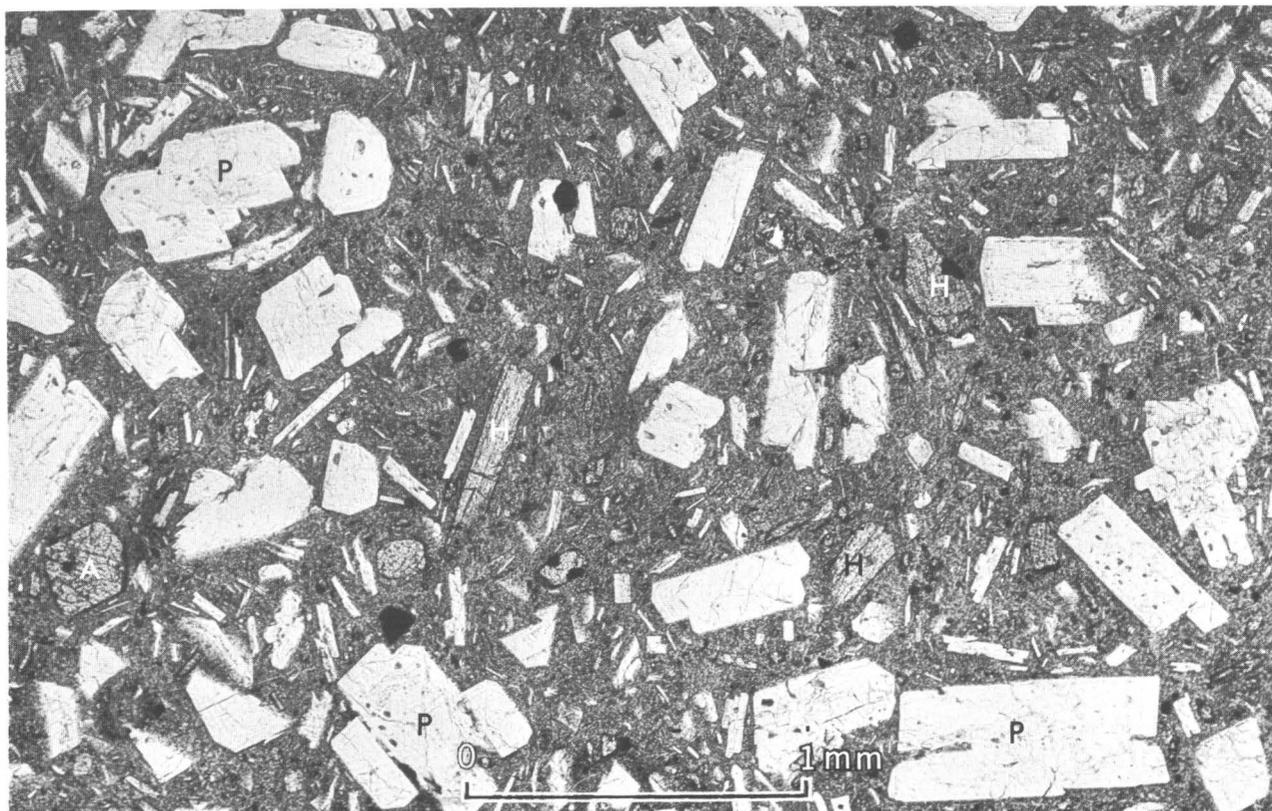
FIGURE 57.—Photomicrograph of porphyritic augite-hypersthene andesite from Mount Rainier. Phenocrysts of plagioclase (P) and smaller prismatic phenocrysts of hypersthene (H) and augite (A) occur in a hyalopilitic groundmass of dark glass and tiny microlites. Base of intracanyon flow in Stevens Canyon, at Sunbeam Falls.

rims of hypersthene or augite. These relations, together with the fact that olivine occurs only in the glassy quickly chilled parts of many flows, shows that this mineral is a common early crystallizing phase in the Mount Rainier magma, but that it was later destroyed by reaction.

Most of the olivine is in the composition range from about Fo_{75} to Fo_{85}, as determined by measuring its $2V$ (Poldervaart, 1950, p. 1073). Some grains are progressively zoned, but the fayalite content rarely increases toward the rim by more than a few percent. Many show no zoning at all. The olivine grains probably ceased to grow before the magma's ratio of Mg : Fe had changed much.

The olivine is generally fresh, but in some flows it is partly altered to an olive-green montmorillonoid, possibly saponite. The alteration occurs mostly along grain margins and cleavage cracks, but occasionally entire crystals are replaced. The alteration is deuteric; the other minerals and glass are fresh. Alteration to iddingsite, common in other Cascade lavas was not found at Mount Rainier.

Brown hornblende is a sparse constituent of a few flows from Mount Rainier. It evidently was unstable during the late stages of crystallization, for all of it is thickly rimmed or completely replaced by dusty aggregates of magnetite or hematite. The hornblende forms phenocrysts that partly enclose hypersthene and augite, but it never occurs as microlites; this suggests that it crystallized as a late intratelluric mineral. The flows that contain hornblende are probably no more silicic than the other pyroxene andesites. Nor is hornblende restricted to flows of any particular age; it is sparsely developed in a few of the oldest lavas from Mount Rainier, as well as in some of the youngest.

The hornblende probably crystallized in batches of magma with higher-than-normal water vapor pressure (P_{H_2O}). Boyd (1939, p. 387) shows experimentally that in chemical mixtures of appropriate composition pyroxene crystallizes when P_{H_2O} is low, whereas hornblende crystallizes when the P_{H_2O} is higher. A building up of high vapor pressure in the magma just before eruption would explain the late-intratelluric formation of hornblende phenocrysts; the release of vapor pressure after eruption would explain the absence of hornblende microlites.

Magnetite is abundant in nearly all the lavas of Mount Rainier, but in some it has been oxidized to hematite. Magnetite crystallized early (it occurs as inclusions near the centers of the large plagioclase phenocrysts) and also late (it replaces hornblende). Some octahedrons of magnetite are strongly resorbed.

Apatite, in tiny prisms, is a constant minor accessory. Tiny flakes of pale-brown biotite occur sparsely in a few flows. These flakes are sometimes associated with finely granular pyroxene in rims around olivine.

Cristobalite, or less commonly tridymite, lines cavities in some of the more porous lavas found high on the slopes of the mountain.

Small xenoliths of Tatoosh granodiorite are found in some flows, being particularly abundant in the basal flow along the ridge crest northeast of St. Andrews Park. In most places, however, clearly recognizable xenoliths of basement rock are lacking.

Small inclusions of micronorite, distinguished from xenoliths of earlier lava by their coarse hypidiomorphic texture, are abundant in most of the Mount Rainier flows, but their origin is obscure. Their chief minerals are plagioclase (An_{55-54}), hypersthene, magnetite, and in some, augite; their bulk composition is more mafic than that of the enclosing glassy andesite. Their hypidiomorphic texture suggests a plutonic origin, but some features of the minerals themselves make this unlikely: the plagioclase, zoned like the phenocrysts in the lava, is of the high-temperature variety; the hypersthene also is high temperature (monoclinic), and lacks the clinopyroxene exsolution lamellae common in gabbroic hypersthenes. These inclusions are therefore tentatively regarded as large glomeroporphyritic clots.

The olivine andesites erupted from satellite cones at Echo Rock and Observation Rock are pilotaxitic lavas that differ chiefly from the pyroxene andesites in having abundant large euhedral phenocrysts of olivine, some of which are as much as 4 mm in diameter. Augite, hypersthene, and strongly zoned plagioclase (ranging from labradorite in the core to sodic andesine in the rims) form smaller less conspicuous phenocrysts. The groundmass consists chiefly of tiny grains of clinopyroxene, andesine, and magnetite.

MAGMA OF MOUNT RAINIER

As already noted, the pyroxene andesites from Mount Rainier are comparatively uniform in chemical composition; the youngest lava, from Register Rock on Columbia Crest (table 2, No. 5) differs little in composition from the old intracanyon flow at Burroughs Mountain and Yakima Park (table 2, No. 1). The composition of Mount Rainier's magma thus did not change appreciably until late in the volcano's history, when slightly more mafic olivine andesite erupted at Echo Rock and Observation Rock. Nevertheless, the mountain's youngest eruptions, at Columbia Crest, reverted to the normal augite-hypersthene andesite.

Mount Rainier has often been grouped with other large andesitic volcanoes of the Cascade Mountains in generalizations that emphasize the similarities of their lavas. It is true that the andesites of Mount Rainier are similar in composition to the well-known siliceous pyroxene andesites from Mount Shasta (Williams, 1934), Mount Mazama (Williams, 1942), and Mount Jefferson (Thayer, 1937), but nearly all the large volcanoes south of Mount Rainier had rather complex magmatic histories—they erupted lavas ranging as widely as from olivine basalt to dacite and rhyolite, whereas the lavas of Mount Rainier show little variety. Only Mount Baker, north of Mount Rainier, is comparable.

Some inferences can be made about processes that affected the Mount Rainier magma as it rose from its place of origin to the surface. The relevant facts are as follows: 1. A few xenoliths of Tatoosh granodiorite and abundant tiny bits of micronorite (which are probably glomeroporphyritic clots) are the only inclusions in the lava from Mount Rainier, and they show no effects of assimilation. 2. The eruptions at Mount Rainier were not notably explosive, as evidenced by the scarcity of pyroclastics. 3. The lavas did not flow to great distances, and many of the flows are exceptionally thick. 4. Even the very thick flows show no evidence of crystal settling. Points 3 and 4 suggest a somewhat viscous magma, perhaps poor in volatiles. It appears that the Mount Rainier magma rose through a relatively open conduit, not mixing extensively with older rocks, nor concentrating volatiles in a chamber near the surface. The magma at depth showed little tendency to differentiate, even over a considerable span of time.

The origin of the Mount Rainier magma is a matter of speculation. The widely held opinion that andesitic magma evolved from parental basalt is supported, in the southern Cascades, by considerable field and chemical evidence. The close field association of andesite and basalt in this region is well known (Williams, 1932, 1934, 1942, 1944; Thayer, 1937; Verhoogen, 1937; Anderson, 1941, Waters, 1955b): from Mount St. Helens southward these lavas were erupted side by side in many places during the Quaternary; also, many of the andesitic volcanoes were built on a broad foundation of Pliocene and Quaternary olivine basalt. The andesite is linked with basalt, moreover, through intermediate olivine andesite

and basaltic andesite, which is widespread and locally voluminous. Chemically, andesites from the Cascades generally plot on straight-line variation diagrams for cogenetic groups of rocks ranging from basalt to rhyolite.

But Mount Rainier, Glacier Peak, and Mount Baker, the northernmost of the large Quaternary volcanoes, are striking exceptions. They are not floored by basalt, nor did basalt erupt from nearby vents as the andesitic cones grew. Quaternary olivine basalt did indeed issue from numerous vents along the Cascade crest in Washington, chiefly south of Mount Rainier, but these vents are not close to the andesitic volcanoes. The one nearest to Mount Rainier is Tumac Mountain, 20 miles to the southeast.

An alternative hypothesis, which is more consistent with our observations at Mount Rainier but for which we have no proof, is that the andesitic magma formed by fusion of crustal rocks. Crowder (1959) has given a well-documented example of partial fusion of metamorphic basement rocks in the northern Cascades during the Mesozoic. Gneisses and schists derived from geosynclinal sediments and volcanic rocks were converted by anatexis to neomagma, which then became intrusive and ultimately solidified as quartz diorite. The chemical composition of this quartz diorite, listed in table 2, columns 7 and 8, closely approximates that of typical Mount Rainier andesite (table 2); if the neomagma, completely liquified, had risen to the surface, it very likely would have formed pyroxene andesite similar to that at Mount Rainier.

SUMMARY

Mount Rainier volcano, composed dominantly of pyroxene andesite, grew on rugged terrain with at least 4,000 feet of relief, probably mainly in early Pleistocene time. The first clear records of its activity are large intracanyon flows of pyroxene andesite, which blocked and in places obliterated the ancestral valleys of the Mowich, Puyallup, and Grand Park Rivers.

The main cone is built from thin lava flows, erupted from a central vent, and from breccias that were formed by violent stream explosions where the lava became mixed with mud and melt water on slopes covered with snow or ice. Most of the fragmental debris thus formed avalanched or flowed downslope as hot mudflows. Pumice and ash eruptions were infrequent and mostly small; throughout the volcano's history pyroclastic eruptions were greatly subordinate to outpouring of lava.

The volcano reached its maximum height, about 15,000 or 15,500 feet, before the late Pleistocene. At about this time two satellitic volcanoes erupted oli-

vine andesite from its northwest flank. It was then deeply eroded by glaciers, avalanches, streams, and mudflows before the latest upwelling of lava completed the building of the volcano by forming the present almost undissected summit cone. Recent thin pumice deposits, the last prominent one 500 to 600 years old, mantle the surrounding country. Young stream and mudflow deposits floor the upper valleys draining the mountain.

REFERENCES CITED

Although not readily available for reference, three theses are listed with published literature because they are cited repeatedly in text and on maps.

Abbott, A .T., 1953, The geology of the northwest portion of the Mt. Aix quadrangle, Washington: Washington Univ. [Seattle], unpublished Ph. D. dissertation.

Anderson, C. A., 1933, The Tuscan formation of northern California: California Univ. Pubs., Dept. Geol. Sci. Bull., v. 23, p. 215–276.

———— 1941, Volcanoes of the Medicine Lake Highland, California: California Univ. Pubs., Dept. Geol. Sci. Bull., v. 25, no. 7, p. 347–422.

Axelrod, D. I., 1950, Studies in late Tertiary paleobotany, 6, Evolution of desert vegetation in western North America: Carnegie Inst. Washington Pub. 590, p. 215–306.

Bemmelen, R. W. van, 1949, The geology of Indonesia, v. 1A, General Geology: The Hague, Govt. Printing Office, 732 p.

Boyd, F. R., 1959, Hydrothermal investigations of amphiboles, *in* Researches in geochemistry: New York, Wiley & Sons, p. 377–396.

Carithers, Ward, 1946, Pumice and pumicite occurrences of Washington: Washington Div. Mines and Geology Rept. Inv. 15, 78 p.

Cater, F. C., 1960, Chilled contacts and volcanic phenomena associated with the Cloudy Pass batholith, Washington, *in* Short papers in the geological sciences: U.S. Geol. Survey Prof. Paper 400–B, p. B471–473.

Chaney, R. W., and Axelrod, D. I., 1959, Miocene floras of the Columbia Plateau: Carnegie Inst. Washington Pub. 617, 237 p.

Coombs, D. S., 1954, The nature and alteration of some Triassic sediments from Southland, New Zealand: Royal Soc. New Zealand Trans., v. 82, p. 65–109.

Coombs, D. S., Ellis, A. J., Fyfe, W. S., and Taylor, A. M., 1959, The zeolite facies; with comments on the interpretation of hydrothermal syntheses: Geochim. et Cosmochim. Acta, v. 17, p. 53–107.

Coombs, H. A., 1936, The geology of Mount Rainier National Park: Washington Univ. [Seattle] Pub. in Geology, v. 3, no. 2, p. 131–212.

———— 1939, Mt. Baker, a Cascade volcano: Geol. Soc. America Bull., v. 50, no. 10, p. 1493–1510.

Crandell, D. R., 1963, Surficial geology and geomorphology of the Lake Tapps quadrangle, Washington: U.S. Geol. Survey Prof. Paper 388–A.

Crandell, D. R., Mullineaux, D. R., and Waldron, H. H., 1958, Pleistocene sequence in southeastern part of the Puget Sound lowland, Washington: Am. Jour. Sci., v. 256, p. 384–397.

Crandell, D. R., and Waldron, H. H., 1956, A recent volcanic mudflow of exceptional dimensions from Mt. Rainier, Washington: Am. Jour. Sci., v. 254, p. 349–362.

Crowder, D. F., 1959, Granitization, migmatization, and fusion in the northern Entiat Mountains, Washington: Geol. Soc. America Bull., v. 70, p. 827–278.

Curtis, G. C., 1903, Secondary phenomena of the West Indian volcanic eruptions of 1902: Jour. Geology, v. 11, p. 119–215.

Emmons, S. F., 1879, The volcanoes of the Pacific Coast of the United States: Am. Geog. Soc. Jour., v. 9, p. 45–65.

Eskola, P., 1939, Die metamorphen Gesteine, p. 263–407, in Barth, T. F. W., Correns, C. W., and Eskola, P., Die Entstehung der Gesteine: Berlin, Springer, 422 p.

Felts, W. M., 1939, A granodiorite stock in the Cascade Mountains of southwestern Washington: Ohio Jour. Sci., v. 39, p. 297–316.

Fisher, R. V., 1957, Stratigraphy of the Puget group and Keechelus group in the Elbe-Packwood area of southwestern Washington: Washington Univ. [Seattle], unpublished Ph. D. dissertation.

———— 1958, Definition of volcanic breccia: Geol. Soc. America Bull., v .69, p. 1071–1073.

———— 1960, Classification of volcanic breccias: Geol. Soc. America Bull., v. 71, p. 973–982.

Fiske, R. S,. 1963, Subaqueous pyroclastic flows in the Ohanapecosh Formation, Washington: Geol. Soc. America Bull., v. 74, p. 391–406.

Fiske, R. S., and Matsuda, T., 1964, Submarine equivalents of ash flows in the Tokiwa Formation, Japan: Am. Jour. Sci., in press.

Fremont, J. C., 1845, Report of the exploring expedition to the Rocky Mountains in the year 1842, and to Oregon and North California in the years 1843–44: Washington, Blair and Rives, 693 p.

Fuller, R. E., 1925, The geology of the northeastern part of the Cedar Lake quadrangle, with special reference to the deroofed Snoqualmie batholith: Washington Univ. [Seattle], unpublished M.S. thesis, 96 p.

Fyfe, W. S., Turner, F. J., and Verhoogen, J., 1958, Metamorphic reactions and metamorphic facies: Geol. Soc. America Mem. 73, 259 p.

Gard, L. M., Jr., 1960, Suggested source of Miocene volcanic detritus flanking the central Cascade Range, Washington, in Short papers in the geological sciences: U.S. Geol. Survey Prof. Paper 400-B, p. B306–307.

Goodspeed, G. E., Fuller, R. E., and Coombs, H. A., 1941, Metasomatism of a coaly sediment into an igneous appearing rock: Jour. Geology, v. 49, p. 190–198.

Grant, R. Y., 1941, A John Day vertebrate fossil discovered in the Keechelus series of Washington: Am. Jour. Sci., v. 239, p. 590–593.

Grater, Russel K., 1948, The flood that swallowed a glacier: Nat. History, v. 57, p. 276–278.

Hague, Arnold, and Iddings, J. P., 1883, Notes on the volcanoes of northern California, Oregon and Washington Territory: Am. Jour. Sci., 3d ser., v. 26, no. 153, p. 222–235.

Hay, R. L., 1952, The terminology of fine-grained detrital volcanic rocks: Jour. Sed. Petrology, v. 22, p. 119–120.

Holmsen, Per, 1952–53, Landslips in Norwegian quick-clays: Geotechnique, v. 3, p. 187–194.

Hopson, C. A., Waters, A. C., Bender, V. R., and Rubin, Meyer, 1962, The latest eruptions from Mount Rainier volcano: Jour. Geology, v. 70, p. 635–647.

Kaaden, G. van der, 1951, Optical studies on natural plagioclase feldspars with high- and low-temperature optics: Univ. Utrecht, Ph. D. dissertation [private printing], 101 p.

Kautz, A. V., 1875 [Account of attempted ascent of Mount Rainier, 1857], repr. from Overland Monthly, May, 1875 in Meany, E. S., 1916, Mount Rainier—A record of explorations: New York, MacMillan Co., p. 73–93.

King, A. T., 1871, On the discovery of actual glaciers on the mountains of the Pacific Slope: Am. Jour. Sci., 3d ser., v. 1 p. 157–167.

Kuno, Hisashi, 1954, Study of orthopyroxenes from volcanic rocks: Am. Mineralogist, v. 39, p. 30–46.

Landes, H., 1905, Field notes on Mount Rainier: Mazama, v. 2, no. 4, p. 220–223.

Lipson, J., Follinsbee, R. E., and Baadsgaard, H., 1961, Periods of orogeny in the western Cordillera: New York Acad. Sci. Annals, v. 91, p. 459–462.

Matthes, F. E., 1914, Mount Rainier and its glaciers: Natl. Park Service, p. 1–48.

Nockolds, S. R., 1954, Average chemical compositions of some igneous rocks: Geol. Soc. America Bull., v. 65, p. 1007–1032.

Norton, W. H., 1917, A classification of breccias: Jour. Geology, v. 25, p. 160–194.

Oebbeke, K., 1885, Uber das Gestein vom Tacoma-Berg, Washington Territory: Neues Jahrb. Mineralogie, Geologie, u. Paläontologie, v. 1, p. 222–226.

Padang, M. N. van, 1939, Ueber die vielen tausend Hügel im westlichen Vorlande des Raoeng-Vulkans (Ost-Java): De Ingenieur in Nederlandsch-Indie, v. 6, no. 4, p. 35–41. [Cited from English summary by Van Bemmelen, 1949, p. 194–195.]

Poldervaart, A., 1950, Correlation of physical properties and chemical composition in the plagioclase, olivine and orthopyroxene series: Am. Mineralogist, v. 35, p. 1067–1079.

Reusch, Hans, 1901, Nogle optegnelser fra Vaerdalen: Norsk geol. undersögelse, No. 32, Aarbog 1900, p. 1–32. [English summary, p. 218–226.]

Richey, J. E., 1932, The Tertiary ring complex of Slieve Gullion (Ireland): Geol. Soc. London Quart. Jour., v. 88, p. 776–894.

Rigg, G. B., and Gould, H. R., 1957, Age of Glacier Peak eruption and chronology of post-glacial peat deposits in Washington and surrounding areas: Am. Jour. Sci., v. 255, p. 341–363.

Rosenqvist, Ivan Th., 1952–53, Considerations on the sensitivity of Norwegian quick-clays: Geotechnique, v. 3, p. 195–200.

Russell, I. C., 1897, Glaciers of Mount Rainier: U.S. Geol. Survey Ann. Rept. 18, pt. 2, p. 349–415.

Smith, G. O., 1897, The rocks of Mount Rainier: U.S. Geol. Survey Ann. Rept. 18, pt. 2, p. 416–423.

———— 1900, The geology of Mount Rainier: Mazama, v. 8.

———— 1903, Description of the Ellensburg quadrangle [Washington]: U.S. Geol. Survey Geol. Atlas, Folio 86, 7 p.

Smith G. O., and Calkins, F. C., 1906, Description of the Snoqualmie quadrangle [Washington]: U.S. Geol. Survey Geol. Atlas, Folio 139, 14 p.

Smith, G. O., and Mendenhall, W. C., 1900, Tertiary granite in the northern Cascades: Geol. Soc. America Bull., v. 11, p. 223–230.

Smith, R. L., 1960, Ash flows: Geol. Soc. America Bull., v. 71, p. 795–842.

Snavely, P. D., Brown, R. D., Roberts, A. E., and Rau, W. W., 1958, Geology and coal resources of the Centralia-Chehalis district, Washington: U.S. Geol. Survey Bull. 1053, 159 p.

Speight, R., 1917, The geology of the Banks Peninsula: New Zealand Inst. Trans., v. 49, p. 365–392.

Tertsch, H., 1942, Zur Hochtemperaturoptik basischer Plagioklase: Mineralog. Petrog. Mitt., v. 54, p. 193–217.

Thayer, T. P., 1937, Petrology of later Tertiary and Quaternary rocks of the north-central Cascade Mountains in Oregon, with notes on similar rocks in western Nevada: Geol. Soc. America Bull., v. 48, p. 1611–1652.

Tolmie, W. F., 1833 [Diary of W. F. Tolmie, excerpts Aug. 28 to Sept. 5, 1833, describing first attempts to ascend Mount Rainier], in Washington Hist. Quart.: v. 1, no. 1, p. 77–81.

Turner, F. J., 1947, Determination of plagioclase with the four-axis universal stage: Am. Mineralogist, v. 32, p. 389–410.

Verhoogen, J., 1937, Mount St. Helens, a Recent Cascade volcano: California Univ. Pubs., Dept. Geol. Sci. Bull., v. 24, p. 263–302.

Warren, W. C., 1941, Relation of the Yakima basalt to the Keechelus andesitic series: Jour. Geology, v. 49, p. 795–814.

Waters, A. C., 1955a, Geomorphology of south-central Washington, illustrated by the Yakima East quadrangle: Geol. Soc. America Bull., v. 66, p. 663–684.

———— 1955b, Volcanic rocks and the tectonic cycle, in Poldervarrt, Arie, ed., Crust of the earth: Geol. Soc. America Spec. Paper 62, p. 703–722.

Waters, A. C., 1961, Keechelus problem, Cascade Mountains, Washington: Northwest Sci. v. 35, p. 39–57.

Wentworth, C. K., and Williams, Howel, 1932, The classification and terminology of the pyroclastic rocks: Natl. Research Council Bull. 89, Rept. Comm. Sedimentation, p. 19–53.

Williams, Howel, 1932, Geology of the Lassen Volcanic National Park, California: California Univ. Pubs., Dept. Geol. Sci. Bull., v. 21, no. 8, p. 195–385.

———— 1933, The geology of Tahiti, Moorea, and Maiao: B. P. Bishop Museum Bull. 105, p. 19–21.

———— 1934, Mount Shasta, California: Zeitschr. Vulkanologie, v. 15, p. 225–253.

———— 1941, Calderas and their origin: California Univ. Pubs., Dept. Geol. Sci. Bull., v. 25, no. 6, p. 239–346.

———— 1942, The geology of Crater Lake National Park, Oregon, with a reconnaissance of the Cascade Range southward to Mount Shasta: Carnegie Inst. Washington Pub. 540, 162 p.

———— 1944, Volcanoes of the Three Sisters region, Oregon Cascades: California Univ. Pubs., Dept. Geol. Sci. Bull., v. 27, no. 3, p. 37–84.

Williams, Howell, Turner, F. J., and Gilbert, C. M., 1955, Petrography: San Francisco, Calif., Freeman & Co., 406 p.

Willis, Bailey, 1883, Canyons and glaciers, A Journey to the ice fields of Mount Tacoma: The Northwest [mag.], v. 1, no. 2; repr. in Meany, E. S., 1916, Mount Rainier—A record of explorations: New York, Macmillan Co., p. 142–149.

Wise, W. S., 1959, Occurrence of wairakite in metamorphic rocks of the Pacific Northwest: Am. Mineralogist, v. 44, p. 1099–1101.