FRONTISPIECE

Olallie Butte, viewed from the southwest across an arm of Monon Lake. The oversteepening effect of glacial action on the lower slopes is quite pronounced.

THE GENERAL GEOLOGY OF THE NORTH SANTIAM RIVER SECTION OF THE OREGON CASCADES

by

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Chapter 1

INTRODUCTION

The district discussed in this paper is located in North-western Oregon, about fifty miles south of the Columbia River. The section investigated extends eastward from the Willamette River a few miles south of Salem to the Cascade Range summit between Mount Jefferson and Olallie Butte. It follows the North Santiam River, which is near the 44° 45' parallel. The area includes portions of the Salem, Lebanon, Stayton, Mill City and Mt. Jefferson quadrangles of the Topographic Atlas of the United States and part of a fifteen minute quadrangle west of the Mill City sheet which was mapped by the author.

Brief accounts of observations in the Oregon Cascades which were published toward the end of the last century were so general in character as to have little bearing on the present problems. The first detailed study in the Cascades was that of Diller and Patton[1] in 1902. They deciphered the history of Mt. Mazama which is now so well known. In 1914 Washburne[2] suggested the possibility of a fault along the east side of the Willamette Valley, but stated none need be postulated if the lavas are younger than the sediments, as is the case.

In 1916 Ira Williams published two papers on the Cascades. The first[3] is a description of certain parts

1. Diller, J.S. & Patton, H.B.; Geology & Petrography of Crater Lake Nat'l Park: Surv. Prof. Paper 3, 167 pp, 1902
2. Washburne, C.; Reconnaissance of the geology and oil prospects of Northwestern Oregon: U.S.G.S. Bull. pp 8, 14 1914
3. Williams, Ira; Some little known scenic pleasure places on the Cascade Range in Oregon: Min. Res. Oregon vol. 2 No. 1 114 pp, 1916

of the range, mainly from a scenic point of view, with a geologic background. This paper contains many good illustrations. The second deals with the Columbia Gorge[1] and describes the salient features of that famous section.

Dr. E. T. Hodge, perhaps the chief investigator of the Cascade Range in Oregon has published several papers, but only two are directly concerned with the North Santiam District. A paper on Mount Jefferson[2] was based on rather brief field studies but gives a good general idea of the volcanic history of the peak and the surrounding area. The unconformable relations of the different lava series were recognized and the glacial geology described but as the investigation was limited to the High Cascades, the magnitude of the post-Miocene unconformity was not perceived. In 1928[3] Hodge set forth views regarding the section of the Range between the McKenzie and North Santiam Rivers. The young lavas were designated as the Mount Jefferson formation; and structure was suggested as the factor determining the north-south valley occupied by the North Santiam and McKenzie Rivers. The steep folds at Detroit and in the South Santiam River at Cascadia were noted, the inference being that both are part of the same structure. The rhyolite rocks of the Cascade Range were correlated with the Clarno Eocene, the tuffs with the John Day formation, and the basaltic lavas with the Columbia

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1. Williams, Ira; The Columbia River Gorge: its geologic history: Min. Res. Oregon Vol. 2, No.3, 1916
2. Hodge, E.T.; The geology of Mt. Jefferson: Mazama, Vol. 7, No. 2, pp. 25-55, 1925
3. Hodge, E.T.; Framework of Cascade Mountains: Oregon: Pan. Am. Geol. Vol. 49, No. 5, pp. 341-356, (1928).

River Lavas. It was stated that west of Detroit the lavas rest unconformably on the tuffs. Later detailed work in the Mt. Hood region is as yet unpublished.

Chaney[1] on the basis of studies in the Ashland region, considered that the rocks range from Clarno to Mascall, and stated that the southern part of the range was not elevated before the upper Miocene.

The most recent general paper on the Oregon Cascades was published by Callaghan[2] of the U.S. Geological Survey. He pointed out the linear character of many features of the Range and indicated that south of Mt. Hood it could be divided longitudinally into two parts; the Western Cascades and the High Cascades.

The Western Cascades consist of lavas ranging from basalts to rhyolites, which have been intruded and deformed. They are seperated by a marked unconformity from the High Cascade lavas which are very fresh basalts and andesites. Callaghan includes some new analyses of lavas and intrusive rocks from the central portion of the range.

The present investigation is the indirect result of a three months reconnaissance of the Western Cascades in the summer of 1931 with Dr. Callaghan. The structure of the McKenzie section was known to be homoclinal, whereas the structure of the Columbia River section consists of a series of gentle folds. The problem of correlating the structure of

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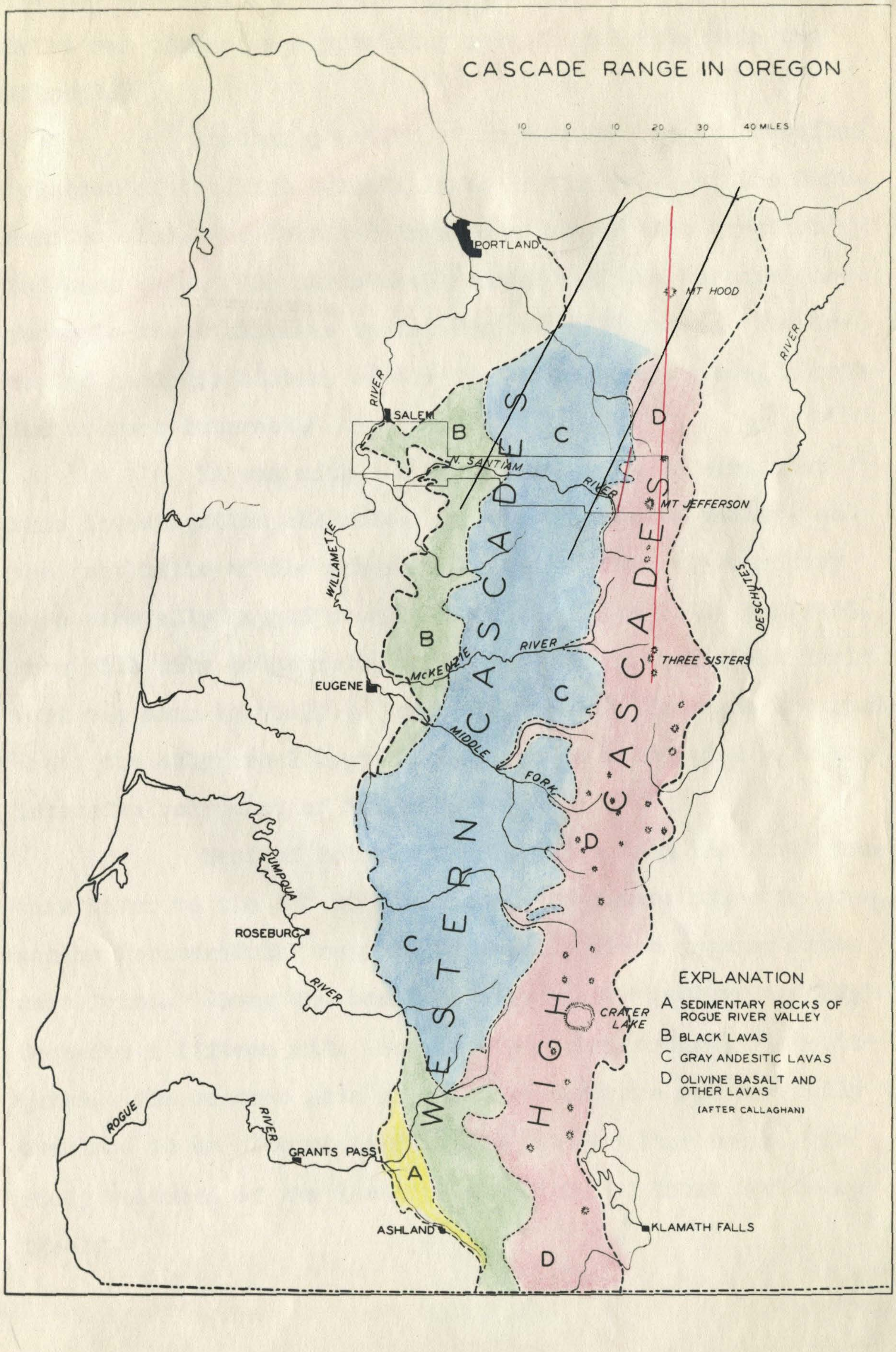
1. Chaney, R.W.: Suggestions regarding the age of the Southern Cascade Ranges: (Abst) Bull. Geol. Soc. Am. Vol. 41, p. 147 1930
2. Callaghan, Eugene: Some features ofthe volcanic sequence in the Cascade Range in Oregon: Trans. Am. Geophys. Un, 14th An. Meeting, pp 243-249, 1933

PLATE 3

Index map of Cascade Range in Oregon, showing divisions according to Callaghan. The area outlined is included in this report. The axes of the principal Western Cascade anticlines are shown in black; the Cascade fault and structural axes of the High Cascades are shown in red.


CASCADE RANGE IN OREGON
10 0 10 20 30 40 MILES
PORTLAND
MT HOOD
SALEM
RIVER
N. SANTIAM
RIVER
MT JEFFERSON
WILLAMETTE
DESCHUTES
RIVER
McKENZIE
RIVER
THREE SISTERS
EUGENE
MIDDLE
FORK
UMPQUA
ROSEBURG
RIVER
CRATER
LAKE
ROGUE
RIVER
GRANTS PASS
ASHLAND
KLAMATH FALLS
WESTERN CASCADES
HIGH CASCADES
A
B
C
D
EXPLANATION
A SEDIMENTARY ROCKS OF ROGUE RIVER VALLEY
B BLACK LAVAS
C GRAY ANDESITIC LAVAS
D OLIVINE BASALT AND OTHER LAVAS
(AFTER CALLAGHAN)

the two sections was recognized, and the North Santiam district was chosen as a promising area in which to make the attempt.

Another question of importance was the detailed relation of the High Cascade lavas to the rocks of the Western Cascades. No detailed investigation of this question had been made. The structural relation of the Western Cascades to the Willamette Valley was not well known. The detailed geologic history of the region was, of course, a problem of much interest.

It was with the above questions in mind that this investigation was made. In the Willamette Valley, and the foot hills of the range, 300 square miles of territory were carefully mapped areally from the Salem Hills eastward. From Mill City to Detroit, within the range, sufficient field work was done to yield an accurate idea of the general structure: the only areal mapping done was in connection with the intrusive body west of Detroit.

East of Detroit the district from the North Santiam River to the 44° 50' parallel, 250 square miles in area, was rather carefully mapped, especially where younger lavas were found. Along the boundary between the Western and High Cascades a fifteen mile long section was thoroughly investigated. The western side of the High Cascades was carefully examined in an attempt to find any buried older rocks. No study was made of the lavas or structure of Mount Jefferson itself.

Resume of General Geology and Physiography

The Cascade Range can be divided longitudinally into two distinct ranges which have been named the Western Cascades and the High Cascades[1]. (See Plate III) The boundary between the two ranges in the North Santiam River section roughly follows a line through Breitenbush Hot Springs and the north-south stretch of the North Santiam River.

The Western Cascades are about fifty miles wide east of Salem; if the Salem Hills, which are structurally part of the range, be included, the width is sixty miles. The general elevation of the eastern half of the range is between forty-five hundred and five thousand feet, and the relief between three and four thousand feet. The Salem Hills and the foothill slopes west of Stayton are practically all below one thousand feet high. The region is one of heavy rainfall; as a consequence the forest cover is dense and outcrops are poor, except where erosion is unusually rapid or the timber has been cleared off by man or fire.

The geology of the Western Cascades is relatively simple. The rocks are Oligocene-Miocene lavas and tuffs, andesitic in composition, cut by hypabyssal intrusives. The flows and pyroclastics have been folded into gentle folds, namely, two synclines and two anticlines, generally striking north of east; dips on the limbs average three to five degrees. The folds, in order from west to east are: the Willamette syncline, Mehama anticline, Sardine syncline, and Breitenbush anticline. The last-named fold is very asymmetrical; the western limb dips fifty degrees, the eastern

1. Callaghan, E., Some features of the volcanic sequence in the Cascade Range in Oregon: Trans. Am. Geophysical Union, 14th An. Meeting, pp 243-245, 1933

limb about twelve degrees. Whereas the western border of the range is characterized by two or three degree dip slopes extending down to the Willamette Valley floor, the eastern margin is an eroded fault scarp, steep in places, now buried under the High Cascade lavas or their equivalents.

The width of the High Cascades is about twenty-five miles; the crest-line is about ten miles east of the western edge and averages more than fifty-five hundred feet in elevation south of Olallie Butte. This post-Miocene range is built up of basaltic lavas rich in olivine and retaining their original structures. In the main the lavas dip in both directions from the north-south crest of the range, and the line of modern volcanic peaks. Old plugs exposed along this axis are indicated as the sources of the flows. Glaciated dip slopes deeply trenched by glacial troughs dominate the topography south of Mt. Jefferson, whereas north of that peak the crest widens to give the range the aspect of a plateau glacially dissected along its margins. South of the right-angle bend in the Santiam River northeast of Mt. Bruno (hereafter referred to as the Santiam Elbow) the western limit of the High Cascades proper is the Santiam River Valley itself; north of the elbow no physiographic boundary is evident.

The author wishes to express his gratitude to Dr. J. P. Buwalda of the California Institute of Technology for his help and criticism in interpreting the structural and physiographic portions of this paper. The writer is also greatly indebted to Dr. Ian Campbell of the same institution for his aid in the petrographic work and for several days spent in the field. Thanks are also due Dr. W. M. Davis and Dr. E. T. Hodge for their criticism and kind cooperation. Dr. Eugene Callaghan's criticism and suggestions during the course of the study are especially appreciated.

CHAPTER II

STRATIGRAPHY

The formations of the Western Cascades range in age from Pleistocene (not including alluvial valley deposits) to Lower or Middle Oligocene, and possibly to Eocene. Detailed stratigraphic investigation is rendered difficult by the discontinuity of outcrops, by the petrographic uniformity of the whole section, and by the variable structures of the volcanic ejectamenta and lavas. The absence of fossils everywhere except in the marine Oligocene sediments renders accurate correlations impossible at this time. No fossils except fragmentary leaves and petrified wood have been found east of Stayton. A six inch water worn boulder containing Oligocene marine (?) fossils is reported to have been found in low terrace gravels along the North Santiam River about six miles east of Detroit.

Eocene

Two occurrences of rhyolite are tentatively correlated with the upper Clarno Eocene, and two others are suggestive of Clarno age. The two latter occurrences, on the north side of Minto Mountain and the north side of the Sentinnel Hills ridge, are very similar. In both localities the rocks are thick, pink to white flow rhyolite dipping eastward under a covering of Minto lavas. In the Sentinel Hills area the upper twelve hundred feet is known to ba a homogeneous rhyolite section; the lower six hundred feet is very poorly exposed and may contain other rock types. A plug of porcelainous white rhyolite forms a prominent knob at thirty-eight hundred feet elevation on the Sentinel Trail. The exposed Minto Mountain section is thinner, contains green tuff not far from the top, and probably contains andesitic rocks in the lower portions.

Other rhyolite, generally more massive, crops out at several places along the Santiam Highway from the vicinity of Boulder Creek almost to Tunnel Creek, a distance of five and a half miles. Massively bedded green tuffs are apparently interbedded with the rhyolite. Some flow rhyolite crops out in the Breitenbush River east of Fox Creek in a highly disturbed zone which suggests faulting. It may have been brought up from below by the disturbance.

The Minto and Sentinel Hills rhyolite is considered as possibly Eocene because such rhyolite is common to the upper part of the Clarno. Rhyolite occurs in the Western Cascades however, and abundantly in some parts, although no rocks of this type occur in the North Santiam River section proper, except east of Detroit at low elevations. Rhyolite is present in considerable thickness in the Quartzville district about fifteen miles southwest of Detroit, so may occur in other nearby parts of the range. The high elevation of the eastern exposures (five thousand feet in the Sentinel Hills and forty-two hundred feet in Minto Mountain) precludes stratigraphic continuity with the Breitenbush rhyolite for two reasons; the low elevations at which the latter occur (two thousand feet), and the prevailing eastward dips. There is no evidence for duplication of beds by faulting. These structural considerations and occurrences of rhyolite in nearby parts of the range suggest that the Minto and Sentinel Hills rhyolites may be part of the post-Eocene sequence. The Breitenbush rhyolite is probably Eocene lava brought up by the folding in the Breitenbush anticline.

Oligocene-Miocene

The separation of Oligocene and Miocene formations in the Oregon Cascades is as yet very incomplete. At the western edge of the Range, rocks belonging in the two periods are separated by a distinct unconformity which has not been traced east of the crest of the Mehama anticline. Within the Range no fossils have been found, nor has good evidence of any stratigraphic break been located in the volcanic sequence. East of Detroit the Sardine and Breitenbush series are separated purely on a lithologic basis. The relations of the Oligocene-Miocene formations are diagramatically shown in the follwoing table.

Table I

Correlation chart of the Oligocene-Miocene formations of the Western or Older Cascades.

	Western	Central	Eastern
Miocene	Fern Ridge tuffs		
	Stayton lavas	Sardine	Series
Oligocene	Illahe - Mehama		
			Breitenbush Series

The Illahe and Mehama Formations:

The Illahe is the only formation in the North Santiam Cascade section definitely known to be Oligocene in age. The Oligocene beds underlie the Salem Hills and are exposed in places in the hills north of Turner, particularly along the low scarp which faces the alluvial floor of the Willamette Valley. They are made up of fairly well bedded tuffaceous marine sediments, varying in coarseness from pebbly conglomerates to chalky massive white ash; (Plate IIIA); silty sandstones are most abundant. The regional dip of the formation is to the northeast, but local dips vary greatly. As exposures of the Illahe are scarce because of its poor resistance to weathering, an estimate of its thickness is only little better than a guess. As Schenck[1] reported Eocene rocks a few miles south of Independence, the base of the formation is probably not far southwest of the Salem Hills. A half mile is probably a liberal estimate for the local thickness of the Oligocene.

Marine fossils are abundant in places in the true Illahe Formation. The invertebrate fauna includes Macrocallista pittsburgensis (Dall), Molopophorus gabbi Dall, Nucula (Acila) shumardi Dall, and Tellina eugenia Dall, all of which are characteristic of the Pittsburg Bluff and Eugene Formation assemblages. The latter formations, with which the Illahe is correlated, are generally regarded as Lower or Middle Oligocene.

1. Schenck, H.G.; Stratigraphic Relations of the Western Oregon Oligocene formations; U.C.Publ. Dept. Geol. Sci; Vol.18, pp 39-48, 1928

PLATE 4

A. Fine waterlaid tuff of the Illahe formation two miles northwest of Turner in road cut east of Battle Creek. Marine fossils were found in this exposure. Concentric weathering is remarkably developed near the three foot rule.

B. Agglomerate of the Mehama Formation in Thomas Creek about one and one half miles east of Schimonek Bridge. The boulders are labradorite andesite. Note person for scale.

PLATE 4.

A.

B.

East of Turner the Illahe Formation coarsens and contains no marine fossils. Continuous tracing of the formation eastward is impossible, but it apparently grades into the terrestrial Mehama volcanic deposits. Intermingling of shaly material with coarse conglomerates and agglomerates is well shown in Thomas Creek northeast of Scio. Plate 4b shows an exposure of very massive agglomerate, which lies between thin shaly beds. The relations suggest near-shore deposition.

The maximum exposed thickness of the Mehama volcanics is about six hundred feet, in the crest of the Mehama anticline. The section is composed mostly of fine white or greenish tuffs, some of which contain fossil wood. In the hill a mile southeast of Mehama, however, tremendously coarse agglomerate is well exposed. A fragment of a flow fifty by twenty-five by ten feet indicates that the source of the pyroclastics was near at hand.

The laterally gradational character of the Illahe, and the similarity of the structural relations of both the Illahe and the Mehama formations to the overlying Stayton lavas strongly indicate their contemporaneity. They are therefore regarded as stratigraphically equivalent. The Oligocene shoreline was not far east of Marion, and very close to Turner, north of which place it swung northeastward. The Mehama volcanics are very probably equivalent to the Eagle Creek formation in the Columbia Gorge section.

The Stayton Lavas:

The Stayton lavas form a cover on the Illahe Formation in the Salem Hills, and in the foothills of the Cascades underlie the dip slopes up to an elevation of fifteen hundred feet. Near the crest of the Mehama anticline they merge with the Sardine series. The maximum thickness of the lavas in the area investigated is about four hundred feet, but their thicknesses increase northward. The flows have been completely removed in many plaues, but their average remaining thickness is about two hundred feet. They are mainly basaltic in composition, but thin rhyolites may possibly occur in the Salem Hills.

A marked unconformity separates the Stayton lavas from the Illahe-Mehama formations. The average dip of the lavas is not over three degrees; in the Oligocene formations the average dip is ten degrees, the greatest, ninety. The lavas were poured out on an erosion surface of moderate relief. The magnitude of the unconformity and the thickening of the lavas toward the northeast suggest a Miocene age. The Columbia River Basalts have been traced to the southern edge of the Estacada quadrangle[1], and it is altogether likely that the Stayton lavas are actually the thin border of that series of flows. In this paper the Stayton lavas are considered as Middle Miocene.

Fern Ridge Tuff:

The Fern Ridge formation consists of andesitic

1. Barnes, F.F., and Butler, J.W.; Unpublished master's thesis, Univ. of Oregon, 1930.

tuffs and conglomerates. It varies from zero to about twelve hundred feet in thickness, the latter figure being attained northeast of Mehama near the House Mountain scarp. Thin patches and low isolated hills of it occur down the foothill slopes into the eastern edge of the Stayton sheet. The ridge crests northwest of House Mountain (Plate XXII A) owe their remarkable uniformity to the resistance of the capping tuff beds. The basal beds of the series, where not tuffs, are fine pebble conglomerates, and even sandstones, whereas most of the conglomerates higher up contain large boulders.

The Fern Ridge materials lie conformably on the Stayton lavas, and very likely are part of the same general volcanic sequence. They cannot be differentiated east of the Mehama anticlinal axis, so are regarded as eruptive and erosion products deposited contemporaneously with the upper lavas of the Sardine series. The boulders in the conglomerates are lithologically very similar to the Sardine lavas, as such an hypothesis would require. The formation is undoubtedly Miocene, probably Upper Middle.

The Sardine Series:

The Sardine series derives its name from Sardine Mountain, where a four thousand foot section of lavas is well exposed. The lavas are mainly calcic andesites, but dacites also occur. Intercalated tuffs, though by no means rare, are less abundant than flows. This thick succession of lavas, rising westward partly accounts for the eminence of Monument Peak and Mt. Horeb above the general landscape to the west. The whole section along the river from Mill City to Halls shows low southeasterly dips; the thickness of the series must be not less than six thousand feet if an average dip of three degrees is assumed. The calculated thickness east of the synclinal axis, where dips have been accurately determined, agrees closely with the above figure. The base of the Sardine series is necessarily more tuffaceous because of the intergradation of the Sardine series with both the Mehama and Breitenbush series which are predominantly fragmental.

The stratigraphic relations of the Sardine series to the formations to the west require that the upper portion be Miocene, and the basal portion Oligocene. Such a correlation implies continuous vulcanism through the Upper Oligocene, Lower and Middle Miocene. This may very well have been the case, with the result that no definite break would occur in the series. Careful search revealed no unconformity, so for the present it is assumed there is none. The diorite intrusion at Halls must be post-Middle Miocene, probably Upper Miocene.

Breitenbush Series:

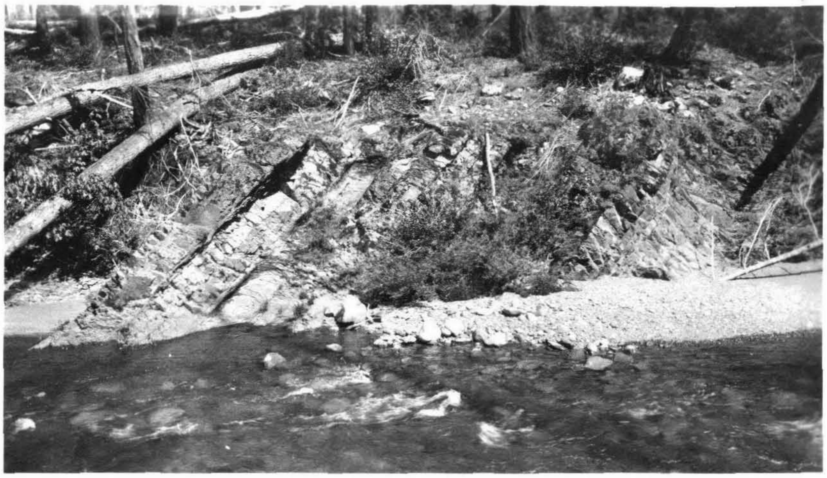
As noted above, the Breitenbush tuffs grade upward into the Sardine lava section. The minimum thickness of the tuffs is about seventy-five hundred feet, making a total for both of about thirteen thousand five hundred feet. Almost the entire type section - the Breitenbush River from the center of section 36, T. 9 S., 5 E. to Breitenbush Hot

PLATE 5

A. Tilted thinly bedded waterlaid tuff of the Breitenbush Series exposed in the Breitenbush River east of French Creek.

B. Lense of pink rhyolitic tuff in the bed of the Breitenbush River west of Byar's Creek. Massive agglomerate forms both banks of the stream. Lensing out of the tuff is quite plainly shown.

PLATE 5.



A.

B.

Springs - is composed of green tuffs or tuffaceous agglomerates, with very few flows. The upper part of the section is in general much better bedded (See Plate IV A.) than the lower parts near and east of the Breitenbush anticlinal axis. Some beds are sandy or conglomeratic and waterlaid; many are composed of large angular pumice fragments in a green matrix showing absolutely no bedding, and obviously were deposited by eruption. The basal portion of the Breitenbush series contains rhyolitic flows and tuffs, especially in the North Santiam section east of Dry Creek. These were discussed above in connection with the Eocene. In the Santiam the structure is very indefinite, and minor faults occur. Exposures are too poor and scattered to permit very detailed search for a stratigraphic break, and one may be present between the rhyolitic and andesitic facies of the series.

The utter lack of fossils in the Breitenbush tuffs leaves only structural evidence for purposes of correlation. The gradational relation to the Sardine series, so similar to the comparable relation of the Mahama volcanics, suggests that the upper portion of the tuffs may be Oligocene of the same general age as the Mahama-Illahe. The base of the section may include the upper portion of the Clarno, thus accounting for the fhyolitic character of the rocks.

In conclusion, the volcanic sequence in the Western Cascades compares well with the known sequence in Eastern Oregon. The base of the Breitenbush series is rhyolitic, and is correlated with the Clarno, also rhyolitic. The main Breitenbush series is tuffaceous, corresponding to the

John Day formation. Overlying the tuff series is the Sardine lava succession which is petrologically similar to, and correlated with, the Miocene Columbia River basalts.

Pliocene-Pleistocene

The Pliocene-Pleistocene succession of lavas is separated from the Western Cascade section (Miocene and pre-Miocene) by a marked unconformity. This depositional break extended from the cessation of Western Cascade vulcanism, possibly at the end of Middle Miocene time, to the beginning of High Cascade activity in the Pliocene. During this time dioritic intrusion, mineralization, folding and uplift of the range, and erosion to a relief of over a half mile occurred. This stratigraphic break is recognizable the entire length of the Cascade Range in Oregon.

Most of the post-Miocene sequence is petrographically homogeneous. Light gray olivine basalts and agglomerates predominate. True tuffs are comparatively uncommon. In the uppermost part andesitic rocks are common, and petrographic varieties are more numerous. In a single group of flows, some may be very thin, - ten feet or less; - while others are very thick, three hundred feet or more.

The Outerson Series:

The Outerson series comprises the oldest flows of the Younger or High Cascade basalts. The lavas were poured out on the eroded eastern margin of the Western Cascades from volcanic centers in Outerson Mountain, Mt. Bruno and unknown sources. Reddish agglomerates are very abundant.

In the east side of Mt. Bruno below forty-eight hundred feet elevation there is a three hundred foot section of finely laminated tuffs, probably lacustrine deposits, with intercalated basalt flows. Structures at the base of some of the flows suggest pillow-structure. The sediments vary from almost white to greenish gray and black. A few leaf fragments and a conifer bract were found: diligent search might discover a moderately large flora. The fine beds grade upward into cross-bedded lapilli, and both are covered with four or five hundred feet of thin lava flows. The type section east of Outerson Mountain is composed entirely of agglomerates, flow-breccias, thin flows, and some tuffs. The series is at least three thousand feet thick locally.

The Battle Ax Lavas:

The Battle Ax basalts were extravasated from a volcano whose vent was approximately where Battle Ax now stands. The section is almost entirely lava flows, and has been very badly dissected and worn down. Although the greatest remaining thickness is only a little over one thousand feet, the lavas occur through a vertical range of three thousand feet in a rather limited area, so they must have been much thicker originally. As these lavas were nowhere found in contact with the lavas of the High Cascades proper, their correlation is problematical.

The Minto Basalts:

The Minto basalts are named from Minto Mountain, in the eastern end of which (Lizard Ridge) they are well exposed with one of the intrusive plugs where they originated. The entire succession is normal gray olivine basalts and agglomerates. Flows of this series originated mainly along the axis of the High Cascades, and form the bulk of the range. They lie unconformably on the Outerson series and older rocks, but in many places cannot be distinguished with certainty from the former. The thickness of the Minto lavas is at least three, and probably over four thousand feet along the crest of the range, decreasing to the east and west.

The Pigeon Basalts:

The Pigeon basalts, on whose surface Pigeon Prairies are situated, consist of sixteen hundred feet of thin, light gray flows which partially filled the North Santiam valley west and south of the Elbow. (Plate X A) The thickness of the series is the greatest at the Elbow, indicating proximity to the source, and thins toward the west and south. The basalts once extended west well beyond Boulder Creek, and at least as far south as Marion Creek. These are the youngest lavas in the Santiam section of the Cascades, with the possible exception of Olallie Butte and the associated effusive rocks which to date cannot be differentiated from the Minto series.

Correlation

At present it is only possible to place the High Cascade lavas within the Pliocene and Pleistocene periods. Whether the Outerson series is Lower or Upper Pliocene is unknown, nor is there any means at hand whereby the Pliocene and Pleistocene may be distinguished. The various unit series have been described in order from oldest to youngest; all are post-Miocene and pre-Wisconsin, but where each comes in that interval is uncertain. The Battle Ax lavas may belong either with the Outerson of Minto lavas; on the basis of physiographic evidence they are certainly not younger than the latter. The comparatively slight dissection of the Pigeon basalts implies an early Pleistocene age.

Glacial and Post-Glacial Deposits

Glacial deposits of three different ages are found along the North Santiam valley. Till two hundred feet thick, of probable early Pleistocene age is present above and below Mill City. Scattered outcrops of till in the vicinity of Detroit very likely belong to another glacial epoch. Morainic deposits of Wisconsin age are widespread east of Tunnel Creek and Breitenbush Hot Springs. Most of the valleys of the region contain some filling or terraces of river gravel; probably none of these gravels is of pre-Pleistocene age.

CHAPTER III

STRUCTURAL GEOLOGY

The division of the Cascade Range in Oregon into two fairly distinct north-south units has already been pointed out; discussion of the structural features of the range can be carried out most effectively by keeping this division in mind, and considering the two units separately.

A question of possibly greater importance in a region of volcanics rather than marine deposits , is the relation of original structures to those of orogenic origin. Whereas most initial dips of sedimentary deposits are probably less than three degrees, initial dips in volcanic deposits may be the angle of repose of fine ash, which reaches thirty degrees or over. On the other hand, such high angles of inclination would occur only locally, and would be very diverse in direction, as around volcanic cones. Subsequent folding would accentuate some dips and correspondingly lower others, unless the axis of folding coincided with the axis of deposition.

Great lava flows, such as those of the Columbia River formation evidently flowed great distances on nearly flat surfaces, and can be treated essentially as are marine deposits. Tuffs, spread over a wide area might be similarly regarded unless evidence to the contrary were at hand. In the Western Cascades, the prevailing low dips of the lavas, and their reversal of dip on the opposite limbs of the same fold strongly suggest that the lavas are of the low-dip fissure-flow type, especially as there is no evidence that they were poured out on the crests of the present folds. It is the opinion of the writer that initial dips are a relatively unimportant factor in the Western Cascades, but probably entirely control the structure of the High Cascade and related lavas.

The Western Cascades

As briefly mentioned above the Western Cascades are a thick mass of volcanic deposits of various types. These deposits have been thrown into a series of folds trending northeast-southwest, the folding being more intense toward the southeast. The Western Cascades proper consist of two anticlines and one syncline. In order from west to east these are: the Mehama anticline, a broad, low, symmetrical fold; the Sardine syncline; and the Breitenbush anticline, a sharply asymmetrical fold characterized by dips of fifty degrees in the western limb and twelve degrees in the eastern limb. The eastern limb of the Breitenbush anticline is cut off by the almost north-south Cascade fault. In this discussion, the Willamette River, south of Salem is considered the western boundary of the Western Cascades, although the Salem Hills are topographically distinct from the range proper. This necessitates the treatment of the Willamette syncline, a very open fold west of the Mehama anticline, as a Western Cascade structure. All of the Western Cascade folds appear to plunge toward the northeast with varying steepness. The High Cascade structures, on the other hand trend nearly north-south and are apparently the unmodified original structures of the lavas. The Western Cascades, being geologically older than the High Cascades in every way, will be considered first. Details of their structure are most logically discussed in order from west to east.

PLATE 6

The Eola Hills, the structural northward extension of the Salem Hills, viewed from the northern end of the Salem Hills. The homoclinal structure of the hills is quite evident, and some landslide topography may be seen along the western scarp. Illahe Hill is in the middle distance on this side of the Willamette River. The bench in the foreground is formed on the basal flow of the Stayton Lavas. The view is slightly west of north, Salem being just beyond the right edge of the picture.

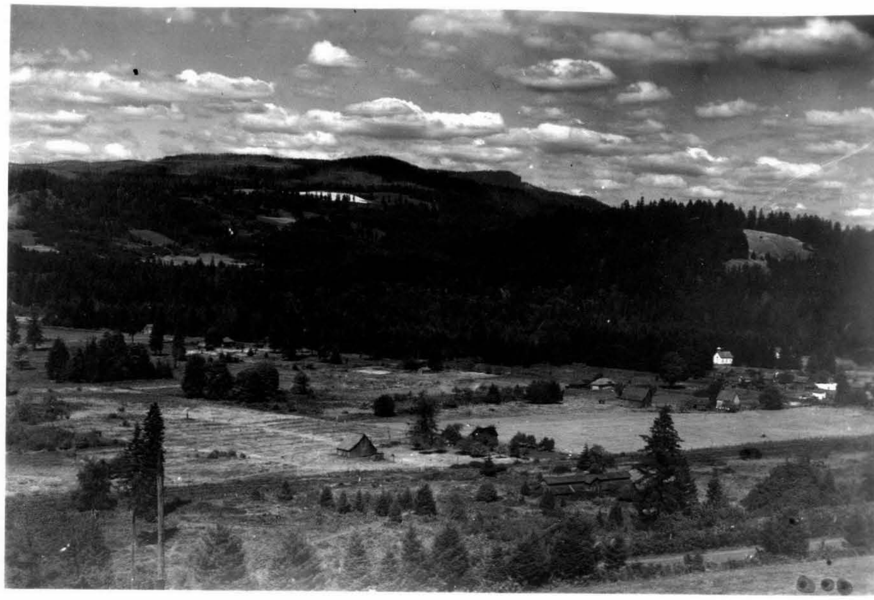
PLATE 6.

PLATE 7

A. The House Mountain Scarp as seen from the terrace southwest of Lyons. The junction of the North Santiam and Little North Santiam Rivers is directly below the light colored field on the hillside. The broad terraced floor of the North Santiam River Valley lies in the foreground. The view is northeast, parallel to, but west of the axis of the Mehama anticline.

B. McCully Peak and Thomas Creek valley as seen from the west. The structural surface on the Stayton lavas is visible either side of the peak which is composed of Fern Ridge tuff, and marks the crest of the Mehama anticline. The width of the valley is due to sapping of the Mehama fragmental rocks which underlie the Stayton lavas.

PLATE 7.



A.

B.

The Willamette syncline is a very broad open fold plunging very gently about N 30^{o}E. The western limb is formed by the Salem Hills (Plate VI), which are capped with a thin covering of Stayton lavas dipping 0^{o} to 3^{o} east or northeast. The eastern limb is represented by the northwestward dipping Miocene formation in the foothill slopes of the Cascades proper north and east of Stayton (Plate 22 A). The nose of the fold has been cut through by the North Santiam River; the lavas west of Marion undoubtedly swing around and are continuous with those near Shellburn. A low east-west crosswarp is located just north of Turner where the lavas are cut through by the valley of Mill Creek; the lavas can almost be matched flow for flow across the gap, proving their continuity. The slight structural low between this warp and the nose of the fold has been named the Stayton Basin. It is about sixteen miles wide from the east foot of the Salem Hills to Stayton. North of this warp, the syncline apparently widens very rapidly.

The Stayton lavas, which are the index horizon in the Willamette syncline, lie unconformably on the Illahe and Mehama formations. The Illahe dips average about ten degrees,- are generally northeasterly, but vary greatly in direction and amount. The steepest dip, 45^{o}, was observed at the basal contact of the Stayton lava in a roadcut on the Pacific highway. The Mehama formation likewise shows many structures at variance with those of the lavas. Whereas north of Scio the lavas dip one to three degrees northwest toward the Stayton basin, the Mehama structures dip irregularly eastward. Such directions of dip persist east of Stayton also, about halfway to Mehama. This persistency of dips in the

Oligocene formation implies that the Willamette synclinal axis is somewhat west of the axis of a steeper Oligocene structure. Ancient landsliding could scarcely account for such dips over so large and area.

The Mehama anticline seems to follow older structure more closely. The axis strikes about N.40°E., runs just west of McCully Peak, through the westernmost gap in the ridge between the North Santiam and the Little North Santiam Rivers, and somewhat east of House Mtn. (See Plate VII). The fold is very gentle, with few dips over seven degrees in the Mehama formation; however there is a highly disturbed area southeast of Mehama. The hill in section 17, T. 9 S., R.2 E. is composed of a very coarse agglomerate dipping forty-seven degrees north. In the Little North Santiam the tuffs are folded to verticality and cut by several small dikes. The limited extent of the disturbance points to a local center of vulcanism.

Dips in the Miocene lavas are very low. Northwest of House Mountain the dip averages one to two degrees northwest. In Mill City basaltic flows are exposed sloping not ~~over~~ three or four degrees eastward.

East of Mill City dependable structural evidence can be secured only by tracing single flows for some distance along the river; this procedure at best gives only apparent dips. In the gorge east of Niagara, where exposures are best, the structure has been obscured by mineralization. The intercalated tuffs, which normally would be eroded away leaving the flows as ribs on valley walls, have been so indurated that they are as resistant as the lavas, with the result that topographic indications of the structure are almost entirely lacking. The fragmentary evidence, however, indicates a general, very low dip upstream in the gorge east of Sardine Creek. The horizontal lavas in the south end of Sardine Mountain above the diorite contact represent the bottom of the Sardine syncline.

Dikes are not uncommon in the east limb of the Mehama fold. In the Little North Santiam River east of the center line of R. 2 E. there are several large basaltic dikes which are traceable for short distances along the river and which attain thicknesses of fifty feet. These dikes may have been feeders for the Sardine flows. Small dikes are numerous between Gates and Mayflower Creek, but are not large enough to have been of importance as feeders. The strike of such intrusives is usually between north and northwest. They cannot as a rule be traced beyond the limits of the river bed.

The diorite plug at Halls is apparently located exactly in the bottom of the syncline. The contacts are very poorly exposed, except in the river. The northern contact appears to be almost flat in places. In the narrow upper branches of the first creek east of Mayflower Creek the andesitic rocks extend down the divides for some distance above the intrusive rock in the creek bottoms. It is impossible to show this relation on the map because of the lack of topographic detail. Dioritic dikes and stringers are very common between the main plug and a point five hundred feet west of Mayflower Creek. Just west of the creek, two vertical dikes, twenty-five and one hundred twenty-five feet thick, respectively, and striking N. 50° W., are associated with a small pyritized and sericitized zone. East of the main intrusive no dikes were observed; this suggests that the main axis of the intrusion at depth is west or northwest of the exposed portion.

Intrusion of the diorite disturbed the lavas very little, if at all. Flows visible above the contact in Sardine Mountain, as stated above, are horizontal and undisturbed. Mineralized and breeciated zones are common within the intrusion near its margins, and in inclusions of country rock. The undisturbed condition of the country rock suggests that intrusion took place by assimilation or stoping, rather than by forcible injection.

The character of the folding in the east limb of the Sardine syncline is well shown in the eastern end of the Dome Rock ridge. From the trail up Byars Peak, the structure can be visibly followed from a horizontal attitude at Dome Rock to the fifty degree dips indicated on the map (Plate II) just below the break in slope at the east end of the ridge. The flows have indubitably been folded with the Breitenbush tuffs. The axis of the sharp folding has not been well located south of Detroit. The structure in the central isolated hill appears to be steep; in Blowout Creek below Blowout Lake tuffs appear to be nearly horizontal. The north slopes of the Blowout Cliff mountain are covered with a very dense growth of young firs, but the valley pattern of the eastern half strongly suggests in clined strata. North of French Creek the western edge of the folding goes between Byars Peak and Marten Buttes, and probably under Battle Ax.

The Breitenbush anticline has many aspects of a domical fold. Along the Breitenbush River west of Fox Creek the dips are all to the west and vary from fifty degrees or slightly over to twenty degrees or less as the axis is approached. East of Fox Creek, which is very close to the crest of the fold, the dips are uniformly about twelve degrees in southeasterly directions. In the vicinity of Gold Butte, where the structure is well shown in several exposures, the dips are northward and average about ten degrees. Further evidence of domical, or northward plunging anticlinal structure is the very limited exposure of rhyolite in the Breitenbush River as compared with the width of the rhyolitic zone in the North Santiam. That the rhyolite is carried down to the east is indicated by its absence in the North Santiam River south of the Elbow, at least as far as the Bachelor Trail crossing. To the knowledge of the writer, no geologic work has been done in the Western Cascades between the North and South Santiam Rivers east of Quartzville, but the occurrence of rhyolites in the South Santiam at Cascadia, as suggested by Hodge[1], indicates that the Breitenbush structure is a long anticline which plunges to the north of the North Santiam River.

There has been much fracturing along the axis of the Breitenbush fold. The rhyolite exposures in the Breitenbush River occur only in connection with a fractured zone three tenths mile wide, the eastern edge of which is about a quarter mile west of F Fox Creek. In this zone, very numerous andesitic dikes striking between N.10° W. and N. 30° W. cut brecciated red flow rhyolites and tuffs. The whole section is so jumbled that structural details mean nothing. The entire area northeast of Humbug Creek has been more or less mineralized. Several prospects in the Elk Lake area have been worked from time to time but evidently with small success. The best prospect is located a half mile southeast of Dunlap Lake at an elevation of forty-one hundred feet on the trail up the ridge north of Gold Butte. The prospect is in al-

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1. Hodge, E. T.; Framework of the Cascade Mountains in Oregon: Pan. Am. Geol., vol.49, No. 5; p.346, 1928.

tered tuff apparently (the tunnel is caved) along a vein containing some vein quartz , mainly of the cryptocrystalline variety. Bleaching and reddening of rocks is quite common in this general area. The mineralization may well be related in a general way to the fracturing. In the North Santiam River there is no noticeable mineralization, but two small faults and numerous slickensides were noted. The faults occur about seven tenths of a mile east of the McCoy Creek trail. There a block of an andesite flow fifty feet thick and two hundred feet long abuts against green tuff at both ends with north-south fault contacts. Movement was largely vertical.

The Minto Mountain and Sentinel Hills rhyolite, although well within the limits of the High Cascades, belongs structurally with the older rocks. The flows dip fourteen or fifteen degrees eastward in Minto Mountain and similarly in Sentinel Hills. In the sentinel Hills and intrusive mass of white rhyolite at least one hundred yards across implies that these lavas may have been very limited in extent. They are covered unconformably by the westward dipping Minto basalts.

The eastern margin of the Western Cascades is very problematical. In many respects it has the character of a linear eastward-facing scarp running almost due north and south, from Collawash Mountain northeast of Breitenbush Hot Springs, through Outerson Mountain and along the east side of Mt. Bruno. The scarp has everywhere been buried by the Pliocene-Pleistocene rocks, and is therefore accessible for investigation in a few places only. In Collawash Mountain the scarp is very steep, and can be rather easily traced. As shown in section F - F' (Plate II), it has the slope of a steep mountainside.

In the north side of Outerson Mountain the upper contact of the tuffs is obscured by landsliding, but where it can be localted it indicates a slope of about six hundred feet per mile. In Mt. Bruno the eastward face appears to be very gentle. (See section H - H', Plate II) That the scarp has been cut into by the present Santiam River south of the Elbow is indicated by the occurrence of old andesites along the base of Minto Mountain under the Pigeon lavas (See section H - H', Plate II).

Explanation of such a scarp by faulting would be rather simple, were it not for the high elevation of the buried rhyolite in Minto Mountain and the Sentinel Hills. This rhyolite is east of the scarp and presumably on the downthrow side of the fault yet occurs at elevations fully as great as the main skyline of the Western Cascades. Four alternate hypotheses may be advanced to explain the scarp character of the eastern edge of the andesite series and the occurrence of the rhyolites. They are;

1. The scarp is an eroded fault scarp, the down thrown side being to the east. The rhyolite is considered as isolated hills of Clarno age standing above the general surface of the downthrown block. The throw of the fault is at least half a mile.

2. The rhyolite was accumulated locally on the eastern edge of the range, which was then faulted down.

3. The rhyolite is part of the Western Cascade sequence, hence is not Eocene, and occurs as outliers separated by erosion from the main range. The scarp is purely an erosional feature.

4. The scarp is the result of erosion and thinning of the f formations toward the east. The rhyolite masses by this hypothesis are Eocene hills east of the mountain range.

Discussion:

1. If the rhyolite hills were Clarno dropped a half mile with referance to the present range, they must have stood a half mile above it before faulting. This would require a great thickness of Eocene, and some would be expected west of the scarp, underlying the younger andesites at fairly high altitudes. Such is not the case.

2. This hypothesis has much in its favor. It explains the linear character of the scarp and the presence of andesitic rocks under the rhyolites at Minto Mountain. It also explains the similarity in the dips in the rhyolites and the andesites along the Santiam south of the Elbow. The valley west of Minto Mountain would be essentially a fault line valley.

3. This differs from (2) only in that no fault is postulated. It seems rather difficult to conceive an erosional process that would completely remove all but two isolated remnants of the eastern margin of a range like the Western Cascades, and leave such a linear feature. This concept is especially difficult to sustain when it is noted that the rocks dip from fifteen to twenty degrees eastward and are of types rather resistant to erosion. It explains the similarity of structure of the rhyolites and andesites.

4. This hypothesis has the same objection as (3) in accounting for the scarp and the isolated character of the rhyolite exposures. A further objection is the lack of evidence that the formations do thin appreciably in this vicinity, although they perforce must thin somewhere to the east. In addition, it would be a remarkable coincidence that the dips of the Eocene and later rocks are so similar when the formations are probably separated by a nonconformity.

Hypothesis 4 seems untenable from almost any angle. No. 3 has less serious objections, but fails to account plausibly for the scarp and the disappearance of the older rocks east of the scarp. It does explain the structure of the rhyolite hills. No.1 is objectionable for the reasons stated. No. 2 seems most reasonable. It explains the nature of the scarp, the similarity of the structure of the rhyolite and the andesites of the older range, and the general absence of older rocks east of the scarp. Furthermore, the Outerson volcanic center is located about where the fault would pass. According to this hypothesis, the rhyolites may be Oligocene or Miocene, but cannot be Eocene. Burial of the critical contacts under the Pigeon basalts and glacial debris at the western base of Minto Mountain precludes definite determination of the actual structure even with exhaustive study in this area; investigation elsewhere along the scarp may yield more positive information.

The High Cascades

The High Cascades appear to be a rather simple accumulation of lavas which can be separated structurally into three or four main groups. As above enumerated, these groups are the Outerson, Battle Ax, Minto, Pigeon, and possibly later Olallie (and Mt. Jefferson) lavas.

The Outerson lavas were poured out from sources located near, or on the east scarp of the Western Cascades. For this reason the main dips of this series are eastward, at angles varying from two or three to twenty-five degrees, depending in part on the steepness of the underlyingportion of the scarp. That the main eruptive center ~~eruptive center~~ was just east of Outerson Mountain is shown by the presence of a typical vent complex of intrusive plugs and dikes, very steep and irregular dips, and glassy flows

That a minor center occupied the site of Mt. Bruno is indicated by thirty degree dips in agglomerates in the western ridge. The source of the thick series of gently east dipping flows in Coffin and Bachelor Mountains is unknown, but can not have been very far away.

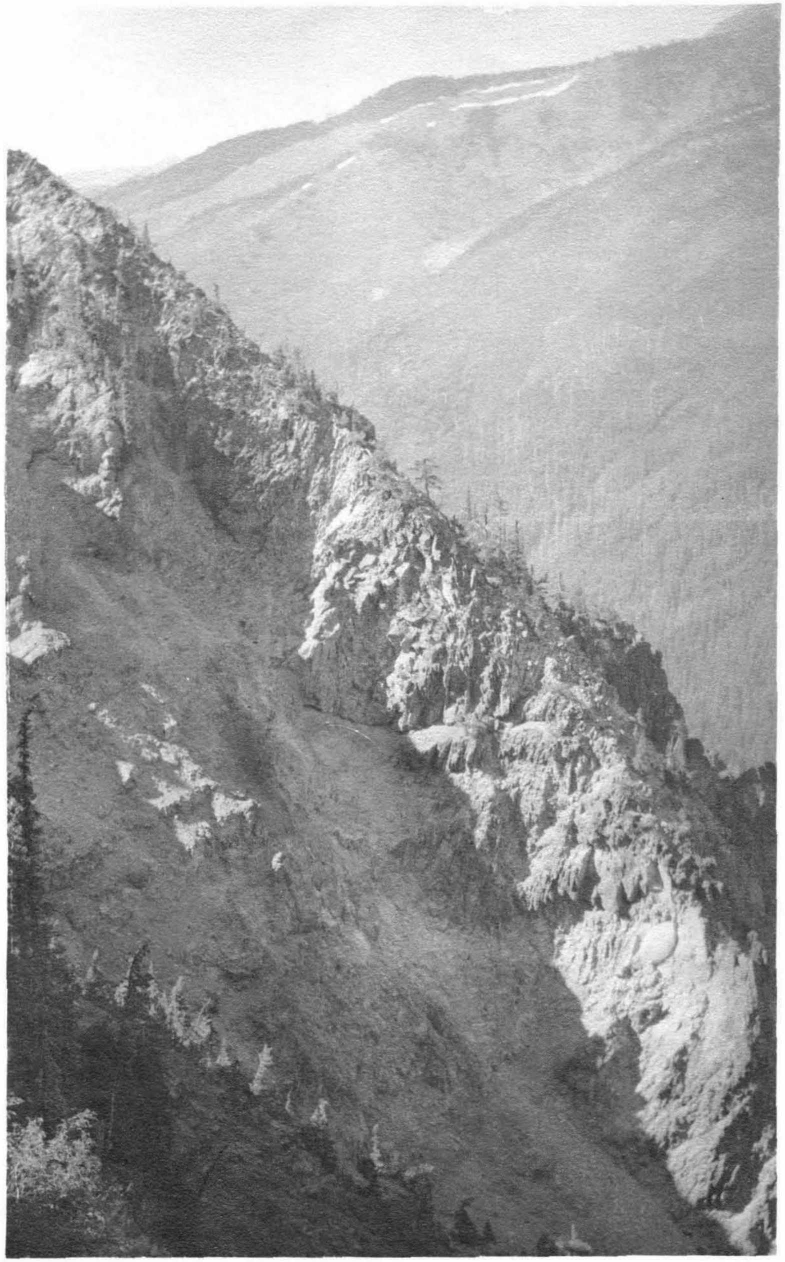
The Battle Ax lavas were extravasated over a mature topography from a single volcano. The flows near the vent dip twenty to thirty degrees but flatten rapidly away from it. The flows at some distance from the vent are very thick, three hundred feet or more, and have the dip of the underlying surface of the old rocks. The parent plug was of the same type as the Minto plugs. The thick flows are very platy near the top and rudely or perfectly columnar below. Battle Ax seems to have been the southernmost cone of a chain of volcanoes which extended northward in a direct line toward Mt. Hood. This line appears to closely follow the western edge of the belt of high-angle dips.

The Minto lavas dip east and west from the crest of the High Cascades. Near the crest the dips average five or six degrees; they are practically flat where the lavas lap over the Outerson or older rocks , as in Bald Butte. The lavas flowed around the accumulation of Outerson materials giving rise to topography such as that found in Woodpecker Ridge and the ridge southeast of Cheat Creek(See Plate II). Probably Minto Mountain was such an obstacle that the valley west of it was not completely filled; these Minto lavas did not apparently enter the Santiam valley west of the Elbow, because they were blocked by the Outerson deposits which were con tinuous across the present val-

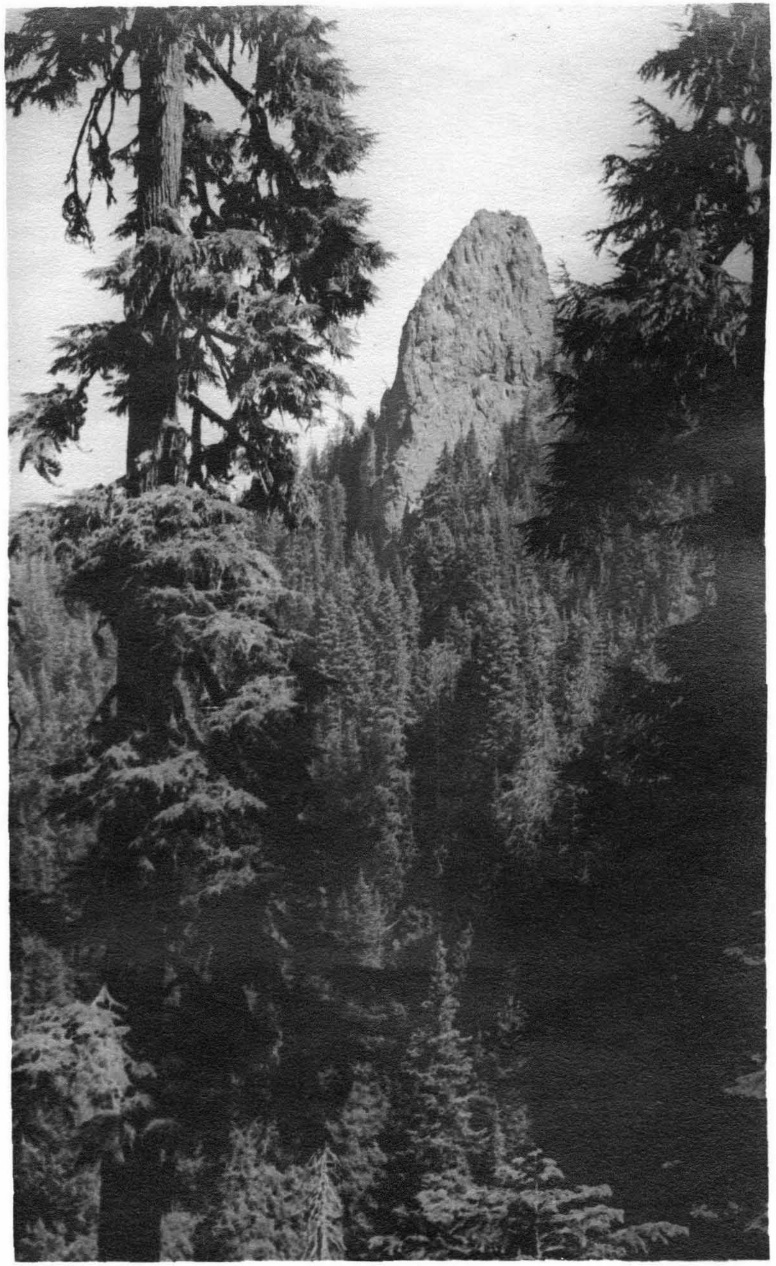
PLATE 8

A. View of basalt flows in the east side of Grizzly Peak. The pinnacles at the lower right are part of the intrusive mass from which the flows issued. The rock at the extreme upper left is also a fine facies of the intrusive mass. The slopes of Mt. Jefferson are visible in the background across the valley of Hunt's Creek.

B. Spire Rock, one of the intrusive plugs of the Outerson Complex, as viewed from the trail to the southwest.



A.



B.

PLATE 9

View south from Outerson Mountain up the North Santiam River valley. The truncated spur ridges of the High Cascades are visible to the left; the central flat is the Pigeon Prairie surface, to the right of which lies the present stream valley of the North Santiam River. The eastern face of Mt. Bruno crosses the western (right) half of the picture.

PLATE 9.

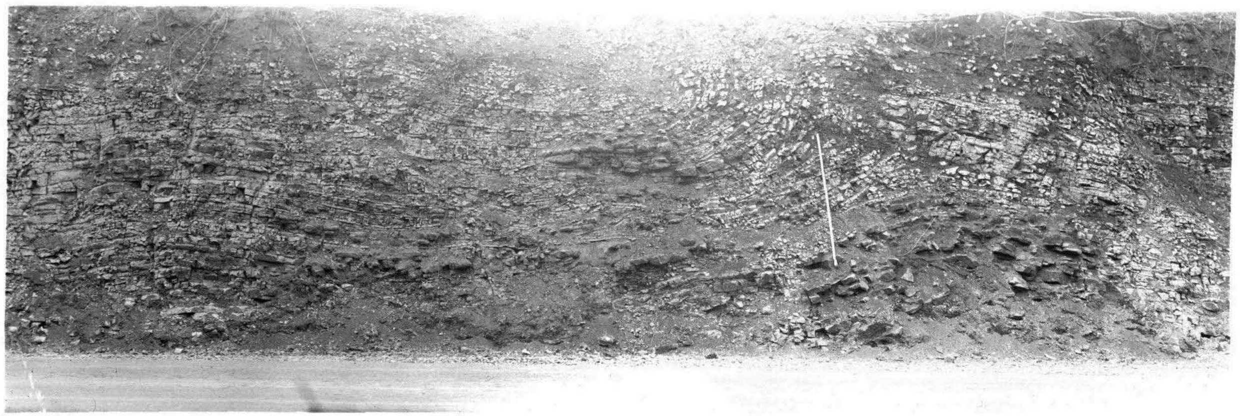
PLATE 10

A. The North Santiam River valley at the Elbow, viewed from Minto Mountain. The tip of Mt. Bruno is visible at the extreme left; the present youthful valley of the river bending around it. The broad surface of the Pigeon Basalts, cut across by the valley of Whitewater Creek, is seen to merge with the slopes above. The higher peaks to the right are the high-standing Outerson deposits; the western end of Woodpecker Ridge is visible in the lower right.

B. Platy andesite on the road cut one half mile north of Stayton. The rule is three feet long.

PLATE 10

A.



B.

ley of Whitewater Creek. The structure of the lavas is easily discerned because of the manner in which the thicker flows form ribs along the sides of the glaciated valleys. Very thick flows occur in the Minto series but are not common. Most of the lavas are very thin. In a fifteen hundred foot exposure in the western end of Bald Butte there are at least fifty flows and half of the section is flow breccia. Where the Minto lavas lie on the thin bedded lavas of the Outerson series, the two series are almost impossible to distinguish because of the minor irregularites inherent in such deposits and the petrographic identity of the flows.

The sources of the Minto Lavas were plugs of the type occurring in Park Butte and the east end of Lizard Ridge. At the latter place the rocks can be traced continuously from a light-colored granular micro-gabbro to the normal porphyritic gray flow basalt. The flows c an be seen issuing from the parent mass. (See Plate VIII A) The valley of Hunts Creek is cut through the plug so only the finer grained shell, as it were, is left. The Park Butte plug is roughly circular and about one-third mile in diameter. The prominence of the butte resulted from its origin. The Lizard Ridge intrusive appears to be linear , about a mile long parallel to the line of Hunts Creek, and does not crop out on the east side of the valley. There may be two plugs close together; one just east of the low pinnacle south of Grizzly Peak, the other including the spectacular lava capped pinnacles west of Hunts Cove. Hodge[1] reported Miocene lavas cut by the later

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1 Hodge, E. T.; The Geology of Mt. Jefferson. Mazama, vol. 7, No. 2, p. 32, 1925.

feeder dikes in the western side of Hunts Creek valley. These were said to be overlain by the Pliocene lavas (Minto of this paper), which filled valleys in the Miocene flows. The writer failed to find evidence of the unconformity and considers the entire section to belong to the Minto series. The structure of the Minto lavas, except for local peaks like Mt. Jefferson and Olallie Butte controls the physiography of the High Cascades.

The Pigeon Basalts are merely an accumulation of lavas in the North Santiam River Valley. As shown in section G - G'(Plate II) the valley they filled was eroded partly in Outerson, partly in pre-Outerson rocks, and was very broad in comparison with the present valley from the same point (Plate X A). South of the Elbow the valley was cut in rocks of all the earlier series, was probably not less than a mile and a half wide opposite Woodpecker Ridge and narrowed rapidly to the south. (See Plate IX) The lowest elevation these lavas reach now is a little less than two thousand feet, but they extended at least two or three miles farther west originally. They are therefore considerably younger than the Minto series, and indicate a renewal of vulcanism in the general vicinity of the postulated fault bounding the Western Cascades.

Regional Structural Relations

The Western Cascade structures in the North Santiam and Columbia River sections are very similar and in structural alignment. The northwestward projection of the Mehama anticline nearly coincides with the Eagle Creek anticline of the Columbia Gorge; furthermore, both folds are very broad and gentle. The similar projection of the Breitenbush anticline lies in the vicinity of the Ortley and Bingen anticlines. The two last-named folds are sharp, but smaller than the Breitenbush structure and, taken together may be regarded as its equivalent. The southwestward projection of the Mehama anticline crosses the South Santiam River not far from the steep structures east of Cascadia. This structural alignment of the three sections strongly suggests that the folds of the Western Cascades are persistant for long distances, and cut diagonally across the main south-north trend of the Western Cascades.

Whereas the Western Cascade folds trend northeast-southwest, the veins and associated dioritic dikes of the entire Western Cascade system in Oregon strike northwest. This discordance of trends, the northward plunge of the main folds, and the Turner warp suggest that the period of dioritic intrusion and associated mineralization was not contemporaneous with the main folding but followed minor warping and fracturing along northwest-southeast axes. This implies a ninety degree rotation in the direction of application of compressive forces, between the periods of folding with concomitant weakening of forces. The contact relations of the Halls diorite as mentioned above indicate emplacement by stoping and assimilation and suggest that adjustment of compressive forces was nearly if not quite complete before intrusion occurred. Since the intruded rocks are considered as Middle Miocene in age, the folding and intrusion are believed to have occurred during Upper Miocene time, probably toward the end of the Miocene period, simultaneously with the Miocene orogeny of the Washington Cascades.[1]

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1 Smith, G. O.; Geology and Physiography of Central Washington. U. S. Geol. Sur. Prof. Pap. 19, p. 22, 1903.

Faulting of the Basin Range type has been described in the Klamath Lake region by Gilbert[1], and the scarp south of Crater Lake is commonly regarded as a fault scarp. It seems not unlikely therefore that faulting of this typemay determine the distribution of the High Cascade vents. The location of the Outerson and Pigeon vents along the Cascade scarp is a case in point.

Close relation of volcanic activity and faulting has been noted in many parts of the world. The linear distribution of the High Cascade volcanic centers is very suggestive of such a relation. The main peaks of post-Minto age as well as the Minto plugs are restricted to a narrow belt along the crest of the High Cascades. From Olallie Butte south this narrow belt is almost a single line which is paralleled by the Cascade fault. Toward the north the belt widens and distribution of eruptive centers is more irregular, implying weakening of the structural control. This implication is borne out by the general absence of fractures in the Columbia Gorge.

This system of Basin Range type fractures diagonally intersects the northeast trending Western Cascade folds, completely cutting them off south of the Breitenbush River. The change of tectonic activity from folding, mainly along northeast trending axes to normal faulting along a north-south axis implies a time interval of some magnitude. It is suggested that the Cascade Scarp was formed in Pliocene time after the range produced by the late Miocene diastrophism had been considerably reduced by erosion.

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[1] Gilbert, G. K.; Studies of Basin Range Structure; U.S.G.S. Prof. Pap. 153, pp. 76 - 85, 1928.

The last uplift along the Cascade fault occurred in pre-Outerson time and probably tilted the entire region to the west as a single block. The arch type of uplift described by Smith and Willis in the Washington Cascades[1] does not appear to have occurred in the Cascade Range in Oregon. If such upwarping has occurred the evidence is very obscure. The Western Cascades exhibit few signs of renewal of uplift in Pleistocene time such as is outlined by Willis in Washington.[2]

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1 Smith, G. O. and Willis, Bailey,;Contributions to the Geology of Washington. U. S. Geol. Survey Prof. Paper 19, pp. 39,85-86, 1903.

2 Idem, p. 92.

CHAPTER IV

PETROGRAPHY

The general andesitic nature of the lavas in the Cascade Range has been recognized for sometime, probably largely as a result of Diller and Patton's monograph on Crater Lake.[1] It should be noted, however, that this investigation did not include any of the older Western Cascade lavas of the Rogue River valley, and hence is representative of the High Cascades only. Aside from the Crater Lake district detailed study of the Range has been limited to the structure of the Columbia Gorge section, having been of secondary interest. The post-Columbia River Basalt lavas of the Cascade have usually been dismissed as andesitic. Detailed petrographic studies in the Mt. Hood region by Dr. E.T. Hodge and his assistants are as yet unpublished.

Callaghan[2] was the first investigator to systematically discuss the petrography of the Western Cascade lavas as distinct from those of the High Cascades. He pointed out their variation from high calcium andesites or low pyroxene basalts to rhyolites, and in a general way indicated the distribution of the various types. He also, for the first time, described the nature of the later intrusives in the lavas and pointed out their chemical similarity to the Miocene intrusives of the Washington Cascades. He only briefly touched upon the High Cascade lavas, but indicated their basaltic nature.

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1. Diller, J.S. and Patton, H.B.; Geology and petrography of Crater Lake Nat'l Park: U.S.G.S. Prof. Paper, No. 3, 1902
2. Callaghan, E., op. cit., p. 243

In this paper the Western Cascade and High Cascade lavas are discussed separately because of the homogeneity of each of the groups, their marked age differences and contrast of field appearance.

Illahe, Mehama and Breitenbush Series

A detailed investigation of the sediments of the Illahe and Mehama formations was not made. However, some microscopic and hand specimen studies indicate the general volcanic nature of the Illahe sediments. They consist essentially of plagioclase, quartz and small rock fragments with pyroxene and amphibole in minute amounts and occasional mica flakes, all in a fine matrix containing some carbonate. The rock fragments appear to be andesitic or basaltic. In the coarser beds rhyolite pebbles are not rare.

The Mehama formation also appears to be dominantly andesitic. The agglomerates in Thomas Creek and east of Mehama are petrologically very similar to those occurring in the upper Sardine Series. The tuffs are more acid and in places are rhyolitic. Most of the tuffs are comparatively soft, but tuff forming cherty nodular masses containing plant stems occurs in the Thomas Creek drainage over the divide southwest of Lyons. This material is full of glass shards.

The Breitenbush series is very similar to the Illahe and Mehama in composition. The waterlaid sediments are composed of andesitic debris, and the flows are andesites, (Plate XI A) excepting the rhyolites at the base. Most of the true tuffs higher in the series (Plate XI B) are more acid than the flows, but still andesitic, although a small lense of rhyolitic tuff was observed in the Breitenbush River a mile east of Canyon Creek.

The Oligocene seems to have been a period of transition from the rhyolitic eruptions of the upper Clarno Eocene to the andesitic and basaltic eruptions of the Miocene. It is only natural that the Illahe and Mehama formations should show the same characteristics as the Breitenbush, because they were probably in large part derived from the latter, or deposited simultaneously from the same sources. The location of the centers of eruption is as yet unknown.

Miocene Basalts and Andesites

The division of the chief Western Cascade lavas into the black and dark gray varieties, as suggested by Callaghan,[1] is well exemplified in the North Santiam section. The Stayton lavas are the "black lavas" of Callaghan, whereas the "gray lavas" include the Sardine effusives.

The Stayton Lavas:

The Stayton lavas are dominantly black, aphanitic to finely porphyritic rocks and according to field classifications are basalts. The porphyritic varieties, which only rarely contain phenocrysts more than a millimeter long, are commonly somewhat porous, where not vesicular. The finer flows are in many cases glassy and characterized by blocky structure.

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1. Callaghan; op. cit., p. 243

A few medium to light gray, andesitic appearing flows occur in this series, especially east of Mill Creek Gap. These light gray lavas are very porous and contain small yellowish secondary carbonate crystals in the porous spaces. At Stayton the series also contains a thick very platy dark gray andesite flow. (See Plate 10 B)

In the Salem Hills, and as far east as Shaw, the lavas weather to bright red soil which is in most places very deep. East of Shaw, on the other hand, where the lavas are practically bare, the soil is black and powdery. Just why this change in weathering takes place is not clear. Evidently the black soil is a chemically unaltered basalt powder while the red soil is clearly the product of complete chemical weathering.

In thin section the black basalts show a loose diabasic texture of labradorite feldspar laths, rarely as much as a centimeter, but usually about 0.25 mm long, and granular augite in a glass groundmass. The ratio of phenocrysts to groundmass varies from about 1:3 to 4:1. The labradorite shows little zoning. The augite is of the normal diopsidic variety. The feldspar-augite ratio is large, except in the very glassy rocks, where it may be nearly 1:1. In the more completely crystallized specimens augite does not usually constitute more than fifteen or twenty per cent of the rock. Magnetite commonly occurs as irregular blebs or dendrites, amounting to two or three percent of the slide; it does not seem to contain large amounts of titanium. Small siderite grains and brownish yellow chlorite are common as weathering products. The black color of these rocks is due to the black of deep brown glass base, rather than to ferromagnesian material. The very glassy types are fully as dark as the coarser ones. The black lavas according to the above data are classified as otherwise normal basalts which have a low magnesium and iron content.

The gray lavas are almost identical in composition with the black basalts. They are very porous and the feldspars are more loosely arranged. The feldspar is labradorite, rarely present in two generations. Augite is the phenocryst pyroxene, while minute needles and drops of pyroxene too small to be determined occur abundantly in the glass base. Colorless glass constitutes from ten to fifteen percent of the rock. Siderite is a common secondary constituent of these lavas, probably because their porosity gives free access to weathering agents. Their light color is due to their porosity, and the transparency of the glass base.

Some definitely andesitic flows occur in the Stayton lavas. The platy flow at Stayton is a hypersthene bearing augite andesite. In it phenocrysts of augite, hypersthene and calcic andesine occur in an exceedingly fine groundmass of sodic andesine. Augite in medium sized to small rounded grains is the dominant pyroxene, but scattered hypersthene prisms are present. Much fine magnetite and pyroxene also occur in the groundmass. In a very similar, but more glassy porphyritic flow on the summit of Prospect Mountain (south of Thomas Creek) hypersthene is also common as phenocrysts with augite and zoned plagioclase. Hypersthene was not observed in any other Stayton flows.

Sardine Series

The lavas of the Sardine series are dark gray, coarsely porphyritic to aphanitic, and are included in the gray lavas of Callaghan. They differ from the Stayton lavas in that they very rarely, if ever, show good diabasic texture, but are characterized by glomeroporphyritic texture; namely, aggregations of feldspar and pyroxene grains in a fine groundmass in most cases showing good flow structure.

The phenocryst feldspars vary from sodic bytownite ($Ab_1 An_3$) to medium labradorite ($Ab_2 An_3$) and in rare cases attain lengths of a centimeter. The average size is not over one or two millimeters. Zoning from calcic to sodic labradorite is not uncommon. The groundmass feldspar is more acid than the phenocrysts in any given section, but is for the most part sodic labradorite, or, in a few cases, calcic andesine. Plagioclase constitutes over seventy-five percent of most of the sections studied.

Primary ferromagnesian minerals are augite, hypersthene and olivene. Augite is the predominant phenocryst pyroxene, while hypersthene occurs sparingly, especially in the upper flows of Sardine Mountain. Occasional pseudo-

PLATE 11

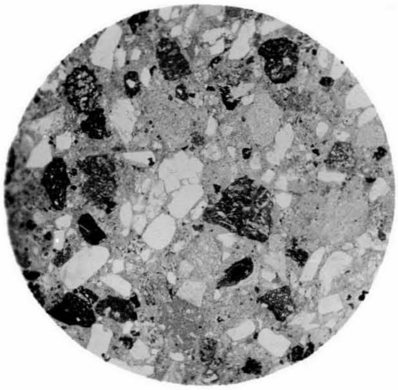
A. Waterlaid green tuff of the Illahe Formation, 3.7 miles east of Detroit on the Santiam Highway. The variety of fragments is apparent. Plain light, 26x, S. 103

B. Fine green tuff from same locality as A. The eruptive nature of this specimen is proved by the abundance of delicate glass shards. Plain light, 26x, S. 104

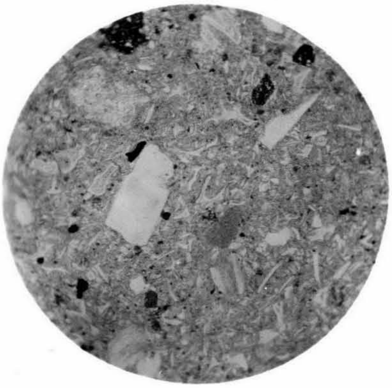
C. Altered dacite of the Sardine Series showing porphyritic texture. A devitrified zone surrounds the partially resorbed quartz phenocryst in the center. Crossed nicols, 26x. Sp. S. 111.

D. Chlorite pseudomorphic after olivine in a dike rock from the Sardine Series. Kernels of olivine may be seen. A reaction rim of granular augite surrounds the chlorite. Plain light, 62x, Sp. S. 72.

PLATE II



A.



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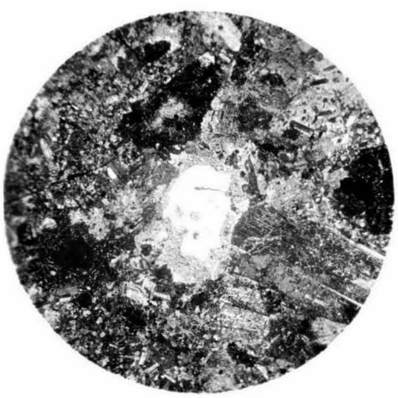

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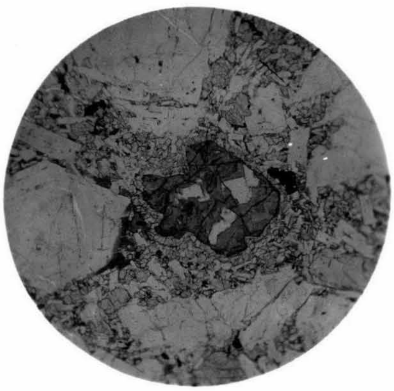

D.

morphs of chlorite, carbonate and quartz after olivine were noted in a few slides. The groundmass pyroxene in all sections occurs as minute prisms and droplets; it may be pigeonite, but was too fine to be determined. The total average content of ferromagnesian minerals is less than twenty-five percent.

Primary quartz (Plate XIC) was observed only in the lavas near the Halls diorite plug. The quartz occurs as comparatively large phenocrysts which have been partially resorbed and fractured. Secondary quartz is present in many flows. Glass, where present in the base is colorless or very light brown, and in small amounts.

The Sardine lavas show many of the effects of weathering or incipient alteration. The ferromagnesian minerals are very commonly partially or entirely chloritized and the feldspars show incipient kaolinization. No olivine, as such, was found in the flows.

The large dikes in the Little North Santiam River are identical in texture and composition with the flows, except that they are somewhat coarser, and contain much olivine in several instances. The olivine remains only as kernels surrounded by chloritic material, which in turn is rimmed with fine granular augite. (See Plate XID) Undoubtedly the augite rims were formed around olivine phenocrysts as a result of magmatic reaction,[1] while the olivine was chloritized after solidification of the rock. Olivine grains and pseudomorphs were observed in all stages from almost perfect crystals to practically completely resorbed grains. Very commonly all the

1. Bowen, N.L.; The Evolution of the Igneous Rocks: Princeton Univ. Press, pp. 54-62, 1928

olivine is gone, leaving a chlorite pseudomorph rimmed by augite. The small dikes are similar to the normal Sardine flows.

The Sardine lavas have the character of normal flow basalts except that the feldspar-pyroxene ratio is larger than in the latter. The low percentage of pyroxene (15 - 25%) indicates a low iron magnesium content. This is borne out by chemical analysis.[1] The rocks may be considered low-augite basalts or high calcium (labradorite) andesites. Where primary quartz is present, they border on dacitic types, with higher silica and soda content.

Rhyolites

The rhyolites found in this section of the Cascades show all the characters of such rocks. They are pinkish to white, porphyritic, and show good flow structure. (Plate XIIA) In the Sentinel Hills section many are lithoidal or glassy and contain lithophysae.

Pyroxenes are absent, having been replaced by chloritic material. Glass is the main constituent, and commonly forms ninety percent of the slides. It is clouded with submicroscopic dust particles, and in most cases partially devitrified. Some of the flows contain numerous inclusions: one flow in the Santiam section was observed to contain obsidian stringers over six inches in length. An obsidian flow occurs in the Sentinel Hills section. The rhyolites of the various localities are very similar though possibly of

1. Callaghan, E. op. cit., p. 246

different ages.

Halls Diorite

The intrusive mass at Halls, southwest of Sardine Mountain, is a little over a mile across, and is roughly rectangular in outline. The rock varies in appearance from a fine grained purplish gray, porphyritic rock at the edge, to a medium grained almost white porphyritic rock in the interior. There seems to be no definite textural gradation from center to edge except at the very edge of the mass. Small aplitic dikes cut irregularly through the intrusive body.

In thin section the dark border phase was found to consist of zoned phenocrysts of andesine in a groundmass of granular feldspar, quartz and accessory minerals. Biotite is common but not abundant as minute flakes, while orthoclase occurs as interstitial material in the groundmass. Orthoclase also replaces the plagioclase phenocrysts, in places almost completely. Quartz is abundant in the groundmass, but probably does not constitute more than five percent of the rock.

The coarsest phase of the rock, (Plate XIIB) exposed in Cumley Creek, contains large hornblendes and phenocrysts of sodic andesine (Ab65 An35), more or less replaced by orthoclase. A few small orthoclase grains occur in the groundmass with quartz and plagioclase, as well as interstitial orthoclase. Epidote is scattered irregularly through the rock as replacements in plagioclase and is probably secondary.

Smaller offshoots of the main mass, such as the dikes west of Mayflower Creek, differ from the latter mainly in ratio of phenocrysts to groundmass, and ferromagnesian minerals. Augite appears to have been the original pyribole, but is now almost completely altered to green hornblende or chlorite (penninite). Orthoclase partially replaces the andesine phenocrysts and is present as interstitial groundmass material. Epidote is abundant, and in places is replaced by quartz. In one slide myrmekite was found present in great abundance.

Although the ferromagnesian minerals vary from place to place, the mineral composition of the plug is quite uniform. Quartz probably does not amount to over five percent of the rock, nor does orthoclase. According to Johannsen's classification[1] the rock is a tonalite, or possibly quartz diorite.

Where intruded by the diorite or small dioritic stringers, the lavas were epidotized and chloritized, the pyroxenes and calcic plagioclases being attacked first. With more intense alteration quartz, sericite, pyrite and hematite were introduced. In hand specimen the main effect appears to have been bleaching which has not destroyed the original texture of the rock. In one slide of a highly altered facies the feldspars have been almost completely replaced by chlorite and sericite, and aggregates of quartz and epidote with small amounts of sericite and sodic plagioclase. The two mineral assemblages seem to be very distinct. The glassy base of the

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1. Johannsen, A.; A descriptive petrography of the igneous rocks: vol. 1, p. 155. Univ. of Chicago Press, 1931

PLATE 12

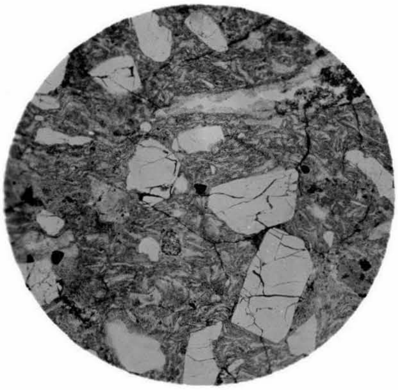
A. Flow rhyolite occurring near McCoy Creek. The phenocrysts are albitá-oligoclase, the fractures containing hematite. Flow structure is visible in the groundmass. Plain light 26x, Sp. S. 113.

B. The coarse facies of the Halls diorite porphyry. Large phenocrysts of twinned hornblende and andesine lie in a groundmass of quartz, plagioclase and orthoclase. Crossed nicols, 26x, Sp. S147.

C. Pseudomorph of sericite after plagioclase in altered tuff, now composed of sericite, hematite and pyrite. From mineralized zone just west of Mayflower Creek. Crossed nicols, 62x. Sp. S. 90

D. Glassy basalt from the Outerson series containing phenocrysts of labradorite and augite in black glass base. The sharp white and black line cutting the largest feldspar is a fracture in the section. Plain light, 26x, Sp. S. 145.

PLATE 12



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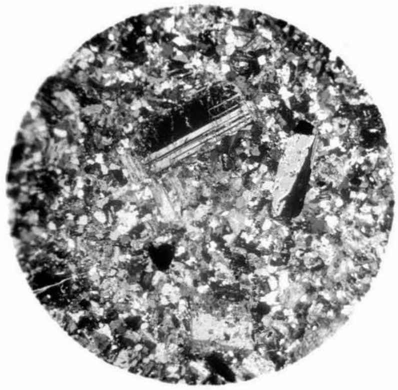

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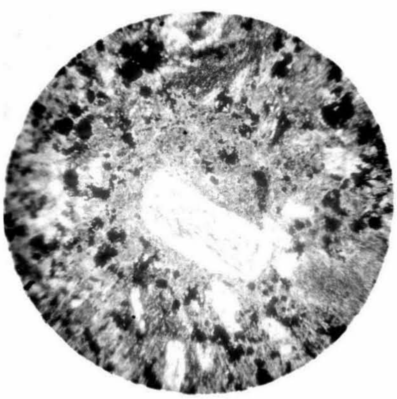
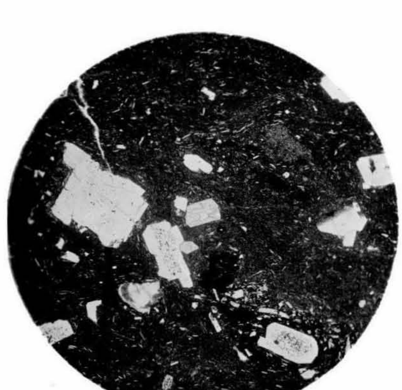


C.



D.

flows was greatly devitrified (Plate XIC) and in many cases partially replaced by small patches of granular secondary quartz. The most intense mineralization produced a rock composed entirely of sericite, pyrite and hematite. (Plate XIIC) A veinlike mass of this rock occurs in association with two large diorite dikes west of Mayflower Creek.

HIGH CASCADE LAVAS

The High Cascade lavas are characteristically light gray, very fresh porous rocks. They commonly contain much olivine. Some of the series are so similar petrographically that separate descriptions of them would be needless repetitions. Thus, the Pigeon basalts are almost identical with the bulk of the Minto flows, and cannot be distinguished from the latter in hand specimen. Generally speaking, there is more variety in the Outerson series and in the later lavas on the summit highland than in any other part of the Pliocene-Pleistocene section.

Outerson Series

The Outerson lavas are basalts ranging in color from black to the normal light gray of the young basalts. The textures vary accordingly from hyalopilitic to loose diabasic or parallel. The finer rocks are denser, show good flow structure, and are largely glass. The coarser rocks are generally lighter colored, more porous, and may or may not show good flowage.

The mineralogy of the lavas is very simple. The feldspars range from bytownite (Ab15 An85) to sodic labradorite (Ab50 An50). Olivine occurs in almost all the flows, is common in most of them, and abundant in many. Hypersthene is much less common than olivine, and when present occurs as phenocrysts. Augite is the dominant pyroxene: it occurs as phenocrysts and in the groundmass. In the extremely glassy rocks labradorite and augite are the only minerals distinguishable. (Plate XIID) In the coarser rocks light colored glass is present only in small amounts.

Battle Ax Series

The Battle Ax lavas are all light gray, porphyritic basalts. Very few of the flows show columnar jointing, but are either massive or platy. The thick flows are very platy. These rocks are so similar to the Minto lavas that they were not studied in much detail.

A thin section of the very thick flow in the top of Gold Butte shows two distinct generations of minerals. The largest phenocrysts, bytownite, are 2.5 mm long; the average size is $\frac{3}{4}$ mm. The groundmass minerals average .08 mm in size. The bytownite grains are surrounded by very narrow sharp rims of andesine, indicating that the groundmass feldspars are also andesine. The dominant pyroxene is hypersthene, which is the only phenocryst pyribole, and is abundant in the groundmass with augite. Large streaky pseudomorphs of magnetite and granular augite (?) have replaced the largest hypersthenes. (Plate XIVA) This flow is noteworthy in that it contains an abundance of tridymite filling minute porous spaces. Small crystals can be distinguished in the hand specimen with the aid of a lens. In thin section some of it shows pseudohexagonal twinning. The mineral composition of the rock and the abundance of tridymite make it unique.

Minto Series

As mentioned above, the Minto series is very homogeneous except in the upper part, which probably comprises effusives of a later date. The Minto series is especially interesting however, for the opportunity it affords for tracing flows outward from the actual center of eruption. This is possible in the east end of Lizard Ridge.

The coarse intrusive is a white granular rock containing abundant honey-colored pyroxenes. In thin section two generations of minerals are evident; the average feldspar sizes of the two generations are 0.5 and 0.3 mm respectively. The pyroxenes are 1.2 and 0.2 mm. The texture is that of a fine gabbro. (Plate XIIIA) The feldspar is labradorite, the phenocrysts being more calcic than the groundmass crystals.

Augite occurs as poikilitic plates molded on the feldspar, whereas hypersthene forms medium sized to small subhedral prisms, some enclosing feldspar drops. A few apatite needles were observed, also tridymite as interstitial material between feldspars. (Plate XIIID) The mineral composition of the rock is as follows: labradorite 74.5%, hypersthene 13.25%, augite 9.75%, magnetite 2%, apatite and tridonite less than 0.5%.

The finer intrusive facies shows two generations, but the contrast in grain sizes is much greater. (Plate XIIIB)

The first generation feldspars (Ab3 An7) range from 0.8 to 1.0 mm in length. The groundmass plagioclase (Ab1 An1) averages 0.25 mm. Hypersthene takes the place of augite as the phenocryst pyroxene, forming similar poikilitic plates. It is associated with olivine, and may be partly a reaction product. Olivine occurs as much resorbed grains rimmed with granular augite, enclosed as kernels in poikilitic hypersthenes or, rarely, poikilitic augite. Most of the augite is in the form of small grains between the feldspars. A few grains of tridymite occur as a filling in minute porous spaces. The mineral composition is: plagioclase 71.4%; pyroxene 19.6%; (Aug:hyp is about 2:1) olivine 6.6%; magnetite 2.5%. The rock because of its fine texture and gabbroid composition is classified as a micro-gabbro.

The flow equivalent of the intrusive is a platy light gray basalt containing small reddish pyriboles in an aphanitic groundmass. At first glance, under the microscope there are two generations of minerals, but actually all gradations in size occur between the finest groundmass and the largest phenocrysts. (Plate XIIIC) The maximum phenocryst size is 1.4 mm, the groundmass size averages about .08 mm. The plagioclase is medium labradorite (Ab2 An3) the largest grains being finely zoned within the labradorite range, and much scalloped by resorption. Olivine is abundant as large grains surrounded by granular augite; some is enclosed in plagioclase. Hypersthene is the dominant pyroxene, occuring as subhedral poikilitic plates, commonly associated with

PLATE 13

A. Coarse facies of Grizzly Peak intrusive, showing gabbroid texture of the rock. Labradorite, hypersthene and augite are the principal minerals present. Crossed nicols, 26x. Sp. S. 129

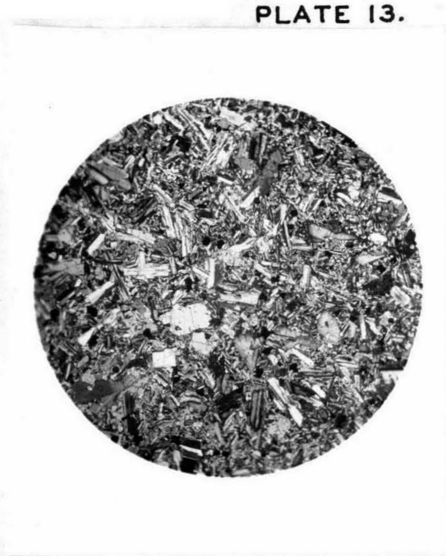
B. Border facies of Grizzly Peak intrusive showing two generations of minerals. An olivine phenocryst lies below and to the left of center. Crossed nicols, 26x. Sp. S. 130.

C. Platy flow basalt facies near the Grizzly Peak intrusive. Porphyritic texture is very pronounced, the center phenocryst being olivine. Crossed nicols, 26x. Sp. S. 131.

D. Interstitial tridymite between feldspars in the coarse facies of the Grizzly Peak intrusive. Both augite and hypersthene are present. Plain light, 62x. Sp. S. 129.

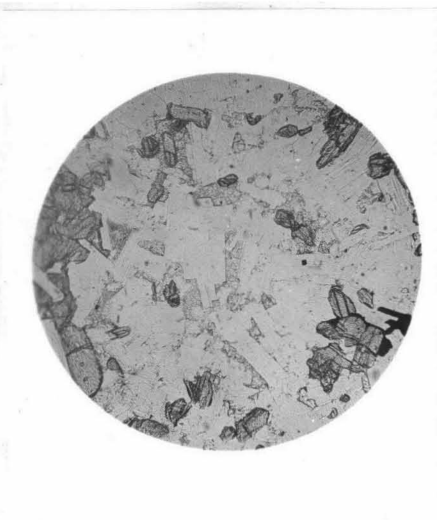
PLATE 13.

A.



B.

C.



D.

olivine and magnetite. Augite is present mainly as fine stout prisms evenly distributed in the groundmass or as rims surrounding olivine, but a few very ragged poikilitic plates were observed. The augite seems to have been the last mineral to crystallize. This rock is almost identical with the Pigeon Basalts, which will not be separately discussed.

The most interesting feature of this sequence, aside from the textural variation, is the change of augite from the phenocryst pyroxene in the coarse intrusive to the groundmass pyroxene in the flow facies. It appears to be explained by some resorption-recrystallization process. The absence of olivine in the coarse rock, and its comparative abundance in the flow is probably to be explained by the reaction principle as set forth by Bowen.[1] In the rapidly chilled flow rock complete equilibrium of the magmatic liquid and crystals floating in it was not established before solidification, with the result that complete resorption of the olivine and calcic labradorite did not take place. Hence the resorbed olivines and zoned feldspars which are present. In the plug, which cooled slowly, equilibrium was established, the olivine being completely resorbed, and the feldspar being essentially unzoned. The presence of free silica (tridymite) in the plug implies that although the Minto basalts contain olivine, they are saturated with respect to silica. The occurrence of tridymite in this plug and in the thick Battle Ax flow suggests that slow cooling of a large mass is necessary for the separation of tri-

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1. Bowen, N. L.; op. cit. p. 54-62

dymite from the magmatic liquid. The absence of even opaline silica in the thin flows bears this out. The presence of the tridymite of course indicates that the lavas and plugs were largely solid though still at relatively high temperatures when the silica crystallized. After formation of the tridymite cooling was probably rather rapid through the inversion temperature of tridymite to quartz, otherwise some inversion would be noted.[1]

The lavas on the High Cascades summit upland are of many varieties. Plagioclase ranging in composition from andesine (Ab3 An2) to labradorite (Ab3 An7), is the only essential mineral. Accessory minerals include augite, hypersthene, olivine and hornblende, in addition to the ubiquitous magnetite. Any two or three of the ferromagnesian minerals may be present in a given rock. Hornblende was not observed with olivine, but that is the only apparent limit on the number of possible combinations.

The commonest rock type is a medium gray basalt spotted with olivine and light colored plagioclase phenocrysts. (Plate XIVB) The plagioclase phenocrysts are calcic labradorite or bytownite, usually well zoned. Olivine is abundant as grains of all sizes in various stages of resorption. Hypersthene and augite also occur as phenocrysts. The groundmass usually shows good flow structure, and is composed of fine feldspar, pyroxenes, and light colored glass. This rock forms massive flows devoid of regular jointing, and very resistant to glacial erosion.

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1. Sosman, R. B.; Properties of silica: Am. Chem. Soc., Mon. Ser., No. 37, p. 76

The true andesites are light pink to white, commonly very porous, and contain glistening black hornblende or pyroxene crystals. The texture of these rocks is very porphyritic: the phenocrysts are sometimes almost a centimeter long while the groundmass is too fine to be determined, as shown in Plate XIVC. The plagioclase is calcic andesine showing some zoning. Hypersthene, augite and hornblende, green or basaltic, commonly occur in the same rock. In general the hypersthene prisms are very clear cut, even showing terminal faces. Only rarely does it show marked resorption effects. Augite, on the other hand, as a rule shows rounded or ragged outlines due to resorption. The outlines of many augites and hornblendes, particularly the green variety of the latter, are very similar. The chief difference in the occurrence of the two minerals is that the hornblende is always surrounded by a smudgy rim of magnetite dust, as in Plate XIVD. It is inferred that the hornblende originated by reaction of augite with the magmatic liquid in accordance with Bowen's discontinuous reaction series.[1]

The basaltic flows were apparently much more liquid than the andesites at the time of extrusion. Many of the basalt flows are persistant and can be recognized by their texture or color over comparatively wide areas. The andesites, on the other hand, are rather restricted areally, and tend to form plug-like masses. Pyramid Butte (Plate 36A) is such a plug. Double Peaks are formed of very massive andesite flows and a thick flow breccia. The very porous nature of the ande-

1. Bowen, N. L.; op. cit., p. 58 et seq.

PLATE 14

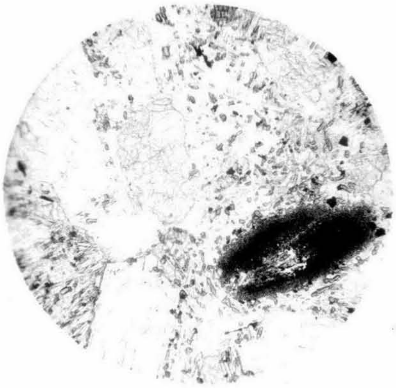
A. Tridymite, above and to the left of center, in the Battle Ax lava capping Gold Butte. The irregular black mass is a granular aggregate of augite and magnetite pseudomorphic after olivine (?). Plain light, 62x. Sp. S. 141.

B. Gray porphyritic olivine basalt flow west of Olallie Lake. Flow structure is evidenced by the orientation of the large gray olivine and twinned labradorite phenocrysts, some of the latter showing zonation. Crossed nicols, 26x. Sp. S. 149.

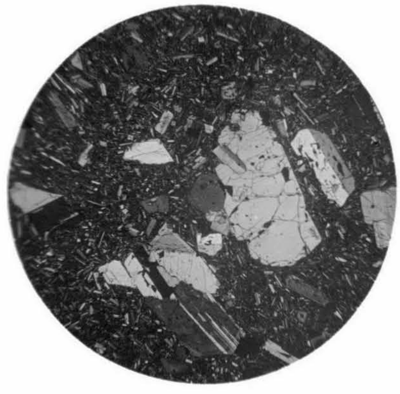
C. Relations of augite, hypersthene, hornblende and magnetite in andesite from Pyramid Butte. A clean cut hypersthene lies just below the center of the picture, and a resorbed augite grain about one half inch above. Just right of the center is a kernel of hornblende rimmed with magnetite. Plain light, 26x. Sp. S. 164.

D. Relations of augite, hornblende and magnetite in andesite from the Sentinel Hills. A magnetite rimmed hornblende lies in the center of the picture, below it lies an augite phenocryst showing some resorption. The large light area at the upper left is a hole in the slide. Plain light, 26x. Sp. S. 150.

PLATE 14.



A.



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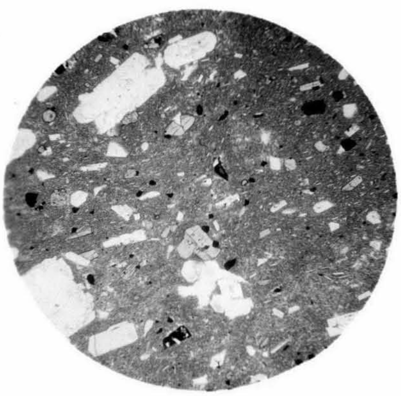


C.

D.

sites suggests a large content of volatile gases; presence of mineralizers is also indicated by the hornblende.

Summary

The Western Cascade lavas are dominantly labradorite andesites or low-pyroxene basalts. A few dacites occur, also olivine basalts. The tuffs are in the main andesitic. Two Rhyolite Series occur in the eastern part; one is probably Eocene, the other is later Tertiary and interbedded in the andesitic series.

The High Cascade lavas are very light gray basalts, usually containing olivine or hypersthene. Andesites occur in the upper part of the Minto Series, but were not observed elsewhere.

CHAPTER V

GLACIAL GEOLOGY

Three distinct periods of glaciation are represented in the North Santiam River basin. The three glaciations, from earliest to latest, are the Mill City, Detroit, and Tunnel Creek, and are named from the localities where their deposits are best exposed. They are correlated with the Sherwin, Tahoe, and Tioga, respectively, in the Sierra Nevada. Correlations with the Sierra glacial stages are based on comparative erosion of the moraines; destruction of glacial erosional forms, and on the fact that glacial stages, the glaciers in this district were all of the alpine type. Certainly the Wisconsin and probably the earlier glaciers also, headed on the western slopes of the High Cascades and moved westward down the walleys of the Western Cascades. During the earlier glaciations extensive ice streams undoubtedly also originated in the Western Cascades, but during the last glacial epoch they were confined to very small areas. The Mill City glacier trawersed two-thirds the width of the Western Cascades; the Wisconsin ice barely entered the eastern side.

The Mill City Glaciation:

The Mill City glaciation was much earlier than the others, and far more extensive. Till and lake silts are exposed along the river from Gates to a point almost three miles west of Mill City. The best till exposures are at Mill City, opposite the Hammond Lumber Company mill, and in the north bank of the river below the bridge. (See Plate VIIA). The vertical banks are exposed twenty-five feet of till covered with a thin veneer of river gravel. Part of the river bed is till, part old volcanics, but the till probably does not extend far below river level. The total thickness of the till is about two hundred feet. North of Mill City reddened morainic deposits are exposed at about 1000' A.T. The best vertical section is in a creek in section23, T.9 s., R.2E., where the till is almost continuously exposed for two hundred feet. The till rests on polished tuff in the river at 725', and the upper limit is a sloping bench about 500' wide at an elevation of 925'.

The till is medium gray in color, very compact, and similar to many continental moraines in fineness of matrix and polishing of boulders. It contains unweathered striated and facetted boulders of all sizes and representative of practically all rock types of the Santiam district. Because of its induration and habit of forming vertical exposures, the till has been mistaken for agglomerate by previous investigators, and at first was mapped as such by the writer.

Except in the road cut above Mill City, where the till has been weathered, the Mill City glacial deposits are very fresh. The Moraines west of Mill City also show the effects of erosion and weathering and in road cuts the matrix is sandy and brown, all except the larger boulders being completely rotted. Where the Moraines deposits are above the level of the main valley fill they clearly show the effects of weathering. Their freshness in exposures along the river banks is probably to be explained by the fact that until very recently, geologically, they lay below river level where groundwater movement would be exceedingly slow in such compact fine material.

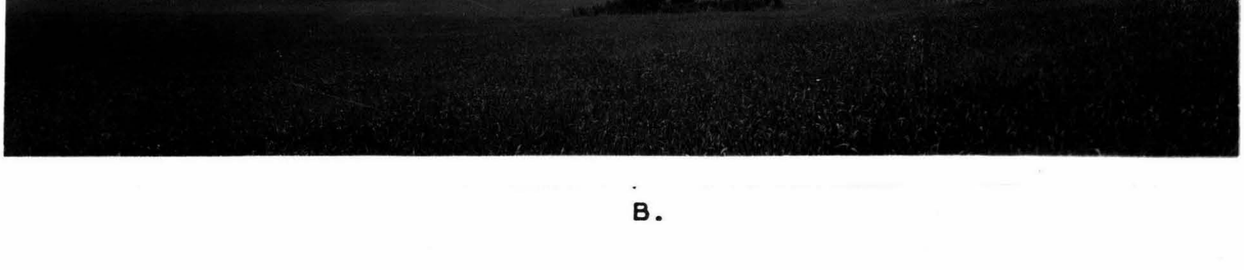
PLATE 15

A. Bluff of Mill City glacial till west of Mill City. The large boulders in the river have apparently fallen from the bank. The upper six feet of the section consists of river gravel.

B. The Mill City glacial moraines four miles west of Mill City. The ridge crossing the view from left to right is a large lateral moraine. In the middle distance is an intermorainal depression, to the right of which rises another small lateral moraine occupied by the orchard. View looking west from a point just below the Mill City road.

PLATE 15.

A.



B.

Outwash gravels and varved silts are associated with the morainal deposits in many places. At Mill City there are two or three small pockets of highly contorted varved silts in the till, and the section shown in PlateXVA grades into outwash gravels with steeply inclined bedding three-tenths of a mile west of Mill City. West of Gates varved silts are exposed continuously for a third of a mile in the north bank of the river. Their average exposed thickness is twenty feet, but they are gently folded in place, extend below water level, and are overlain by heavy river gravels, so that their total original thickness was probably considerably more. Downstream they end against a bedrock dam, and upstream wedge out on top of outwash gravels and till as shown in Plate xvla (Fig.30.)

The fact that the varved silts overlie the delta gravels, instead of fingering into them, indicates that the gravels were first deposited in a shallow lake as the ice retreated; readvance of the ice then raised the morainic dam the varved silts being deposited during the second recession of the ice front. No Mill City moraine is exposed in the river east of Gates.

The only definite topographic expression of the Mill City moraines is west of Mill City. The above-mentioned bench presumably represents a small remnant of lateral moraine. The best morainal topography is found on the broad perched flat in secs.22and 23, T 9s, R2 E. The long narrow east-west ridge (to the left in Plate xv 13) immediately above the river is a lateral moraine lying on tuff. The till is probably not over fifty feet thick. North of this ridge, and below the road where it turns northwest, are two small parallel lateral moraines about twenty feet high which are separated by two imperfectly drained tandem kettle-holes two hundred feet across. The southerly one of these is occupied by the orchard

PLATE 16

A. The glacial section in the North Santiam River west of Gates. Till can be distinguished along the water's edge. It is overlaid by outwash gravels which to the right of the picture show foreset ledding. The dark mass lying above the gravel is composed of moist varved silts.

B. A large glacial erratic in the North Santiam River gorge one half mile east of Mayflower Creek. The striae on the upper surface run at an angle to the shadows. The hammer indicates the size of the water-worn material with which it is associated.

A.

B.

to the right in Plate XVB. Between the large lateral and the small ones is a roughly triangular flat which narrows westward to a sharp V-gully. The flat, still imperfectly drained, is a partially filled intermorainal depression. Above the road to an elevation of about one thousand feet are many irregular humps and depressions suggesting morainic forms. Glacial origin rather than landslide origin is proved by the presence of boulders of oliving basalt not known to occur west of Detroit. The broad flat northwest of the moraines which ends abruptly above the terraced valley floor of the Little North Santam River was evidently an overflow channel into that valley when the ice blocked the main North Santiam valley. Glacial gravels containing facetted boulders and lying on white tuff at about 840 feet on the road down to the Little North Santiam River support this hypothesis.

Although the absolute western limit of the Mill City glacier is not known, the absence of till west of the moraines described suggests that the ice did not go much farther; Mehama may be considered a safe western limit. The presence of outwash gravels near Mill City indicates that the moraine there was a recessional deposit. Contortion of the varved silts might be due to oscillation of the ice front, with consequent over-riding of recessional deposits, or simple slumping. The Gates silts clearly indicate a morainally dammed lake, and are the farthest east deposits positively identifiable as Mill City in age. An occurrence of possible Mill City till lies in the saddle south of the 2200' granodiortte knob between Cumley and Kinney Creeks, in section 18, T. 9 S.,R. 5 E. The saddle is broad and contains a marshy area five acres in extent. Smooth topography which contrasts sharply with the normal rough topography on either side, deep sandy soil, and large olivine basalt boulders all point to glacial action. Since the Mill City ice must have been quite thick in the narrow gorge, the till pocket may have been some distance above the bottom of the glacial valley.

The absence of till east of Gates, coupled with the lack of a glacial profile in the gorge east of Niagara is significant. The glacial trough was undoubtedly very narrow because of the resistance of the mineralized rocks; therefore a comparatively small amount of post-glacial down-cutting might destroy the U-profile, The fact that the prpfile has been destroyed, and the lack of till in the river east of Gates prove that the glacial valley profile was somewhat steeper than that of the present river. These phenomena and their implications are considered below. (See page 64.)

The best indications as to the age of the Mill City moraines are their relations to the terraces of the Santiam Valley. The only morainic topography remaining is on the bench now 200feet above the North Santiam River, and what may have been a drainage channel now ends over 150'feet above the Little North Santiam River terraces. The "hung-up" position of the moraines is due more to lateral planation of the streams than to down-cutting: that is shown by the occurrence of till in the river. All traces of moraines on the south side of the valley have been destroyed. The mile-wide terraced floor of the main river was formed in post-Mill City time, because the gravels of the floor lie on the till. The floor, however, is not merely on gravel terrace deposits, but east of Mill City smoothly truncates tilted lava flows. At Gates the river was superposed on a lava ridge as a result of its meandering. Minor terraces from ten to twenty-five feet high are numerous south and west of Mill City, each indicating a temporary halt in the down - cutting process to the present river level. Whether the till is older than the hundred foot terrace south of Lyons has not been proven, but it may well be. Although destruction of the glacial profile in the gorge might have been accomplished very rapidly, planation of massive lava flows, and formation of the later terraces would require much time. It is suggested that the Mill City glaciation is the equivalent of the Sherwin epoch of the Sierra Nevada.

The Detroit Glaciation

Very little evidence is left of the Detroit glaciation, The only till exposures are; an outcrop in the Santiam River just above the Quartzville trail bridge, a large mass of gray till in river gravels exposed in the road cut at the east end of the Breitenbush River highway bridge, and till in place in a road cut a half-mile southwest of Berry. The exposures are all very small; the till is very similar to the Mill City in compactness, amount of matrix, freshness, and faceting of boulders.

Varved lake silts are well exposed in railroad cuts and the west bank of the Santiam River over a [illegible] mile stretch downstream from a point about two hundred yards above the Detroit Ranger station. The varves are very fine, some being almost paper thin, and show regular coarser banding about every six inches. (See Plate 17a) In very few places are they undisturbed, but are usually highly folded, contorted, and even brecciated in places. Their stratigraphic thickness is not less than forty feet, and may be more. The crumpling may have been caused by slumping, but the severity would suggest overriding of ice as the cause. Above the ranger station the silts give way to heavy river gravels or outwash. The lake filled by the silts must have been nearly a mile long, but not over one quarter as wide. The till near the Quartzville trail bridge is presumably part of the orainic dam.

The lack of exposures of the Detroit till renders eterminations of its age and extent very difficult, if not impossible he height of the till exposure below Berry (1450') shows that the ice xtended into the gorge; probably as far as Halls, possibly to ardine Creek. At Halls the valley has the appearance of an aggraded, odified glacial trough. Moreover, 0.45 mile east of Mayflower reek, and a good hundred feet above the river, the beautifully triated glacial erratic shown in Plate xvl 13, is jammed in a very arrow water-worn channel..Such a large boulder would lose its striae efore travelling far, even in a powerful stream; its freshness in pite of its exposed position precludes placement during Mill City ime. Although not conclusive proof of direct glacial deposition, ts presence strongly suggests that the Detroit glacier extended some distance into the gorge. Moraines would disappear almost as fast as built in such a valley, so their absence is not unexpected. The north bank of the gorge, particularly from Sardine Creek east is more or less plastered to heights of 100 to 150' above the river with very poorly assorted materials which appear to be valley train with admixed talus. In places this material very closely resembles glacial till, but no well facetted boulders were found, nor have any glaciated bedrock surfaces been discovered in the gorge.

The Detroit glaciation was extensive, and it must have greatly modified the topography of the High Cascades. The main essentials of the present glacial topography were probably produced during this time. The higher ridges descending westward from the High Cascade summit regions plainly show the effects of glaciation, but some have comparatively thick soil covers, indicating pre-Wisconsin glaciation; Minto Mountain and Woodpecker Ridge are cases in point. Also, the higher flats such as Grizzly Flats and Bingham Basin do not show the severity of ice action commensurate with their not show the severity of ice action commensurate with their topography. The inference is that they were given their main size and form by the Detroit ice, and were left untouched or only slightly modified by the later Tunnel Creek ice. The same considerations would hold true for the major valleys, but to a lesser extent.

Physiographic evidence is the basis for separating the Mill City and Detroit glaciations. The absence of a glacial profile in the gorge, especially below Mayflower Creek, necessitates downcutting since Mill City time, yet the Detroit till occurs below the present river level. If the Detroit till is considered as a recessional moraine of the Mill City glaciation, either upwarping of the gorge section or glacial gouging on a large scale is required. There is no other evidence for either. Moreover, the valley train material is too poorly sorted and too angular to have been derived from the Tunnel Creek or Breitenbush moraines, and is too low in the gorge to have been deposited in Mill City time, even if that type of material could remain in such a narrow valley for so long a time. The Detroit till is accordingly correlated tentatively with the Tahoe glaciation of the Sierra Nevada.

The Tunnel Creek Glaciation

The mor aines and physiography of the last glaciation, the Tunnel Creek, are for the most part, well preserved. The terminal moraine in the North Santiam Valley is 10.8 miles east of Detroit at an elevation of about 2000'. The morainic topography is perfectly preserved across the full width of the valley floor, except where the river has cut a rather narrow gap. The morainal belt is about a quarter mile wide with a maximum height of one hundred feet. The till as exposed along the highway is shown in Plate XVII B; it is typical alpine-type till, with the rock fragments showing very few facets or striae.

The terminal moraine in the Breitenbush Valley has been almost obliterated. Till is exposed in a roadcut just below Cleator Bend at 2200 feet elevation, a mile and a half west of Breitenbush Springs post office. No morainic topography was observed; river terracing and planation, caused partly by landsliding farther downstream have removed all but a little of the till.

Recessional moraines are well preserved in several places; in the North Santiam valley proper opposite Woodpecker Ridge at 2250 feet, and north of Independence Prairie at 2500 feet; on the Pamelia Creek trail at 2400 feet and probably at 3000 feet just below Hiedeike Creek (a branch of Milk Creek); a beutiful hooked lateral moraine was noted on Milk Creek just below the Skyline Trail at 4400feet. The only recessional moraine noted in the Breitenbush drainage is at 2900 feet in the valley of the South Fork. Most moraines of this type havedoubtless been destroyed by stream action, as considerable thicknesses of till are exposed in the cut ban ks of many streams. The glaciers on Mt. Jefferson are the last remnants of the once powerful local glaciers

T he valley profiles of the High Cascades are typical-

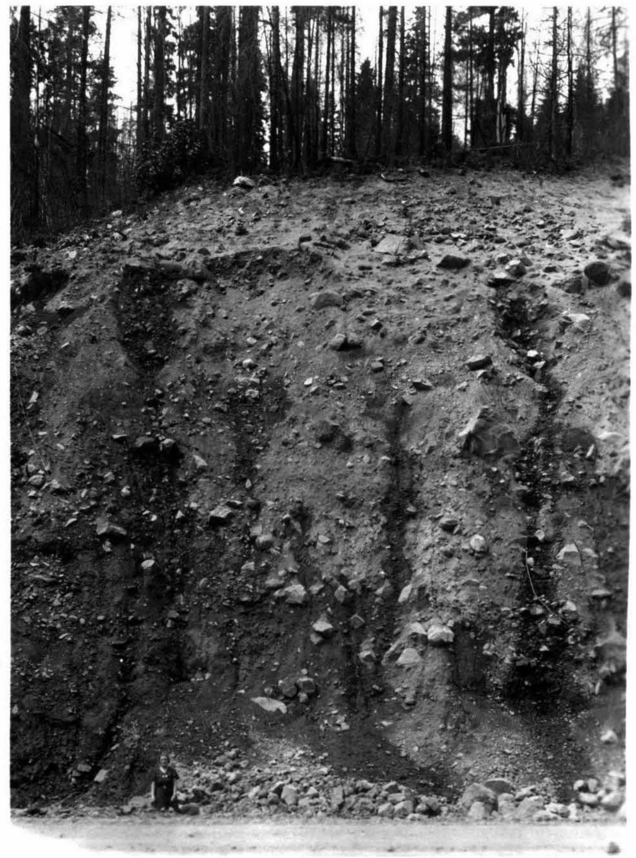
PLATE 17

A. Varved glacial silts one half mile downstream from the Detroit Ranger Station. Cyclical sedimentation is clearly shown by the coarser bands about six inches apart. Individual varves are scarcely visible.

B. Glacial till of the Tunnel Creek moraine a short distance west of Tunnel Creek. The angularity of the erratics is striking.

PLATE 17.

A.



B.

ly glacial, as shown in Plate XVIII A. Indications of the depth of the ice are scant, but if the valleys were well filled, it appears that the ice should have advanced farther down the Santiam River valley than it did. It seems probable therefore that the valleys were only partially filled; thin bench glaciers occupying the higher areas such as Grizzly Flats and Bingham Basin would account for their mild glaciation. On the other hand, the main upland in the vicinity of Olallie Butte was covered by a sheet of ice about 1000feet thick, which overrode some of the larger hills and all of the smaller ones and left a landscape similar to that of the Lake Superior region. Rock-basin lakes are fully as numerous as morainally dammed lakes. Striae indicate that the ice moved northward past Olallie Butte and fanned out westward toward the edge of the summit highland. Probably a similar fanning occurred along the eastern margin. The breadth and elevation of the summit highland must have accounted for the local thickness of the ice sheet. Excepting the summit highland area, the Tunnel Creek ice probably only slightly modified the topography left by the Detroit glaciation.

Jefferson Park (Plate XIX) is a remarkable example of a flat glaciated divide and is worthy of special mention. The park is at an elevation of about 5800feet, has a area of over one thousand acres, and is over one and one-half miles long from east to west. North of it Park Butte ridge stands a thousand feet higher, and Mt. Jefferson towers almost a mile above it on the south. The main section of the park is so flat that only the drainage shows the location of the actual divide. Eastward the park slopes gently down to the cirqued head of the Whitewater River valley. The west end of the park is more stepped down to an elevation of about 5500', where it is hung above the valleys

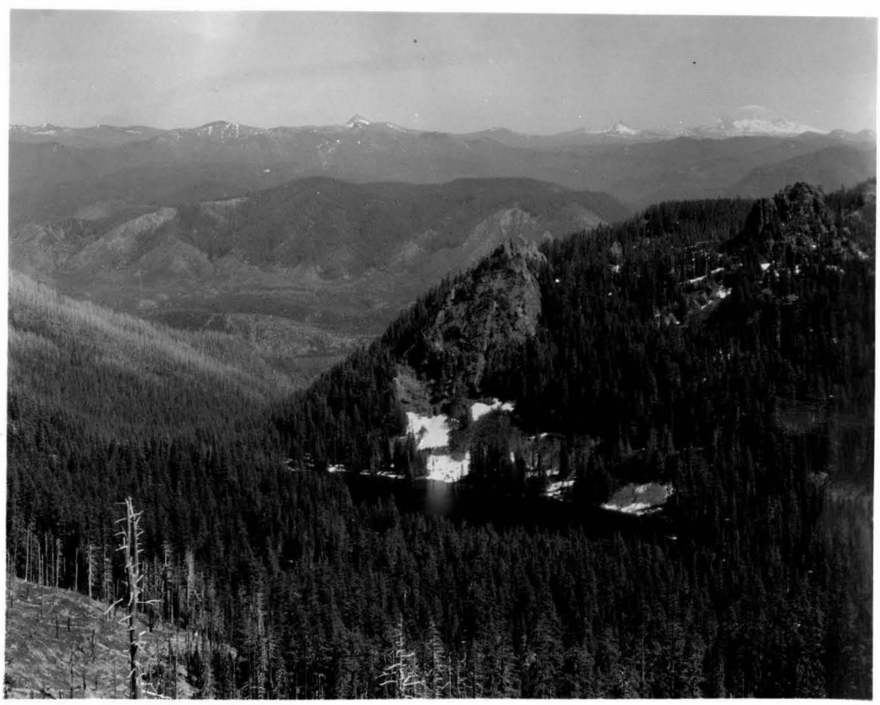
PLATE 18

A. View east down the glaciated valley of Whitewater River from the head of the cirque at the east end of Jefferson Park. The high cliffs of Lionshead are clearly visible at the left.

B. Tumble Lake as seen from Sardine Mountain. The cliffs are formed by massive flows of the Sardine Series. The higher valley level of the Detroit Basin is visible. Three Fingered Jack, Mt. Washington, and The Three Sisters are visible on the sky line beyond the ridge formed by the Outerson deposits.

PLATE 18.

A.



B.

PLATE 19

A. Mt. Jefferson and Jefferson Park from the divide east of Park Butte. Park Butte is at the extreme right. The observer is looking south from a point one thousand feet above the floor of the park.

B. Mt. Jefferson and Jefferson Park as seen across Russell Lake. The trees grow on low basalt ridges which are probably formed by the edges of inclined lava flows.

A.

B.

below. The western hanging , however, is due to cliff resession in lavas, rather than cirque formation. The flatness of the divide is explained by the impinging of extensive glaciers on the north slopes of Mt. Jefferson against Park Butte ridge, with consequent splitting of the ice in to two main east-west glaciers; the directions of striae support this hypothesis.

Contrasting sharply with the extensive glaciation of the High Cascades are the small locally glaciated areas of the Western Cascades. Cirques are by no means rare; Tumble Lake below Sardine Mountain (Plate XVIII B) is in one, and Mary (Macy) Creek heads in a cirque east of Coffin Mountain. Other cirques are known to exist in this general section. In the Elk Lake district is situated the largest area of Wisconsin glaciation west of Breitenbush Hot Springs. There a rather extensive glacier went more than four miles down the valley of Elk Lake Creek and spilled over into the Humbug Creek drainage in two places. The western gap or spillway is near the head of Elk Lake; the eastern one is at the head of Dunlap Creek, where Dunlap lake is perched(See Plate XX). Below the western gap the ice reached to 3600' and below Dunlap Lake to 3200. The lower part of Humbug Creek was terraced by the outwash gravels. At least two well preserved moraines are present in the Elk Lake valley. One dams the lake and the second is a quarter mile farther east below the meadowshown in Plate XX. The head of the valley is a cirque c ut deeply into the old Battle Ax volcano; the northeast slopes of Battle Ax were also heavily glaciated.

The freshness of the moraines and surfaces affected

PLATE 20

Elk Lake Valley as seen from Battle Ax. The western glacial overflow gap is seen at the extreme right; Dunlap Lake lies in the eastern gap. One recessional moraine lies at the foot of the lake, and a second just beyond the open meadow. Gold Butte, directly below Mt. Jefferson, is capped with Battle Ax lava, as is the hill west (to the right) of Dunlap Lake. A north-dipping Miocene lava flow may be seen directly above Dunlap Lake. Olallie Butte and Mt. Jefferson stand conspicuously above the High Cascade crest line. The view is somewhat south of east.

PLATE 20

PLATE 21

A. Rolling topography of the Salem Hills near Summit School. The valley in the foreground nearly parallels the strike of the Stayton lavas. The view looks toward the northwest.

B. Mill Creek gap one half mile north of Turner. The observer is standing on the eastern wall looking northwest at the western bluffs. Mill Creek follows the base of the bluff.

PLATE 21

A.

B.

by the Tunnel Creek glaciation leaves little doubt as to the age of the ice. This glaciation is correlated with the Tioga of the Sierra Nevada and the Wisconsin of the Mississippi Valley.

CHAPTER VI

Physiography.

The physiographic dissimilarity of the Western Cascades and High Cascades especially from Outerson Mountain southward is no less pronounced than the structural hiatus. Since the present forms of each of the two divisions is directly related to its structure and age, individual treatment of each part is required in a complete physiographic analysis of the Cascade Range in Oregon.

Western Cascades

The Western Cascades exhibit a rather close relationship between physiographic form and structure. The characteristics of the range are those of a deeply eroded mass of folded rocks of varying resistance, the range having acquired its present height by normal faulting along the eastern margin. Physiographically, the Western Cascades may be divided into three longitudinal belts: the western belt of dip slopes; the high central belt of massive lava flows deeply incised by steep-walled valleys; and the eastern belt of tuffs which is characterized by rather open valleys.

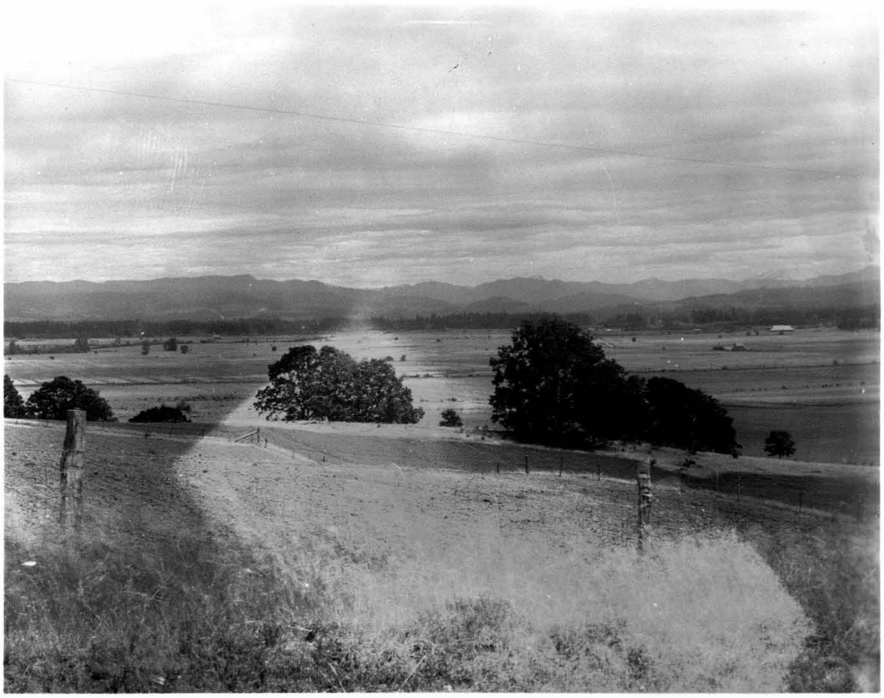
Westward from the crest of the Mehama anticline dip-slope topography is predominant. Below 1000feet elevation the Stayton lavas almost everywhere immediately underlie a thin soil cover(Plate 22B); from 1000 to 2000 feet benches formed by ressitant beds of the Fern Ridge Tuffs are common; above 2000 feet the surface of the ridges is remarkably uniform dipping about one and one-half degrees northwest(See plate 22A). The highest elevation which these surfaces attain is about four thousand feet at the House Mountain Scarpwhich is due to erosion on the crest

PLATE 22

A. A view east across the Stayton Basin to the valley of the North Santiam River. The House Mountain scarp and the dip slopes are visible to the left also dip slopes east of Kingston to the right (over the barn roof). The N. Santiam valley lies in the center of the view. Photograph taken from a point about one mile north of Parrish Gap.

B. Dip slopes on the Stayton lavas southeast of Kingston. The hill in the center distance is composed of Fern Ridge tuff, and tuff occupies the foreground. On the skyline the dip slopes north of the North Santiam River Valley are dimly visible. The view is northeastward from the southwest corner of the Mehama Sheet.

PLATE 22.



A.

B.

of the Mehama anticline. The Salem Hills, as shown in Plate V, are characterised by a surface formed on the eastward dipping S Stayton lavas.

The degree of dissection of the structural surfaces outlined above varies greatly. The structurally flatter parts of the Salem Hills exhibit a rolling landscape which shows little relation to structure(See Plate XXIA). Where the dip of the lavas is steeper as west of Marion, cuesta topography is very pronounced. The western margin of the Salem Hills is an erosion scarp characterised by much landsliding in the Illahe sediments. The foothill slopes of the Western C scades are well preserved below elevations of 1100 feet. Where the upper lava surface attains greater heights exposure of the easily eroded Mehama volcanics permits undercutting and widening of the major valleys. Both the North Santiam River valley (Map, Plate I), and Thomas Creek (Plate VII B) widen out in this way. The smaller streams northeast of Stayton have cut into but not through the lavas, and have incised deep valleys into the Fern Ridge tuffs. The valleys in the lavas are deepened mainly by headward recession of waterfalls on individual flows. The Fern Ridge tuffs are deeply dissected by sub-parallel valleys trending normal to the strike of the formation. The uniformity of the interfluve crests is attributed to structural control rather than tilting of a former peneplain as suggested by Smith[1]. The uppermost portion of the House Mountain scarp is somewhat stepped where one or two massive lava flows

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1 Smith, W. D.; Physical and Economic Geography of Oregon; the Willamette Valley. Commonwealth Rev.; Univ. of Oregon., vol. 7, no.4, p.158,1925.

PLATE 23

A. A gravel terrace partially stripped from the basalt surface, south of Kingston. The soil cover in the foreground is less than six inches deep. The observer is facing northeast.

B. A view south across the old Little North Santiam valley segment northwest of Mehama. The Ridge is composed of Stayton lava, with a remnant of Fern Ridge tuff remaining on top.

PLATE 23

A.

B.

occur although for the most part it is a very irregularly dissected mountain face.

North of the section under consideration the Cascade Range foothills slope directly down to the alluvial floor of the Willamette Valley. In the North Santiam section however the Stayton basin is separated from the main Willamette Valley by the Salem Hills and the Turner Warp. The basin has been breached in two places: at its northwestern corner by Mill Creek gap, shown in plate XXI B, and on its southern side by the main valley of the North Santiam River. The basin is a very low alluvial fan built by the North Santiam River. That the basin was formerly filled to a greater depth is shown by the well preserved gravel terraces which border its eastern margin (Plate I). These terraces which stand nearly one hundred feet above the basin floor, have been reduced by lateral planation on the part of the North Santiam River, and also by stripping from the surface of the underlying Stayton lavas, as shown in Plate XXIII A.

The widening of the valley of the North Santiam River in the Mehama volcanic and tuffaceous facies of the Sardine series is very pronounced as far east as Minto (see Plates I and II). The floor of the valley is veneered with terrace gravels through which the river has cut a narrow channel about fifty feet deep from Mehama eastward. Terraces from ten to fifteen feet in height are not uncommon, but with the exception of high terraces such as those south and east of Lyons, shown in Plate I, the greatest depth of gravel is probably not more than fifty feet. At Gates there are two main terrace levels, one at 985 feet about fifty feet above the river, the other about 125 feet above the river.

The lower terrace level is due partly to gravel filling, partly to planation of tilted lavas.

Further evidence of much lateral planation by the N North Santiam River is shown by thefact that intercision of the Little North Santiam River occurred at some long past date. The intercision occurred northwest of Mehama causing abandonment of the segment of valley shown in Plate 23B. The valle y floor is now three hundred feet above the main valley level and indicates a mile shortening of the Little North Santiam River.

The highest part of the Western Cascades is the belt in which the Sardine Series occurs. In this part of the range, the ridge crests average 4500 to 5000 feet in elevation,the relief being between three and four thousand feet. The valleys are narrow and steep-walled, the ridges being sharp crested. Scattered glacial cirques occur above thirty five hundred feet. In the North Santiam River Valley unmistakable evidence of a wide valley stage is preserved. As shown in Plate XXIV the old valley narrows rapidly eastward and has a lon gitudinal profile several times steeper than the present valley which is sharply incised in its bottom. At the end of the southern spur of Sardine Mountain evidence of this valley is very scanty. The steep longitudinal profile and rapid narrowing of the valley suggest that it was formed by a small stream which headed on a drainage divide south of Sardine Mountain.

The eastern belt of tuffs is characterised by less rugged topography than the central lava belt. The valleys are more open although by no means broad. The general elevation of the ridge crests is nearly the same as in the central zo ne probably because of greater distance between large streams, and the greater elevation of those streams near their heads. Landslides are very common. True hogback topography,(shown in Plate

PLATE 24

The North Santiam River valley west of Sardine Mountain. The broad valley stage, rising rather steeply eastward is visible. McCully Peak shows on the dimly distant sky-line left of the center. Spurs of Monument Peak are visible to the left, spurs of Rocky Top at the right. The incised character of the present gorge is shown at the lower left.

PLATE 24

26A) is developed on the steeply in clined beds in the west limb of the Breitenbush anticline; elsewhere there is a complete lack of correlation between topography and structure.

Where lavas of the High Cascade sequence occur on the tuffs, as in the vicinity of Battle Ax and along the eastern margin of the Western Cascades, they dominate the landscape. The flows,usually massive or platy, commonly form cliffs two or three hundred feet in height(Plates 26 B, 28B). Thin bedded flows give rise to a very sharp crested ridge triangular in cross-section. This kind of ridge, shown on Plate 28A, is best exemplified by Battle Ax itself, hence is for convenience named the Battle Ax type. Although the Battle Ax lavas do not form eminences above the general Western Cascade skyline, the Outerson deposits form a much dissected elongate mass on the e astern margin of the Western Cascades. This mass, which south of the North Santiam River rises f ully one thousand feet above the tuffs to the west, is shown in pPlates XVIII B, XXII A.

The Battle Ax lavas are probably the chief reason for the difference in the present character of the North Santiam and Breitenbush River Valleys. Whereas the Santiam Valley west of Boulder Creek has been continuously eroded since the uplift of the Western Cascades, the Breitenbush Valley was eroded to a fairly broad stage, then filled with lavas, and has now been partly re-excavated. The lower limit of the lavas was 2200 feet or less, only six or seven hundred feet above the present valley bottom. The massive flows now form cliffs, unscalable in places, along the north side of the valley, from French Creek to Humbug Creek. East of the lava outpourings the valley widens at the junction of

Humbug Creeks, then narrows again in the mineralized zone east of East Humbug Creek.

The Detroit Valley seems to be an anachronism. All the Valleys above it are fairly narrow, and the valley below it is in part a gorge in which the stream is not yet at grade. The width of the valley, two miles, can scarcely be attributed to weak rocks, becausethe western side is cut in the Sardine Lavas. Small remnants of high level terraces sloping about three degrees southwestward occur at 2800 and 1700 feet either side of Tumble Creek, which was diverted almost a quarter of a mile southwest behind the lower terrace. Boulders of olivine basalt were found on the higher bench. The isolated hill in the center of the valley seems to have been part of an old floor which is well preserved southeast of the North Santiam River. This level, cut on bedrock, is only slightly dissected by the streams from Blowout Cliffs, yet stands a good two hundred feet above the river. This latter flows through a narrow valley which may have been glacially excavated, and which separates the hill from the surface to the south across the river. The bedrock surface appears to rise slightly toward the riverward margin, suggesting that the hill was the north divide of the valley. Northeast of this hill, and west of the Breitenbush River there are extensive gravel terraces at an elevation of about 1500 feet. These coarse river gravels are well exposed in the Breitenbush road just north of Detroit. The depth of gravel suggests that the Santiam once flowed around the north side of the hill; glacial moraines may have caused the initial diversion from the old channel. That the bedrock surface is Pleistocene is proven by the exposures of Detroit till at the bottom of the Santiam channel; the shelf is probably a remnant of a Pleistocene immature valley stage not preserved in the gorge.

The topographic effect of the young lavas which were poured out on the Western Cascades has been pointed out above with reference tothe Breitenbush Valley. In addition to filling the Breitenbush Valley the Battle Ax lavas established a system of roughly radial drainage centering at Battle Ax. The streams were superposed on the older rocks, and have now cut deep valleys into them. East Humbug Creek and the Breitenbush River apparently skirted the base of the volcano.

The Outerson volcanic accumulations caused a distinct narrowing of the pre-Pigeon North Santiam River Valley, as is to be expected. This constriction is well shown by the small width of the Pigeon basalts between Mount Bruno and Outerson Mountain as compared withtheir breadth south, and especially west of the Elbow. The steepness of the east side of the Coffin group probably is due to severe erosion by the trunk glacier which moved northward along the North Santiam Valley.

In the Breitenbush drainage there is a slight physiographic evidence of the eastern margin of the Western Cascades. The divides north and south of the river contimue from the old lavas to the younger. Breitenbush Hot Springs is situated in a surprisingly broad valley at the junction of the North and South Forks of the Breitenbush River and Devils Creek. The broadness of the valley is probably due to the presence of a body of bedded sandy tuff in the Breitenbush Series. Such sediments crop out along Devils and Skunk Creeks and undoubtedly underlie the large landslide area south and west of the Hot Springs. The north and west limits of the valley are determined by resistant tuffs and lavas somewhat mineralized; the southern side by the cliffs of Outerson Mountain and Timber Butte; and the eastern edge by the cliffed ridge ends of Devils Peak and Breitenbush Mountain.

That the valley has been developed to its present form since Minto time is proven by the occurrence of Outerson lavas in Bald Butte and the base of all the ridges to the south. If the valley had been present it would have been filled with Outerson deposits; or if cut in the Outerson debris would have been filled with Minto lavas. The scarp was essentially intact when the Outerson eruptions occurred; this is proven by the basal contact of the Outerson series. That the Minto lavas went condidderably farther west is proved by their presence, seven hundred feet thick, in Mansfield Mountain. The present broad valley at the hot springs, if eroded out of solid lavas of the type in Breitenbush Mountain, is in compatible with the youthful character of the valley west of Scorpion Creek. It becomes reasonable, however, if the main mas was relatively easily eroded sediments with a thin capping of later volcanics. From the base of the old scarp east, the present streams are considered as consequent; from the base of the scarp to Scorpion Creek they are probably superposed.

High Cascades

South of Jefferson Park the High Cascades are essentially, as described above, an elongate accumulation of lavas rising to a definite crest along the line of volcanic peaks. The physiography of the west side only is included in this paper. The original lava slopes have been greatly lowered and dissected streams heading well toward the crest of the Range as shown by Plate XXX. The valleys are steep walled and terminate in cirques, or broad glaciated basins. The westward

PLATE 30

The panorama of the High Cascades from Woodpecker Hill to Pine Ridge, as seen from Bachelor Mountain. The ridge in the foreground is the eastern spur of Mt. Bruno, east of which the Santiam River flows northward. From north to south the main features are: the valley of Whitewater Creek; Woodpecker Hill, composed of Outerson deposits; Woodpecker Creek; Woodpecker Ridge, composed of Minto lavas; Mt. Jefferson, rising above the valley of Pamelia Creek; Minto Mountain; Minto Creek and Bingham Basin; Bingham Ridge; Marion Creek valley; and Pine Ridge.

PLATE 30

flowing streams are tributary to the northward flowing North Santiam River whose course was originally determined by the intersection of the surfaces of the Outerson series and the Min to series. The Minto Lavas completely buried the isolated rhyolitic hills but flowed around the Outerson deposits in Woodpecker Ridge and east of Boca Cave. Small valleys were formed at the margins of the flows; hence the similarity of the upper courses of Woodpecker and Cheat Creeks. The Santiam River which presumably m marks the original western limit of the Minto Lavas is regarded as the western boundary of this apart of the High Cascades.

North of Jefferson Park the High Cascades assume the plateau character already outlined in the discussion of structure. The western edge of the original surface of the lavas was fully as high as the ridges of the Western Cascades, so there is no e evident physiographic boundary between the two ranges. The source of most of the lavas south and west of Olallie Butte was in the vicinity of Park Butte. This is shown by the dips in the lavas and also by the northwestward dourses of the two forks of the Breitenbush River. The elevations of both features increase uniformly to a maximum of 7000 feet in Park Butte, which is known to be a volcanic neck. The Park Butte volcano probably surpassed the present Mt. Jefferson in grandeur, and unquestionably dwarfed it in bulk. Nobth of Olallie butte the highland elevation decreases uniformly to about 3500 feet at the head of the Oak Grove fork of the Clackamas River in the Mount Hood quadrangle. The valley of Cub Creek, a tributary of the Clackamas River which was beheaded by the Breitenbush River, marks the westernedge of the summit highlands. Rhododendron Ridge, which extends northward from Collawash Mountain, was apparently built up by flows from sources located close to the course of the postulated Cascade fault, and is fully as high as the summit up highland.

Quite probably Olallie Butte and Mount Jefferson were built simultaneously, after the Minto Lavas had been deeply eroded. Both are greatly eroded steep cones set on the older broad mass of the Minto Lavas. The present form of Olallie Butte is not its original outline, but is the result of glacial scour around its base, and normal erosion above the ice level. Its base once extended at least a mile farther west, and the eastern side was deeply eroded by glaciers.

The Pigeon flows straddle the boundary between the Western and High Cascades. Their maximum thickness, 1600 feet, is at Tunnel Creek; their maximum width, almost three miles, is attained a mile west of Timber Butte. The valley they filled was broadly mature in the Breitenbush series, considerably constricted in the Outerson series, and narrowed rapidly from the Elbow south. The old valley bottom is probably only a little above the level of the present Santiam River as the basal contact of the basalts west of the Elbow is nowhere more than 250 feet, and in most places is not more than 100 feet above the river. South of Whitewater Creek the debris from the steep western ends of Woodpecker Ridge and Minto Mountain has buried the Pigeon basalts to such and extent that only the broad benched surface, seen in Plate X A, and occasional small basalt outcrops prove their presence. The southernmost remnant of the flows is the 400 foot knob south of Independence Prairie Ranger Station. It is west and south of the pre-Pigeon channel.

PLATE 31

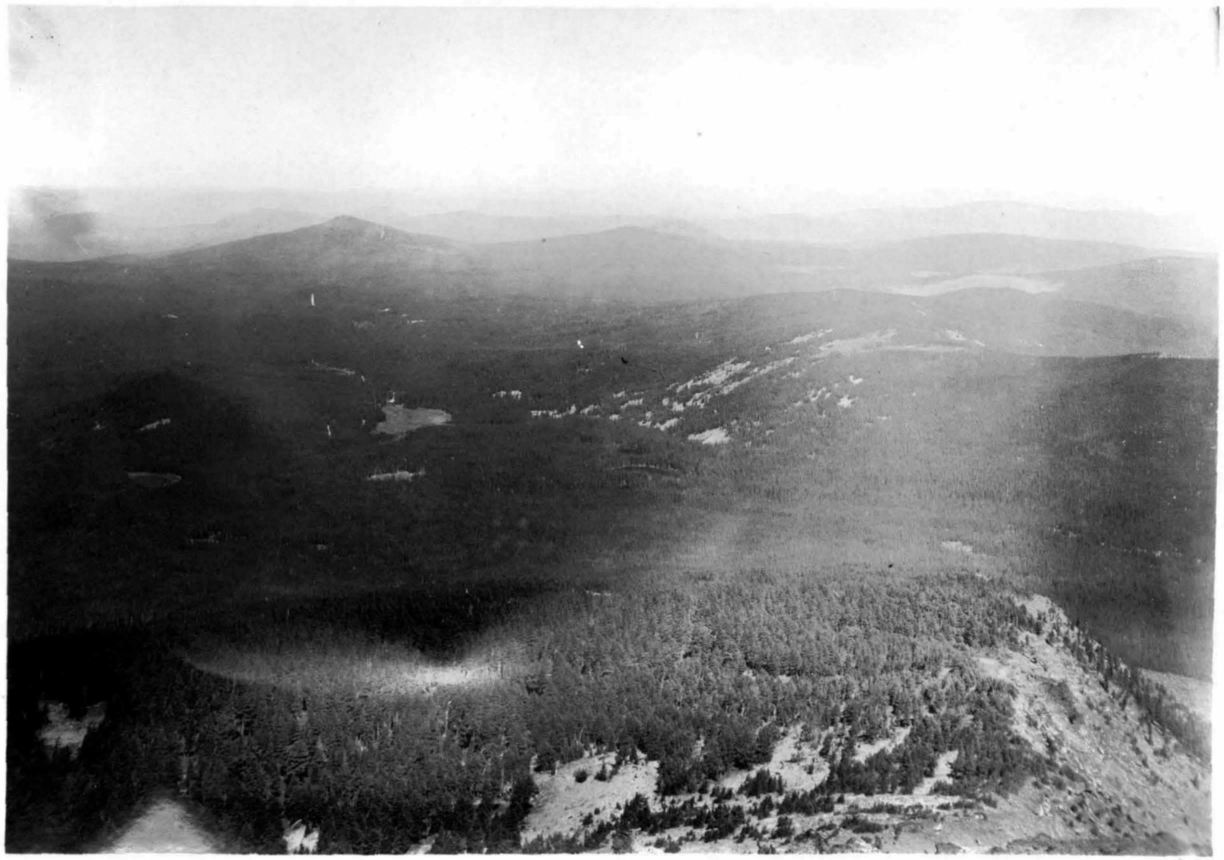
The view west from the summit of Olallie Butte. The plateau character of this part of the High Cascades is apparent. The valley of the North Breitenbush is situated to the left of center; Bald Butte is directly west, to the right of Twin Peaks in the middle distance. On a clear day over forty lakes may be seen from this peak.

PLATE 31

PLATE 32

The view north from the summit of Olallie Butte along the crest of the High Cascades. The Pinhead Buttes, to the west of a line to Mt. Hood, are two slightly eroded volcanic cones built up on the plateau surface.

PLATE 32.



The entire range shows the effects of glacial action action in Wisconsin or earlier times. The high ridges were probably heavily glaciated during the Mill City and Detroit glaciations, but were little affected by the Wisconsin ice. The valleys ho wever , were fairly well filled by the Wisconsin ice and have typical glacial cross-profiles, as shown in Plate XVIII-A. The valley floors are all veneered with glacial till or gravels to such an extent that bedrock is rarely exposed in their lower courses. Milk and Pamelia Creeks are conspicuously aggraded below an elevation of 4000 feet. Pamelia Lake was probably dammed by debris from Milk Creek, which heads in the Milk Creek glaciers on the west side of Mount Jefferson.

Derangement of Drainage

It is not at all surprising, in regions of volcanic activity to find evidences of disordered drainage. In a simple volcanic range the drainage is everywhere consequent. If such a range were built upon an older one, the previous drainage would be locally obliterated and might be deranged at great distances by blocking of major drainage lines.

In the High Cascades the drainage is consequent. The only major drainage change noted in them was the beheading of the Cub Creek branch of the Clackamas River by the Breitenbush River. The capture occurred southeast of Bald Butte, and diverted the upper six miles of Cub Creek. The Cub Creek valley was fully a mile wide at the point of capture, and now stands about eight hundred feet above the Breitenbush River valley. (See Plate XXXIII) The Cub Creek level shown in Plate XXXIV is prominently displayed along both sides of the North Fork of the Breitenbush River, and is almost intact for a distance of three miles west of Breitenbush Lake. The capture must have occurred during Detroit glacial time or earlier, and was probably caused, or at least aided, by ice crossing the former divide west of Cub Creek. The situation is judged to have been similar to that east of Elk Lake during the Wisconsin epoch.

In the Western Cascades drainage changes were probably more profound. The low level of the Battle Ax Lavas and Pigeon Basalts prove that the eastern side of the range had been eroded essentially to the present depth at

PLATE 33

The beheaded valley of Cub Creek as viewed from Breitenbush Mountain. The lower level valley of the Breitenbush River can be seen crossing the view from right to left. Bald Butte is to the left; Mt. Hood is visible over the shoulder of Sisi Butte, which is a partially dissected cone. The ridge in the lower right hand corner is a spur of Breitenbush Mountain.

PLATE 33.

PLATE 34

A view northwest down the valley of the North Fork of the Breitenbush River above the point of capture of the Cub Creek. The breadth of the Cub Creek stage is well shown in contrast to the depth of the valley cut by the Breitenbush River.

PLATE 34.

PLATE 35

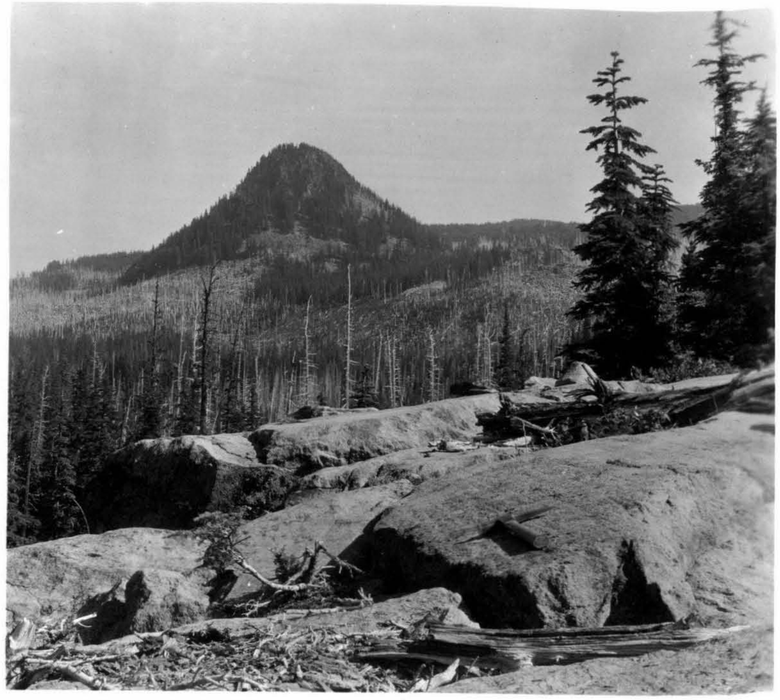
Typical High Cascade topography about three miles west of Breitenbush Lake. Ruddy Hill, composed of red agglomerates, is situated in the center top of the view. Westward dipping lava flows are visible at the upper left. The ridge beyond the lake is composed of massive olivine basalt and shows the effects of glaciation.

PLATE 35.

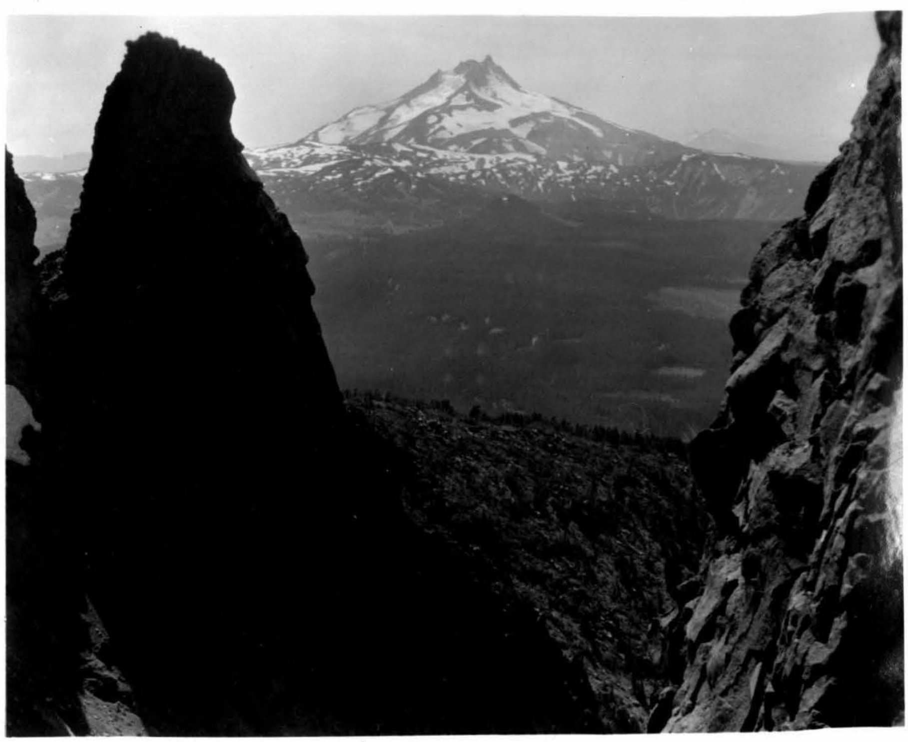
PLATE 36

A. Pyramid Butte, an intrusive andesite plug, seen from the northwest. Glaciated massive olivine basalt is shown in the foreground.

B. Mt. Jefferson as seen from the east side of Olallie Butte. The crags are probably intrusive masses of the old plug, but may be very thick flows. The Three Sisters are visible to left of Mt. Jefferson. Breitenbush Butte is directly below Mt. Jefferson.



A.



B.

least as early as Middle Pleistocene time, and the Santiam Valley developed to maturity. If the Detroit Basin had been present in anything like its present form when the Battle Ax Lavas were poured out, it should have been filled at least partly; however, no olivine basalts are known to occur in place west of French Creek. The broadening of the Santiam Valley from west to east strongly suggests a valley developed by an eastward flowing stream which crossed the Cascade fault scarp and maintained its course in spite of the Outerson eruptions, but turned south at least as far as Woodpecker Ridge, and possibly beyond Minto Mountain. Incomplete damming by the Outerson series would partially account for the widening of the valley just to the west. The topography buried by the Battle Ax Lavas was mountainous, but of the type which would occur in the upper portions of a valley like that occupied by the pre-Pigeon Santiam. The absence of basalts in the Detroit Basin suggests that in pre-Battle Ax time the main valley was the Santiam east of Detroit, and was joined by a minor branch or branches from the southwest, where the Detroit basin is now located. Blowout Creek may have joined the pre-Pigeon Santiam (hereafter referred to as the East Santiam) here.

The wide valley stage in the North Santiam Valley west of Sardine Mountain indicates on the other hand, a rather small stream (the West Santiam) which even in maturity had a high gradient, and probably headed near the south spur of Sardine Mountain. The present valley is a youthful valley, almost a gorge, whose stream is not yet at grade; moreover this valley has been cut down enough in post Mill City time to remove all positive evidence of the powerful glacier which passed through. In addition, the upper portion of the North Santiam which now occupies the valley of the former East Santiam has cut a youthful valley very little below the level reached by the latter, although it must be a much more powerful stream.

In brief, the evidence points to two drainage systems in the Western Cascades in pre-Outerson time. The West Santiam was small east of Mill City; the East Santiam may have been larger but in any event developed a mature valley which opened out through the Cascade fault scarp into the Eastern Oregon of that time. The Outerson eruptions constricted the valley near the now buried scarp, and forced the stream southward past the Outerson deposits in Woodpecker Ridge. Beyond that point its course is unknown. The Battle Ax Lavas caused only local changes in the drainage so are regionally unimportant.

When the Minto Lavas were extravasated, they must have dammed any eastward drainage from the Western Cascades, including the East Santiam. Their superposition of a short section of the Breitenbush River has already been noted. South of Outerson mountain, they formed the east side of the trough along which the North Santiam River flows northward. The rhyolites in Minto Mountain probably were high enough to prevent the total filling of the fault line valley to the west (possibly occupied by the East Santiam); hence the drainage of the area immediately to the south was northward. The East Santiam must have been ponded to the height of the lowest divide, which happened to be that into the West Santiam drainage. The divide was probably not over 2500 feet above tide.

As soon as the ponding occurred, several things must have happened. First, Minto Lake began to fill up with gravel and detritus and the divide began to wear down. As soon as the lake was filled, material started over the divide; the latter was ratherrapidly cut down, and the level of the filling correspondingly lowered with the formation of terraces. The high terraces west of Detroit may well be remnants of such deposits. The presence of olivine basalts on them proves the valley was filled to that level (2800') in post-Outerson time, and in view of the above mentioned relations of the Outerson series, probably in post-Minto time. Filling to such a depth would afford opportunity for superposition of the later streams across spur ridges; the central hill southwest of Detroit is presumed to have originated in this manner.

The width of the valley filled by the Pigeon basalts along the foot of the High Cascade slopes indicates a not inconsiderable period of erosion between the eruption of the Minto Lavas and the later flows, a time certainly long enough for the removal of the lake filling. The thickness of the Pigeon Lavas strongly indicates that they completely blocked the Santiam drainage to the south, as the divide into the McKenzie drainage at present is only one hundred feet higher than the top of the Pigeon flows, and has been built up by post-glacial flows. Again a lake must have formed, whose height was determined by the divide, and the Santiam drainage from the Elbow south went into the McKenzie. During glacial times, possibly first during the Mill City glaciation, certainly during the Detroit, ice passed over the basalts, which are thin bedded and would be eroded rapidly by ice action. Since ice movement was probably relatively fast through the North Santiam Valley and very slow midway between the North Santiam and McKenzie outlets, rapid lowering of the basalt dam at the Elbow and reëstablishment of the North Santiam drainage resulted. Hope of recognition of traces of the lake filling is small because of the severity of the glaciations in this area, the abundance of late glacial gravels and the dense forest cover.

It thus appears that the North Santiam River, though young as rivers go, has had a history greatly complicated by the volcanic eruptions at various times during the Pliocene and Pleistocene. Its history may well have been nearly duplicated by other streams which cross the western Cascades.

HOT SPRINGS

Numerous hot springs occur in the vicinity of Breitenbush Hot Springs, and two health resorts have been built up. The individual springs are small and scattered over an area of about a square mile. The total flow of the springs is estimated at one hundred thousand gallons per day.[1] The temperature of different springs varies between 125° and 194°, about half of the total flow being at the higher temperature. The water contains large amounts of sodium chloride, calcium, and iron carbonates, also some sulfate radical and silica. Sodium chloride is by far the most abundant mineral constituent.

Deposition of carbonate from the spring waters has caused accumulation of the travertine shown in Plate XXXVIIA and cementation of river gravels to form a massive conglomerate. Many small springs issue from the contact between the conglomerate and the underlying andesite flow. One such small spring is shown in Plate XXXVIIB. The main point of geological interest in the springs is the fact that the presence of the springs denotes continuing igneous activity in this part of the Cascade Range.

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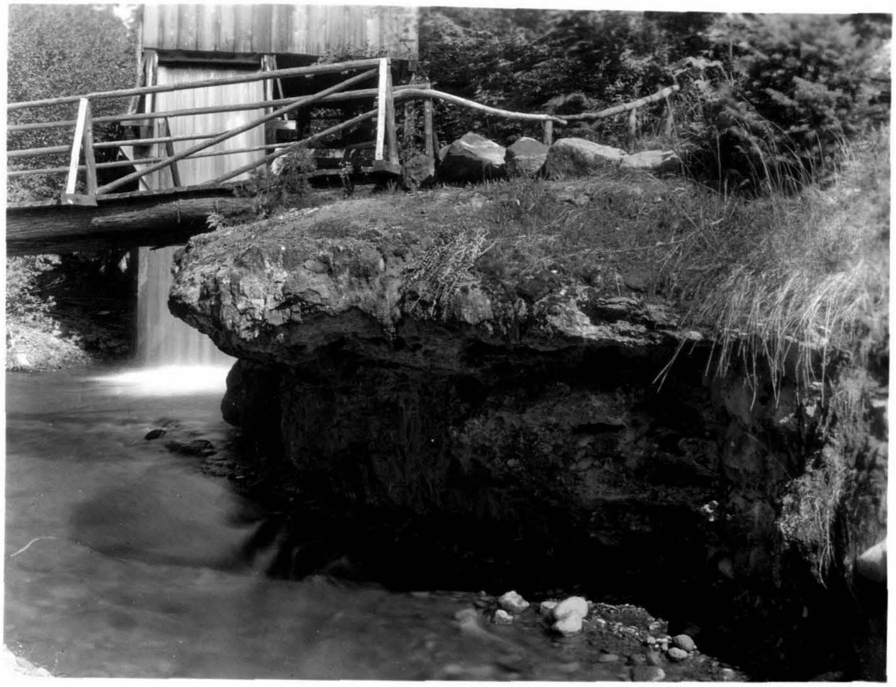
1. Bruckman, H. M.; Personal Communication

PLATE 37

A. Travertine hot spring deposits and cemented river gravel at Bruckman's Resort, Breitenbush Hot Springs. The overhanging ledge is travertine; boulders may be seen in the shadow beneath.

B. Cemented river gravels lying on andesite below Bruckman's Resort. A small spring issues left of the hammer head. The brown limonite stain is dark in the picture.

PLATE 37.



A.

B.

CHAPTER VII

GEOLOGIC HISTORY

The geologic history of the central Oregon Cascades, while complicated in detail because of the intermittent character of volcanic activity, is in broad outline rather simple.

Little or nothing is known of its history before the Oligocene. Apparently shortly after the period of folding at the end of the Eocene, vulcanism began in the Cascade region with eruptions of the explosive type. The great thickness of the Breitenbush tuffs was accumulated while erosion carried much of the debris to the ocean, the shoreline closely following the eastern margin of the Willamette valley. The centers of eruption were probably spread over a rather wide zone. At the end of the Oligocene the gentle folding of the Illahe and Mehama formations occurred. This folding was due to compression which caused local buckling in the soft tuffaceous sediments. The direction of application of these forces is not clear. During this period of buckling the Cascade volcanoes were active so no sharp stratigraphic break occurs within the range as at the western edge.

In the Middle Miocene the explosive activity gave way to extensive extravasation of lava, probably at the same time the Columbia River Basalts were poured out. The sources of these lavas were probably fissures located between House Mountain and the eastern edge of the Sardine Series. The lavas accumulated to a thickness of over six thousand feet, thinning rapidly eastward, and less rapidly to the west.

Flows of the fissure type probably occurred in the general region of the Willamette Valley at this time, as it is not likely that the Stayton lavas all originated in the Cascade Mountains.

Toward the end of the Miocene period volcanic activity ceased, giving erosion full sway. During late Miocene time the folding of the Cascades occurred, probably shortly followed by intrusion of the diorite with concomitant mineralization. The axis of folding was generally northeast-southwest. The steep folding of the tuff series might have been caused by their relative incompetence or by a northwestward acting force which crumpled the east side of the range. In any event, the intensity of the folding decreased toward the northwest.

The period of folding terminated the Miocene, and was probably followed by a period of erosion during which the elevation of the range was considerably reduced. The main drainage divide of the Western Cascades at that time was in the vicinity of Sardine Mountain. The Sardine lavas stood somewhat higher than the tuff area to the east. The elevation of the mountains before uplift by faulting is subject to conjecture as little or nothing is known of the pre-Pliocene landscape.

In Early Pliocene time, tension along the margin of the Basin and Range province and the Columbia Plateau was accompanied by faulting presumably of the normal type, along the eastern side of the Cascades. The Cascades were raised about one half mile along the Cascade fault, and tilted westward. Erosion was rapid in the soft tuffs elevated in the eastern side of the mountains, so valleys were quickly cut down to give a relief of a half mile or more. Soon after the valleys had begun to widen as the streams approached grade, the Outerson volcanic eruptions occurred along the general line of the fault. They partially blocked the East Santiam, causing its valley to be widened.

Soon after the Outerson eruptions ceased, the scene of volcanic activity shifted some eight or ten miles east to the line of modern volcanoes. The Minto lavas were poured out, deeply burying the area east of the Cascade fault. The eastward flowing streams in the Western Cascades were diverted along the western foot of the High Cascades, or reversed in direction. The portion of the East Santiam east of the present site of Detroit was reversed, and forced over the divide into the West Santiam valley. At nearly the same time the Minto lavas reversed the East Santiam, the Battle Ax volcano was built up, filling the Breitenbush valley. Both the post-Minto Breitenbush and reversed East Santiam were now larger than their predecessors by the amount of water they received from the western slopes of the High Cascades. Minto Lake was soon filled with detritus and emptied as the outlet was lowered. During the emptying process, either the Santiam or Breitenbush Rivers was superposed across a mountain spur and in its downward course isolated the hill in the Detroit Basin.

Some time before the Santiam Gorge had reached its present level the Mill City ice advanced from the east, passed through the gorge, and extended almost to Mehama. It was a powerful glacier, and undoubtedly deepened and widened the gorge. Upon its retreat, normal stream downcutting proceeded rapidly with the abundance of water and glacial material

Probably before the Mill City glaciation, and certainly before the Detroit glaciation, the Santiam valley was blocked by the Pigeon basalts. Between the Mill City and Detroit glaciations the Santiam Valley west of the Detroit Basin reached its present depth. This suggests that the Santiam had its full flow, and that therefore the Mill City glacier had breached the Pigeon lava dam. Olallie Butte and Mount Jefferson may have been built about the same time the Pigeon basalts were erupted.

The topography left by the Mill City glacier was probably largely destroyed by the time of the Detroit ice. The latter advanced beyond Detroit, and performed the main sculpturing of the High Cascades. Local pocket glaciers doubtless existed in the Western Cascades during both these early glaciations.

The Wisconsin glaciation apparently followed closely enough after the Detroit so that the cirques of the latter still existed. The Wisconsin ice freshened the cirques and valley profiles but did not affect many of the higher surfaces covered by the Detroit ice. The present cirques in the Western Cascades were probably formed during the Detroit glacial epoch, and freshened during the Wisconsin glaciation.

In post glacial time, in the particular area studied, only normal erosional processes have been in operation. The extensive gravel deposits laid down during the glacial stages have been partially removed and the North Santiam has intrenched itself slightly below its flood plain from Mehama east.

SUMMARY

The section under consideration extends from the Willamette River south of Salem eastward to the summit of the Cascade Range north of Mt. Jefferson. The data secured in the course of this investigation lend support to Callaghan's theory that the Cascade Range in Oregon consists essentially of two longitudinal divisions which he named the Western Cascades and the High Cascades.

The Western Cascade volcanics which range in age from Eocene to Miocene, have been gently compressed into folds trending northeast-shouthwest and intruded in places by dioritic rocks. Structural continuity of the Salem Hills homocline with the Western Cascades is demonstrated. The eastern margin of the Western Cascades is the north-south trending Cascade fault which follows the valley of the North Santiam River, and cuts off the Western Cascade structures. The rocks of the High Cascades are divided into four series. The occurrence of centers of eruption along the Cascade fault, the Cascade fault itself, and the alignment of both the earlier and later High Cascade volcanoes imply that the distribution of High Cascade vents is controlled by a north-south system of fractures, possibly of the Basin Range type, which intersects the Western Cascade structures obliquely.

Structural surfaces are everywhere apparent in the High Cascades, and are well exhibited in the western portion of the Western Cascades. The main mass of the Western Cascades is in the early mature stage of the erosion cycle. The High Cascades have been deeply trenched by glaciated valleys at right angles to the axis of the range.

Three epochs of glaciation are recognized and correlated tentatively with the Sherwin, Tahoe, and Tioga of the Sierra Nevada. All the large glaciers headed on the western slopes of the High Cascades and proceeded down the Western Cascade valleys for varying distances.

It is demonstrated on the basis of structural evidence that the Western Cascade mass was raised essentially to its present position in pre-Middle Pliocene time, and deeply eroded before the earliest High Cascade lavas were erupted. No indication of a widespread erosion surface in this part of the Cascade Range was found. Evidence is given suggesting that before the extrusion of the High Cascade lavas the drainage divide of the Western Cascades was west of Detroit. Eruption of the High Cascade lavas effected integration of the present North Santiam drainage system by damming the eastern streams, forcing them over a low divide to the west. Several other less important drainage changes are described.

Petrographically the bulk of the High Cascade lavas are basalts. Andesitic lavas occur mainly in the younger peaks. The Western Cascade lavas are largely labradorite andesites, although dacites and rhyolites occur.

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THE STRATIGRAPHY AND PALEONTOLOGY

of the

SALEM HILLS, OREGON

Thomas P. Thayer

February, 1934

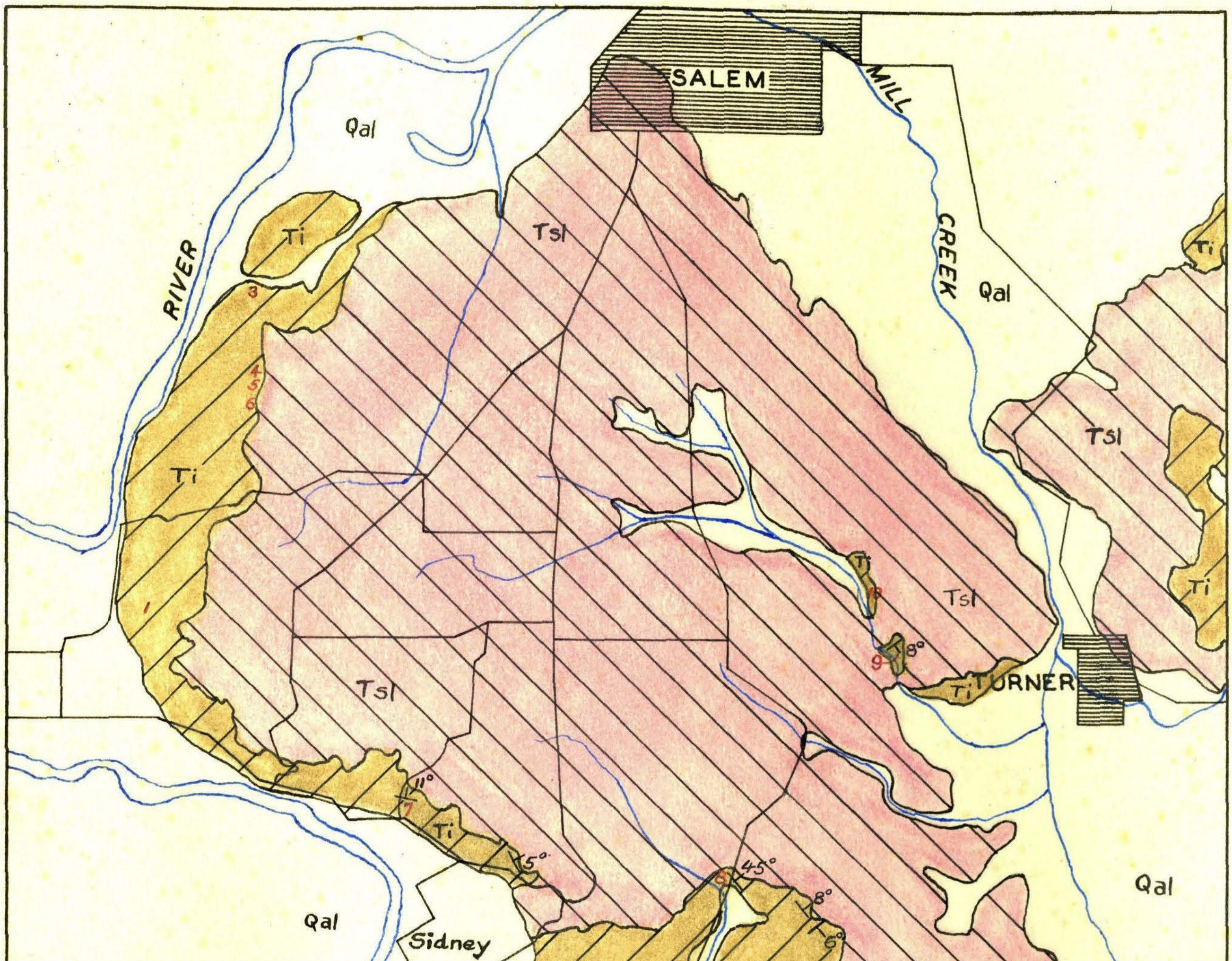
General Geology

The part of the Salem Hills considered in this paper lies south and east of the Willamette River south of Salem, Oregon. The area is in the middle of the Willamette Valley, midway between Eugene and Portland, and occupies the southeastern part of the Salem quadrangle of the Topographic Atlas of the United States.

Structurally the Salem Hills are very simple. They are a warped homocline, maturely eroded, of lavas dipping northeast; the dip varies from 0° to 3°. The lavas lie unconformably on folded tuffaceous marine sediments of the Illahe formation. The angle of dip in this formation is about ten degrees, but dips of forty-five degrees were recorded. The strike of the beds is generally northwest, but varies greatly from place to place. The basal contact relations of the lavas indicate that the pre-lava landscape was in the mature stage of the erosion cycle.

The Illahe Formation

The Illahe formation is best exposed on the western scarp of the hills, but also is revealed in scattered patches east of the hills, where the lava capping has been cut through. In the western portion of the area the Illahe formation is mainly composed of sandy or silty tuffaceous sediments, with occasional thin gravelly waterlaid tuff beds. The maximum pebble size in such beds is less than one half inch; the pebbles are of nearly all types of volcanic rocks, andesites being dominant. Fine sandstone is the predominant type of sediment, some of these beds being very well cemented with carbonate. It is gray when fresh,


SALEM
MILL
CREEK
Qal
Ti
RIVER
Tsl
Ti
Tsl
TURNER
Tsl
Qal
Sidney
Qal
WILLAMETTE
Tsl
Ti
Marion
GEOLOGIC SKETCH MAP OF SALEM HILLS
1 1/2 0 1 2 3 4 5 Miles
EXPLANATION
RECENT
MIOCENE
OLIGOCENE
Qal Alluvium
Tsl Stayton Lavas
Ti Illahe Formation

and brown when weathered. The fragments composing the coarser facies are plagioclase feldspar, quartz, occasional pyriboles and magnetite, in addition to minute rock fragments. The incomplete weathering of the feldspars suggests rather rapid transportation and deposition of the sediments. True shales are entirely absent. The following section a mile north of Sidney is characteristic;

Section at Locality #7

Thickness	
	Basalt
120'	Massive poorly bedded coarse sandstone
3'	Highly cemented medium grained gray sandstone containing abundant pelecypods and some gastropods. Only fossiliferous horizon in section
50'	Fine silty sandstone
12'	Hard waterlaid tuff with half inch fragments
90'	Fine silty sandstone
3'	Coarse waterlaid tuff similar to above
10'	Fine silty sandstone
2'	Coarse waterlaid tuff
25'	Fine soft sandstone
2'	Waterlaid tuff, pebbles to ¼", well cemented
45'	Silty sandstone
1'	Fine dark gray highly cemented siltstone
	Bottom of exposed section
363'	

The formation is much more variable in the eastern part of the area. Very fine white, chalky-textured tuffs are in places interbedded with coarse sandstones and tuffs, the latter having fallen in water but not been reworked. These fine tuffs feel like yeast when wet. Fossils occur even in the coarsest materials.

Northeast of Turner the sediments are more heavily bedded, predominantly pebbly, and apparently contain no fossils. Where the sediments underlying the lavas again appear, ten miles or so east of Turner, they are definitely continental. A flat syncline in the lavas east of the Salem Hills makes it impossible to trace the two facies continuously into each other, but structural relations to the lavas and the type of deposits strongly indicate that the formations are stratigraphically equivalent. Northeast of Scio heavy conglomerates and tuffs interbedded with fine shaly beds are well exposed in Thomas Creek. The relations, and the fact that no fossils were found indicate near shore continental deposition.

Fossils, though not uncommon in the Illahe formation, are well preserved only in the well cemented beds. Most of the remains are casts and molds; specific identification of forms is therefore rather difficult if not impossible. Some thin lenses are almost entirely composed of fragmentary casts, but actual shell material is rarely preserved because of the deep weathering.

Table I

Complete List of Identified Species from the Illahe Formation in the Vicinity of the Salem Hills

Pelecypoda	I. L.	1.	2.	3.	4.
Acila shumardi Dall	1,5,6	x	x	x	x
Leda lincolnensis Weaver	4,11			x	
L. washingtonensis Weaver	4			x	
Macrocallista pittsburgensis Dall	4,10	x	x	x	x

	I. L.	1.	2.	3.	4.
Pitaria clarki Dickerson	7			x	
Solen cf. eugenensis Clark	7		x		
Tellina eugenia Dall		x			
T. cf. pittsburgensis Clark	8	x			
Thracia cf. condoni Dall	4	x	x		x
Yoldia oregana Shumard					x
Diplodonta cf. pariles Conrad	4,6				
Gastropoda					
Molopophorus gabbi Dall	6	x		x	
Natica washingtonensis Weaver	6			x	
Strepsidura lincolnensis Weaver	4			x	
S. washingtonensis Weaver	7			x	

I. L. Illahe localities
1. Pittsburg Bluff
2. Eugene Formation
3. Linconl-Porter horizon, Washington
4. San Lorenzo, Central California

Many of the species of this fauna are highly ornamented forms which are recognized as occurring through a rather limited range in the Lower or Middle Oligocene. The fauna is a warm water tropical type.[1]

Division of the Illahe formation into distinct faunal zones is not possible at this time. The irregular folding of the formation and discontinuity of exposures render such division difficult at best. Furthermore, the present collections are admittedly not sufficiently complete to warrant such a division on paleontologic graunds.

Nuculacea are the most abundant and widespread forms. They are restricted almost exclusively to the fine silty sandstones, although Acila shumardi was found in the pebbly conglomerate at Locality # 1. No forms belonging to this superfamily were found in the coarse sandstone at Locality #7 where Pitaria clarki is so abundant. Gastropods are in general less

1. Dickerson, R.E., Climate and its influence upon the Oligocene faunas of the Pacific Coast: Cal. Acad. Sci. vol. 7, p. 162, 1917.

numerous than pelecypods; they are extremely plentiful but very poorly preserved in lenticular masses at Localities #3 and #4. No gastropods were found east of the Salem Hills proper; this suggests that they lived only in the quieter waters offshore.

Correlation

The distinctness of the faunal assemblage permits rather definite correlation of the Illahe Formation with formations in Washington, California, and other parts of Oregon. The occurrence of Illahe species elsewhere has been indicated in Table I. The similarity of the faunas of the Pittsburg Bluff, Eugene, and Illahe formations would undoubtedly be much closer if complete faunal lists of the three were available.

Schenck has left little doubt as to the correlation of the Pittsburg Bluff deposits with the Eugene Formation.[1] The occurrence in the Illahe Formation of diagnostic species common to both the above formations indicates that deposition was simultaneous in all three areas. The exposures at Holmes Gap which were described by Schenck and identified as Oligocene[2] ~~are~~ undoubtedly belong to the Illahe Formation.

The affinities of the Oregon Oligocene faunas to those of Washington and California has been indicated by Clark.[3] He found that forty percent. of the species found in the San Lorenzo of Central California occurred also in the Middle Oligocene of Oregon and Washington. He found also that the fauna as a whole was entirely distinct from the Eocene (Tejon), whereas less

1. Schenck, H. G., Stratigraphic relations of the Western Oregon Oligocene formations: U. C. Publ. Dept. Geol. Sci vol. 18, p. 38,
2. Idem, p. 41.
3. Clark, B. L., The San Lorenzo series of Middle California: U. C. Publ. Dept. Geol. Sci. vol. 11, p. 95, 1918.

than ten percent. were found in the Miocene or younger beds.

In view of the above data there can be little doubt that the Illahe Formation of the Salem Hills is to be correlated with the Eugene and Pittsburg Bluff Formations in the Middle Oligocene. These formations in Oregon are in turn closely related in time to the San Lorenzo Formation of Middle California and the Lincoln-Porter Horizon of Washington.

The Stayton lavas which overlie the Illahe Formation are regarded as Middle Miocene. The magnitude of the unconformity between them and the Illahe Formation indicates that the lavas are considerably younger than the marine sediments. Furthermore, Middle Miocene Columbia River lavas have been traced to the south edge of the Estacada quadrangle in the west side of the Cascade Range. The Stayton lavas thicken toward the north, and are very likely the thin border of the Columbia lavas.

Fig. 1. Fine white tuffaceous facies of Illahe Formation west of Turner. Three foot rule for scale.

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