UNITED STATES
DEPARTMENT OF THE INTERIOR
U.S. GEOLOGICAL SURVEY

QUATERNARY GEOLOGY OF THE BELLEVUE AREA

IN BLAINE AND CAMAS COUNTIES, IDAHO

by

Dwight Lyman Schmidt

Open File #625
1962
62-0120

Prepared partly in cooperation
with the U. S. Atomic Energy Commission

This report is preliminary and has not been edited for conformity with U.S. Geological Survey standards. Some of the nomenclature used in this report does not conform with that in use by the Survey.

QUATERNARY GEOLOGY OF THE BELLEVUE AREA IN
BLAINE AND CAMAS COUNTIES, IDAHO

by

Dwight L. Schmidt

ABSTRACT

The Bellevue area covers about 350 square miles of a foothill belt between the Rocky Mountains to the north and the Snake River Plains to the south. Complexly deformed impure quartzites and limestones of the Mississippian Milligen and Pennsylvanian-Permian Wood River formations were intruded by large bodies of quartz diorite and granodiorite along regional structures trending northwesterly; the intrusions are part of the Cretaceous Idaho batholith. Erosional remnants of the Challis volcanics, dominantly latitic to andesitic in composition and early(?) to middle Tertiary in age, rest unconformably on the older rocks. A sequence of Pliocene rhyolitic ash flows and basaltic lava flows unconformably overlies the Challis and older rocks and is in turn unconformably overlain by olivine basalt of late Pliocene or early Quaternary age. The main valleys of the area, partly erosional and partly structural in origin, are underlain by late Quaternary olivine basalt flows (Snake River basalt)

and intercalated lacustrine, fluvial, proglacial sediments.

The Big Wood River, the master stream of the area, flows southward through a narrow steep-sided valley in the mountainous country north of the Bellevue area and debouches into a broad alluvial valley, the Wood River Valley, in the foothill belt. The valley has the shape of an isosceles triangle with a ten mile long, east-west base consisting of a ridge of Pliocene volcanics which separates the valley from the Snake River Plains to the south. The river now flows through a narrow gap in the southwest corner of the triangle. A similar, but wider, gap around the east end of the ridge was formerly occupied by the river. The river has been shifted back and forth between these two gaps at least four times during an interval in which six late Quaternary basalt flows erupted in the Bellevue area. Two of the flows caused direct diversion of the river and another was influential in bringing about a diversion on an aggradational fan upstream from the lava dam.

During each of two late Pleistocene stages of glaciation in the mountains north of the Bellevue area, large volumes of proglacial outwash caused the Big Wood River to build a large aggradational fan in the Wood River Valley. Just prior to the Bull Lake stage the river, flowing out the east gap, was blocked but not diverted by the youngest basalt flow in the Bellevue area. During the proglacial aggradation, the river shifted widely on its fan and spilled alternatively out both the east and west gaps. After the Bull Lake stage, the west gap had an advantageous base level relative to the lava-blocked east gap, and the river cut down in the west gap. After the second, Pinedale, proglacial aggradation in the

Wood River Valley, the west gap still maintained an advantageous base level, and the river again cut down in the west outlet valley where it remains today.

Periglacial deposits completely dominate the sidestream valleys of the Bellevue area. They formed under a rigorous climate during the Pinedale stage, when slope erosion accelerated by frost activated processes caused aggradation of valley floors by local detritus. Even at present the larger sidestreams are so choked with detritus that the streams have not regained control of their valley floors.

Recent basalt, comparable in age to the younger flows of the Craters of the Moon National Monument, spread from a rugged, cratered vent several miles south of the Bellevue area. Using degree of weathering, erosion, and soil development as a basis of comparison, this flow provides an end point for estimating the relative ages of the six late Quaternary flows in the Bellevue area.

QUATERNARY GEOLOGY OF THE BELLEVUE AREA IN BLAINE AND CAMAS COUNTIES, IDAHO

by

DWIGHT L. SCHMIDT

1961

TABLE OF CONTENTS

Appendix consisting of four documentation figures is available from author on request: DLS, USGS, Box 25046, MS 913, Denver, CO 80225

ILLUSTRATIONS

Documentation figures 40, 41, 42, and 43
of the Appendix are available from
the author on request

DLS, USGS Box 25046, MS 913
Denver, CO 80225

QUATERNARY GEOLOGY OF THE BELLEVUE AREA IN BLAINE AND CAMAS COUNTIES, IDAHO

INTRODUCTION

The Bellevue area is located in south-central Idaho in a foothill belt between the Sawtooth Mountains to the north and the Snake River Plains to the south (fig. 1). In spite of an altitude of about 5000 feet and a semi-arid climate, the valley floor of the Big Wood River, the principal stream, supports an abundant growth of grains and foliage plants where irrigation water is available. The surrounding rolling foothills, extending up to 6000 feet, are covered with sage and grass, and high peaks commonly over 7000 feet along the northern border of the area have forested north-facing slopes. The principal town in the area is Bellevue, about 4 miles south of Hailey, Idaho.

Complexly deformed Carboniferous sedimentary rocks, cut by granitic bodies associated with the Idaho batholith, underlie most of the northern part of the area. Scattered erosion remnants of early and middle Tertiary volcanics of intermediate composition, the Challis volcanics, rest unconformably on the granitic and older sedimentary rocks in the foothill area. Pliocene silicic and basaltic volcanics unconformably

FIGURE 1. Index map of Idaho showing the location of the Bellevue area, Blaine and Camas counties, Idaho.

overlie the Challis and older rocks and are in turn overlain unconformably by late Pliocene or early Quaternary olivine basalts which make up most of the southern part of the area. Finally, the main valleys, which are in part eroded in the older rocks and in part of structural origin, are partly filled by a late Quaternary sequence of olivine basalt flows and associated sediments.

Radioactive placer deposits were discovered by prospectors in gravel along Camp Creek in the fall of 1953; the radioactivity is due to uranium-bearing thorite. Preliminary sampling by the U. S. Bureau of Mines was so favorable that in May and June, 1954, the Bureau drilled and blocked out the principal deposits on Camp and Rock creeks (A. F. Robertson and R. M. Storch, 1955a and 1955b). That same year, from June through August, the writer investigated the geology of the placer deposits for the U. S. Geological Survey on behalf of the Raw Materials Division of the U. S. Atomic Energy Commission. The study was continued in the summer of 1955 and for several months in subsequent years.

In 1954 the geologic mapping was largely restricted to the modern and ancient gravels of Camp and Rock creeks and to the granitic rock from which the radioactive minerals were derived. It became evident in the course of this work that, because deposition of the placer gravel was caused in part by the blockage of drainage lines and in part by climatic changes associated with glacial stages, an understanding of the origin of the placer deposits required a comprehensive study of the Pleistocene history. The study was extended in 1955 and later years to the Wood River Valley where the complex interrelationships between

Quaternary volcanism and sedimentation could be worked out with the aid of numerous drill logs of water wells in the valley and could be tied to glacial stages in the mountains to the north.

This thesis is limited to the academic aspects of the Quaternary geology. The older rocks, from the Carboniferous sedimentary rocks through the Pliocene silicic volcanics, are briefly described to make the geologic map self-explanatory (pl. 1) and to provide a geologic setting for the discussion of Quaternary geology. Local stratigraphic names have been assigned to 2 units of Cretaceous intrusive rock, 4 units of the Pliocene volcanics, and to 8 units of the Quaternary basalt flows. Application of the principles of stratigraphy to the volcanic units and use of the concept of grade and base level control of deposition of the associated sediments provide a key to the solution of the Quaternary history.

Sincere appreciation is expressed to Professor J. Hoover Mackin for help in all stages of the investigation; to Professors P. Misch, J. D. Barksdale, and V. S. Mallory for critical reading of the manuscript; and to the other members of the geological faculty of the University of Washington for discussion of special problems. The writer was ably assisted in the field in 1954 by Paul L. Williams. The cooperation of R. M. Storch, A. F. Robertson, and other personnel of the U. S. Bureau of Mines and of R. O. Smith and other personnel of the Ground Water Branch of the U. S. Geological Survey is gratefully acknowledged. Thanks is extended to many friends, especially to Bill Brooks, for the many kindnesses during my stay in Hailey.

PRE-QUATERNARY ROCKS

Carboniferous sedimentary rocks

A sequence of Carboniferous sedimentary rocks cropping out in the eastern part of the Bellevue area has been described by Hewett (Umpleby, Westgate and Ross, 1930, p. 209) and Anderson (Anderson, Kiilsgaard and Fryklund, 1950, p. 2) and is part of a body of Carboniferous rocks which extends far to the east. Lindgren (1900, p. 193) named these rocks the Wood River formation, but Umpleby, Westgate, and Ross (1930, p. 25) restricted that term to an upper unit consisting of calcareous and quartzitic sandstones containing small amounts of conglomerate, shale, and dolomitic limestone; an underlying unit of fine-grained argillaceous rock was called the Milligen formation. The rocks of this eastern area, designated "Carboniferous sedimentary rocks, undifferentiated," on plate 1, are about half Wood River formation and half Milligen formation; they are reproduced on plate 1 from Anderson's map (Anderson, Kiilsgaard, and Fryklund, 1950, fig. 2).

The age of the Milligen formation is considered to be Mississippian by Westgate and Ross (Umpleby, Westgate, and Ross, 1930, p. 29, 34) and is probably largely Lower Mississippian according to Skipp (1961, p. 376) (pre-middle Mississippian, Thomasson, 1959, p. 1687). The age of the Wood River formation is Middle and Upper Pennsylvanian and Lower Permian (Desmoinesian-Wolfcampian) according to Bostwick (1955, p. 941).

A western area of Carboniferous sedimentary rocks extending from the headwaters of Croy Creek, across the head of Rock Creek to Reed Creek is essentially a roof pendant surrounded on all sides except the north by the Idaho batholith. The rocks are typically exposed on the west flank of Kelly Mountain where a monotonous section, about a mile thick, consists of quartzites and limestones containing varying amounts of calcareous, siliceous and argillaceous admixtures. Most of the quartzites are highly siliceous, and some of the limestones, especially the impure varieties, are silicified. The fresh rocks are generally white to medium gray and weather white to light gray and rarely pink. They are commonly massive or thick-bedded, but there are thin layers in places. On the basis of the dominantly quartzitic and calcareous lithology, the rocks in the western area are probably part of the Wood River formation; there are some small areas of Milligen lithology.

The bedding is generally steep to vertical and strikes N. 30° W., paralleling the granite contact west of Kelly Creek. Silicified breccia zones a few inches to many feet thick commonly parallel the bedding and suggest that the section is faulted. In some places the bedding attitudes are erratic, changing by 90° in strike and dip over distances of a few tens or hundreds of feet.

The rocks in the western area in general are more indurated and siliceous than the rocks in the eastern area; the difference is probably caused by contact metamorphism associated with the Idaho batholith. The outcrop pattern along a large part of the intrusive contact indicates that the contact generally dips gently eastward, suggesting that in many places the batholith is at relatively shallow depths. Fine-grained mica schists and tactites are common in a belt as much as several hundred feet wide along the contact.

A detailed stratigraphy for the Milligen and Wood River formations has not been established in the Bellevue area because distinctive units within the formations have not been found. In working with these formations in the mineral-rich area around Hailey and Bellevue, Hewett, Umpleby (Umpleby, Westgate and Ross, 1930), and Anderson (Anderson, Kiilsgaard and Fryklund, 1950) were chiefly concerned with detailed local problems at certain mines, and where a stratigraphy was established at one mine, it could not be extended to nearby mines, much less extended to larger areas. Fryklund (Anderson, Kiilsgaard and Fryklund, 1950, pl. 3), mapping on a scale of 1:2400, about 3 miles east of Kelly Mountain, subdivided a one square mile area into 7 map units, and found that most contacts between these units are faults. In the absence of distinctive marker beds, correlation across the faults was impossible.

Cretaceous intrusive rocks

General statement.--Intrusive rocks underlie the largest single area of outcropping rock in the Bellevue area. Two types, quartz diorite and granodiorite according to the Johannsen classification, make up 95 percent of the exposed part of the intrusions. The quartz diorite intrudes Carboniferous sedimentary rock forming a single intrusive body.

The granodiorite intrudes both the Carboniferous rock and the quartz diorite, forming what is believed to be a single intrusion exposed principally in three areas separated by roof pendants of Carboniferous rock or by younger volcanic rocks. Alaskite, pegmatite, aplite, lamprophyre, and hydrothermal quartz-vein deposits cut the quartz diorite and granodiorite; all of these rocks are genetically related to a single intrusive episode, namely, to some phase of the emplacement of the Idaho batholith. Later, porphyry dikes of probable Tertiary age (Anderson, Kiilsgaard and Fryklund, 1950, p. 5, 6) independently intruded the older rocks, but they are minor features and will not be discussed.

The occurrence and distribution of heavy accessory minerals in the intrusive rocks have been studied in detail as a part of the investigation of radioactive placer deposits because the placer minerals originate largely in the granodiorite, but the results of the study will be reported in another paper. For the present, only a brief description of the granitic rocks is given.

The granitic rocks weather readily to a granite sand and are not well exposed. The rock of the rare, ledge outcrops is either obviously atypical or cannot be assumed to be typical. The most reliable fresh rock comes from mine adits and mine dumps throughout much of the area; few of the mine workings are open, from which hardrock specimens can be collected in place.

Croesus quartz diorite.--The Croesus quartz diorite is named for the Croesus Mine west of Bellevue from where typical rock was collected and chemically analyzed by Waldemar Lindgren. In hand specimen it is a gray, uniformly textured rock with an average larger grain size of 3 millimeters (range, 1 - 4 mm). It is moderately foliated in many places. The rock is a pyroxene(augite-hypersthene)-hornblende-biotite quartz diorite with [redacted] ilmenite, magnetite, sphene, allanite, apatite, and zircon (Anderson, Kiilsgaard, and Fryklund, 1950, p. 4). Modal compositions given by Hewett (Umpleby, Westgate and Ross, 1930, p. 212) and Anderson (Anderson, Kiilsgaard, and Fryklund, 1950, p. 4), and a normative analysis by Lindgren (1900, p. 195) are shown in table 1 and are plotted for comparison on a modified Larsen ternary diagram in figure 3. Deuteric and hydrothermal alteration products generally include minor amounts of epidote, chlorite, sericite, and sulfides. In small areas, a lighter-colored facies of the quartz diorite contains 5 to 10 percent potassium feldspar and no pyroxene or hornblende; biotite is abundant.

The Croesus intrusive body extends about 5 miles along a generally northwesterly regional structural trend in the Carboniferous sedimentary rock and has an average width of 2 1/2 miles. In the Minnie Moore Mine the quartz diorite overlies interbedded argillites, quartzites, and limestones of the Milligen formation dipping 30° to 50° westward, and it cuts the bedding at a small acute angle suggesting a sill-like relationship (Umpleby, Westgate and Ross, 1930, pl. 28). Farther north the quartz diorite cuts the Carboniferous sedimentary rock at high angles. In general the intruding magma took advantage of complex, pre-intrusive structures in the sedimentary rock. Post-intrusive, northwest trending faults displace the quartz diorite-Carboniferous sedimentary rock contact with moderate down throw to the northeast (Umpleby, Westgate and Ross, 1930, p. 235).

Hailey granodiorite.--The Hailey granodiorite is named for the Hailey Gold Belt located in the headwater area of Camp, Croy and Rock creeks. Relatively fresh specimens of the Hailey granodiorite are found on the dumps of the Tip Top Mine (sec. 17, T. 1 N., R. 17 E.) as well as in mine adits and dumps throughout the Gold Belt area. More typically disintegrated rock is well exposed in road cuts as, for example, along the Croy-Camp Creek road west of Kelly Gulch. The rock is light gray in color, essentially directionless and has a uniform granular texture with an average larger grain size ranging from 2 to 5 millimeters. The rock is a hornblende-biotite granodiorite containing magnetite, sphene, apatite, zircon, allanite, and uranothorite as accessory minerals and containing chlorite, epidote, muscovite, sericite, anatase, rutile, and pyrite as later alteration minerals. Modal compositions of the granodiorite from different localities in the Bellevue area and by various authors are given in tables 1 and 2.

In thin sections and on polished rock slices stained with sodium cobaltinitrite, the Hailey granodiorite appears to have been partly recrystallized endomorphically under the influence of late magmatic and postmagmatic solutions (Anderson, 1942; Gilluly, 1933). Large crystals of plagioclase, exhibiting many delicate growth zones, are clearly relicts of an early crystallization. These crystals have been reconstituted to a sodic andesine composition; the original growth zones in most crystals have been either altered to a pseudozonation or obliterated. Twinning is complex, discontinuous, and commonly not sharp. That the crystals have been subjected to pressures and crushing is shown by bent twin planes, by small parallel fractures filled with quartz, potassium feldspar and epidote, and by a general roundness of shape.

Microcline and microperthite are the common forms of potassium feldspar. In the relatively small amount of rock low in potash, the potassium feldspar occurs as intergranular wedges and small grains or rarely as large porphyroblasts, but where the potash content is near the average amount, the potassium feldspar occurs as an equigranular mosaic with quartz and plagioclase, in general, at the expense of the early plagioclase. Quartz also occurs either intergranularly or as large aggregates which are part of the quartz-feldspar mosaic; these aggregates are recrystallized cataclastic grains exhibiting strong undulatory extinction.

Both quartz and potassium feldspar clearly replace plagioclase. Commonly several linear elements, millimeters to centimeters long, can be seen in most thin sections along which plagioclase crystals have been truncated and along which quartz and potassium feldspar have crystallized; such linear elements are obliterated within several millimeters or centimeters along their length by a more random crystallization of quartz and potassium feldspar. These linear elements are clearly relict fracture planes caused by shearing or fracturing which aided the endomorphic recrystallization. Myrmekite is abundant between potassium feldspar and plagioclase and further suggests that potassium feldspar has replaced plagioclase. Relationships suggest that a large part of the potash, some part of the silica, and perhaps some soda were metasomatically introduced during endomorphism; these constituents probably originated from deeper in the consolidating mass.

Hornblende is commonly subhedral and largely biotitized, or it is euhedral and little altered suggesting that two different generations of hornblende crystals are present. Biotite varies from shredded, ragged grains, which are bent and twisted, to fresher-appearing, euhedral grains. In many places it is partly altered to chlorite. Scattered crystals of sphene are either anhedral or perfectly euhedral and are as much as 5 millimeters long. The euhedral sphene lies across the relict fractures and evidently crystallized relatively late. Allanite is also late; in places long, slender euhedral crystals have grown across or through the late quartz-potassium feldspar mosaic. Magnetite occurs in roundish grains commonly in association with biotite crystals or hornblende-biotite clots suggesting that it is in part late and related to the biotitization of hornblende.

Anhedral crystals of epidote and quartz, as large as 2 millimeters, fill late fractures cutting across plagioclase crystals and the adjacent quartz-potassium feldspar mosaic. The epidote also occurs as very fine-granular aggregates in much later shear zones. Rarely, relict muscovite-coated joints, in which the muscovite crystallized in open space, are healed and in places obliterated by coarse-granular quartz.

In weathered outcrops the Hailey granodiorite typically appears to be essentially directionless in structure and homogeneous in texture and composition. However, a comparison of hand specimens and stained rock slices collected throughout the Bellevue area, indicates that structures and textures vary from directionless granular to foliated gneissose. All degrees of recrystallized cataclasis and shearing can be found.

The Hailey granodiorite extends, presumably as a single intrusive mass, from the Picabo Hills across the entire southwestern half of the Bellevue area to about 10 miles northwest of the area; it is separated from the main mass of the Idaho batholith by a strip of Carboniferous sedimentary rock about one mile wide. During the present study, modal analyses representing 13 typical granitic rocks, collected along a 25-mile-long section from the Princess Blue Ribbon Mine in the northwestern part of the area to the south-central Picabo Hills, show a wide variation about an average granodioritic composition; this variation about the average is shown in two ternary plots of different mineral systems in figures 2 and 4 (table 2). The modal values are also plotted against distance along the section in figure 5; seemingly systematic variations in mineral content occur.

The interpretation of the wide variation in composition supports the hypothesis of endomorphism, either an endomorphic recrystallization or an endomorphic metasomatic crystallization with addition of potash, some silica and perhaps some soda. "Typical specimens" as applied to the rocks used for modal analyses means that the most granitic-appearing, stained rock slices (as compared with several hundred slices of all varieties of the Hailey granodiorite examined) were used after being assured that these samples were representative of the outcrop area from which they were collected. This sampling excludes those less potassic and (or) foliated rocks which even if not as abundant are representative of some outcrop areas and which may be more closely related to the composition of the original magma.

The Hailey granodiorite, then, has a much wider range of composition than is suggested by the 13 "typical specimens," and it has a history which is far more complex than the average composition of the 13 "typical specimens" might suggest. For example, when the average mode (E, table 2) is recalculated and plotted on the ternary system, albite-potassium feldspar-quartz, the plot falls in the minimum melting trough (fig. 6), but this need not mean a crystal-"magmatic" liquid equilibrium, as is commonly suggested (Bowen, 1954, p. 11), for several reasons. (1) The experimental minimum melting trough represents one set of P-T conditions, namely 1000 bars and 740°C; that this set should fit conditions during the consolidation of the Hailey granodiorite seems fortuitous. (2) The wide scatter of individual modes about the average Hailey granodiorite as noted in figure 6 does not suggest a systematic convergence to a common eutectic. (3) The complex endomorphic replacement textures of the Hailey granodiorite did not result from a crystal-"magmatic" liquid equilibrium; rather they probably represent a more restricted equilibrium between crystals of the consolidated magma and postmagmatic solutions.

Various authors have given what they considered representative modal compositions of other parts of the Hailey granodiorite. These are given in table 1, and the estimated average modes are plotted in figures

2 and 3 for comparison with the modes determined in this study.

The shape of the Hailey granodiorite intrusion was largely determined by complex pre-intrusive structural features in the Carboniferous sedimentary rock. The major structure controlling the eastern margin of the main body is a northwest trending, steeply dipping fault which north of Croy Creek along Kelly Gulch forms a major intrusive fault. South of Croy Creek the intruding magma spread laterally probably along a generally flat structure and intruded the Croesus quartz diorite west of Bellevue. Along the head of Diebenow Creek the intrusive contact dips about 10° eastward under the quartz diorite, but west of Diebenow Creek the granodiorite contact rises steeply between the quartz diorite and a large roof pendant of Carboniferous sedimentary rock in the eastern headwaters of Rock Creek. In the Picabo Hills the contact is again a steeply dipping intrusive fault.

Alaskite.--Small, widely spaced alaskitic bodies occur in the Hailey granodiorite, more abundant bodies occur along intrusive contacts between the granodiorite and the Croesus quartz diorite and between the granodiorite and the Carboniferous rock, and in places swarms of dikes trending north-northeast extend discontinuously for thousands of feet into the adjacent quartz diorite. For example, abundant large dikes cut the Croesus quartz diorite west of Scorpian Gulch (Anderson, Kiilsgaard and Fryklund, 1950, p. 5). The bodies within and at the granodiorite contact are roughly pod-shaped or irregular masses a few feet to many feet across, whereas, the dikes in adjacent rocks are inches to tens of feet wide and may be hundreds of feet long. The alaskite is white and leucocratic and has a mixed granular texture which varies from fine- to coarse-grained. Contacts with the granodiorite are compositionally and texturally gradational and commonly vague; in the adjacent country rock the dikes of alaskite have sharp contacts. Plagioclase (sodic andesine) and quartz are essential minerals, and potassium feldspar chiefly replaces the plagioclase along irregular, yet commonly well defined, fronts which in many places cut across older plagioclase-quartz textures (for example, an andesine-quartz, graphic granite texture) and structures. A few large biotite, sphene and other accessory-mineral crystals occur in the coarse-grained facies of the alaskite.

The alaskitic dikes cutting the Croesus quartz diorite and the Carboniferous rocks were probably developed along fractures caused by stretching of the roof at the time of the intrusion; essentially quartz and plagioclase, derived from the consolidating magma during a late magmatic stage, were originally emplaced in these fractures. On the other hand, the pods of alaskite along the contacts and within the granodiorite were probably developed at sites of internal stress where incipient strain resulted in mechanical and (or) chemical segregation of late magmatic, felsic constituents before long fractures could form in the consolidating crystal mush. The time of the formation of the alaskite is believed to be prior to the endomorphic potassic feldspathization because potassium feldspar occurs as a replacement of the early plagioclase. In fact, the lack of primary potassium feldspar in the alaskite is strong evidence that the initial magma contained little potash.

Aplite and pegmatite.--Aplite and pegmatite dikes occur throughout the granodiorite, where in general they are inches to many feet wide and extend for several feet to many tens of feet in length. They are more or less uniformly spaced, tens to hundreds of feet apart, throughout the granodiorite body, but in some places they occur in swarms in which they are several times as abundant. The dikes generally trend northeasterly. The rock forms a resistant surface float and stream gravel, but probably because of intense fracturing, the dikes rarely crop out.

The rock is characteristically colored pink by feldspars. Potassium feldspar is an essential mineral and is dominant relative to plagioclase and quartz. Locally large crystals of any of the accessory minerals common to the granodiorite may occur in the pegmatite.

Aplite and pegmatite occur together in most dikes although rarely are the two systematically zoned. In some dikes the aplite has been brecciated and recrystallized. In other dikes aplitic streaks appear to be finely crushed pegmatite which has been recrystallized suggesting that the grain size of some aplitic rock may be controlled by shearing. Contacts between aplite-pegmatite and wall rock may appear sharp in hand specimen, but they are irregular on the scale of the grain size of the aplite.

The aplite and pegmatite are later than the alaskite. Fractures were probably opened by stretching in a wide peripheral zone by magma at a greater depth at a time when the peripheral zone was solid enough to take long fractures. This time was probably late endomorphic because, although potassium feldspar is an essential mineral in the aplite-pegmatite, the endomorphic recrystallization in the granodiorite was no longer active. The aplite-pegmatite material was probably derived from a less consolidated part of the intrusive mass below the level of the rock now exposed in the Bellevue area.

Lamprophyre dikes.--Well defined lamprophyre dikes, inches to several feet wide, cut the granodiorite in many places throughout the mass but are perhaps most abundant in the Hailey Gold Belt. The rock is dark gray to greenish black, is fine-grained, and commonly contains a few phenocrysts or xenocrysts of mafic minerals, quartz, or feldspar.

The dikes are closely associated with a northwest shear structure (most commonly about N. 60° W.) which occurs throughout the granodiorite in the Bellevue area and which is especially associated with quartz vein deposition in the Gold Belt. The lamprophyre is later than the most prominent shearing and is both earlier and later than the quartz-vein deposition.

Age.--The Hailey granodiorite and the Croesus quartz diorite are much younger than Lower Permian, the youngest Carboniferous sedimentary rock which was strongly deformed prior to being intruded by the granitic magmas. The granitic rocks are much older than middle Tertiary (Oligocene (?)-Miocene(?)), the age of the overlying volcanics which are separated from the granitic rocks by a major unconformity. A lead-alpha age of about 114 million years has been determined for zircon from the Croesus quartz diorite. This age is well within the range of the probable mean age of the Idaho batholith, 108± 12 million years, which was determined from individual ages of 16 rocks collected from different places in the batholith (Larsen and others, 1958, p. 51).

Hypothesis of late magmatic and postmagmatic events.--Seven distinctive phases of late magmatic and postmagmatic fracturing are recognizable along with a proposed late magmatic, endomorphic, and hydrothermal history of the Hailey granodiorite; the sequence of fracturing, numbered consecutively below, provides a good relative time marker. The intruding Hailey magma was probably quartz dioritic or calcic-granodioritic in composition. (1) The alaskitic dikes in the Croesus quartz diorite and Carboniferous sedimentary rocks were emplaced along fractures caused by upward pressures from the intruding Hailey magma. (2) The magma consolidated to a nearly rigid crystal mush in which stresses resulted in incipient strains and segregation of alaskite within the granodiorite. (3) During the completion of the consolidation, stresses resulted in an intensive breaking along innumerable, short, discontinuous fractures which were healed by a coarse-granular recrystallization of potassium feldspar and quartz with the aid of late magmatic and postmagmatic solutions. A large part of the potash, some of the quartz, and some soda were probably introduced from the less consolidated magma at depth during this phase. This constituted the endomorphic recrystallization of the Hailey granodiorite. Where the dynamic fracturing (shearing) was not fully endomorphically recrystallized to a directionless granular texture, perhaps because of a lack of solutions and (or) potash (metasomatic constituents), the Hailey granodiorite is still, at present, foliated and contains less potash. (4) At the end of the endomorphic phase, the Hailey granodiorite, representing a peripheral cooling zone, was rigid enough to take long fractures, and into these fractures aplite and pegmatite were emplaced probably from below. (5) Continued shearing stresses resulted in breccia zones, inches to feet wide, in which only the finest part of the crushed matrix was recrystallized, probably because temperatures were lower and (or) recrystallizing solutions were less abundant. In some of these breccia zones, a few euhedral porphyroblasts of potassium feldspar and roundish aggregates of quartz grew alongside of cataclasts of older minerals. (6) Later shears, spaced a few feet apart in the Hailey Gold Belt, cut the granodiorite throughout the Bellevue area; schistose, crushed rock within a fraction of an inch to inches of a central shear plane was basified through a process by which alkalis and silica were leached by passing hydrothermal solutions, presumably at appropriate pressures and temperatures (Morey and Hesselgesser, 1952; Kennedy, 1955, p. 498). This is an early hydrothermal phase. In prospect pits and on stained rock slices, every gradation from intensely foliated, basified rock to lamprophyre-like mobilized rock can be found associated with these shears. It is suggested that, in a few more intensely sheared zones, dynamic shearing actually generated enough heat to partially melt the crushed, basified rock and that such mobile material formed early lamprophyre dikes. (7) Still later quartz and metalliferous minerals were deposited along some of these basified shears and along new fractures to produce quartz veins. This is a late hydrothermal phase which occurred when temperatures and (or) pressures were too low for alkali leaching. At this same time more lamprophyric melt, probably generated by shearing, leaching, and melting at a lower depth under favorable P-T conditions, was intruded along the quartz-vein structures.

Tertiary Challis volcanics

A sequence of Tertiary volcanic and sedimentary rocks, resting unconformably on the pre-Tertiary rocks, has been tentatively correlated with the "Challis volcanics" of central Idaho by Anderson (Anderson, Kiilsgaard, and Fryklund, 1950, p. 7). In the Bellevue area the sequence is divided into a lower, dominantly tuffaceous sedimentary unit and an upper, dominantly volcanic unit.

The sedimentary unit is typically exposed 1 1/2 miles upstream from the mouth of Bullion Creek where it rests on the Carboniferous rocks. The unit is probably several hundred feet thick and is made up chiefly of tuffaceous sandstone interbedded with shale and conglomerate. The rock is commonly gray to tan and consists largely of pyroclastics of intermediate composition with lesser amounts of locally derived quartzitic and granitic detritus. Conglomerates containing cobbles and boulders of quartzitic and granitic rock are abundant in the Democrat Creek drainage, the next drainage north of Bullion Creek, north of the border of the Bellevue area. Plant stem impressions and silicified wood are characteristic of the unit. Thin volcanic flows are intercalated with the sedimentary layers.

The overlying volcanic unit is typically exposed along the Magic City road, east of Magic Reservoir, where it is at least 200 feet thick and may be more than twice as thick. The section consists of many lava flows and volcanic breccias, some of which may be ash flows; the individual flows have not been studied. The lithologic range is from brownish red to purple latite(?) to grayish green or black andesite(?) and basalt. Most flows show a well developed fluidal or compaction foliation. The rocks are porphyritic, averaging about 20 percent of phenocrysts (range, 5% to 50%) whose average length is about 1 millimeter, but in some flows the phenocrysts are as large as 5 millimeters. The latite(?) is characterized by phenocrysts of euhedral plagioclase and hornblende; pyroxene and biotite are less abundant. The groundmass is aphanitic to glassy and is generally stained a moderately bright yellow by sodium cobaltinitrite suggesting that considerable microcrystalline and cryptocrystalline potassium feldspar is present. In many of the andesitic rocks, pyroxene is the only conspicuous abundant phenocryst. Lindgren (1900, p. 196), Hewett and Ross (Umpleby, Westgate, and Ross, 1930, p. 35, 215), and Anderson (Anderson, Kiilsgaard and Fryklund, 1950, p. 6) have described these volcanics north of Croy Creek as consisting dominantly of reddish porphyritic biotite latite, red-brown hornblende andesite, and grayish green to black augite andesite with a few occurrences of black pyroxene basalt, olivine basalt, and light-colored rhyolite.

The gross distribution of the Challis volcanics in the Bellevue area is shown on the geologic map (pl. 1); the specific distribution of the tuffaceous sedimentary unit and the volcanic unit where these were mapped separately is indicated by letter symbols. North of Croy Creek large areas of both units occur. In the headwater area of Rock and Croy creeks, many small erosional remnants consist mostly of the sedimentary unit. The largest areas of Challis volcanics occur in the western part of the Picabo Hills and in the hills east of Gannett. The area east of Gannett has been mapped only along the valley side; it is part of a much more extensive area of Challis rocks to the east and north.

The Challis volcanics are readily distinguished from younger volcanics by alteration and deformation. Of three types of alteration, deuteric, hydrothermal, and weathering, the hydrothermal alteration is the most intensive. The rock is deeply weathered only where the presently exposed rock had been subjected to one or more earlier periods of weathering.

Hydrothermal alteration occurs to some degree in most places but is most intensive in the rocks on upper Rock Creek where the entire groundmass and phenocrysts of many rocks are altered to secondary minerals such as calcite, zeolites, opal and chalcedony, and clay minerals. In many places the rock is dark green where a celadonite-like mineraloid is dispersed throughout the rock. These secondary minerals commonly fill small voids and fractures which have penetrated the entire rock where the alteration is most intense.

Deuteric alteration is most noticeable in the thick volcanic flows in the Timmerman Hills where the hydrothermal alteration has not obscured the rock changes. Hornblende and biotite phenocrysts in places are altered to iron oxides and other secondary products; pyroxenes are altered to antigorite-like mineraloids; and feldspars, especially the cores of coarser-grained, earlier-formed phenocrysts, are partly altered to clay minerals. Devitrification of the groundmass is generally complete, and in silicic varieties vapor-phase crystallization is marked by conspicuous crystal-lined voids.

In spite of complex local faulting, the over-all structure of the Challis volcanics across the Bellevue area is a gentle southward slope, which averages 65 feet per mile in a north-south section from Bullion Creek to the Timmerman Hills. Even without knowing the direction of the average depositional slope, this one section would suggest that there is little post-Challis downwarping toward the Snake River Plains. There is no evidence showing whether or not the Challis volcanics have been downfaulted into the structural basins such as Camas Valley or the southern end of the Wood River Valley.

Locally, the Challis volcanics are complexly faulted; attitudes, most readily measured in the sedimentary rocks, vary greatly; dips are flat to vertical, and strikes may be in any direction. Where the sedimentary rocks are in depositional contact with the older rocks, the strikes are mostly parallel to the contact, but the dips are rarely systematic, for example, along the normal contact on Bullion Creek the dips vary from 40° to 75° away from the contact. The coarse conglomeritic, basal sediments indicate that the Challis was probably deposited on a rugged topography and that some part of the variability of strikes and dips may be due to locally steep depositional slopes. Numerous, close-spaced joints of two or more intersecting sets form a characteristic crisscross pattern seen on aerial photographs (1:20,000); this pattern is useful for photo-interpretation.

The Challis volcanics are much younger than the Idaho batholith and are older than the overlying, Pliocene, silicic volcanics. On the basis of fossil plants, Lindgren (1900, p. 197) considered the sedimentary rocks along Bullion Creek to be correlatives of the Payette formation of western Idaho of Miocene(?) age. From a much larger fossil-plant record of the Germer member in the upper part of the Challis volcanics near Challis and Salmon, Idaho, Ross considers the Challis volcanics to be probably of Oligocene age and possibly as young as early Miocene (Ross, 1937, p. 65-68). For the present, however, no direct correlation can be made with the "Challis" volcanics in the Bellevue area.

Pliocene Magic volcanic rocks

General statement.--The southern half of the Bellevue area is underlain in part by a sequence of Pliocene silicic and basaltic flows named the Magic formation for the Magic Reservoir. On the geologic map (pl. 1), these rocks are divided into three silicic members and one basaltic member. A total of about 2500 feet of section has been measured, but because the thickest part of each member lies in broad valleys cut in earlier members, the aggregate thickness of the volcanics in any one place at present is only about 1000 feet (fig. 7). At the end of the volcanism, the pile of volcanics probably was much thicker; the total stratigraphic section may have been well over 5000 feet.

Two of the silicic members consist of numerous ash flows (Smith, 1960a, p. 800; Ross and Smith, 1961, p. 3) which range from deposits emplaced at high temperatures (Mansfield and Ross, 1935) to some emplaced at intermediate or low temperatures (Smith, 1960a, p. 831). Individual ash flows are 10 to 100 feet thick, each typically having a black vitrophyric lower zone of dense welding and a stony upper zone of partial welding (Smith, 1960b, p. 154); the lowermost and uppermost zones of no welding are rarely exposed and in some places may not be present (fig. 9). A pyroclastic origin for the ash flows is indicated by the vertical zonation of individual ash flows, by conspicuous eutaxitic structures seen in outcrops where large autoliths of pumice have been compacted, and by a well developed compacted-shard structure in thin sections of the vitrophyric zones.

The Magic units generally slope gently southward and pass under the Snake River Plains south of the Picabo Hills; for example, their average dip along a 10 mile north-south distance through Gannett is 1° (about 90 feet per mile). The dip is probably in part depositional and in part due to deformation. Locally the units are downdropped on step faults of small displacement toward topographic lows such as Camas Valley, the Wood River Valley, Silver Creek valley, and the Snake River Plains. The faulting suggests that part of the Magic volcanic rocks may have collapsed into Camas Valley and the southern Wood River Valley.

No fossils have been found in the Magic volcanics in the Bellevue area. The volcanics are tentatively considered to be Pliocene because their lithology and general stratigraphic position are similar to that of presumably Pliocene silicic volcanics elsewhere around the border of the Snake River Plains (Mount Bennett rhyolite, north of Mountain Home, "early Pliocene," Malde, 1959, p. 272; Shoshone Falls andesite at Shoshone Falls (rhyolitic chemical composition given by King, 1878, p. 592), "Miocene(?)," Stearns, 1955, p. 463; Walcott tuff near American Falls, "upper Miocene to late Pliocene," Stearns, 1956, p. 19; rhyolitic ash flows in southeastern Idaho, post-upper Miocene, Mansfield, 1952, p. 52; rhyolitic volcanics near Jarbidge, Nevada, "Pliocene," Coats, 1957, p. 303; Owyhee rhyolite in southwestern Idaho, "upper Miocene or lower Pliocene," Kirkham, 1931, p. 579).

Picabo tuff.--The Picabo tuff, the oldest member in the Magic volcanic sequence, underlies large areas in the Picabo Hills and is divided into a lower unit, tuff A, and an upper unit, tuff B (fig. 7). Both units consist of many ash flows and of several tuffaceous sedimentary layers. In places typical vertical zonation is lacking, and the individual ash flow is probably part of a group of flows with a complex cooling history, that is, the ash flow is part of a "cooling-unit complex" (Smith, 1960a, p. 800). A layering of resistant and nonresistant rocks tends to cause hard layers to crop out and develop a bench and slope topography which is typical for the unit. The fresh rock is variously colored lavender or brown and weathers to brown or red-brown. It contains less than 5 percent phenocrysts, predominantly andesitic plagioclase; quartz, clinopyroxene, sanidine, magnetite and pyrite are much less abundant.

Tuff A is exposed in a section about 500 feet thick, without a base, on the Silver Creek road 2 miles east of Highway 93. The harder rocks making up tuff A are densely welded and have a closely spaced foliation formed by the compaction and welding of dominantly fine-grained glass shards at a relatively high temperature. Long, flat, narrow gas cavities, in places 100 times as long as wide, define a diagnostic lineation which lies in the plane of the compaction foliation and nearly parallels the axis of a broad valley filled by the tuff; it was developed by viscous movement during and after the latest stage of compaction of the tuff.

Tuff A presently occurs in an area, 4 miles wide, across the central part of the Picabo Hills where it fills an ancestral broad valley eroded in pre-Picabo rocks.

Tuff B is exposed in a section measuring about 450 feet, without a top, on a prominent hill two miles east of Gannett. In contrast to the harder rocks of A, those of tuff B are distinctly less welded and have a granular fragmental texture with the exception of the basal vitrophyric zone of individual ash flows where the fragments are compacted to a closely spaced eutaxitic structure. In the principal ash flow of tuff B, a layer within the vitrophyric zone consists of 25 to 50 percent lithophysal cavities which have rough, angular walls and range from an inch to a foot in diameter but average 2 inches in diameter (fig. 8). This particular layer commonly forms a cliff 5 to 70 feet high which is diagnostic of tuff B.

The basal part of tuff B is a thick layer of white tuffaceous sediment which is poorly exposed in the pass between the Wood River Valley and Sonnars Flat. It is identified elsewhere only by abundant landslides developed on the unconsolidated sediment; such landslides are mapped along the sides of the Wood River Valley both to the north and to the east of the pass (pl. 1).

A broad valley, which is eroded in the rocks of tuff A, was largely filled by the tuffaceous sediments, so that the overlying ash flows of tuff B spread out on an adjacent rolling upland where the tuff now caps many of the higher hills of the Timmerman Hills (pl. 1).

The source of the Picabo tuff is not known. The westernmost occurrence, an erosion remnant on Little Rock Creek, is probably part of tuff A. A thick section of Picabo tuff underlies the eastern Picabo Hills but has not been studied. Units equivalent to tuffs A and B are well exposed east of Priest Station where individual ash flows are thicker and more densely welded than in the central Picabo Hills.

<u>Moonstone rhyolite</u>.--Moonstone Mountain, consisting of what is here named Moonstone rhyolite, stands in bold relief about 1000 feet above the surrounding lowlands on the north edge of Camas Valley. The mountain is probably a partially eroded, endogenous volcanic dome, that is, it grew by expansion of viscous lava from within (Williams, 1932, p. 53) as is indicated by a well exposed concentric flow layering which dips toward the center, steeply at high altitudes and gently at low altitudes (figs. 10 and 11). Dips are as low as 10° north on the south flank of the

mountain below the 5200 foot contour. To the southeast one lobate structure extends out beyond the dome and probably represents a short tongue of viscous lava. Numerous sets of joints cut the rock, but an approximately radial set which is vertical and strikes perpendicular to the flow layering is the most conspicuous (fig. 10).

The Moonstone rhyolite typically consists of as much as 50 percent phenocrysts with an average length of 5 millimeters. The phenocrysts are, in order of abundance: fresh-appearing sanidine; rounded, resorbed quartz; large altered crystals of plagioclase mostly replaced by potassium feldspar and small unaltered crystals of andesine; pigeonite, partly altered to antigoritic material; variously altered, dark brown oxyhornblende; and accessory magnetite, zircon, apatite, and allanite. Devitrification and vapor-phase crystallization (Smith, 1960b, p. 152) are intense in most rocks and are made evident by cristobalite-adularia intergrowths and tridymite in pore spaces. The rhyolitic composition was determined by chemical analysis (H. A. Powers, U. S. Geological Survey, written communication, 1960). Rarely xenoliths consist of medium-grained granitic rock types and coarsely porphyritic basalt.

The Moonstone rhyolite varies considerably in texture and possibly in composition within Moonstone Mountain; the change may be caused either by greater deuteric and devitrification alteration in the central part or by a gradual change in composition of the extruded material. The rock exposed near the base of the mountain (hill 5360, fig. 10) is dark brown, dense, glassy, and highly porphyritic, contains abundant altered plagioclase phenocrysts, and at least in places is autobrecciated.

Supposedly, this is an early basal part of the dome. The rock making up the main part of the mountain is light gray to light purple and weathers white or brown. It is much less dense, less coarsely porphyritic, contains few or no altered plagioclase phenocrysts, and is so highly devitrified that the flow layers merge into a nearly massive rock. Between these two rock types is a transitional zone several hundred feet thick in which the flow layering is best developed and consists of darker, more glassy layers alternating with light colored, devitrified layers.

Rock identical to the Moonstone rhyolite underlies an area of at least 100 square miles in the eastern part of the Mount Bennett Hills; it is more than 500 feet thick. Only a few square miles of this rock occur on the south edge of the Bellevue area (pl. 1) where it is typically exposed in the mouth of the canyon in the northeast corner of section 9, 2 1/2 miles southwest of Magic Resort.

In most outcrops in the Mount Bennett Hills, the Moonstone rhyolite is highly devitrified, light colored and massive. In a few places a faint layering is due to vague flat planes or joints, tens of feet apart, of unknown origin. Locally in steep-sided canyons, a foliation, consisting of probable flow layers inches to feet thick, forms swirls and rolls 100 to 1000 feet across. In these places a strong east-west and a weaker north-south joint set is etched out by erosion. These structures resemble, on a small scale, the flow layering and jointing in Moonstone Mountain. A black, glassy, in part scoriaceous(?) top, 10 to 30 feet thick, grades downward to the light-colored rock. The glassy top is only partly designated on the geologic map (pl. 1); with detailed mapping it should be a good stratigraphic marker.

The source of the Moonstone rhyolite in the Mount Bennett Hills is believed to be Moonstone Mountain, but it seems anomalous that such magma, crowded with phenocrysts, could have a low enough viscosity to flow a minimum of 12 miles from the vent where similar magma was so viscous as to form an endogenous dome. Even if the source of the magma in the Mount Bennett Hills were buried below the flow on the south edge of Camas Valley, a minimum flow distance of 8 miles would be required. On the other hand, if the mechanism of flow was that of a nuée ardente, then Moonstone Mountain could well be the source and the glassy autobrecciated rock making up hill 5360 could well be the actual vent rather than a basal phase of the extrusive dome as suggested above.

Three other extrusive domes, exhibiting the same strong concentric foliation structures as occur on Moonstone Mountain and consisting of a lavender-gray, low-density, porphyritic rock, are located along Highway 93 near the south border of the Bellevue area; a small part of one dome appears on the geologic map (pl. 1), one mile east of the highway. The rock of the domes can be distinguished from the Moonstone rhyolite on the basis of phenocryst abundance, size, and composition.

Square Mountain basalt.--Numerous flows of Square Mountain basalt with an aggregate thickness of more than 600 feet are typically exposed west of Square Mountain in the canyon of Camp Creek (fig. 12). Most of the flows are characterized by a blocky jointing habit which commonly develops into poorly defined columns. In places rosette structures many feet across are due to a concentric platy jointing caused by cooling anomalies induced by the proximity of the valley side.

The rock is bluish black on fresh surfaces and weathers to a dark reddish brown. In thin section it is seen to be composed of unaltered, euhedral laths of plagioclase, 0.2 by 1 millimeter in size, and anhedral augite, 0.3 millimeter in diameter, set in a felty-textured, dark brown to opaque groundmass.

The Square Mountain basalt is contaminated with as much as 20 percent of rhyolitic material; large xenocrysts of quartz, plagioclase, and potassium feldspar are resorbed and altered but are otherwise identical with the phenocrysts of the Moonstone rhyolite and may represent mixing of basaltic and rhyolitic magmas in depth prior to the extrusion of the basalt. The uniform distribution and alteration of the xenocrysts in the extruded basalt further support this hypothesis. The presence of a large amount of microscopic or submicroscopic potassium feldspar in the groundmass is indicated by a moderate yellow stain produced by the sodium cobaltinitrite test on polished rock slices.

Glassy-clear quartz xenocrysts, constituting 1 per cent or less of the basalt, have been rounded by resorption and range from 1 to 20 millimeters in diameter but average about 2 millimeters. Euhedral xenocrysts of plagioclase, constituting as much as 10 percent of the rock, have been altered to phantom crystals as a result of reaction with the basaltic lava. They average 5 millimeters in length and are made up of ordered, oriented aggregates of small reconstituted plagioclase crystals about 0.2 millimeter long which delicately outline the euhedral shape and zonal structure of the original plagioclase. The edges of some phantom crystals are partially disintegrated where the small reconstituted crystals were stirred into the basaltic lava. In many respects the phantom plagioclase crystals resemble those described by Fenner (1944, p. 1092) and Wilcox (1944, p. 1066, pl. 5) in Yellowstone National Park (Boyde, 1961, p. 403). Potassium feldspar xenocrysts, making up less than 1 percent of the rock and ranging from 1 to 10 millimeters in diameter, have partially reacted with the basaltic lava, chiefly by resorption of an outermost altered zone resulting in a rounded shape.

White gneissic xenoliths, ranging from one inch to several feet in diameter and averaging six inches in diameter, are typically scattered at wide intervals in the quartz basalt. Several clusters of these xenoliths on Square Mountain above Tolmie Creek are about 10 feet high by 10 feet wide by 100 feet long and consist of 50 to 70 percent xenoliths set in a basaltic matrix (fig. 14).

A conspicuous gneissic structure in the xenoliths (fig. 13) is due to layers of brown glass alternating with layers of colorless glass. The glasses, constituting about 20 percent of the rock, are interstitial to large, aligned crystals of quartz, potassium feldspar (microcline?) and plagioclase; the plagioclase is least abundant and commonly is almost entirely replaced by potassium feldspar. Relicts of clinopyroxene rimmed by orthopyroxene or of orthopyroxene alone are associated with the brown glass, indicating that the brown glass marks mafic layers in the original gneiss, and suggesting that the brown color of the glass is due to slight contamination by mafic constituents in a glass of dominantly quartzo-feldspathic composition. In many places the glasses are devitrified.

The interstitial melt mingled with the basaltic lava only near the margins of the xenoliths.

These gneissic xenoliths probably represent a coarse-grained, pyroxene-bearing granitic gneiss of the granulite facies formed in a deep-seated, dry environment where biotite or hornblende were unstable. During their upward rise, the greater temperature and heat of the basalt caused selective intergranular melting in the gneiss, and upon the extrusion and rapid cooling of the basalt, the interstitial melt in the xenoliths froze with only incipient crystallization. It is speculated that the gneissic xenoliths were a part of the wallrock enclosing the rhyolitic magma which contaminated the basalt. This wallrock gneiss may be similar to the rock which was melted to produce the rhyolitic melt.

Two other types of xenoliths occur rarely. One xenolith of granitic rock consists of andesine and quartz, 1 to 5 millimeters in diameter, and dispersed dark clots of intensely altered and partially fused material representing former mafic minerals. This xenolith probably originated from an intermediate depth between the deep granitic gneiss and the presently exposed Hailey granodiorite because the quartz and feldspar grains in the xenolith are not fused as in the granitic gneiss and because the xenolith has a very different composition from that of the Hailey granodiorite. The other type of xenolith, represented by a specimen about 9 feet in diameter, is a gabbroid rock consisting of an open mesh-work of large labradorite laths, as much as 10 millimeters long, with interstitial, partially resorbed clinopyroxene, devitrified glass, and about 20 percent large voids. That the labradorite crystals are not resorbed indicates that the labradorite was stable in the basaltic magma. This xenolith suggests that a gabbroic body exists at a depth greater than that of the granitic gneiss.

The Square Mountain basalt has been mapped over an area 12 miles north-south by 8 miles east-west (pl. 1); it extends west beyond the border of the area. In the Mount Bennett Hills it conformably caps the black glassy top of the Moonstone rhyolite wherever the contact has been observed. North of Camas Valley, however, it fills an ancestral canyon of Camp Creek which almost coincides with the present but steeper canyon. Above this canyon it lies on a flat upland eroded on granite, and at Moonstone Mountain it laps up on the rhyolitic dome. It appears that, following the extrusion of the Moonstone rhyolite, Camp Creek cut a canyon which was in turn filled by Square Mountain basalt. That the time interval of canyon cutting was short is suggested by the fact that the basalt directly overlies the black glassy top of the Moonstone rhyolite in the Mount Bennett Hills.

The Square Mountain basalt is broken into numerous blocks by normal faults of small displacement, possibly 10 to 100 feet. In the Mount Bennett Hills the faults trend east-west, whereas, north of Camas Valley, step faults of larger displacement drop the basalt toward the valley; for example, two such faults are mapped between Moonstone and Square Mountain where a section of basalt over 600 feet thick was dropped into Camas Valley. On the other hand, part of the thick section on the Camas Valley side may be due to original deposition into an already existing Camas structural depression, and the thin section of basalt only one mile to the north may never have been more than tens of feet thick.

The location of the vent is not known, but it may be in the vicinity of the north or west side of Moonstone Mountain where the basalt is thick, where major structures are present, and where the xenoliths are most abundant. Likewise the Moonstone eruptive center is nearby so that if a genetic relationship does exist between the rhyolite and the quartz basalt, as suggested above, then, their respective vents might be expected to be near each other.

Poison Creek tuff.--The Poison Creek tuff is typically exposed along lower Poison Creek where it consists of many ash flows and at least one pumice fall with an aggregate thickness of several hundred feet. The rock is crystal-poor, lapilli-rich and varies from black densely welded tuff to orange nonwelded tuff. Phenocrysts make up 5 to 10 percent of the rock, and the average larger phenocrysts are about 1 millimeter in diameter. The phenocrysts consist of quartz, sanidine, andesine, hypersthene, clinopyroxene, and accessory magnetite, zircon, and apatite. The pyroxenes are largely altered to antigorite-like material. A rhyolitic composition is indicated by chemical analysis of a specimen collected near the center of section 9, T. 2 S., R. 17 E., in the Mount Bennett Hills (H. A. Powers, U. S. Geological Survey, written communication, 1960). The Poison Creek tuff is divided into a lower unit, tuff A, and an upper unit, tuff B.

Tuff A is typically exposed in the north canyon wall of Camas Creek about 1000 feet west of Hot Springs Landing where it is more than

200 feet thick. It appears to be a single thick ash flow which has a flat-lying sheeting consisting of vaguely defined layers tens of feet thick. The rock is light purple and is densely welded, and it contains many streaks and irregular blotches of highly devitrified, commonly porous rock which probably represents lapilli-size, pumice and lithic fragments which were compacted and plastically deformed and even dragged out into swirly, flow structures during the welding of the tuff.

Locally, wide-spaced flow layers bend and curve in all directions commonly on radii of 10 to 20 feet. For example in an area, 500 feet wide, one mile west of Hot Springs Landing, foliation layers, accentuated by a 1/2 inch wide platy jointing, are bent into large roll structures but have an average upward trend. The meaning of these internal structures is not understood; this particular locality could be the site of a vent of at least part of the Poison Creek tuff.

Tuff B is typically exposed in a section which is more than 300 feet thick along Highway 68, at and east of Poison Creek. It consists of many ash flows each of which is as much as several tens of feet thick; the thickness of individual ash flows thins toward the top of the section. Each flow has a thin white ashy base; a relatively thick, black vitrophyric zone which is characterized by a granular, spherulitic and perlitic texture; a pink partially welded zone; and a white nonwelded ashy top. The degree of welding, in addition to decreasing upward in each flow, also in general, decreases from the base to the top of the tuff B section. The number and size of autolithic fragments in each flow generally increase upward in the section.

The lowermost ash flow of tuff B has a vitrophyric zone which at least locally is unusually thick; where the highway crosses Poison Creek the vitrophyre is about 60 feet thick. This anomalous thickness appears to be caused by the filling of a locally rough topography cut in tuff A. The uppermost ash flows have well developed eutaxitic textures formed by compacted fragments identified as collapsed pumice lapilli. The uppermost unit of tuff B is an orange pumice-lapilli, air-fall deposit, tens of feet thick, which at least in part was reworked by water.

The Poison Creek tuff unconformably overlies the Square Mountain basalt; contacts of tuff B in the Clay Bank Hills and contacts of tuff A in the Mount Bennett Hills rest on either Square Mountain basalt or Moonstone rhyolite (pl. 1). Xenoliths of Square Mountain basalt in the Poison Creek tuff likewise support this age relationship.

Structurally, the Poison Creek tuff is broken by numerous north-south and east-west trending faults which commonly displace the various units as much as several tens of feet. Faults of larger displacement have dropped large areas of the tuff into the Camas structural basin.

The location of the source vent of the Poison Creek tuff is not known; a possible source west of Hot Springs Landing is suggested above, but one of the crystal-poor extrusive domes just south of the border of the Bellevue area is also a possible source. Unweathered xenoliths of Square Mountain basalt high in the Poison Creek tuff suggest that the vent is at least within the area of distribution of the basalt. The thick, lapilli-size, air-fall deposit at Poison Creek suggests a nearby source.

QUATERNARY STRATIGRAPHY

Quaternary basalts

General statement.--The Snake River basalt was named by Lindgren (1900, p. 18) in the Boise district, and the name was extended to the whole Snake River Plains region by I. C. Russell (1902, p. 38, 59). In the central Snake River Plains, Stearns (1938, p. 56, 63) divides that part of the Snake River basalt which overlies the Hagerman beds of upper Pliocene age into two groups of earlier and later Pleistocene age, and a third group of Recent age. Each group consists of the products of numerous eruptions. Along the canyon of the Snake River he distinguishes "eleven local Pleistocene formations" which probably correspond to the later Pleistocene group.

In the Bellevue area the Snake River basalt has been divided into two units in the course of this study (pl. 1). The older unit, the Clay Bank basalt, has been tilted southward and eroded to form a small part of the belt of foothills between the mountain terrane of older rock to the north and the Snake River Plains of younger basalt to the south. The younger unit, the Bellevue basalt, occupies valleys cut in the foothill belt and extends southward to make up the Snake River Plains. The Bellevue basalt constitutes a succession of basalt units which erupted at intervals from different vents in and around the Bellevue area; it includes Recent basalt which makes up Stearns' third group (1938, p. 94).

There is as yet no correlation between the basalt section in the Bellevue area and the section described by Stearns to the south except for the Recent basalt.

The Clay Bank basalt is not subdivided although it consists of many flows. The Bellevue basalt is subdivided into units which are from oldest to youngest, the Hay basalt, Wind Ridge basalt, Myrtle basalt, Sonnars basalt, Macon basalt, Priest basalt, and the Burmah basalt. Each unit consists of a single flow or sequence of flows believed to have issued from the same vent during a single eruptive period. Assignment of member status to these units means that the Clay Bank basalt and Bellevue basalt have formational rank and that the Snake River basalt is considered a group.

Clay Bank basalt.--The Clay Bank basalt is typically exposed in the Clay Bank Hills one mile east of the southern end of Magic Reservoir (sec. 8 and 9, T. 2 S., R. 18 E.); other good outcrops can be seen along the 400 foot high, west-facing scarp overlooking Magic Reservoir. About 100 feet of basalt can be measured in single outcrops, and a total thickness of several hundred feet or more is probable when factors of erosion and deformation are considered. The sources and directions of flow of the various units are not known. At the type locality the basalt is broken into fault blocks about 1/4 mile wide which are gently tilted to the south. On the steep west slope of the Clay Bank Hills the basalt is complexly broken into fault slices and landslide slumps.

The lithology of the Clay Bank basalt is similar to that of the Bellevue basalt discussed in the next section. In bold outcrop the Clay Bank basalt is not noticeably more weathered than the Bellevue basalt except on the average the Clay Bank basalt is less porous because products of incipient weathering tend to fill a primary pore texture common to Snake River basalts. On the other hand some basalt that had been buried and which is now exposed in beach cuts on Magic Reservoir is thoroughly weathered to a gray clayey mass. It is assigned to the Clay Bank formation even though age relations are complicated by faulting and slumping. The deep weathering probably occurred during a time when both climate and topography were different from those of the present.

The age of the Clay Bank basalt is known only to be younger than the Pliocene silicic volcanics.

Bellevue basalt.--In addition to each of the 7 named members of the Bellevue basalt which crop out in the Bellevue area, well logs record that still other flows conformably underlie the named members in the deeper parts of the valleys of the Big Wood River and of Silver Creek near the southern edge of the Bellevue area. These are considered non-designated members of the Bellevue formation.

The Bellevue basalt is a porphyritic olivine basalt with a conspicuous diktytaxitic texture. Scattered phenocrysts of olivine are as large as 2 millimeters, and feldspar laths reach a length of 2 millimeters but are most commonly less than 1 millimeter. Pyroxene phenocrysts are rarely seen in hand specimen except where they are concentrated in glomeroporphyritic masses with feldspar laths.

Diktytaxitic texture (Fuller, 1931, p. 116); Mackin, 1961 is the most distinctive feature of the basalt. It is characterized by an all-pervasive system of irregularly shaped, angular pores into which feldspar, pyroxene and olivine crystals project or form an open crystal mesh-work across the voids. The size of the pores averages a fraction of a millimeter and ranges from microscopic to 1 millimeter in different parts of a single flow or in different flows.

The color of freshly broken basalt is black to gray; however, the uppermost parts of some flows are light gray due to an abundance of small feldspar crystals. The dark color of most outcrops is caused by a black or dark gray lichen cover.

Most of the Bellevue basalt is of the pahoehoe type, characterized by a particular surface pattern which is best seen and examined on aerial photographs. The pattern is caused by many small pressure ridges and pressure domes which commonly are not aligned and bear no relation to the direction of the flow. Somewhat older flows exhibit linear elements etched out by erosion and accentuated by stipplings of vegetation; these are surely related to the original flowage of the basalt. On still older flows, loessic and residual soil obliterates the surface rock patterns, but as the consequent and subsequent microdrainage lines develop, the original flow directions are even more strongly accentuated.

Aa, cinders, and basalt breccia have not been seen on outcrops of the basalt. Several well logs from the Silver Creek valley record a few feet to tens of feet of "broken basalt and cinders" which suggests that Aa may be associated with some flow fronts.

The base of some Bellevue flows consists of bulbous glassy basalt a few inches thick which in some places is altered to palagonite. The first foot or so of basal basalt is irregular in texture and structure, and commonly contains various flow-distorted and altered zones. The main bulk of the flow is dense basalt containing sparse, scattered holes which are angular and irregular in shape and 1 millimeter to several centimeters in diameter. The upper several feet of the flow is highly vesicular.

Columnar jointing is developed in all the flows; the columns are poorly symmetrical and average 2 to 3 feet in diameter. A distinct threefold vertical joint system has been seen at only a few outcrops; it consists of a coarsely jointed basal part, a heterogeneously jointed to platy jointed central zone, and a columnar jointed upper part.

In bold outcrop the basalt weathers brown, forming a rind about 1 millimeter thick. Where the rock is covered by a thin soil, the rind is several centimeters thick, and where the cover is thick, the weathered rind is also thicker. The difference in depth of the weathered rind clearly expresses the moisture-holding effect of the soil cover.

The Hay member consists of a single flow and is typically exposed in a highway cut 3 1/2 miles southeast of Gannett along State Highway 23 near Hay Station for which the flow is named. A well formed vent, containing a crater about 50 feet deep is located one mile north-northeast of the highway cut. The vent is at an altitude of about 4900 feet and from it the surface of the flow slopes downvalley at 32 feet per mile for about 8 miles whereas the surface slopes upvalley at 45 feet per mile for about 3 miles. About 2 square miles of Hay basalt are exposed; most of the flow is buried under alluvium. The original dimensions are estimated as about 12 miles long, 2 miles wide, and 60 feet thick (figs. 32 and 39), making a volume of about 1/3 cubic mile.

The Wind Ridge member is named for an inconspicuous ridge about one mile south of Stanton Crossing (secs. 21 and 28). It consists of two flows; both probably erupted from the same vent at an altitude of 5240 feet on the south valley side two miles south of Stanton Crossing. Both belong to the same eruptive period, although, sufficient time elapsed between eruptions for about 150 feet of sediment to be deposited by the Big Wood River where it was dammed by the lower Wind Ridge flow.

The lower flow is typically exposed in a 50 foot river-cut scarp on the south valley side at Mahoney Flat. The upper basalt is best exposed 1/2 mile southwest of Stanton Crossing on the south wall of the Big Wood River canyon. At this place the upper basalt overlies Big Wood River gravel which rests on granitic rock; this same gravel overlies the lower Wind Ridge basalt to the west.

Both Wind Ridge flows filled the west outlet valley of the Big Wood River from Stanton Crossing to Hot Springs Landing, a distance of 3 to 4 miles, and the lower flow, at least, extended into Camas Valley, probably to beyond Magic Resort. The upper basalt extended about a mile farther upvalley to Stanton Crossing (fig. 39). Below Rock Creek the two flows cannot be distinguished. The average width of the Big Wood River valley into which the basalt spread was probably about a mile (fig. 33). Fifty feet of section of the lower basalt are exposed at the type locality and another 90 feet are estimated to extend below the base of the outcrop (fig. 39); whether or not the total 140 feet represent a single flow or a sequence of lower Wind Ridge flows is not known. The upper flow is about 40 feet thick at the type locality. The volume of the Wind Ridge member is estimated to be 1/8 cubic mile.

The Myrtle member consists of an exposed upper flow and at least one buried lower flow. The lower flow is inferred from evidence seen near Magic Dam; its existence is also suggested by a drill log (well, 2S-18E--18aa1, fig. 39). The upper flow is named for Myrtle Point on Magic Reservoir and is typically exposed in the upper half of the Big Wood River canyon at the spillway of Magic Dam.

The upper flow was extruded from an inconspicuous vent at an altitude of about 5050 feet on the east slope of the Mount Bennett Hills, 3 1/2 miles southwest of Magic Dam and south of the Bellevue area. Part of the basalt spread upvalley with a slope of about 55 feet per mile for 5 miles to Myrtle Point, 1 1/2 miles north of Magic Dam (fig. 39). The largest portion of the lava spread down the ancestral Big Wood River valley with a slope of 90 feet per mile for an unknown distance; 2 1/2 miles southeast of the vent, the Burmah flow covers the flow. The basalt thins from about 50 feet at Magic Dam (fig. 34) to 30 feet one mile north of the Dam, and it is not exposed 1/2 mile farther where the sudden widening of the river valley suggests that this was the northern edge of the flow. The area and volume are not calculated because of insufficient data.

The Sonnars member is named for Sonnars Flat in the southeastern corner of the Bellevue quadrangle; the flat extends east and south of the quadrangle boundary, as does the basalt. Two flows are designated on the geologic map (pl. 1) and south of the Bellevue area these two flows merge with still other flows. The basalt is so situated, between the Big and Little Wood rivers and the Timmerman-Picabo hills, that it is not dissected and vertical exposures of more than a few feet are nonexistent.

The vent of the westmost flow is an inconspicuous flat dome, 20-30 feet high and about 1/4 mile in diameter; and it is located on Sonnars flat, 1 1/2 miles east of U. S. Highway 93 (sec. 11). The other vent is located on the south slope of the Picabo Hills, 1 1/2 miles north of Mud Lake (sec. 15, T. 2 S., R. 19 E.).

About 8 square miles of Sonnars basalt are exposed in the Bellevue area, and at least 30 additional square miles of the Sonnars complex are exposed south of the Bellevue area. The volume has not been calculated.

The Macon member is named for Macon Flat at the eastern end of Camas Prairie. The Macon flow is typically exposed everywhere along the Camas Creek canyon from Blaine to the mouth of Poison Creek and specifically at the gaging station opposite the mouth of Willow Creek.

The flow has an inconspicuous vent at an altitude of 5330 feet on the south edge of Camas Valley, 11 miles southwest of Hot Springs Landing and 6 miles due south of Blaine. The surface of the basalt slopes about 40 feet per mile to the east, north, and west extending the full width of Camas Valley from Fairfield to Magic Reservoir. The total initial surface area is estimated to be 100 square miles, of which about 65 square miles is exposed on the surface at present. The thickness is estimated as about 100 feet; therefore, the total volume of basalt is about two cubic miles, by far the largest flow in the Bellevue area.

The Priest member is typically exposed 3 miles northeast of Priest Station on the banks of Silver Creek (one mile east of Picabo). The basalt is not trenched by Silver Creek so that vertical exposures of more than a few feet do not exist.

A cratered vent at an altitude of about 5,000 feet on the east side of the Picabo Hills, 2 miles south of Picabo was the source of the flow. In an upvalley direction the basalt surface slopes 90 feet per mile for about 3 miles to its distal end and in a downvalley direction the surface slopes 70 feet per mile. South of the border of the Bellevue area, the flow merges with other flows of similar age. An estimated average thickness of about 100 feet and an area of about 12 square miles mean that the volume of Priest basalt in the Bellevue area is about 1/4 cubic mile. This is probably only a small fraction of the total volume erupted from the Priest vent.

The Burmah member occupies the Big Wood River valley south of the Bellevue area and is by far the youngest and most impressive flow in the surrounding area. It is named for Burmah Station about 16 miles north of Shoshone (sec. 25, T. 3 S., R. 18 E.).

The vent of the Burmah basalt, locally referred to as Black Butte, is 4 1/2 miles south of Magic Dam and rises 400 feet above the surrounding older lava plain (Harrington, 1948, p. 464). The crater is about 1/2 mile across and over 100 feet deep with near vertical walls. The flow spread up the Big Wood River valley about 1 1/2 miles and down-valley 17 miles to Shoshone and another 16 miles to Gooding. Its width varies from 1 to 4 miles. The basalt has been studied only to compare the weathering and erosion of its crater with others in the Bellevue area.

The time interval represented by the Bellevue basalt includes a part of the Pleistocene, probably the late Pleistocene, and all the Recent. The two youngest members, the Burmah and the Priest, are the only flows for which an approximate absolute age has been determined. On the basis of weathering and erosion, the Burmah flow is comparable in age to the younger flows at Craters of the Moon National Monument, 40 miles east of Picabo, where the age of the youngest flow is estimated to be 1650 years B. P. as based upon tree-ring counting technique (Murtaugh, John G., 1961, Univ. of Idaho M. S. thesis; Stearns, 1938, p. 100). If by comparison the Burmah flow is estimated to be 3 to 5 times older than the youngest flow at Craters of the Moon, then the Burmah flow would be 5,000 to 8,000 years old.

The age of the Priest basalt is approximately fixed by its relationship to glacial deposits. It is probably slightly older than the Bull Lake stage; it is certainly older than the latest glaciation because, in the vicinity of Picabo, Pinedale outwash overlies the Priest basalt. Evidence favoring a pre-Bull Lake age is fully developed in the geologic history section and is summarized as follows. The Priest basalt dammed the ancestral Big Wood River in its eastern outlet near Picabo; the river spilled over the dam; somewhat later, during the Bull Lake stage, glacial aggradation in the Wood River Valley diverted the river into the present western outlet at Stanton Crossing.

The relative age of the members of the Bellevue basalt cannot be determined by the law of superposition, but within broad time intervals the degree of weathering, erosion, and the development of a soil cover are reliable guides to age. These features were studied first on detailed aerial photographs (1:20,000 scale) and in every case were checked by field examination of the respective flows. The erosion of the vertical-walled craters of the Hay, Priest, and Burmah flows is an especially reliable guide to age.

Quaternary sediments

General statement.--Twelve units of unconsolidated sediments have been mapped in the Bellevue area. The units fall into three groups: (1) deposits impounded by lava dams, (2) deposits formed by proglacial or periglacial aggradation, and (3) deposits caused by normal erosional processes. Formal stratigraphic names are not assigned to these sedimentary units. They are discussed in chronological order; those associated with the various Snake River basalt members are identified by reference to the appropriate member, and the glacial sediments are identified by reference to age and source.

Sediments associated with the Clay Bank basalt.--Fluvial sediment, probably mostly gravel, occurs below one of the Clay Bank flows which forms a caprock 1.7 miles southeast of Magic City. Hard clayey silt, silty angular sand and gravel, and clean angular sand and fine gravel are associated with deeply weathered Clay Bank basalt along the east shore of Magic Reservoir both north and south of Magic City. In general, sediments associated with the Clay Bank basalt are abundant and as widespread as the basalt even though the sediments are rarely exposed in outcrop; gravel float is commonly the only indication of the presence of the sediments.

Thicknesses of ten to several tens of feet of sediment are believed to occur between exposures of the different flows. The pebbles contained in these sediments are derived locally from granitic rocks and Challis volcanics. In general the sediments are deeply weathered; most pebbles are decomposed, and the silts and sands are partly indurated by clay.

The physiographic picture presented by these few facts is that, while the Clay Bank flows spread chiefly from the south, probably up an ancestral Big Wood River valley, sediments were derived locally from a foothill area to the east and intercalated with basalt on the east side of the ancestral valley where the Clay Bank basalt is exposed at present.

Old gravel of unknown origin.--The only sedimentary unit in the Bellevue area for which there is no clue of origin or relative age is a deeply weathered gravel resting on granitic rock in the lower Rock Creek valley. This gravel is typically exposed in a prospect pit on hill 5158 about 1 1/2 miles northeast of the Rock Creek Ranch. This hill and the western end of hill 5236, one mile south, are the only places where the gravel has been found. The total area is about 1/2 square mile. The thickness of the gravel was more than 200 feet which is the present thickness of gravel on hill 5158 after extensive erosion.

About 40 percent of the pebbles are Carboniferous sedimentary rock, 60 percent are aplite, pegmatite and quartz derived from granitic rock and a few are volcanic rock, including some Square Mountain basalt. About 75 percent of the total volume is a sandy matrix derived largely from granitic rock. At the prospect pit the average diameter of larger pebbles is 2 inches, and the maximum diameter is about 1 foot. A few boulders of resistant rock, several feet in diameter, are lag on the eroded top of the deposit.

The gravel is younger than the Square Mountain basalt which is the youngest rock type identified in pebbles in the deposit. The gravel appears to be older than the Bellevue basalt because the gravel on hill 5153 is at too high an altitude to have been graded to any known base level controlled by any of the Bellevue basalts, yet the possibility exists that the gravel has been uplifted since its deposition and in this way may be associated with one of the Bellevue basalts.

Pediment deposits.--Pediments formed chiefly by lateral corrasion are typically developed at the confluence of Kelly Gulch and Croy Creek, and on the East Fork of Rock Creek (pl. 1). A veneer of gravel, about 20 feet thick, is composed dominantly of resistant Carboniferous sedimentary rock which is the bedrock for the entire headward part of both drainages and 10 percent alaskitic rock which is derived from dikes cutting the Carboniferous bedrock. Angular to subrounded boulders are as much as several feet in diameter. The Kelly pediment is cut on sedimentary rock, and the East Fork pediment is cut in part on granitic rock and in part on sedimentary rock. Both pediments are deeply dissected, as much as 100 feet (fig. 15).

A pediment on Little Rock Creek, probably formed partly by lateral corrasion and partly by interstream erosion processes, is developed entirely on granitic bedrock. Twenty percent of its pebbles are resistant sedimentary rock derived from a small fraction of a square mile of Carboniferous bedrock at the head of the drainage basin. The other pebbles consist of 60 percent aplite and pegmatite, and 20 percent volcanic rock. The pebbles are subangular to subrounded and have an average larger diameter of 3 inches and a maximum diameter of about one foot.

The pediments in the Bellevue area probably represent a stable or slowly degrading period of erosion; at the end of that period they were probably more prevalent than is indicated by the few scattered remnants. The Kelly and East Fork pediments are relatively well preserved by a veneer of gravel consisting of resistant rock types. The Little Rock Creek pediment is less well preserved by a veneer of gravel consisting of few resistant rock types. Other pediments, probably widely developed on the extensive granitic terrane, are not preserved because there was no veneer of gravel. That such pediments did exist is suggested by even-crested ridges of interstream areas, especially areas in the vicinity of larger streams; for example, the area east of the Little Rock Creek pediment (fig. 40, appendix). In a granitic terrane pediments form by interstream erosion processes rather than by lateral corrasion because aplite and pegmatite are the only sources of gravel and their amount is much too small to form a continuous gravel veneer; hence, there is no protective cover of gravel. Dissection of such armorless pediments would cause rapid erosion of the interstream areas with destruction of the flat element which typifies the pediment in general.

On the basis of degree of weathering and dissection the pediments mapped in the Bellevue area are pre-Bull Lake in age. Flint and Denny (1958, p. 156) have tentatively assigned certain pediments associated with glacial deposits in Utah to the Sangamon and (or) older interglacial stages.

Sediments associated with the Bellevue basalt, Wind Ridge member.--The oldest sediments associated with the Bellevue sequence of basalt flows are lacustrine silt and clay deposited in a lake impounded by the lower Wind Ridge flow in the ancestral Wood River Valley. Laminated silt is well exposed in a stream-cut scarp west of Ditto Hill, 0.7 mile southeast of Stanton Crossing (fig. 17). Drill logs indicate that over 100 feet of silt and clay were deposited in the southern part of the Wood River Valley at the time of the spreading of the lower Wind Ridge flow.

"Sand," as much as 50 feet thick where recorded in drill logs, overlies the lacustrine sediments and represents a post-lake, fine-grained alluvial facies which is poorly exposed in a gully, cut in the scarp of the Big Wood River, 0.6 mile southwest of Stanton Crossing. Coarse gravel, 10 to 20 feet thick, overlies the sand north and east of Stanton Crossing but overlies the lower Wind Ridge basalt west of Stanton Crossing. It is best exposed in the south river scarp above the gaging station, 1/2 mile southwest of Stanton Crossing. The gravel was deposited by the Big Wood River after the river re-established its course across the lower Wind Ridge lava dam. The upper Wind Ridge flow in turn overlies this gravel west of Stanton Crossing. Valley-side alluvial deposits, now deeply weathered and dissected, are graded to the top of the upper Wind Ridge flow and are as old as the flow. They are mapped in the stream valley southeast of Ditto Hill.

The high terrace north of Poverty Flat is a composite, valley-side fan made up of several individual fans of sidestreams and of valley-side alluvium. At Diebenow Creek the alluvium is more than 150 feet thick, the height of the terrace (fig. 16). The alluvium, as exposed at the base of the terrace, is dominantly sand and pea gravel; the average diameter of the larger pebbles is 1/4 inch, and the maximum diameter is 1/2 inch. In one small lens of coarser gravel, the average diameter of the larger pebbles is one inch, and the maximum diameter is 2 inches. Most of the sand was derived from granitic rock, and the remainder was derived from sedimentary rock, whereas, the pebbles are dominantly Carboniferous sedimentary rock and about 10 percent aplite and pegmatite.

The fan was probably built in response to the more than 100 feet of upbuilding in the Wood River Valley following the extrusion of the Wind Ridge basalt flows. Sometime prior to Bull Lake time, the fan was dissected on its west side by Diebenow Creek, perhaps as a result of the diversion of the Creek to the west and outlet valley at Stanton Crossing while the Big Wood River was still flowing to the east. The dissected part of the fan was then aggraded with periglacial alluvium of Diebenow Creek during the Bull Lake stage; this periglacial terrace is somewhat lower than the high terrace and is graded to the Bull Lake proglacial outwash in the Wood River Valley. The age of the top of the high terrace is probably somewhat younger than the upper Wind Ridge flow and somewhat older than the Bull Lake stage. The front of the terrace was last trimmed back during post-Pinedale time, so that weathering and dissection of the youthful scarp are not guides to the age of the fan.

Sediments associated with the Bellevue basalt, Myrtle member.--Coarse sand underlies the Myrtle flow along the canyon walls of the Big Wood River, one mile north of Magic Dam. A section about 20 feet thick is exposed about 100 feet below the level of the Magic Reservoir spillway. The deposit consists of coarse sand and contains enough silt to be partially indurated. It is poorly laminated and is capped by a layer of clay, approximately a foot to several feet thick. The sediment was probably deposited in a shallow lake impounded by a basalt flow which may be a lower flow, a submember of the Myrtle basalt.

Thick lacustrine and fluvial sediments stratigraphically overlie the Myrtle flow; they were for the most part deposited upstream from the Myrtle lava dam. The sediments are exposed in cuts as much as 50 feet high along the west shore of Magic Reservoir from Magic Resort to Hot Springs Landing, and for about a mile up Camas and Rock creeks (fig. 18).

A fluvial facies of crossbedded and layered sand containing small lenses of fine gravel and a few layers of silt is exposed in a road cut on Highway 68, 1000 feet west of Rock Creek. The fine gravel consists dominantly of granitic detritus with up to 10 percent Carboniferous sedimentary rocks and minor amounts of volcanic rocks.

A deltaic facies is exposed opposite the mouth of Poison Creek. It consists of 80 feet of layered silty sand with lenses of pea gravel and a 3-foot layer of white clay, all of which are overlain by several tens of feet of foreset sand dipping about 30° westward. Most of the pea gravel consists of locally derived volcanic rock. Sediments of Camas Creek or the Big Wood River are not present.

A lacustrine facies grades from laminated silt and clay at Magic Resort to layered fine sand and silt northward along Magic Reservoir. Water wells at Magic Resort are reported to pass through about 100 feet of "clay and sand."

Sediments associated with the Bellevue basalt, Macon member.--Sediments which overlie the Macon flow are subdivided on the geologic map on the basis of source (pl. 1) into Willow Creek gravel, Camp Creek gravel, Rock Creek gravel, Camas Creek gravel, and valley-side deposits. In the Bellevue area the sediments are chiefly alluvial fans built out over the distal part of the Macon flow, whereas, west of the Bellevue area the sediments consist of alluvial fans built out over lacustrine deposits

(Walton, 1960) laid down in a lake impounded by the Macon flow.

The Willow Creek gravel forms a broad dissected alluvial fan on top of the Macon basalt where Willow Creek debouches from the mountains. The gravel, capped by local slope deposits, is typically exposed on the west bank of Willow Creek along the county road just east of Willow Creek School. Drill logs record 100 feet of gravel overlying rocks of the Macon flow two miles south of the apex of the fan. The gravel thins to the south against the northward sloping surface of the basalt and becomes zero, 3 to 4 miles south of the apex.

The gravel consists of about 25 percent sedimentary rocks, about 75 percent granitic rocks including aplite and pegmatite, and minor amounts of Challis volcanics. The average diameter of larger pebbles is 2 inches, and the largest pebbles are about 8 inches in diameter. All but the most resistant quartzites are deeply weathered; the granitic pebbles crumble to gruss when disturbed.

The Camp Creek gravel was for the most part thoroughly mixed with Willow Creek gravel, but a narrow strip consisting entirely of Camp Creek gravel extends southeastward between the east margin of the Willow Creek fan and the east valley side (pl. 1). Along the southern part of the strip, the Camp Creek gravel is 10 to 20 feet thick and is underlain by Willow Creek gravel, as is indicated by drill logs of the U. S. Bureau of Mines. Northward the Camp Creek gravel thickens to 30 and 50 feet. The gravel contains aplite and pegmatite pebbles, a few pebbles of Square Mountain basalt, and a large amount of granitic sand. The average diameter of the larger pebbles is 2 inches.

The Rock Creek gravel makes up the highest part of the terrace east and west of lower Rock Creek and is poorly exposed in shallow cuts along Highway 68 east of Rock Creek. The gravel is estimated to be 20 feet thick. In one place pebbles are about 85 percent aplite and pegmatite, and 15 percent sedimentary rock. The average diameter of larger pebbles is about 1/2 inch, and the maximum diameter is 1 inch. The relationships between these sediments and the underlying fine-grained fluvial facies of the Magic sediments are not dependent on sediment identification but are more reliably determined by correlation of terrace heights.

The Camas Creek gravel rests on the Macon flow downstream from the Willow Creek fan. In general it is poorly exposed along the west edge of Magic Reservoir from Camp Creek almost to Magic Dam (fig. 19). It is well exposed in a road cut 1000 feet west of Poison Creek on Highway 68, where it overlies Myrtle sand, and in a beach cut at Myrtle Point in the sourthern part of Magic Reservoir. Thicknesses vary from 10 feet at Poison Creek to 20 feet at Magic Resort. At Poison Creek the gravel consists of about half Carboniferous sedimentary rock and half aplite and pegmatite, plus small amounts of volcanic rocks, and at Myrtle Point Carboniferous rock is dominant. The average diameter of larger pebbles is 1 1/2 inches, and the maximum diameter is 3 inches.

The Camas Creek gravel was largely derived from the Willow Creek drainage, and smaller amounts were derived from Camp and Rock creeks. Little or no sediment was derived from that part of the Camas Creek drainage which is upstream from the Willow Creek confluence because transportation of sediment from upper Camas Creek was interrupted by the lake impounded by the Macon flow just west of Willow Creek.

The fifth Macon sedimentary unit is designated "valley-side deposits." The largest deposits occur on the south flank of Moonstone Mountain and on the northeastern slope of the Bennett Hills about one mile southwest of Magic Resort. The small sidestream fans, which constitute most of the deposits in both areas, are composed of volcanic detritus and contain more sandy and silty material than gravel. The basis for associating the valley-side deposits with the Macon flow is that their surfaces are graded to surfaces of mainstream deposits impounded by the Macon flow. Moreover, the degree of weathering and dissection is about the same in the sidestream and mainstream deposits.

Proglacial outwash

General statement.--The Bellevue fan, making up most of the large triangular-shaped valley floor south of Bellevue, is a gravel terrace which can be traced northward to moraines in the upper Big Wood River drainage basin. The moraines were deposited during the latest glacial stage; the gravel is clearly proglacial outwash of the same age. Higher terraces, though fragmental and inconspicuous, can be correlated with an earlier set of moraines on the basis of several lines of evidence. No older glacial deposits are known to occur in the Bellevue area.

Throughout the Rocky Mountains two late Pleistocene glaciations are recognized. In western Wyoming, Blackwelder (1915) named the later, the Pinedale stage, and the earlier, the Bull Lake stage. These stage names either have been directly applied in many places in the Rocky Mountains (Fryxell, 1930; Horberg, 1940; Holmes and Moss, 1955; Richmond, 1960, p. 1372), or they have been used to make tentative correlations (Flint and Denny, 1958).

In the southern half of Idaho, deposits of two glacial stages have been mapped by Capps (1940) in Secesh Basin, Valley County; by Mackin and Schmidt (1956) in Bear Valley and Long Valley, Valley County; and by Anderson (1956) in Lemhi County. Williams (1961(?), in press; 1957, M. S. thesis, University of Washington), in Stanley Basin, about 40 miles northwest of the Bellevue area, correlates two glacial stages with the Pinedale and Bull Lake on the basis of topographic and stratigraphic position, weathering, and dissection of moraines and outwash deposits.

These same factors make it possible to correlate the two Wyoming stages with those in Bear Valley, Long Valley, and the Bellevue area. During a field conference with Mr. Charles Hunt in Long Valley in 1955, it was decided that the correlation was sufficiently definite to use Blackwelder's terminology in south-central Idaho rather than to use local names.

The headwater parts of the Big Wood River drainage basin are intensely glaciated (Umpleby, Westgate, and Ross, 1930, p. 36, 39), and aerial photographs clearly show that morainal deposits of two principal stages are present. Intensive glaciation is to be expected because large areas are above 8500 feet, which is the altitude of the lowest principal cirques of the Pinedale stage. This agrees with data given by Williams (1961(?), in press) in the nearby Stanley Basin where the Pinedale cirque line varies from 8500 feet to 9000 feet depending upon local topographic and meteorologic factors. The largest valley glaciers in the Big Wood River drainage extended down to about 6500 feet altitude.

In the Willow Creek drainage only a small area is over 8500 feet altitude suggesting that Pinedale glaciation was confined largely to cirque glaciers and that Bull Lake glaciation was probably intense enough to form several short valley glaciers.

Bull Lake stage.--The Bull Lake proglacial gravel can be observed in the river scarp on the valley side at Stanton Crossing and in a borrow pit at the foot of Timmerman Hill where the Silver Creek road joins U. S. Highway 93. Excellent terraces are preserved southwest of Hot Springs Landing across Camas Creek (fig. 19), north of Magic Resort across Lava Creek, and between Magic Resort and Myrtle Point; all these terraces exhibit a degree of dissection similar to deposits of Bull Lake age elsewhere in the Northern Rockies. On lower Willow Creek, Bull Lake outwash has been tentatively mapped on the basis of terrace position and degree of dissection (pl. 1).

The Bull Lake terrace between Stanton Crossing and Magic Dam is about 100 feet above the modern Big Wood River, but at the Timmerman Hill outcrop, two miles upstream from Stanton Crossing, the terrace is about 40 feet above the modern stream and only a few feet above the Pinedale outwash terrace. The thickness of the terrace gravel at Hot Springs Landing is about 30 feet and at Stanton Crossing is only 11 feet; a well log (1S-19E-13dd1) east of the Timmerman outcrop records 16 feet of gravel which is probably all Bull Lake. The apparent thinness of what might be expected to be a thick aggradational outwash deposit is explained by the geologic history--the Big Wood River was diverted to its present, western outlet as a result of Bull Lake aggradation, and the terrace gravel at and below Stanton Crossing was deposited by the Big Wood River late in the aggradational cycle.

The pebbles at Stanton Crossing consist of about 80 percent Paleozoic sedimentary rock, 10 percent Tertiary volcanic and 10 percent granitic rock. At other outcrops, Paleozoic sedimentary rock types are similarly abundant. The larger pebbles at Stanton Crossing have an average diameter of 2 inches and a maximum diameter of 5 inches. Pebbles of volcanic and some less-resistant sedimentary rocks are deeply weathered, and some pebbles of granitic rock crumble upon handling. At the Timmerman Hill locality a paleosol has a chocolate brown B horizon, one to two feet thick, which contains deeply weathered gravel in a heavy clay matrix. The paleosol underlies slightly weathered Pinedale loess. In many places, erosion has prevented soil development or has removed any paleosol which may have developed.

The area of the Bull Lake outwash mapped in the Bellevue area is small because the outwash in the Big Wood River valley upstream from the mouth of Reed Creek has been buried by Pinedale deposits and because the outwash from Reed Creek downstream to Magic Dam has been largely removed by erosion except where it has been preserved on rock-defended terraces.

Pinedale stage.--The Bellevue fan, which has its apex at Bellevue, is a large apron of Pinedale proglacial outwash comprising about 75 percent of the Wood River Valley. The outwash gravel is continuously exposed in river scarps along the entrenched Big Wood River from Stanton Crossing northward to beyond Bellevue (pl. 1). The gravel is typically exposed in a borrow pit along U. S. Highway 93, 4 miles south of Bellevue. A set of terraces occupying about 30 percent of the relatively narrow valley floor north of Bellevue is an upvalley extension of the fan (fig. 23).

The terrace scarp of the Bellevue fan is about 30 feet high, and the terrace scarp between Bellevue and Ketchum averages about 40 feet high. The Pinedale outwash is estimated to be about 50 feet thick on the basis of a study of the accordance of surfaces of small tributary terraces with the Big Wood River valley floor and a study of "drowned" mouths of small tributary valleys. The many deep well logs in the Big Wood River valley do not distinguish between the Pinedale and underlying Bull Lake outwashes.

Ten pebble counts of material collected at about equal intervals from Ketchum to Stanton Crossing, average 50 percent Carboniferous sedimentary rock, 25 percent Tertiary volcanic rock, 15 percent early Paleozoic sedimentary rock, and 10 percent granitic rock. In addition, the outwash contains small amounts of interstitial sand and silt. In the ten samples the average diameter of the larger pebbles is 2 inches, and the maximum diameter averages 9 inches; in general, the maximum diameter decreases from 12 inches at Ketchum to 6 inches at Stanton Crossing.

A bar and swale topography on the surface of the outwash indicates deposition by a braided stream; this feature is best seen on high altitude aerial photographs. Depositional structures within the gravel are generally not seen. The gravel is not noticeably weathered except for a few of its granitic pebbles. It is overlain by a few inches to several feet of gray loess.

Periglacial deposits

General statement.--Large areas of alluvium and colluvium are mapped as periglacial deposits, that is, deposits formed under a rigorous Pleistocene climate in which accelerated slope erosion caused aggradation of valley floors and accumulation on the valley sides (Bryan, 1949; Smith, 1949; Denny, 1951; Peltier, 1950). The periglacial landforms so completely dominate the smaller valleys as to suggest that post-Pinedale erosion has been negligible; the relationships correspond closely to those described by Budel in Europe as, for example, in northwest Germany, West Jutland and southern England. Budel's conclusion that the "Holocene geomorphic processes, in contrast to Pleistocene, were so weak and insignificant, that they could not change fossil landscapes . . . we live today . . . on a fossil tundra" (Budel, 1953, translated by Wright and Alt, 1959, p. 4) applies just as well to the small valleys in the Bellevue area.

Debris comprising the periglacial deposits is locally derived, either directly from adjacent valley sides or from short tributary valleys. The material is angular to subangular and contains a high percentage of fines; where the bedrock is Carboniferous sedimentary rock, a rubble of rock fragments is in a matrix of silt; where the bedrock is granitic rock, a silty granite sand contains scattered angular fragments of aplite and pegmatite. Sorting is poor, although washing, generally in the more central part of the valley, gives rise to lenses of silty sand and gravel.

The thickness of the valley-bottom deposits averages about 8 feet where tested by drilling on lower Rock, Reed, and Brock creeks but varies depending upon the size of the valley and the erodibility of the valley-side bedrock under the periglacial conditions.

That the valley bottom deposits are Pinedale in age, rather than Recent, is readily demonstrated in the vicinity of the modern entrenched Big Wood River where the sidestream valley floors are dissected into a set of terraces graded to the Pinedale outwash terrace of the master stream. An older set of terraces in the valleys of many sidestreams is probably also periglacial in origin. On the basis of degree of dissection and weathering, the older terraces are believed to be Bull Lake in age.

Bull Lake deposits.--Extensive deposits of alluvium and colluvium of Bull Lake age and of periglacial origin are mapped along the west side of the Big Wood River Valley south of Poverty Flat and on the slopes of Timmerman and Ditto hills. The deposits of Brock Creek are graded to the Bull Lake outwash terrace of the Big Wood River at Stanton Crossing.

Bull Lake periglacial deposits have been, for the most part, either buried or destroyed by Pinedale periglacial activity.

Pinedale deposits.--The valley of Croy Creek typifies Pinedale periglacial deposition. That the valley floor is choked by alluvium from its own tributaries is shown by the sinuous longitudinal profile of the mainstream (fig. 21). The sinuousness is strikingly shown in a plot of stream slope against distance along the stream (fig. 21, lower curve) which makes it clear that Croy Creek has a steeper slope at the mouth of each of its principal tributaries. On aerial photographs this steepening of grade is seen to be caused by sidestream alluvial fans built out on the Croy Creek valley floor; probably during the last periglacial stage. The fans of the principal tributaries, Kelly, Red Elephant, and Bullion creeks, still cover 50 to 75 percent of the width of the mainstream valley floor. Croy Creek is incised in steep-walled meanders at each of the three confluences, and each tributary stream is likewise incised in its own fan, indicating that the fans are not growing at present and that Croy Creek has not been vigorous enough since the deposition of the fans to shift laterally on its valley floor. Between the major tributaries Croy Creek meanders lazily on a slackened slope; these reaches of the stream are still being aggraded in a feeble attempt by the stream to smooth its profile. The creek has incised its Pinedale valley floor for a distance of only 2 miles from its junction with the Big Wood River.

The Rock Creek valley shows particularly well the domination of the mainstream valley floor by tributary fans (pl. 1). Rock Creek meanders sluggishly along a narrow meadowy course from one side of its valley to the other as dictated by the tributary fans. Although the tributaries are relatively inactive today, Rock Creek is still not vigorous enough to cut laterally to regain control of its valley floor.

Patches of Pinedale deposits, mapped on the headward reaches of all streams (pl. 1), further indicate the widespread effects of the periglacial climate. Catenary-shaped valley bottoms, choked with local wash, alternate with smooth round ridge tops, covered by one a few inches of rock debris or granite sand. The valley bottoms are dry except locally where emergence of springs indicates that water commonly flows within the alluvium.

The most striking periglacial features of the Bellevue area are terraced alluvial fans of minor tributaries that enter the Big Wood River valley (fig. 20). The three examples shown in figure 22, all on the east valley side in the vicinity of Hailey (fig. 23), have drainage basins of 1 to 2 square miles in area and gradients of 700 ±200 feet per mile. The fans are composed of an angular to subangular rubble of bedrock fragments in a matrix of clayey silt. The features are compound, consisting of (1) a high fan which is truncated by the Big Wood River and trenched in turn by its tributary, (2) an intermediate fan which in most places is truncated and trenched, and (3) a low fan, much smaller in size than the other two, which is neither truncated nor trenched. On a few tributaries the intermediate fan was not truncated by the Big Wood River, and as a result the low fan is not present.

The low fan is deposited on the Pinedale outwash terrace of the Big Wood River and is clearly postglacial. The intermediate fan was truncated by the Big Wood River when the river was flowing on its Pinedale valley train. It seems likely, by analogy with the fan building by minor streams elsewhere in the area, that the fan was built during Pinedale time, when the braided Big Wood River had little power to cut back valley side deposits, and that the truncation probably occurred during the waning phase of the outwash aggradation. The same analogy suggests that the high fan is Bull Lake in age. Degree of weathering and dissection of the intermediate and high fans conforms with that of the respective stages.

Postglacial deposits

Big Wood River gravel has been deposited on noncyclic terraces cut at all levels between the Pinedale outwash terrace and the present flood plain from Stanton Crossing northward to beyond Bellevue. The terraces are the result of lateral corrasion during a period of degradation following glacial aggradation.

Deposits of spring-fed streams, a second type of post-Pinedale sediment, underlie almost the entire southern half of the Wood River Valley but because the original outwash surface is still partly preserved the entire area is mapped as Pinedale outwash on plate 1. Many spring-fed streams emerge along an east-west spring line at the latitude of Gannett and flow either west to the Big Wood River or east to form Silver Creek. The spring line, where the groundwater table in the valley alluvium intersects the surface, divides the valley into a dry northern half and a wet, in places swampy, southern half.

North of the spring line a loamy silt, consisting chiefly of loess a few inches thick, overlies the outwash gravel. Little has changed here since the end of glacial time.

South of the spring line, in an east-west belt about 2 miles wide, the spring-fed streams are entrenched 5 to 15 feet below the surface and flow in tight meanders each on its own narrow flood plain. The interstream areas consist of massive, gray clay, 4 to 6 feet thick, resting directly on gravel and overlain by about 2 feet of black loamy soil. The clay and gravel are strongly bleached in an organic reducing environment whereas the overlying soil is not. Because pebble counts, size analyses, and altitude relationships indicate that the gravel is the same as the Big Wood River outwash gravel found north of the spring line, the clay is believed to have been Pinedale loess accumulated on a wet, vegetated surface and simultaneously altered to clay in an environment dominated by humic acid. The top soil is postglacial loess accumulated under drier conditions resulting in part from entrenchment of the spring-fed streams.

To the south the interstream areas wedge out, and the spring-fed streams unite to form Silver Creek. South and east of Hay the entire valley floor becomes a single, very broad flood plain worked by Silver Creek. Overbank silt and oxbow-lake clay, 2 to 3 feet thick, rest on Silver Creek gravel which is reworked Big Wood River outwash. Recent loess, a few inches thick, commonly rests on the older parts of the flood plain.

West of the fan divide, west-flowing spring-fed streams have regraded much of the outwash terrace surface to the level of the modern Big Wood River. East of the fan divide, the amount of regrading by Silver Creek is probably slight and cannot be evaluated accurately from the available topographic maps.

Sidestream alluvium constitutes a third type of postglacial deposit. In most places it is impossible to distinguish this deposit from sidestream alluvium of Pinedale age. Small post-Pinedale alluvial fans can be seen in figure 20 and are mapped in figure 22 where the detailed terrace relationships indicate age. The largest alluvial fans are those of the Seamans and Slaughterhouse creeks at Bellevue (pl. 1); they were deposited after the last swing of the Big Wood River against the east valley side at the end of the Pinedale aggradation. However, much of the sidestream alluvium was probably deposited when slope erosion was still rapid during the waning phases of the periglacial climate rather than being deposited uniformly during the entire Pinedale to Recent interval.

Recent alluvium

The present flood plain of the Big Wood River is mapped as Recent alluvium. It is about 30 feet below the Pinedale outwash surface and makes a strip, about one mile wide, across the Bellevue area. Willow Creek has a flood plain about 100 to several hundred feet wide which is only several feet below its Pinedale outwash terrace. The area of Recent alluvium of the other sidestreams is too small to be mapped at the scale of the geologic map and commonly is too ill-defined to map in the field.

LATE QUATERNARY GEOLOGIC HISTORY

General Statement

The Quaternary history of the Bellevue area, as indicated earlier, is characterized by the spreading of successive basalt flows on the valley floor of the Big Wood River; most of the flows blocked the river, causing lacustrine and fluvial sedimentation upvalley, and some flows caused shiftings of the Big Wood River back and forth between its eastern and western outlets. Independent of the basalt eruptions, outwash gravel from glaciers, occupying the headwaters of the Big Wood River drainage basin, flooded the Wood River Valley during several periods of glaciation; the floods of outwash caused the river to aggrade and at least once shifted it from one outlet valley to the other. Downcutting occurred between episodes of eruption and glaciation, but the net result of many cycles of lava damming, sedimentation, glacial aggradation, and downcutting has been an average raising of the river.

The sequence of events of the known part of the Quaternary history is summarized in table 3 and is presented in sketch maps and generalized sections on a scale of 1:250,000 (figs. 24-31). The interpretations shown in the drawings result from combining surface observations with information from a large number of water wells. The drill logs have been brought together by R. O. Smith and Walton (Smith, 1959; Smith, 1960; Walton, 1960). The surface and subsurface data used as a basis for interpretation are compiled in cross sections along the longitudinal profiles of the Big Wood River, Camas Creek and Silver Creek on a scale of 1:48,000 and with a vertical exaggeration of about 1:40.

Some of the more critical relationships between the flows and the sediments are shown in a profile-section through the east and west outlet valleys of the Big Wood River and across the south end of the Bellevue fan divide (fig. 39). Three other profile-sections, figures 41, 42 and 43, are in the appendix as documentation for the sketch maps and generalized sections used to outline the history, and for use by those concerned with local details. Figure 41 is along Camas Creek, figure 42 is along the modern Big Wood River, and figure 43 is along the east outlet valley. Correlations between surface and subsurface data depend largely on the lithologic descriptions in the drill logs; the correlations are especially reliable where a vertical succession of lithologic units rather than a single unit can be traced between adjacent wells.

The depositional slopes of the basalt flows and alluvial deposits can be measured where these form the surface and also where the flow or deposit is identified in drill logs. Slopes determined in this way are used as a basis for correlation between subsurface units where drill log descriptions alone are not sufficient for purposes of correlation.

Lava dams blocked the Big Wood River where the various basalt flows spread across the river valley. The altitudes of the lava dams were determined on cross-valley sections by extending the profile of the flow from the source vent to the lava dam site; five such sections, plotted on the same scale as the profile-sections are shown in figures 32 through 36. Knowing the altitude of a single lava dam makes it possible to determine (1) the maximum possible height of lacustrine sediments deposited in the lake, (2) the amount of aggradational fluvial sediments deposited above the lake sediments, and (3) whether the river maintained its course or was diverted to the other outlet valley. Knowing the altitude of each lava dam, the relative age of each flow, and some data about the sediment deposited upvalley, makes it possible to reconstruct some of the detailed events associated with the Bellevue basalt volcanism.

Generalized structure contours, drawn on the either still existing surfaces or on reconstructed surfaces of the basalt flows and fluvial deposits on figure 40 (appendix), are based on information from topography, measured sections, and drill logs. Also shown on figure 40 are the locations of the profile-sections, the cross-valley sections, and those water wells from which drill log data are used.

Faulting and warping pose a problem in formulating any hypothesis in which stream and lava depositional slopes in feet per mile are used in correlation and interpretation; this is especially serious in an area in which Quaternary volcanism was presumably accompanied by deformation. However, structural displacements are known to be associated only with the large Macon eruption; mild deformation may have occurred at other times, but as far as is known, it did not enter into the sequence of drainage changes discussed here.

Pre-Hay drainage

The Bellevue fan is triangular in shape, with the town of Bellevue at the apex on the north, and the Picabo Hills at the base on the south. The modern Big Wood River exits through a narrow valley to the west; the broad outlet valley on the east is occupied only by Silver Creek, which rises on the fan. Just prior to the Hay eruption, the broad eastern valley was used by the ancestral Big Wood River (fig. 24); this situation sets the stage for developing the late Quaternary history. The southwestern half of the Bellevue fan may have been foothills at that time, or it may have been a valley floor used by the Big Wood River at some still earlier time. Bellevue flows older than the Hay flow already occupied the east outlet valley south of Picabo; two of these flows are shown in section in figures 24 and 39.

Hay flow

The Hay flow was extruded from its vent at about 4900 feet altitude on the north side of the broad valley floor of the ancestral Big Wood River (fig. 25). The river was blocked where the flow built up against the south valley side at an altitude of 4840 feet as estimated from the cross-valley section through the Hay lava dam and vent, figure 32. The valley floor on which the Hay flow spread consists of gravel with a smooth continuous slope of about 11 feet per mile. The pre-Hay valley floor can be traced eastward (fig. 39) from a gravel aquifer south of Gannett (well 1S-19E-17dc1, fig. 40), under basaltic cinders and a thin basalt tongue at the flow front, under massive basalt as much as 160 feet thick near the vent, and southward to Priest Station where the gravel underlies 30 feet of massive basalt (well 2S-20E-12cd1).

The low slope of 11 feet per mile was probably controlled by a base level imposed by the latest pre-Hay flow south of Priest Station. Upvalley from Hay this influence diminished, and a more normal gradient of 22 feet per mile is assumed in figure 25.

The Big Wood River spilled over from a lake impounded by the Hay flow (fig. 25) to establish or re-establish its course in the west outlet valley. If the present triangular shape of the Wood River Valley had not existed prior to the diversion, it was developed at this time.

Lower Wind Ridge flow

The lower Wind Ridge flow was extruded from the south side of the west outlet valley (fig. 26) and spread north into the outlet valley where it blocked the Big Wood River about 1 to 1 1/2 miles west of Stanton Crossing. With a base at an altitude of 4700 feet and a top at an altitude between 4840 and 4875 feet, the flow was at least 140 feet thick at the lava dam.

A large lake, Wind Ridge lake, impounded between the lower Wind Ridge and Hay flows, filled with silt and clay to a thickness of about 100 feet and to an altitude of about 4840 feet, the height of the spillway across the Hay flow (figs. 26, 39). The lake extended about halfway up the Wood River Valley toward Bellevue, and drill logs in the shallow northern half of the lake indicate an intertonguing of deltaic sand and fine gravel.

As the lake filled with sediments to a water plane at about 4840 feet, the Big Wood River built a fan-delta represented by a sheet of sand as much as 60 feet thick on the fan divide. The fact that this sand extends into both the east and west outlet valleys indicates that the river was switching from one outlet to the other during the aggradational phase. On the west side of the fan near Stanton Crossing, the sand grades upward into coarse gravel which indicates that the Big Wood River eventually established grade across the lower Wind Ridge flow. The gravel crops out in the west outlet valley west of Stanton Crossing, and parts of it are mapped on the geologic map (pl. 1) as sediment associated with the Wind Ridge flow; it is also shown in the profile-section, figure 39.

Upper Wind Ridge flow

The upper Wind Ridge flow probably came from the same source as the lower Wind Ridge flow (fig. 27). Its surface slopes about 90 feet per mile to where it intersected the north valley side at an estimated altitude of 4930 feet (fig. 33). The Big Wood River was blocked by about 60 feet of basalt.

The shallow lake impounded by the flow rose to the altitude of the fan divide at about 4900 feet and spilled across the divide into the

the east outlet valley. Silt and clay, deposited in the lake, are well exposed at an altitude of about 4870 feet along the small creek west of Ditto Hill and are recorded in drill logs (for example, well 1S-18E-13cal) in the southwestern part of the Wood River Valley.

Lower Myrtle(?) flow

Sometime after the diversion of the Big Wood River to the east and before its return to the west, the Myrtle and Macon flows were extruded in the eastern part of Camas valley (figs. 28 and 29). Several lines of evidence indicate that the river did not occupy its western outlet during the Myrtle and Macon volcanism. First, Big Wood River gravels have not been found associated with these flows. Second, the 100-foot difference in altitude between the west and east outlet valleys caused by the upper Wind Ridge lava dam seems sufficient reason for the Big Wood River to have remained in its east outlet valley for a long time--time enough for both the Myrtle and Macon eruption. Third, the present canyons of the Big Wood River below Magic Dam, of Camas Creek above the Camp Creek confluence, and of the Big Wood River at the Rock Creek confluence are all about the same age as judged by their similar and youthful form which is considerably younger than would be expected if the river had cut down immediately after the spreading of the Macon flow, thus suggesting that the down cutting occurred when the Big Wood River was diverted to the west valley long after the Macon flow. The Stanton terrace (figs. 39 and 35, sections D, E), composed of Big Wood

River gravels, clearly marks the return of the river to the west after the spreading of the Macon flow.

Prior to the eruption of the Myrtle flow, an older flow, probably a submember of the Myrtle basalt, blocked Camas Creek presumably near Magic Dam and impounded a lake in which silt and sand were deposited (fig. 28). Part of these sediments crop out between altitudes 4690 and 4720 in canyon walls submerged in Magic Reservoir 1/2 mile north of Magic Dam. (In September 1960 Magic Reservoir was drained to about an altitude of 4690 feet which afforded an excellent opportunity to observe many outcrops normally submerged.) The sediments in the reservoir walls probably correlate with "clay and sand" recorded between altitudes 4680 and 4750 feet in the log of a well (2S-17E-11ab1) two miles west of the canyon on the west valley side. The lava dam of this lower Myrtle(?) flow was probably about 4740 feet, the altitude shown in the cross-valley section (fig. 34); more probably the actual lava dam site was to the south, now buried under the Myrtle flow, at an altitude as high as 4750 feet. Because gravel is not known to cap the clay and sand associated with the lower Myrtle(?) flow, the sedimentation cycle was probably not completed and Camas Creek did not transport its bedload across the flow; however, drill log and outcrop data are lacking across a 2 mile gap in the center of this ancestral valley.

The lower Myrtle(?) flow has not been seen in outcrop, but it might be expected to occur in the modern Big Wood River canyon south of Magic Dam. What may be the lower Myrtle(?) flow is recorded in the log of a water well (2S-18E-18aa1) located at Magic Dam where a basalt-like rock is 46 feet thick, has a top at 4735 feet, and is overlain by the Myrtle flow, but the log description does not clearly distinguish this "basalt" from underlying older bedrock. If the lower Myrtle(?) flow is not present at Magic Dam then the flow probably occurs to the west where it is still covered by the Myrtle flow.

Myrtle flow

The Myrtle flow issued from its vent at an altitude of about 5050 feet on the west valley side and spread 3 1/2 miles northeasterly to the present site of the Magic Dam where it blocked Camas Valley at an estimated altitude of 4790 feet (figs. 28 and 34). Myrtle lake was impounded behind the lava dam. At its eastern end the lake was at least 47 feet deep, this being the thickness of the flow at the spillway, but 4 miles west of Blaine, clays deposited in the lake are at an altitude of 4733 feet, 57 feet below the spillway (section A, fig. 35; figs. 40 and 41, appendix; Walton, 1960, fig. 4). These clays are directly overlain by the Macon flow, indicating that the lake was not completely filled with sediment and that the Camas drainage system was far from integrated when sedimentation in the Myrtle lake was ended by the Macon flow.

At each sidestream, alluvial fans fringed Myrtle lake and spread aprons of sand and silt out into the lake. Sand from the Rock Creek drainage, which is well exposed along the west side of Magic Reservoir (fig. 39; fig. 41, appendix), does not contain gravel and is not overlain by gravel, the implication being that the sand was deposited in a low-gradient fluvial environment or in shallow lake water. Deltaic bedding, dipping 15° to 25° west and exposed only in a single outcrop opposite Poison Creek, indicates that a lake did exist at least part of the time, but the widespread lack of deltaic bedding in other exposures suggests that Myrtle lake may have been ephemeral.

Thick deposits of bedded sand in the Rock Creek valley one mile upstream from Hot Springs Landing imply deposition by a low-gradient stream. A gravel facies associated with such a large amount of sand must have been deposited farther upvalley; perhaps the erosional remnants of the gravel "of unknown origin" (pl. 1) in the Little Rock Creek valley belong to this Myrtle depositional cycle.

Mild warping with uplift at Hot Springs Landing is implied for the Myrtle sand deposits in the Rock Creek Valley and in Camas Valley because the present altitudes and slopes of the deposits are greater than depositional. For example, the sand deposit in the northeastern part of Camas Valley is fan-shaped, sloping southward and westward from an apex at Hot Springs Landing, and the slope of the contact beneath the Macon flow is 31 feet per mile, from 4850 feet at Hot Springs Landing to 4800 feet at Magic Resort and 4790 feet at the Myrtle lava dam; this slope is too great to be entirely due to the fluvial deposition of sand.

Macon flow

The Macon basalt was extruded from its vent at an altitude of 5330 feet on the south edge of Camas Valley (fig. 29) and spread into a lake basin surrounded by steep-gradient alluvial fans as indicated in figure 35 where each cross section shows the flow lapping up on southward sloping fans. At its western extremity, the flow has a steep frontal edge, about 200 feet per mile, suggesting that it advanced into lake water, perhaps in part that of Myrtle lake and in part that impounded by itself (section A, fig. 35; fig. 40, appendix). A deep lake was impounded by the Macon flow at the contact of the flow and the ancestral Willow Creek fan (section B, fig. 35). The slopes of Willow Creek and Camp Creek were slackened by the Macon flow, and these two creeks built a composite alluvial fan, as much as 140 feet thick extending 3 1/2 miles out from the mountain front, in order to re-establish grade (fig. 40, appendix). In so doing, the spillway channel of Camas Creek was moved southward as much as 2 miles up the slope of the Macon flow (section B, fig. 35).

The waters of Camas Creek spilled over the lava dam at about 5000 feet altitude and followed a nearly consequent course, generally along the north and east valley side. Because little or no sediment came from the west across the lava dam, Willow and Camp creeks supplied most of the sediment to Camas Creek along the 10 mile reach between the lava dam and Hot Springs Landing. The original slope of this reach was too low to transport the coarse alluvium of Willow and Camp creeks; these creeks, in addition to building their own composite fan, supplied the material with which Camas Creek built up its slope and developed a broad alluvial valley floor, 1 1/2 miles wide at Hot Springs Landing and 2 miles wide at Magic Resort (pl. 1; figs. 40, 41, appendix).

Movement along the Willow Creek fault dropped the valley floor about 70 feet relative to the north valley side during the time Willow Creek and Camp Creek were building their composite fan, but because movement on the fault was completed before dissection of the Willow Creek fan began, the surface of the fan is not faulted. The north edge of the Macon basalt was broken by the faulting which displaced the frontal parts upward on the valley side where they were buried by alluvium (section C, fig. 35).

After Camas Creek developed its broad alluvial valley floor, the northeast corner of Camas Valley was warped upward; Camas Creek cut down antecedently across the upwarp. Camas Creek in regrading became stabilized about 80 feet below its upwarped gravel. Probably at this same time, Willow Creek entrenched, cutting a wide valley 30 feet below its original fan surface (fig. 40, appendix). Camp Creek, which was no longer controlled by Willow Creek on the Willow-Camp fan, established a consequent course along the eastern edge of the dissected Willow-Camp fan and the north valley side.

Priest flow

In the east outlet valley, the Big Wood River had probably entrenched across the Hay lava dam and had presumably stabilized with respect to a base-level control somewhere to the south, when the Priest flow was extruded from its vent at an altitude of 5000 feet on the east side of the Picabo Hills (fig. 30). Part of the flow spread northeastward about 2 miles forming a lava dam, 110 feet high, with a spillway altitude of about 4820 feet at the north valley side, one mile east of Picabo (fig. 36).

Priest lake, impounded by the flow, was the site of deposition of silt and clay which grade upward to sand and fine gravel (well 1S-20E-27db1). Because the altitude of the Priest spillway (4820) was below the level of the fan divide in the Wood River Valley, which was about 4900 feet at that time, the Big Wood River was not directly diverted to the west outlet valley by the Priest flow. The river continued to flow in the east outlet valley where it aggraded and built up its gradient to about 9 feet per mile, the present slope of the valley between the Priest and Hay stations.

Bull Lake outwash

During the Bull Lake glacial stage, outwash from glaciers in the upper Big Wood River drainage basin caused the river to aggrade its valley floor downstream from the glaciers. On the Bellevue fan the Bull Lake deposits range from 20 to 70 feet in thickness; the river evidently shifted back and forth between the east and west outlet valleys during the period of aggradation (fig. 30). The fact that only sand and fine gravel reached the Priest lava dam means that the river did not continue to flow out through the east outlet valley long enough to completely regrade the east branch of its Bull Lake valley train. The establishment of the Big Wood River in the west outlet valley is indicated by coarse gravel deposits extending with smooth profile from the apex of the Bellevue fan to Magic Dam (fig. 42, appendix).

The first spillover of the Big Wood River through the western outlet was probably caused by earlier erosional lowering of the outlet combined with Bull Lake aggradation on the Bellevue fan. One possible combination is given in figure 30, where erosion of 40 feet in the weak granitic rock in the north abutment of the upper Wind Ridge lava dam (4930 feet) would have lowered the spillway to 4890 feet, and aggradation of 30 feet at the apex of the Bellevue fan, on an average slope of 32 feet per mile, would have raised the river course enough for it to spill to the west. Faulting or warping of the west outlet valley, such as that associated with the Macon flow, may have introduced a third variable to be taken into account in determining the altitude changes involved in the diversion.

In most places the establishment of the Big Wood River in the western course was accomplished without much change in the levels of the valley floors formed earlier by the local streams; for example, in the Stanton terrace at Hot Springs Landing about 30 feet of Big Wood River gravel overlies about 10 feet of Camas Creek gravel indicating that the river, during its aggrading phase, did not rework the full thickness of older gravel.

The Bull Lake outwash surface, represented by the Stanton terrace, is above the Pinedale outwash from Magic Dam to about one mile north of Stanton Crossing (figs. 39, 37). Farther north on the Bellevue fan the Bull Lake outwash passes beneath the Pinedale outwash, but it is not possible to reconstruct the Bull Lake surface with any confidence because Pinedale and Bull Lake gravels are not distinguished in drill logs and because distances of extrapolation are too great to allow use of profiles of sidestream fans to determine the Bull Lake level of the Bellevue fan (fig. 38).

During the Bull Lake stage, periglacial deposits accumulated in sidestream valleys and on the margins of the Bull Lake Bellevue fan.

Pinedale outwash

Outwash caused aggradation in the Big Wood River valley downstream from extensive Pinedale glaciers in the headwaters of the Big Wood River. On the Bellevue fan, the river shifted back and forth across the fan, spilling to the west and to the east, as it built up the entire Wood River Valley floor to a slope of 32 feet per mile (fig. 31). In the east outlet valley the base-level control imposed by the Priest lava dam caused the slope to decrease to 9 feet per mile; gravel was still not transported across the Priest flow. The west outlet valley clearly had the advantage of a steeper over-all slope determined by base-level altitude and distance from the apex of the Bellevue fan.

The Pinedale profile cannot be extended downvalley from Stanton Crossing because terraces are not preserved between Stanton Crossing and Hot Springs Landing. Although terraces are present beneath Magic

Reservoir, they have not been identified.

Periglacial deposits accumulated during the Pinedale stage just as in the Bull Lake stage. As discussed earlier, a "fossil tundra landscape" still persists today in sidestream valleys in the Bellevue area.

Modern Big Wood River

After Pinedale aggradation the Big Wood River cut down about 30 feet below the surface of the Bellevue fan; numerous cut-in-fill terraces at different levels from place to place are evidence of lateral shifting of the river during the period of downcutting. The present valley floor is about one mile wide and has a slope of about 30 feet per mile upstream from Stanton Crossing.

Downstream from Stanton Crossing the river flows through a canyon cut in the Wind Ridge basalt, where the valley narrows to about 400 feet and is about 110 feet deep, and where the river has a gradient of 21 feet per mile. Farther downstream, where it crosses the Myrtle flow, the river has a canyon about 600 feet wide and 160 feet deep.

On the east side of the Wood River Valley, spring-fed Silver Creek flows on a consequent course down the slope of the Pinedale outwash fan; it has not entrenched its valley floor more than a few feet.

REFERENCES CITED

Anderson, A. L., 1942, Endomorphism of the Idaho batholith: Geol. Soc. America Bull., v. 53, p. 1099-1126.

_______ 1956, Geology and mineral resources of the Salmon quadrangle, Lemhi County, Idaho: Idaho Bur. Mines and Geology Pamp. 106, 102 p.

Anderson, A. L., Kiilsgaard, T. H., and Fryklund, V. C., Jr., 1950, Detailed geology of certain areas in the Mineral Hill and Warm Springs Mining Districts, Blaine County, Idaho: Idaho Bur. Mines and Geology Pamp. 90, 73 p.

Anderson, A. L., and Wagner, W. R., 1946, A geological reconnaissance of the Hailey gold belt (Camas district), Blaine County, Idaho: Idaho Bur. Mines and Geology Pamph. 76, 26 p.

Blackwelder, Eliot, 1915, Post-Cretaceous history of the mountains of central western Wyoming, part III: Jour. Geol., v. 23, p. 307-340.

Bostwick, D. A., 1955, Stratigraphy of the Wood River formation, south-central, Idaho: Jour. Paleontology, v. 29, p. 941-951.

Bowen, N. L., 1954, Experiments as an aid to the understanding of the natural world: Phila. Acad. Nat. Sci. Proc., v. 105, p. 1-12.

Boyd, F. R., 1961, Welded tuffs and flows in the rhyolite plateau of Yellowstone Park, Wyoming: Geol. Soc. America Bull., v. 72, p. 387-426.

Bryan, Kirk, 1949, The geologic implications of cryopedology: Jour. Geology, v. 57, p. 101-4.

Budel, Julius, 1953, The periglacial-morphologic effects of the Pleistocene climate over the entire world: translated by H. E. Wright and David Alt, Internat. Geology Review, v. 1, p. 1-16, 1959.

Capps, S. R., 1940, Gold placers of the Secesh Basin, Idaho County, Idaho: Idaho Bur. Mines and Geology Pamph. 52, 43 p.

Chayes, Felix, 1956, Petrographic modal analysis: an elementary statistical appraisal: New York, John Wiley and Sons, 113 p.

Coats, R. R., 1957, Jarbidge quadrangle, Nevada, in Geologic investigations of radioactive deposits--Semiannual progress report, Dec. 1, 1956 to May 31, 1957: U. S. Geol. Survey TEI-690, p. 303-312, issued by the U. S. Atomic Energy Comm. Tech. Inf. Service, Oak Ridge, Tenn.

Denny, C. S., 1951, Pleistocene frost action near border of the Wisconsin drift in Pennsylvania: Ohio Jour. Sci., v. 51, p. 115-125.

Fenner, C. N., 1944, Rhyolite-basalt complex on Gardiner River, Yellowstone Park, Wyoming: a discussion: Geol. Soc. of America Bull. v. 55, p. 1081-1096.

Flint, R. F., and Denny, C. S., 1958, Quaternary geology of Boulder Mountain Aquarius Plateau, Utah: U. S. Geol. Survey Bull. 1061-D, p. 103-164.

Fryxell, F. M., 1930, Glacial features of Jackson Hole, Wyoming: Augustana Library Publications no. 13.

Fuller, R. E., 1931, The geomorphology and volcanic sequence of Steens Mountain in southeastern Oregon: Washington Univ. Pub. in Geology, v. 3, no. 1, 130 p.

Gilluly, James, 1933, Replacement origin of the albite granite near Sparta, Oregon: U. S. Geol. Survey Prof. Paper 175-C, p. 65-81.

Harrington, E. R., 1948, Black Butte, a recent subsidence crater, Idaho: Sci. Monthly, v. 66, p. 451-466.

Holmes, G. W. and Moss, J. H., 1955, Pleistocene geology of the southwestern Wind River Mountains, Wyoming: Geol. Soc. America Bull., v. 66, p. 629-654.

Horberg, Leland, 1940, Geomorphic problems and glacial geology of the Yellowstone Valley, Park County, Montana: Jour. Geology, v. 48, p. 275-303.

Johannsen, Albert, 1939, A descriptive petrography of the igneous rocks: v. 1, 2d ed., Chicago, The University of Chicago Press, 318 p.

Kennedy, G. C., 1955, Some aspects of the role of water in rock melts, in Poldervaart, Arie, Editor, Crust of the earth: Geol. Soc. America Special Paper 62, p. 489-503.

King, Clarence, 1878, Systematic geology: U. S. Geol. Explor. 40th Parallel (King), v. 1, 803 p.

Kirkham, V. R. D., 1931, Igneous geology of southwestern Idaho: Jour. Geology, v. 39, p. 564-591.

Larsen, E. S., 1938, Some new variation diagrams for groups of igneous rocks: Jour. Geology, v. 46, p. 516-520.

Larsen, E. S., Jr., Gottfried, David, Jaffe, H. W., and Waring, C. L., 1958, Lead-alpha ages of the Mesozoic batholiths of western North America: U. S. Geol. Survey Bull. 1070-B, p. 35-62.

Lindgren, Waldemar, 1900, The gold and silver veins of Silver City, De Lamar, and other mining districts in Idaho: U. S. Geol. Survey 20th Ann. Rept., pt. 3, p. 65-256.

Mackin, J. Hoover, 1961, A stratigraphic section in the Yakima basalt and the Ellensburg formation in south-central Washington: Washington Division of Mines and Geology Report of Investigation 19.

Mackin, J. H., and Schmidt, D. L., 1956, Uranium- and thorium-bearing minerals in placer deposits in Idaho in Page, L. R., and others, Contributions to the geology of uranium and thorium by the United States Geological Survey and Atomic Energy Comm. for the United Nations Internatl. Conf. on Peaceful Uses of Atomic Energy, Geneva, 1955, Proc. Internatl. Conf. on Peaceful Uses of Atomic Energy, United Nations Pub. v. 6, p. 587-592: U. S. Geol. Survey Prof. Paper 300, p. 375-380: Idaho Bur. Mines and Geology, Mineral Resources Rept. no. 7, 9 p.

Malde, H. E., 1959, Fault zone along northern boundary of western Snake River Plain, Idaho: Science, v. 130, no. 3370, p. 272.

Mansfield, G. R., 1952, Geography, geology, and mineral resources of the Ammon and Paradise Valley quadrangles, Idaho: U. S. Geol. Survey Prof. Paper 238, 92 p.

Mansfield, G. R., Ross, C. S., 1935, Welded rhyolitic tuffs in southeastern Idaho: Am. Geophys. Union Trans., 16th Ann. Mtg., pt. 1, p. 308-321, Nat. Research Council, Aug. 1935.

Morey, G. W., and Hesselgesser, J. M., 1952, The system H_2O-Na_2O-SiO_2 at 400°C; Am. Jour. Sci., Bowen Volume, p. 343-371.

Peltier, Louis C., 1950, The geographic cycle in periglacial regions as related to climatic geomorphology: Annals Assoc. of American Geogs., v. 40, p. 209.

Richmond, G. M., 1960, Glaciation of the east slope of Rocky Mountain National Park, Colorado: Geol. Soc. of America Bull., v. 71, p. 1371-1382.

Robertson, A. F., and Storch, R. H., 1955a, Camp Creek radioactive Mineral placer area, Blaine and Camas counties, Idaho: U. S. Bureau of Mines RME-3136, 27 p., issued by U. S. Atomic Energy Comm., Tech. Inf. Service, Oak Ridge, Tenn.

________ 1955b, Rock Creek radioactive mineral placer area, Blaine County, Idaho: U. S. Bureau of Mines RME-3139, 25 p., issued by U. S. Atomic Energy Comm., Tech. Inf. Service, Oak Ridge, Tenn.

Robertson, Forbes, 1952, Graphic method of showing differentiation in igneous rocks (abs.): Montana Acad. Sci. Proc., v. 11, 37-39.

Ross, C. P., 1937, Geology and ore deposits of the Bayhorse region Custer County, Idaho: U. S. Geol. Survey Bull. 877, 157 p.

Ross, C. S., and Smith, R. L., 1961, Ash-flow tuffs: their origin geologic relations and identification: U. S. Geol. Survey Prof. Paper 366, 81 p.

Russell, I. C., 1902, Geology and water resources of the Snake River plains of Idaho: U. S. Geol. Survey Bull., v. 199, 192 p.

Skipp, B. A. L., 1961, Interpretation of sedimentary features in Brazer limestone (Mississippian) near Mackay, Custer County, Idaho: Am. Assoc. Petroleum Geologist Bull., v. 45, p. 376-389.

Smith, H. T. U., 1949, Physical effects of Pleistocene climatic changes in nonglaciated areas: Geol. Soc. America Bull., v. 60, p. 1485-1516.

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

_______ 1960b, Zones and zonal variations in welded ash flows: U. S. Geol. Survey Prof. Paper 354-F, p. 149-159.

Smith, R. O., 1959, Ground-water resources of the middle Big Wood River-Silver Creek area, Blaine County, Idaho: U. S. Geol. Survey Water-Supply Paper 1478, 64 p.

_______ 1960, Geohydrologic evaluation of streamflow records in the Big Wood River Basin, Idaho: U. S. Geol. Survey Water-Supply Paper 1479, 68 p.

Stearns, H. T., 1955, Pillar Falls mud flow and Shoshone Falls andesite near Twin Falls, Idaho: Geol. Soc. America Bull., v. 66, p. 463-464.

Stearns, H. T., Crandall, Lynn, and Steward, W. G., 1938, Geology and ground-water resources of the Snake River plain in southeastern Idaho: U. S. Geol. Survey Water-Supply Paper 774, 268 p.

Stearns, H. T., and Isotoff, A., 1956, Stratigraphic sequence in the Eagle Rock volcanic area near American Falls, Idaho: Geol. Soc. America Bull., v. 67, p. 19-34.

Thomasson, M. R., 1959, Paleocurrent and sedimentary structure studies in upper Paleozoic rocks of central Idaho (abs.): Geol. Soc. America Bull., v. 70, p. 1687.

Umpleby, J. B., Westgate, L. G., and Ross, C. P., 1930, Geology and ore deposits of the Wood River region, Idaho: U. S. Geol. Survey Bull. 814, 250 p.

Walton, W. C., 1960, Ground-water resources of Camas Prairie, Camas and Elmore counties, Idaho: U. S. Geol. Survey open-file rept., July 1960, 125 p.

Wilcox, R. E., 1944, Rhyolite-basalt complex on Gardiner River, Yellowstone Park, Wyoming: Geol. Soc. America Bull., v. 55, p. 1047-1079.

Williams, Howel, 1932, The history and character of volcanic domes: California Univ. Dept. Geol. Sci. Bull., v. 21, p. 51-146.

Williams, P. L., 1961(?), Glacial geology of Stanley Basin, Idaho: Idaho Bur. Mines and Geology Pamph., in press.

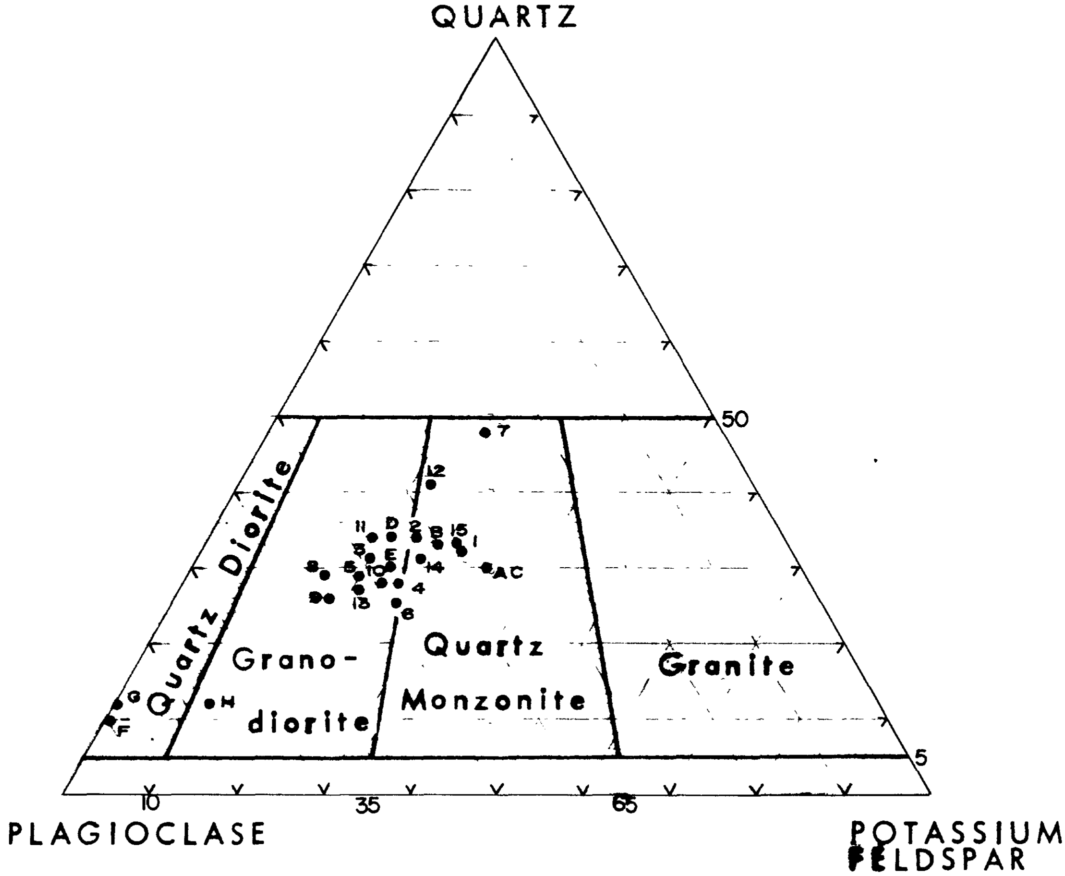


FIGURE 2.--Classification of granitic rocks of the Bellevue area according to Johannsen (1939), but modified so that the granodiorite and quartz diorite fields are separated by the 10 percent potassium feldspar line rather than by the 5 percent potassium feldspar line. The wide scatter of individual modes of "typical specimens" of the Hailey granodiorite indicates the variable complex internal composition of the granitic body and is one factor supporting the hypothesis of endomorphism. Points F, G, and H represent the Croesus quartz diorite. Letters and numbers refer to modes listed in tables 1 and 2.

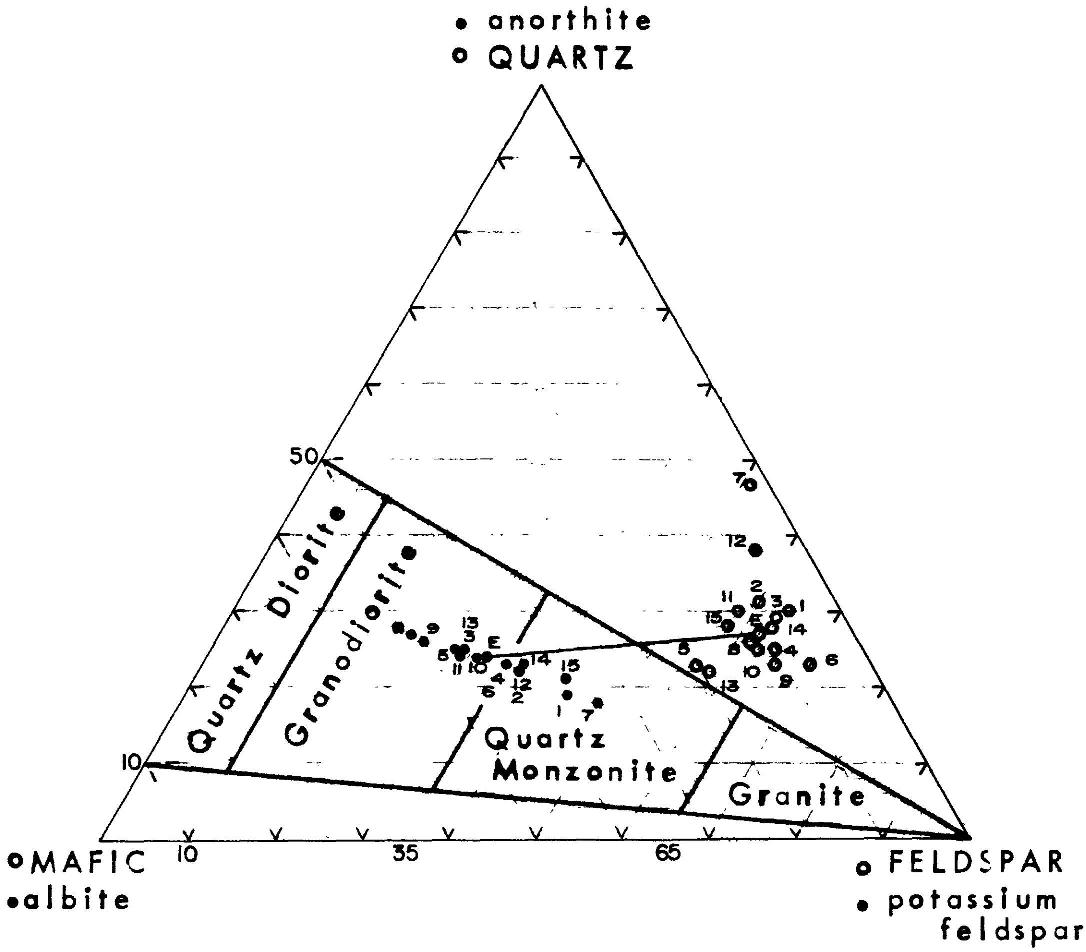


FIGURE 3.--Modified Larsen ternary diagram (Robertson, 1952; Larsen, 1938) with plots of granitic rock modes as published by various authors from different localities in the Bellevue area (table 1). These are compared to the mode of the average Hailey granodiorite (E). Subdivisions denote "order 2" granitic rocks of the modified Johannsen classification. Solid circles refer to the albite-anorthite-potassium feldspar part of the diagram, and the open circles refer to the mafic-quartz-feldspar part.

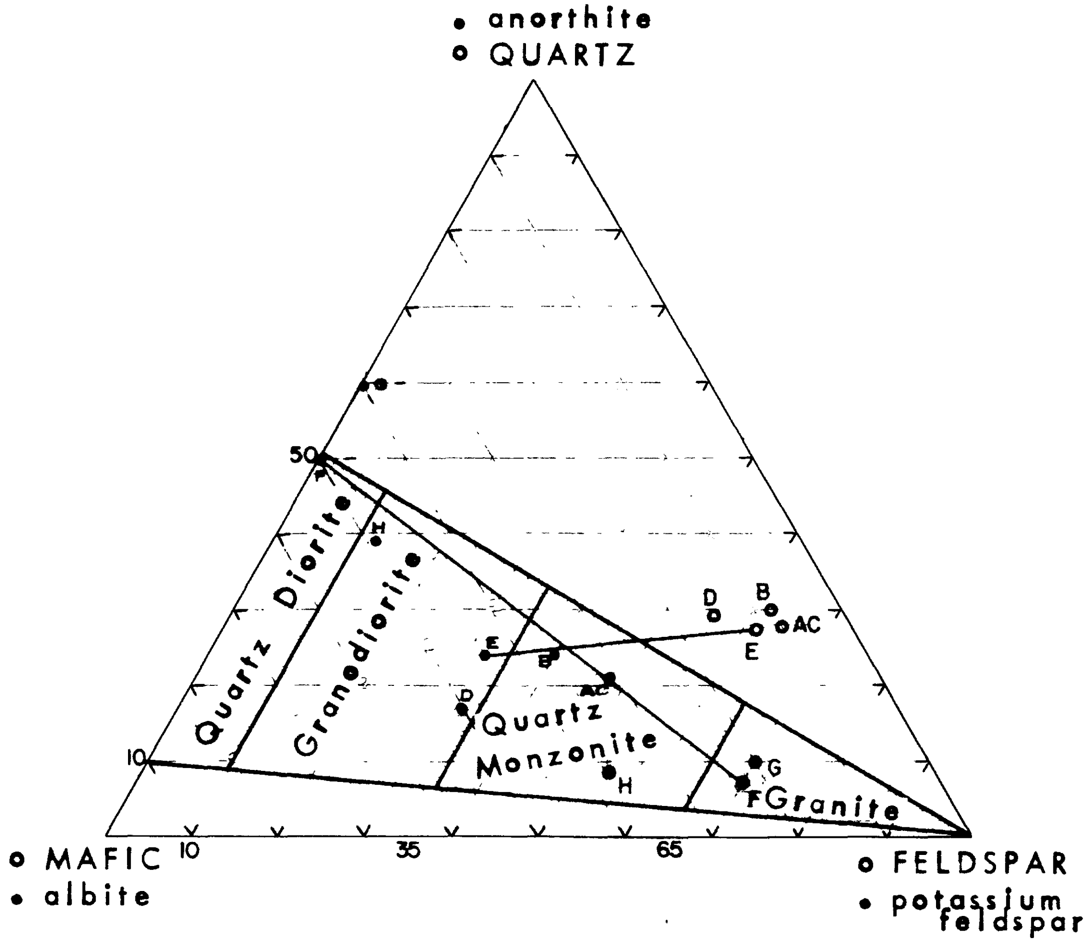


FIGURE 4.—Modified Larsen ternary diagram (see for reference fig. 3) with plots of individual modes of typical specimens of the Hailey granodiorite (table 2) collected along a northwest-southeast section across the Bellevue area. As in figure 2, a wide range of composition is indicated.

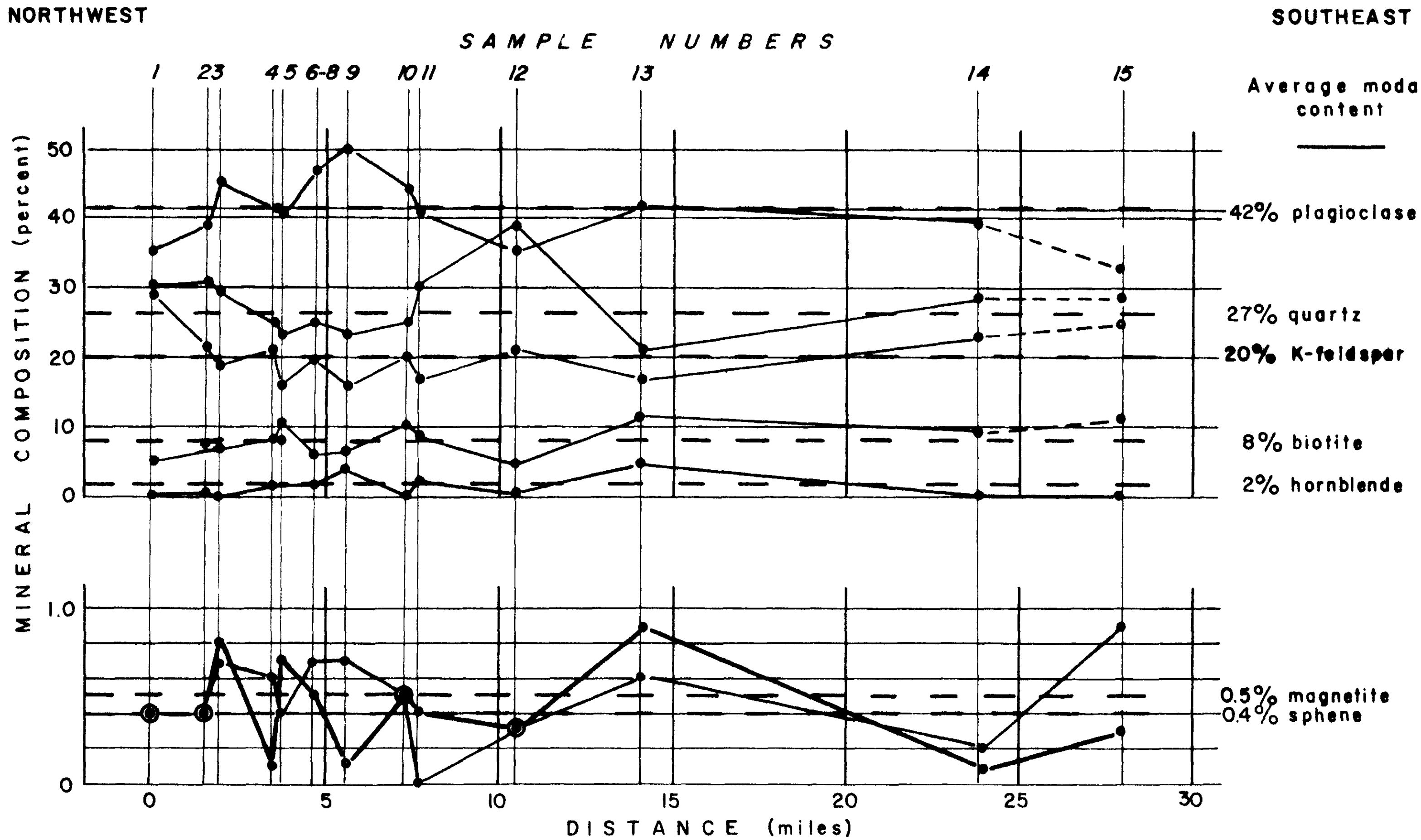


FIGURE 5.--Mineral variation curves of the Hailey granodiorite plotted against distance along a northwest-southeast section across the Bellevue area. The average modal mineral compositions are shown by the heavy dashed lines. The modal data are listed in table 2. The seemingly systematic variations may reflect original differences in the composition of the magma or possible secondary changes caused by endomorphic metasomatism.

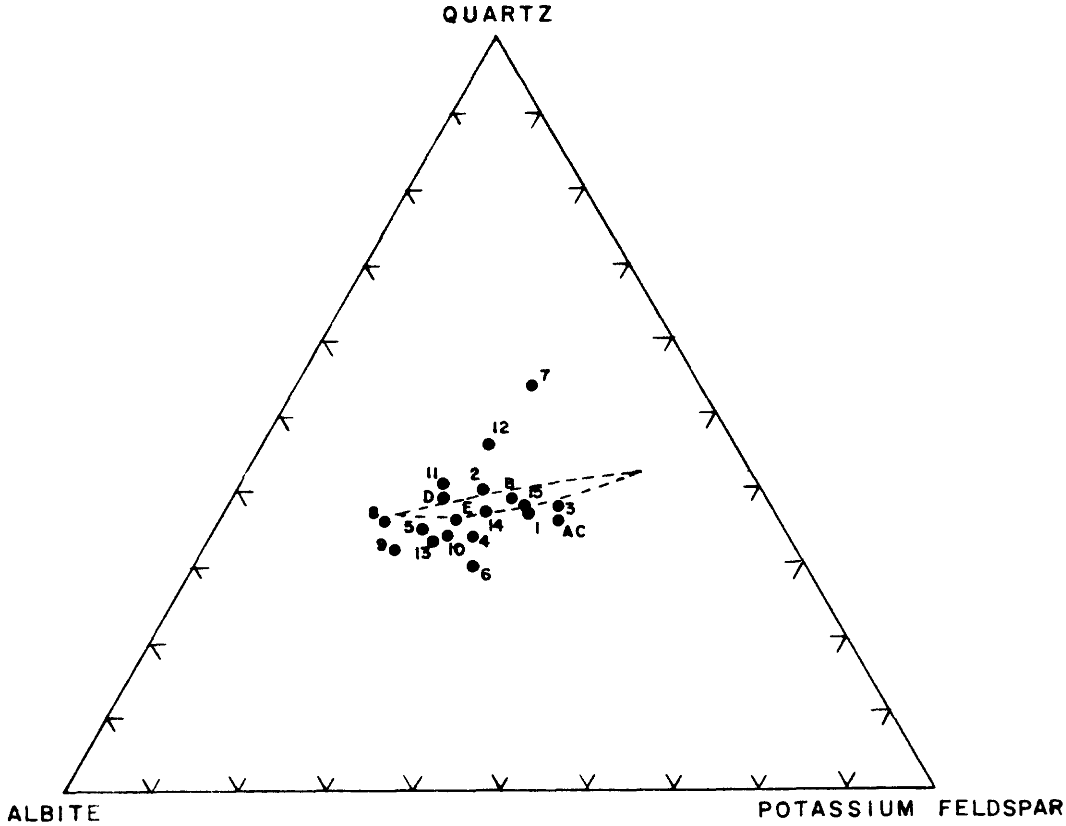


FIGURE 6.—Ternary diagram of the system, albite-potassium feldspar-quartz, with plots of the recalculated modes of the Hailey granodiorite (tables 1 and 2). The wide scatter of points about the minimum melting trough (740° and 1000 bars, Bowen, 1954, p. 11) shows the influence of post-magmatic endomorphic recrystallization which disrupted textures produced by an earlier crystal-magmatic liquid equilibrium.

TABLE 1.—Modal compositions of granodiorite and quartz diorite from the Bellevue area as published by various authors. Average values have been estimated for use in plotting in the ternary diagrams. Rock names are assigned according to the modified Johannsen classification, figure 3. "X" in table indicates presence of mineral.

	HAILEY GRANODIORITE					CROESUS QUARTZ DIORITE		
Locality	A Anderson (1950, p. 4-5) (Diebenow Creek) Quartz monzonite modal	B Anderson (1946, p. 5) (Hailey Gold Belt) Quartz monzonite modal	C Hewett (1930, p. 212) (Diebenow Creek) Quartz monzonite modal	D Lindgren (1900, p. 82, 221) (Idaho-Democrat Mine) Granodiorite normative	E Schmidt (1961) (Princess Blue Ribbon Mine to Picabo Hills (table 2)) Granodiorite modal	F Anderson (1950, p. 4) (West of Bellevue) Quartz diorite modal	G Hewett (1930, p. 212) (West of Bellevue) Quartz diorite ("Quartz calciclase gabbro") modal	H Lindgren (1900, p. 82, 195, 222) (Croesus Mine) Granodiorite normative
Published compositions (percent)								
Quartz	20-35	30	20-35	29.3	27	5-10	10	8.4
Potassium feldspar	32±	20-30	30-45	18.1	20	—	—	7.6
Plagioclase	32± Andesine	30-45 Andesine	25-60 Andesine	36.7 An26	42 An35±	70± And.-Lab.	70± Lab.	46.6 An44
Biotite	5-10	5-10	5-8	9.7	8.0	10	10	}24.1
Hornblende	±	±	0-5	—	2.0	}15±	3	
Pyroxene	—	—	—	—	—		5	
Sphene	Several %	x	2-3	0.9	0.4	x	—	1.3
Magnetite	x	x	—	0.3	0.5	x	x	1.5
Allanite	x	—	—	—	0.04	x	—	—
	Zircon Apatite Ilmenite Epidote Chlorite Sericite	Zircon Apatite Epidote Chlorite Leucoxene	Muscovite	Apatite Chlorite Epidote Calcite Pyrite	Zircon Apatite Epidote Chlorite Sericite Leucoxene	Zircon Apatite Chlorite Epidote Sericite	Apatite	Apatite Chlorite Calcite Pyrite

Estimated averages

	A		B		C		D		E		F		G		H	
Feldspar		64		62		64		56		62		70		70		54
Quartz		28		30		28		29		27		8		10		8
Mafics		8		8		8		15		11		22		20		38
Total		100		100		100		100		100		100		100		100
Albite	20	31	22	36	20	31	28	50	27	44	35	50	28	40	30	49
Anorthite	13	21	15	24	13	21	10	17	15	24	35	50	42	60	24	39
Potassium feldspar	31	48	25	40	31	48	18	33	20	32	0	0	0	0	8	12
Total	64	100	62	100	64	100	56	100	62	100	70	100	70	100	62	100
Quartz	28	36	30	39	28	36	29	39	27	36	8	18	10	26	8	17
Potassium feldspar	31	39	25	32	31	39	18	24	20	27	0	0	0	0	8	17
Albite	20	25	22	29	20	25	28	37	27	37	35	82	28	74	30	66
Total	79	100	77	100	79	100	75	100	74	100	43	100	38	100	46	100
Quartz	28	30	30	33	28	30	29	34	27	30	8	10	10	12	8	11
Potassium feldspar	31	34	25	27	31	34	18	21	20	23	0	0	0	0	8	11
Plagioclase	33	36	37	40	33	36	38	45	42	47	70	90	70	88	54	78
Total	92	100	92	100	92	100	85	100	89	100	78	100	80	100	70	100

TABLE 2.—Modal compositions of 13 typical granitic rocks collected from sample localities along a 25 mile, northwest-southeast section across the main body of Hailey granodiorite (plate 1). Sample 7 is an atypical rock, quartz-rich wall rock, from a vein structure. Sample 15 is a typical specimen from the Idaho-Democrat Mine. One thousand points were counted in an area of 1200 square centimeters on each stained rock slice (Chayes, 1956).

Locality	Princess Blue Ribbon Mine	Champlain Mine	Happy Day Mine	Near Richardson Summit	Near Richardson Summit	Near Treasure Vault Mine	Near Treasure Vault Mine (atypical) (wall rock of quartz vein)	Near Treasure Vault Mine	Hattie Gulch (middle)	Hattie Gulch (mouth)	Hattie Gulch (near mouth)	Head of Little Rock Creek	Uhrig Mine	South-central Picabo Hills	Idaho-Democrat Mine	Hailey granodiorite (average: 1-14 less 7)
Mineral composition (volume percent)	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	x̄
Quartz	30	31	29	25	23	23	47	26	23	25	30	38	22	28	28	27
Potassium feldspar	28	22	18	22	16	24	24	14	16	20	17	21	17	23	25	20
Plagioclase	35	38	45	42	41	45	26	49	50	44	41	35	42	39	33	42
Biotite	5.2	7.5	7.0	8.0	11	5.3	1.5	7.2	6.5	10	8.6	4.5	12	9.1	12	8.0
Hornblende	0.1	0.7	—	1.7	8.1	1.5	—	2.3	3.7	0.4	2.4	0.7	4.8	—	0.4	2.0
Sphene	0.4	0.4	0.8	0.1	0.7	0.1	0.2	0.9	0.1	0.5	0.4	0.3	0.9	0.1	0.3	0.4
Magnetite	0.4	0.4	0.7	0.6	0.4	0.1	0.4	1.3	0.7	0.5	—	0.3	0.6	0.2	0.9	0.5
Allanite	—	—	—	—	—	—	—	—	0.1	—	—	—	0.3	—	0.1	—
Epidote	—	—	—	—	—	—	—	0.1	—	0.5	0.5	—	—	—	—	0.1
Total	99.1	100.0	100.5	99.4	100.2	99.0	99.1	100.8	100.1	100.9	99.9	99.8	99.6	99.4	99.7	100.0
Feldspar	63 64	60 60	63 63	64 65	57 57	69 70	50 51	63 62	66 66	64 63	58 58	56 56	59 59	62 63	58 58	62 62
Quartz	30 30	31 31	29 29	25 25	23 23	23 23	47 47	26 26	23 23	25 25	30 30	38 38	22 22	28 28	28 28	27 27
Mafics	6 6	9 9	8 8	10 10	20 20	7 7	2 2	12 12	11 11	12 12	12 12	6 6	19 19	9 9	14 14	11 11
Total	99 100	100 100	100 100	99 100	100 100	99 100	99 100	101 100	100 100	101 100	100 100	100 100	100 100	99 99	100 100	100 100
Albite	23 37	25 41	29 46	27 42	27 47	29 42	17 34	32 51	33 50	29 45	27 47	23 41	27 46	25 40	21 36	27 44
Anorthite	12 19	13 22	16 25	15 23	14 25	16 23	9 18	17 27	17 26	15 24	14 24	12 22	15 25	14 23	12 21	15 24
Potassium feldspar	28 44	22 37	18 29	22 35	16 28	24 35	24 48	14 22	16 24	20 31	17 29	21 37	17 29	23 37	25 43	20 32
Total	63 100	60 100	63 100	64 100	57 100	69 100	50 100	63 100	66 100	64 100	58 100	56 100	59 100	62 100	58 100	62 100
Quartz	30 37	31 40	29 38	25 34	23 35	23 30	47 54	26 36	23 32	25 34	30 41	38 46	22 33	28 37	28 38	27 36
Potassium feldspar	28 35	22 28	18 24	22 30	16 24	24 32	24 27	14 19	16 22	20 27	17 23	21 26	17 26	23 30	25 34	20 27
Albite	23 28	25 32	29 38	27 36	27 41	29 38	17 19	32 45	33 46	29 39	27 36	23 28	27 41	25 33	21 28	27 37
Total	81 100	78 100	76 100	74 100	66 100	76 100	88 100	72 100	72 100	74 100	74 100	82 100	66 100	76 100	74 100	74 100
Quartz	30 32	31 34	29 31	25 28	23 29	23 25	47 48	26 29	23 26	25 28	30 34	38 41	22 27	28 31	28 33	27 30
Potassium feldspar	28 30	22 24	18 20	22 25	16 20	24 26	24 25	14 16	16 18	20 23	17 19	21 22	17 21	23 26	25 29	20 23
Plagioclase	35 38	38 42	45 49	42 47	41 51	45 49	26 27	49 55	50 56	44 49	41 47	35 37	42 52	39 43	33 38	42 47
Total	93 100	91 100	92 100	89 100	80 100	92 100	97 100	89 100	89 100	89 100	88 100	94 100	81 100	90 100	86 100	89 100

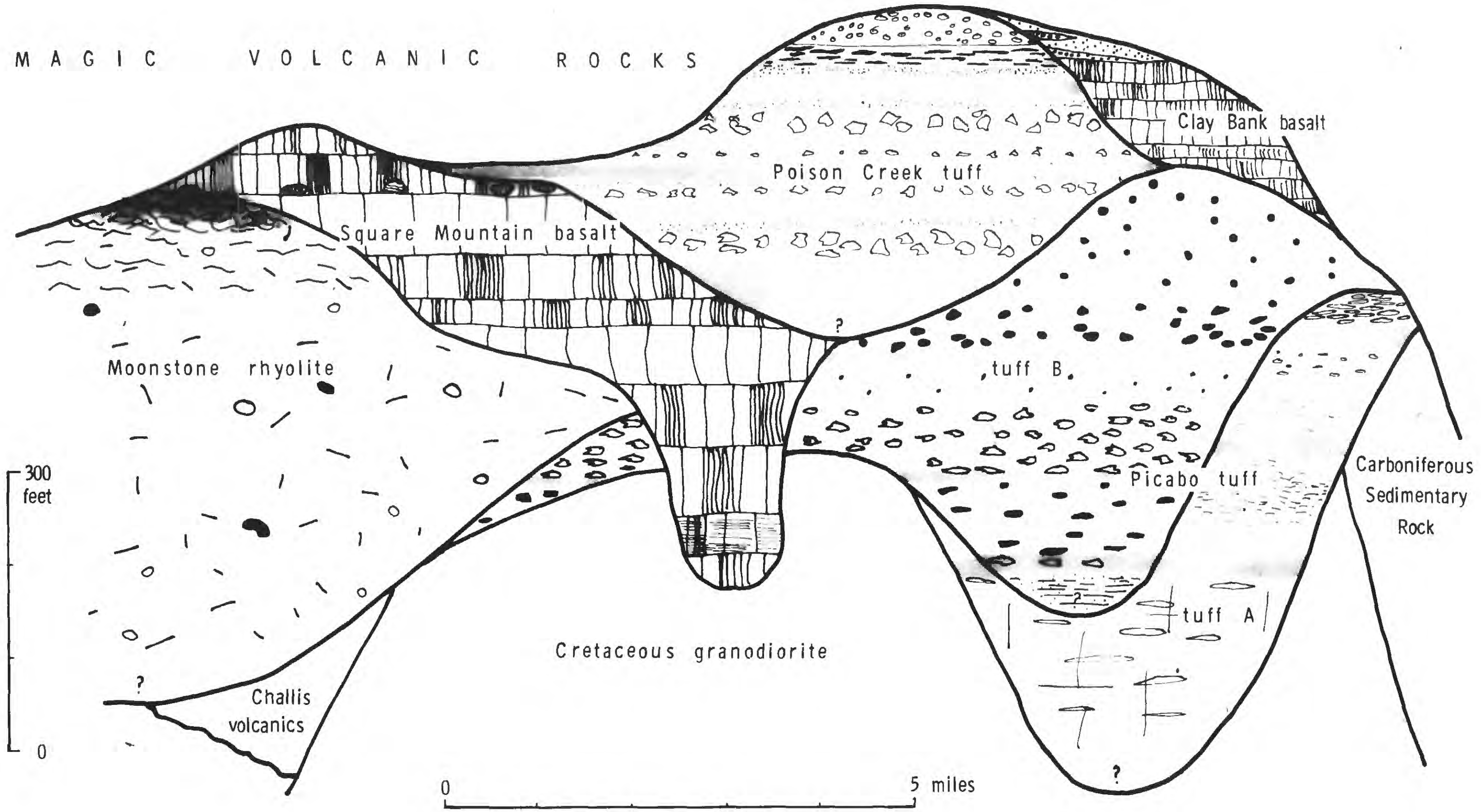


FIGURE 7.--Magic volcanic rocks in a diagrammatic east-west section across the southern part of the Bellevue area: vertical exaggeration is extreme. Each unit rests unconformably on older units indicating that time lapses between the several eruptive series were long enough for the erosion of broad valleys.

FIGURE 8.--Picabo tuff A. Characteristic lithophysal cavity zone (six feet thick) in the basal vitric part of a typical ash flow in tuff A. Below the cavernous rock is the remainder of the vitrophyric zone (one foot thick; above hammer) which is underlain by an indurated ash layer (one to two feet thick; below hammer) making up the base of the ash flow. The upper zones of the flow have been removed by erosion. The outcrop is in the central Picabo Hills.

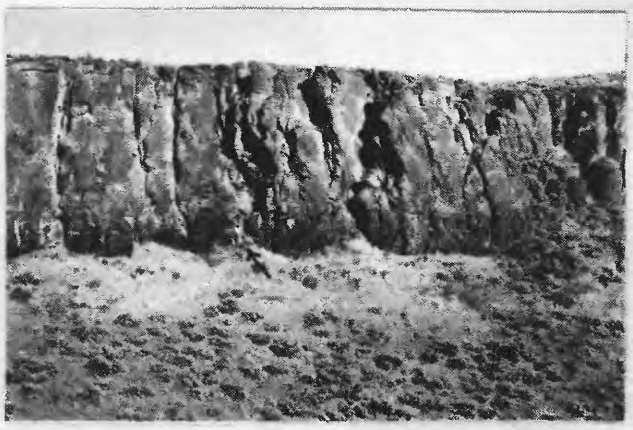


FIGURE 9.--Picabo tuff B. Part of one ash flow showing vertical zonation; 12 feet of basal vitrophyre overlain by 48 feet of less densely welded rock; the top is eroded. The outcrop is 2-1/2 miles east-southeast of Priest Station. Note size of man (arrow) for scale.

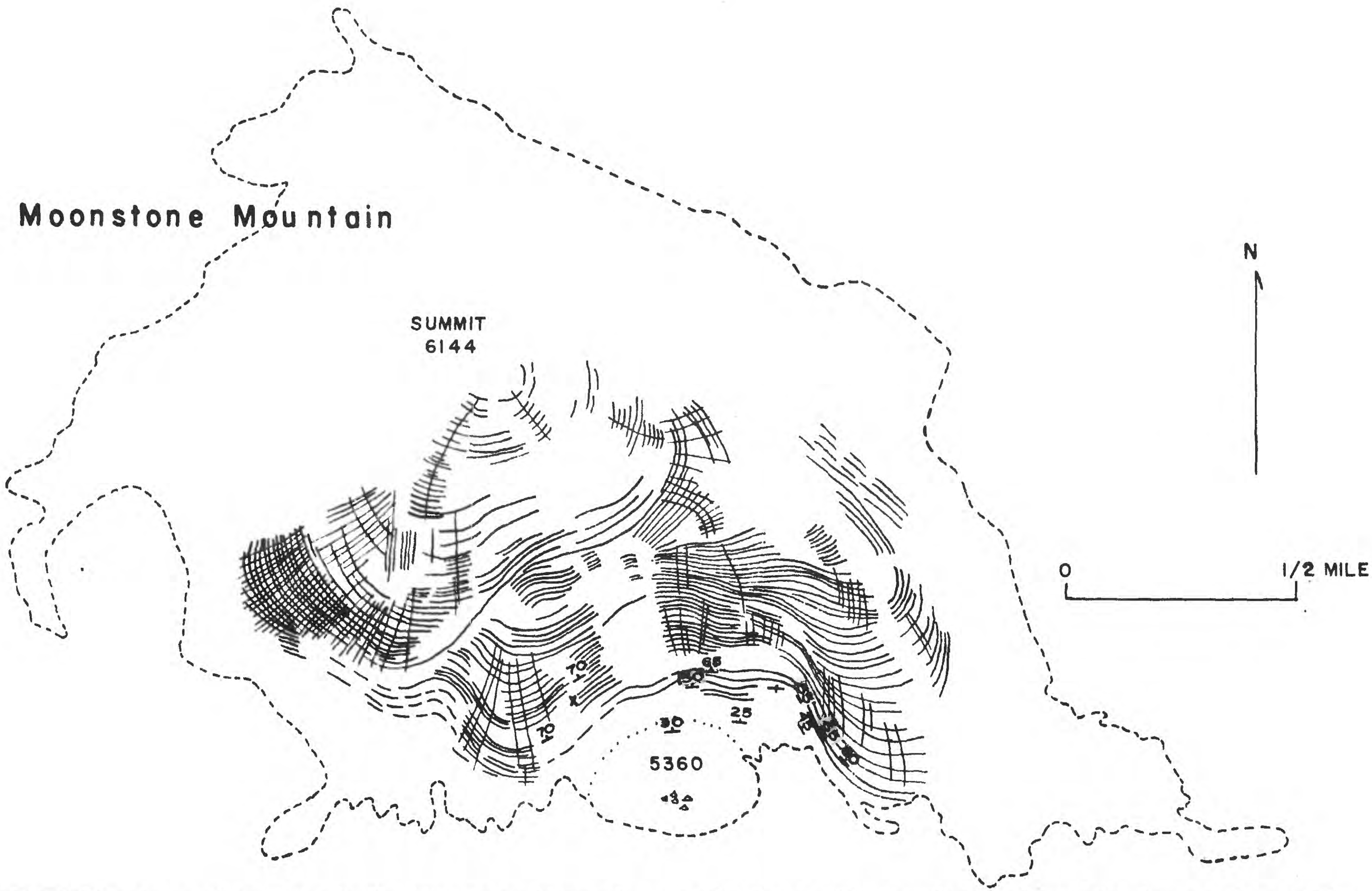


FIGURE 10.--Internal structure in the Moonstone Mountain extrusive dome. Concentric flow layers dipping steeply at high altitudes and shallowly at low altitudes (the change of dip at "x" is shown in fig. 11) are cut by a prominent set of radial joints. One lobate structure extends out beyond the dome to the southeast and represents a short tongue of viscous lava. Hill 5360 however consists of a dark brown glassy breccia which may be an early basal phase or a late vent phase. Younger rocks lap up on the dome everywhere so that its base is not exposed.

FIGURE 11.--Flow layering on the south flank of Moonstone Mountain suggests an endogenous extrusive dome. Dips are toward a central vent to the left; they change from shallow at low altitudes to steep at high altitudes within a vertical distance of several hundred feet. The layering is marked by darker, denser, glassier layers alternating with lighter colored, more porous, highly devitrified layers.

FIGURE 12.--Flows of Square Mountain basalt exposed in about 400 feet of section in Camp Creek Canyon west of Square Mountain.

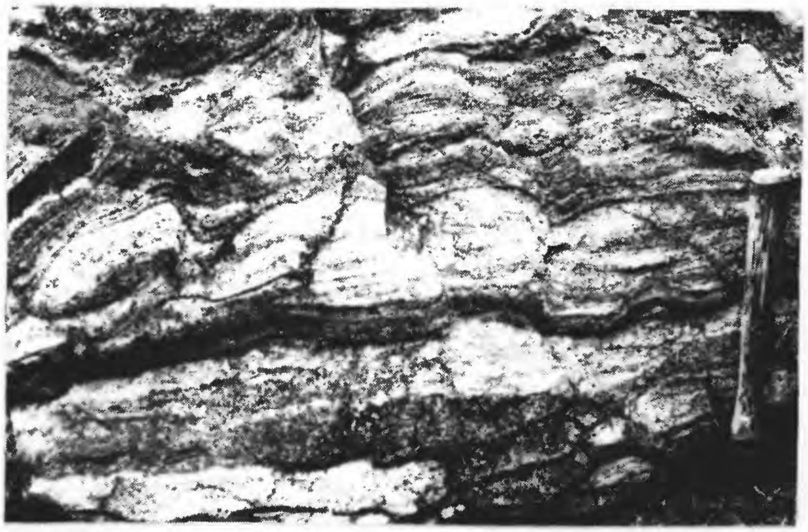


FIGURE 13.--Xenolith of granitic gneiss in the Square Mountain basalt. Brown glassy layers, representing former mafic layers containing relict pyroxene, alternate with quartz-microcline-plagioclase layers containing about 20% clear interstitial glass. The xenoliths represent a deep-seated gneiss probably of the granulite facies which has been brought to the surface by the basalt.

FIGURE 14.--A cluster of xenoliths of granitic gneiss in the Square Mountain basalt occurring on Square Mountain above Tolmie Creek.

FIGURE 15.--Dissected pediment on the East Fork of Rock Creek. The contact between resistant Carboniferous quartzites and limestones, high background and upper half of pediment, and granitic rock, foreground and lower half of pediment, is drawn on the photograph. The sedimentary bedrock is the source of a resistant capping gravel which has prolonged the life of the pediment. The depth of dissection at the foot of the pediment is about 80 feet.

FIGURE 16.--High terrace north of Poverty Flat consisting of valley-side alluvium deposited as the Big Wood River valley floor filled and aggraded behind the Wind Ridge lava dam. View is across the Big Wood River bypass canal and modern valley floor. Mountains in background, to left of center, are Carboniferous sedimentary rocks, and the higher mountains to right of center are granodioritic and quartz dioritic intrusive rock.

FIGURE 17.--Lacustrine sediments associated with the Wind Ridge basalt. Thinly laminated silt and clay deposited in a lake impounded by the lower Wind Ridge basalt and overlain by Big Wood River gravel of Bull Lake age which in turn is overlain by local periglacial alluvium of Bull Lake age. The shovel is 5 feet long. The outcrop is 0.7 mile southeast of Stanton Crossing.

FIGURE 18.--Lacustrine sediments associated with the Myrtle basalt. Layered fine sand (15 feet thick) is overlain by silty clay (2 feet) and silty talus (2 feet). Macon basalt (10 feet thick) overlies the talus but is exposed beyond the top of the scarp. The silty clay has intruded the silty talus; this structure is probably caused by differential load pressures at the time of the spreading of the Macon basalt. The outcrop is on the shore of Magic Reservoir, 1-1/2 miles north of Magic Resort.

FIGURE 19.--Terrace marginal to the Macon basalt. High terrace (Qsmm) to right of Magic Reservoir consists of Camas Creek gravel deposited on top of the eastern margin of the Macon flow (Qbm). The lower terrace (QBm) in the middle foreground is Bull Lake outwash deposited by the Big Wood River. View is south of Hot Springs Landing. Symbols are the same as on the geologic map (pl. 1).

FIGURE 20.--Periglacial sidestream terraces. Undissected low fan (QRs) is latest Pinedale and post-Pinedale, intermediate terrace (QPs) is Pinedale, and high terrace (QBs) is Bull Lake. The foreground surface (QPm) is Pinedale outwash of the Big Wood River one mile south of Hailey (fig. 23).

Page 111 follows

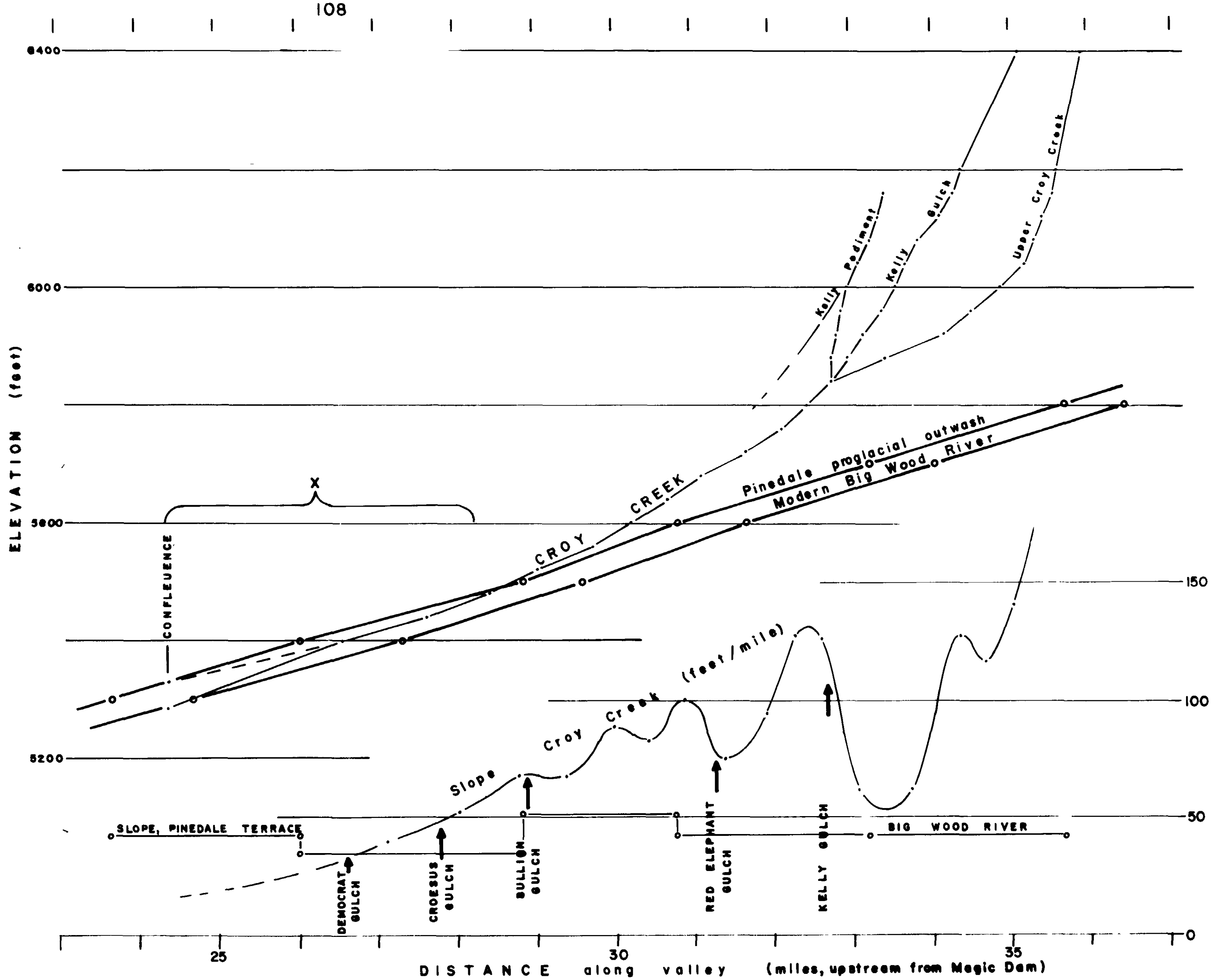


FIGURE 21.--Longitudinal profile of Croy Creek. The valley floor of Croy Creek is graded to the Pinedale proglacial terrace of the Big Wood River; only within 4 miles of the confluence (distance "X") has Croy Creek entrenched its Pinedale valley floor. The low gradient of the Pinedale Valley floor near the confluence (along distance "X") suggests that Croy Creek deposition was not able to keep up with the proglacial aggradation of the Big Wood River. The lower curve, slope plotted against distance, shows the dominating influence of small tributary deposits of periglacial origin on Croy Creek, a principal sidestream. The slope of Croy Creek is steeper at the mouth of each tributary demonstrating that Croy Creek has not yet mastered its valley floor in post-Pinedale time. The imperfect coincidence of maximum slope and tributary confluence is in part caused by widely spaced control.

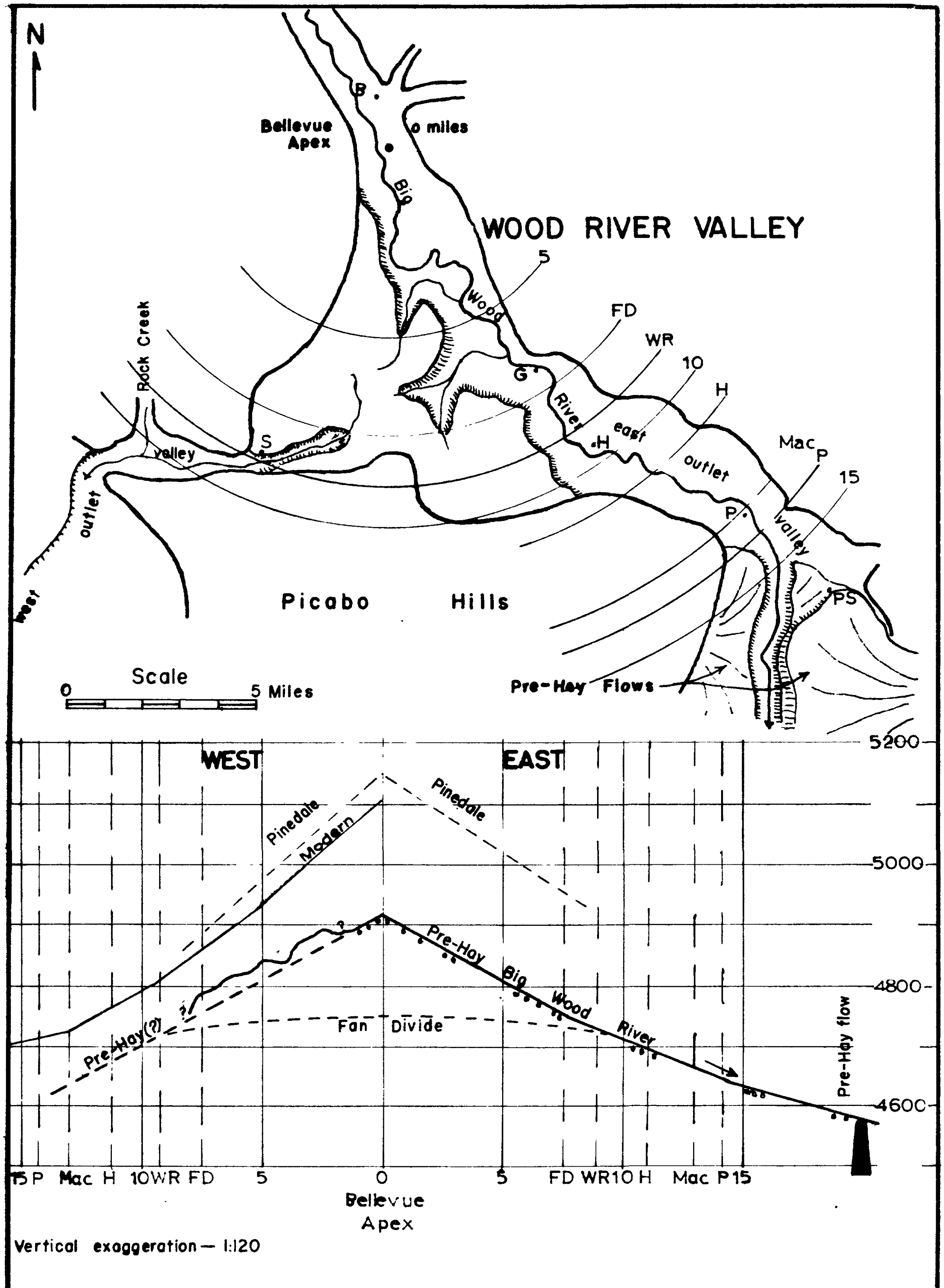
TABLE 3.—Sequence of events during the Late Quaternary. The arrows indicate whether the Big Wood River flowed out its east or west outlet valley at the time of each event.

west valley	← →	east valley
Figure 24	→	Pre-Hay drainage Big Wood River flowing east.
Figure 25	←	Hay flow (Dam 4840±) Big Wood River diverted directly to west.
Figure 26	→	Lower Wind Ridge flow (Dam 4875) Wind Ridge lake, elevation 4840; clay and silt fill to 4840.
	← →	Post-Wind Ridge lake, sand aggradation; Big Wood River spilling both east and west.
	←	Big Wood River entrenched in west valley.
Figure 27	→	Upper Wind Ridge flow (Dam 4930±) Big Wood River diverted directly to east.
Figure 28		Myrtle flow Camas Valley - Big Wood River flowing east. Sonnars flow Sonnars Flat - no mainstream affected.
Figure 29		Macon flow Camas Valley - Big Wood River flowing east.
Figure 30	→	Priest flow (Dam 4820) Priest Lake, elevation 4820±; clay and silt deposition.
	→	Post-Priest Lake, sand aggradation.
	← →	Bull Lake outwash Wood River Valley was aggraded with outwash; river spilled east and west.
	←	Big Wood River entrenched in west valley.
Figure 31	← →	Pinedale outwash Wood River Valley was aggraded with outwash; river spilled east and west.
	←	Big Wood River entrenched in west valley.
	←	Burmah basalt (Dam about 4700±)

FIGURE 24.--PRE-HAY DRAINAGE. The Big Wood River flowed in its east outlet valley, then occupied in part by Bellevue flows older than the Hay flow. It is not known whether the river had occupied its west outlet valley prior to the spreading of the Hay flow.

For reference, the present Wood River Valley is outlined and localities are indicated by letters: B, Bellevue; S, Stanton Crossing; G, Gannett; H, Hay Station; P, Picabo; PS, Priest Station. The arcs are reference distances measured from the apex of the Bellevue fan to: WR, the Wind Ridge lava dam; H, the Hay lava dam; Mac, the Macon lava dam of Rock Creek; P, the Priest lava dam; FD, fan divide.

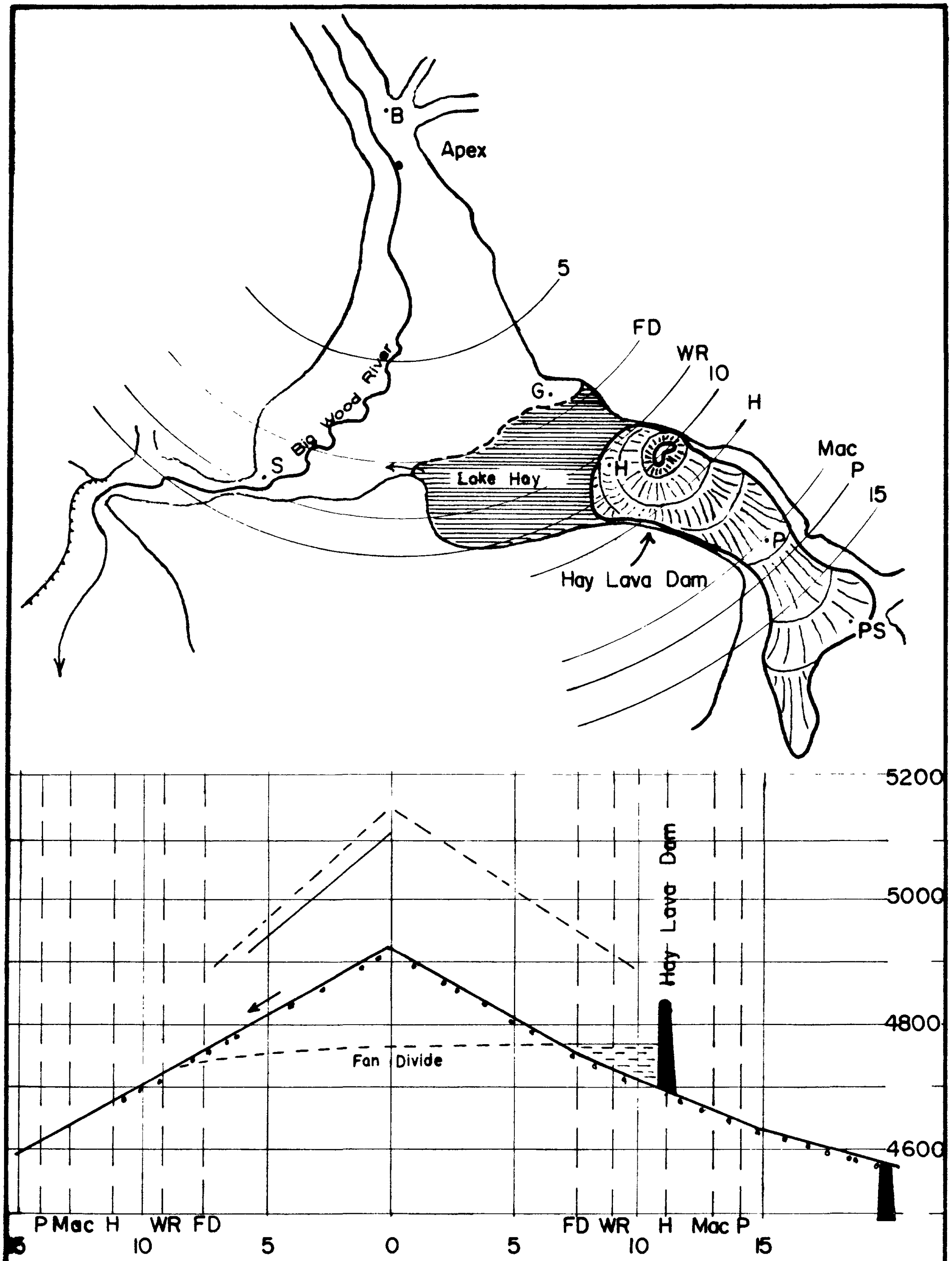
The vertical exaggeration in the generalized sections is about 1:120. The profiles are mostly drawn as several straight-line segments to emphasize those segments which are controlled by some known or estimated value of slope.



PRE-HAY DRAINAGE

FIGURE 24

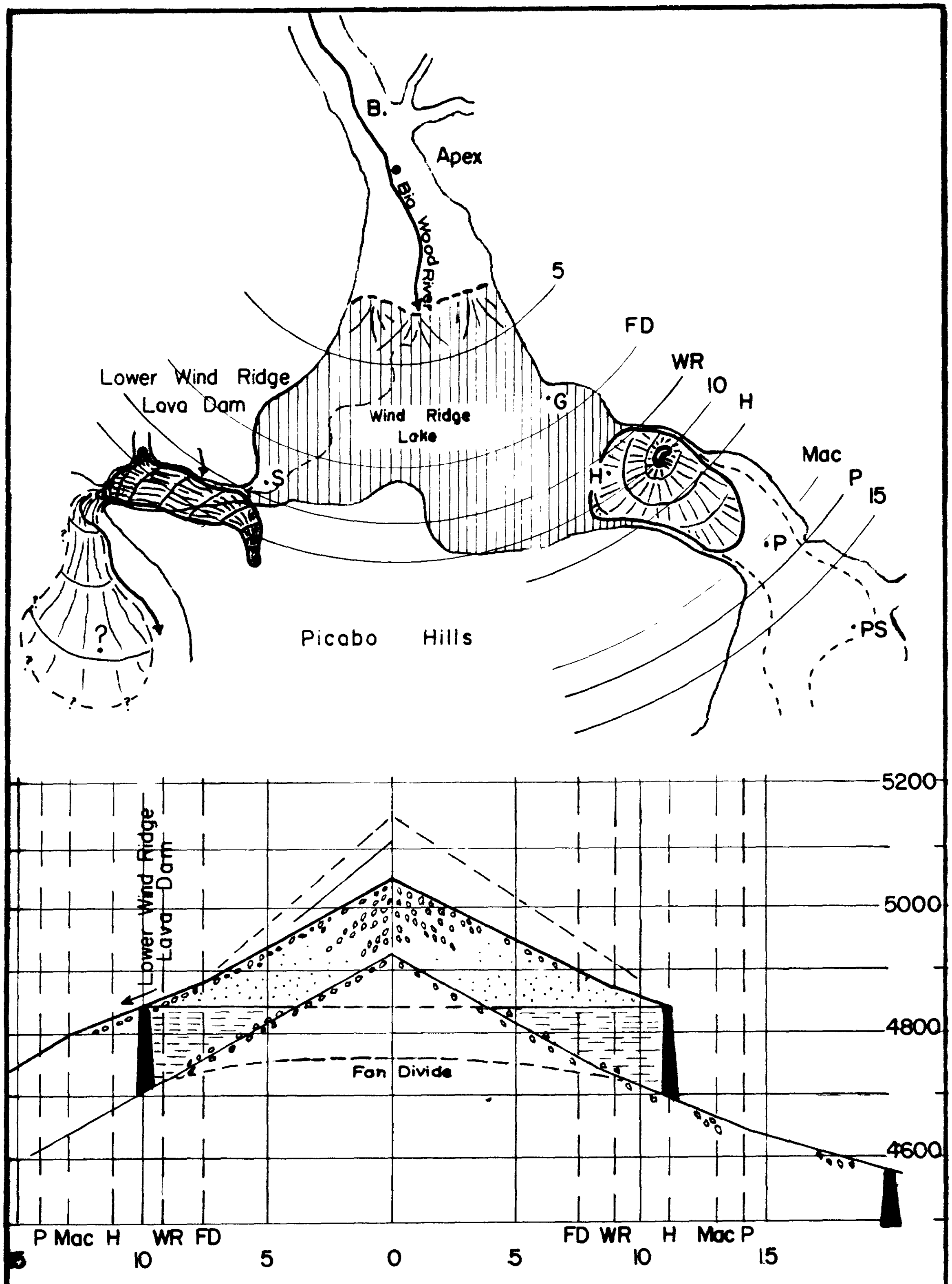
FIGURE 25:--HAY FLOW. The Hay flow blocked the Big Wood River, and the river was diverted across the fan divide to the west outlet valley.



HAY FLOW

FIGURE 25

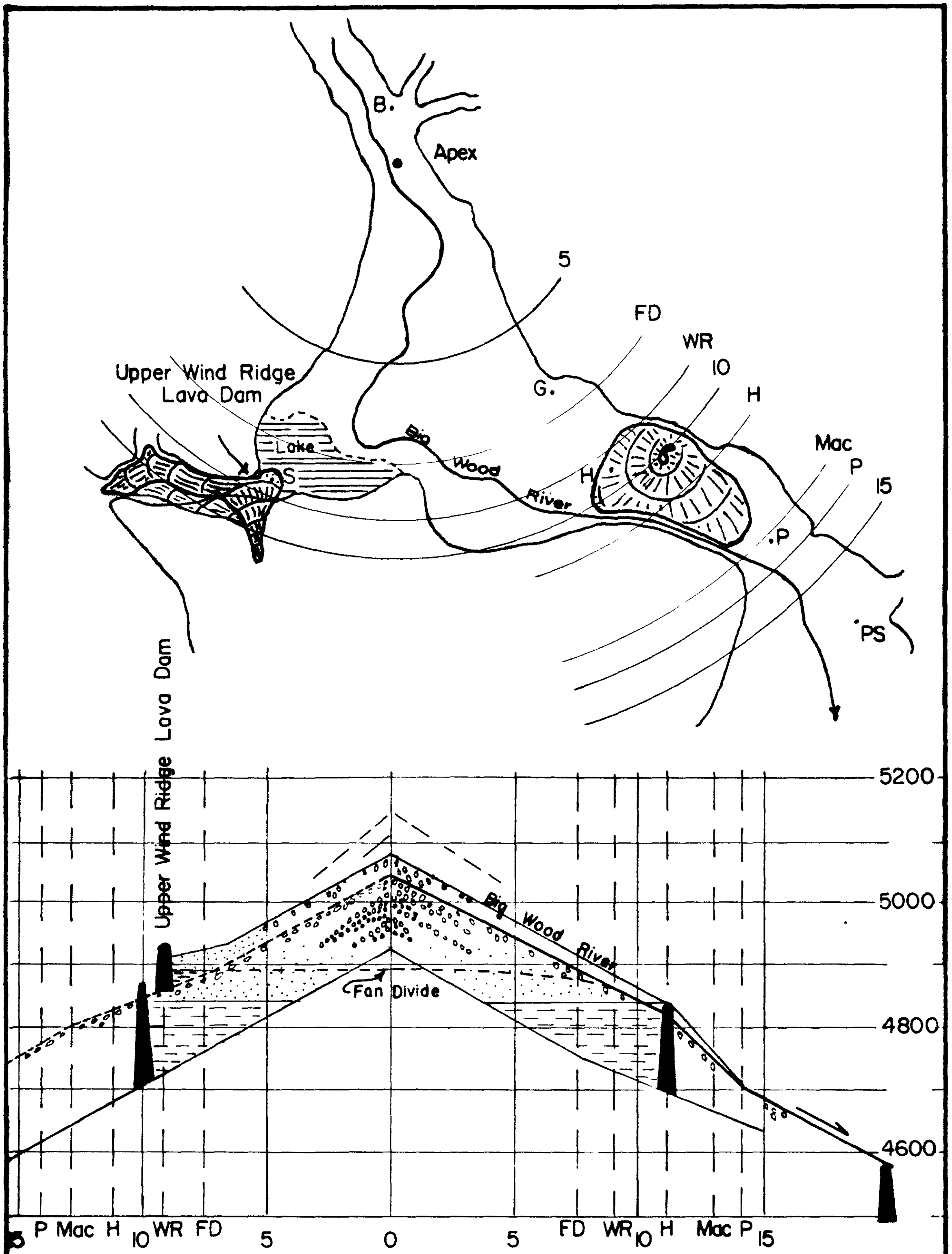
FIGURE 26:--LOWER WIND RIDGE FLOW. The Wind Ridge lake was impounded between the lower Wind Ridge flow and the Hay flow. Upon completion of lacustrine sedimentation and fluvial aggradation in the Wood River Valley, the Big Wood River established its course in the west outlet valley, depositing gravel on top of the lower Wind Ridge flow.



LOWER WIND RIDGE FLOW

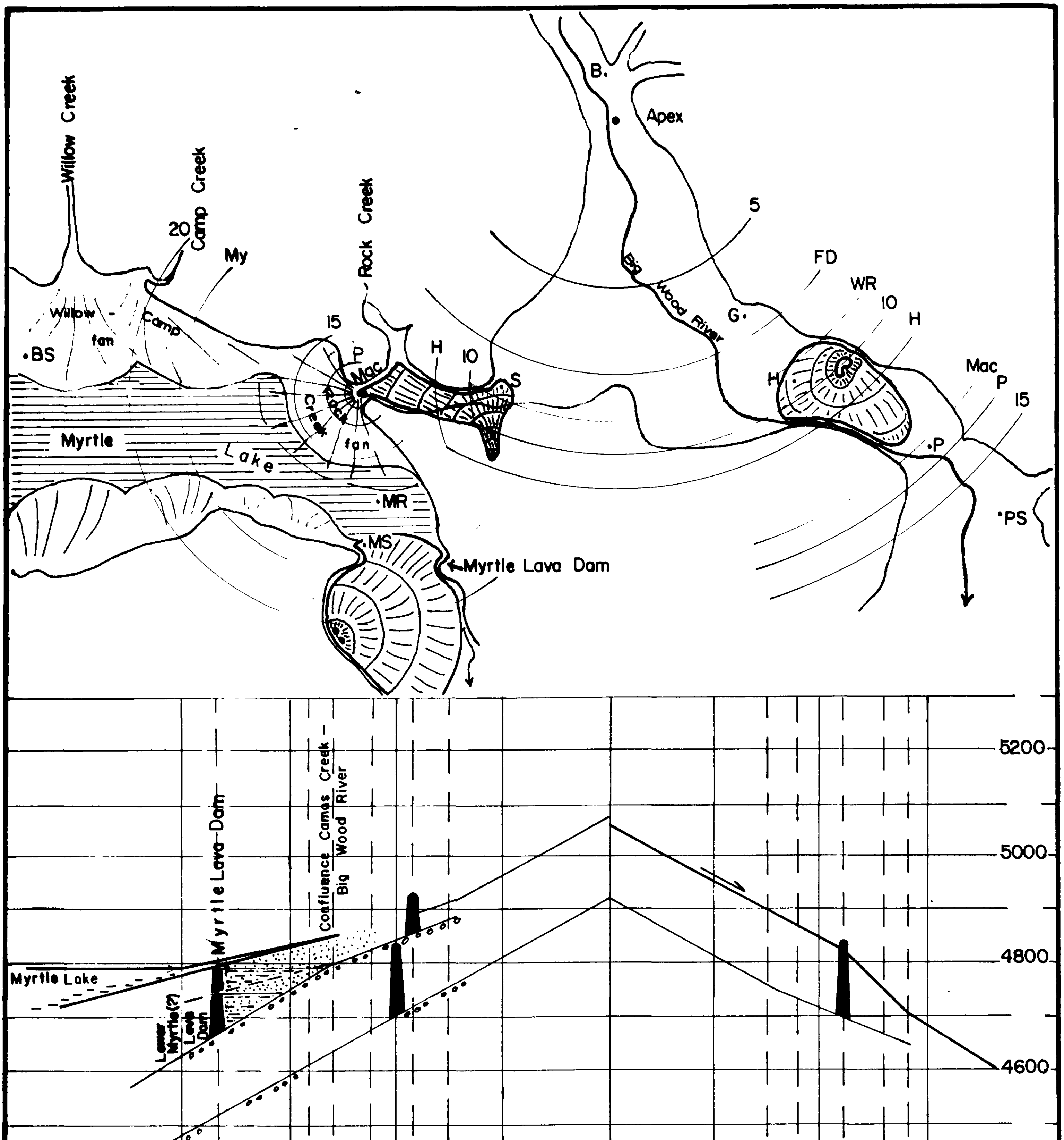
FIGURE 26

FIGURE 27:--UPPER WIND RIDGE FLOW. The Big Wood River was diverted across the fan divide to the east outlet valley where it became entrenched along the margin of the Hay flow.



UPPER WIND RIDGE FLOW FIGURE 27

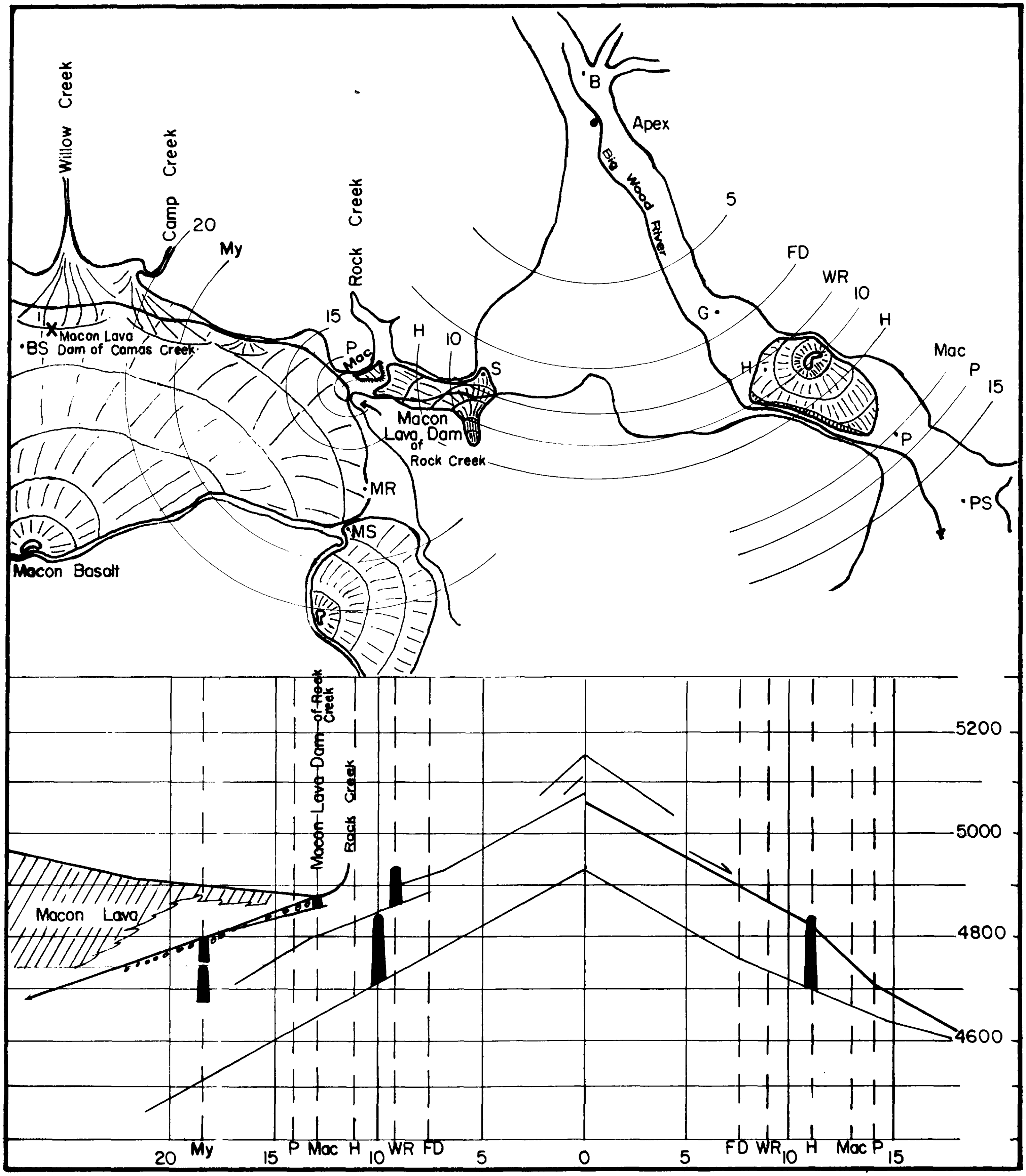
FIGURE 28:--MYRTLE FLOW. Myrtle lake was impounded by the Myrtle flow at about the site of the present day Magic Reservoir dam. The lake was not completely filled by the deposits of the streams draining into Camas Valley before the eruption of the next flow. Additional references: MS, Magic Station; MR, Magic Resort; BS, Blaine Station; My, distance to Myrtle lava dam from the Bellevue apex.



MYRTLE FLOW

FIGURE 28

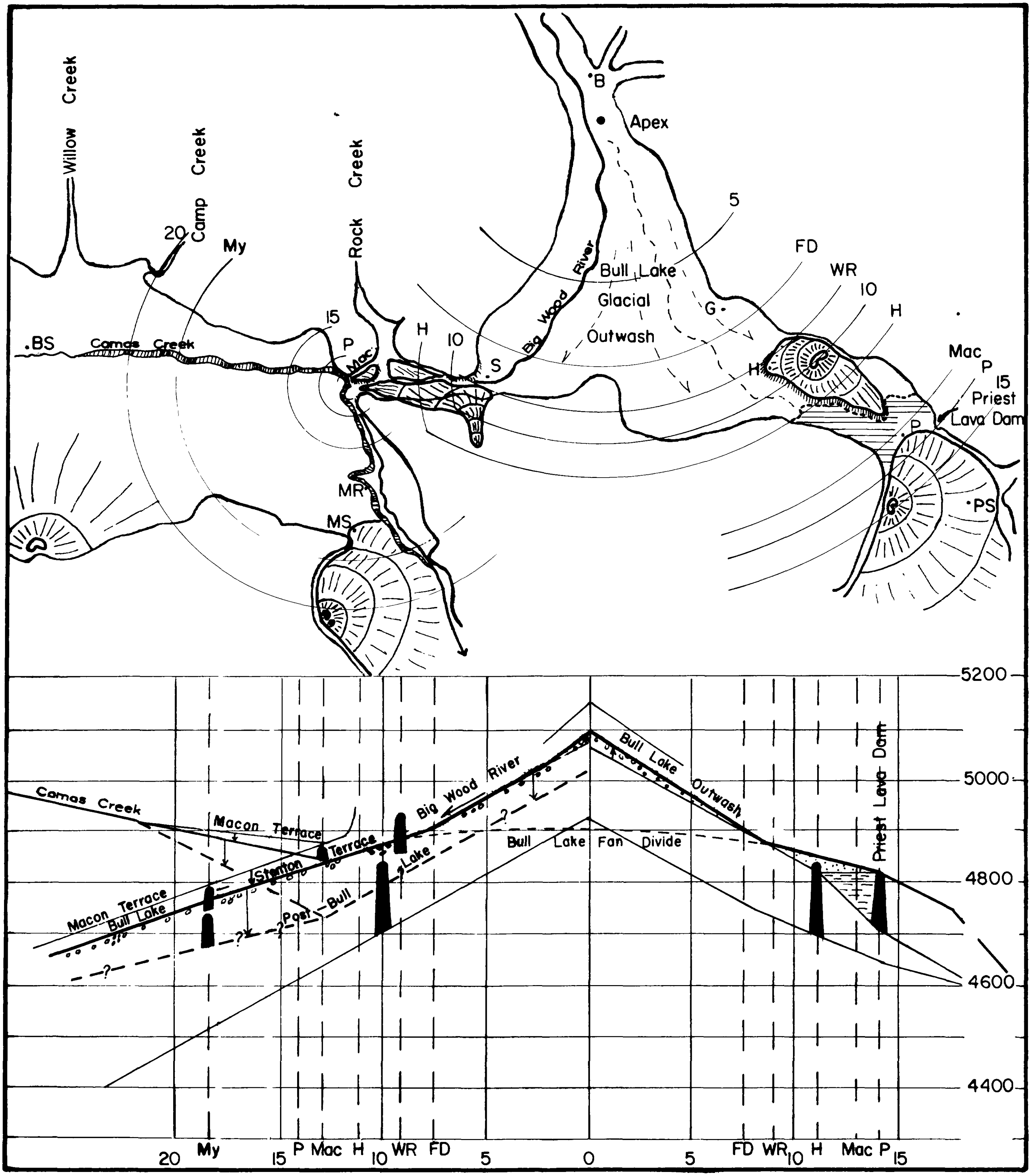
FIGURE 29:--MACON FLOW. The Macon flow spread over the entire eastern part of Camas Valley impounding a large lake west of the map area and disrupting the Camas Creek drainage.



MACON FLOW

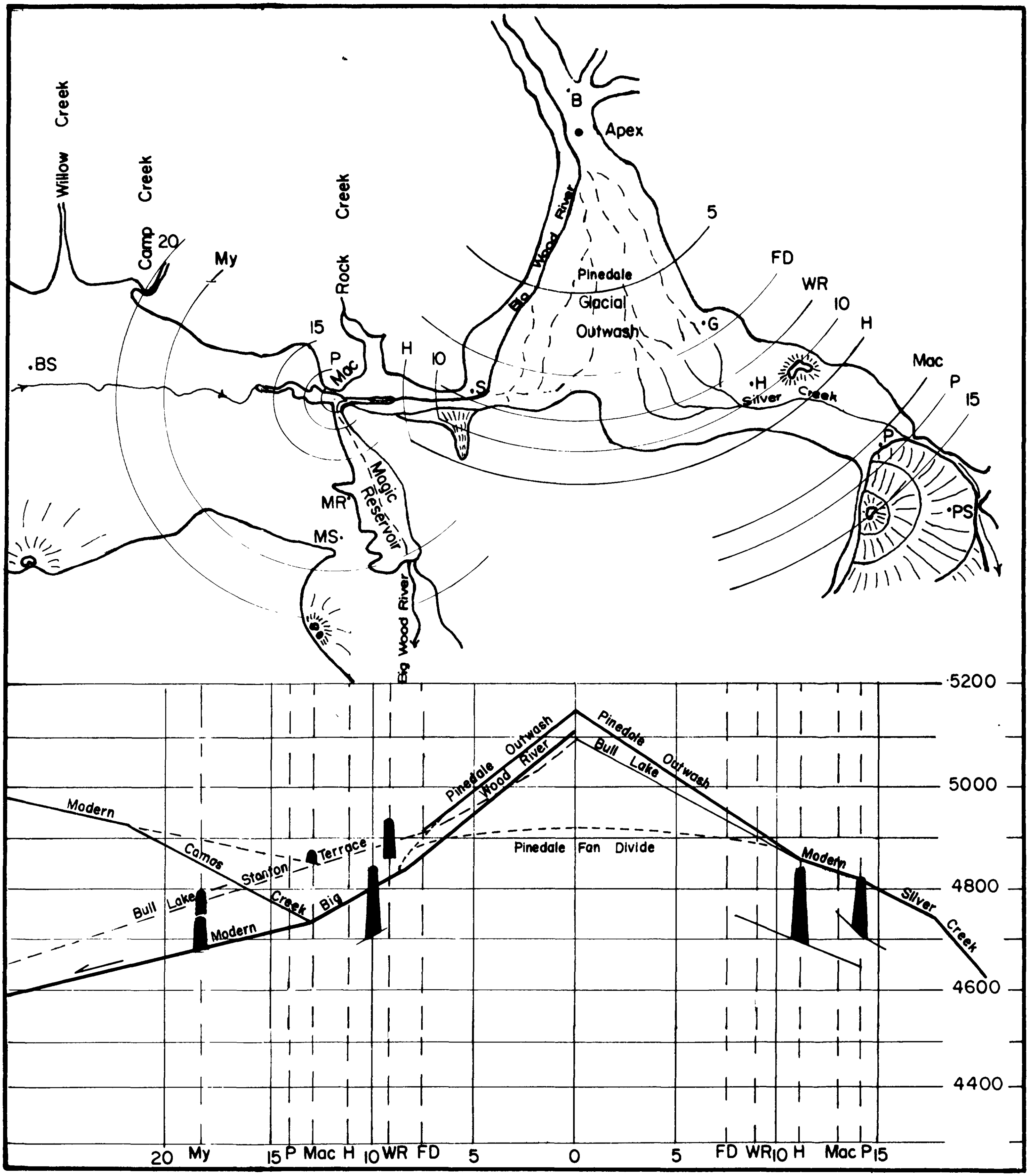
FIGURE 29

FIGURE 30:--PRIEST FLOW AND BULL LAKE OUTWASH. The spreading of the Priest flow impounded a small lake; the Big Wood River continued to flow to the east, filled the lake with sediment, and eventually transported sand and fine gravel across the lava dam. Before the river fully graded itself across the flow, glacial outwash of the Bull Lake stage caused aggradation and building of an outwash fan in the Wood River Valley. The river shifted back and forth on the fan, and at the time of downcutting was in the west valley where it became entrenched across the Wind Ridge, Macon, and Myrtle flows.



PRIEST FLOW AND BULL LAKE OUTWASH FIGURE 30

FIGURE 31:--PINEDALE OUTWASH. During the Pinedale stage, glacial outwash again caused aggradation and the building of the Bellevue fan in the Wood River Valley. The river shifted both east and west on the fan, and the west outlet valley again became the site of downcutting in post-Pinedale time.



PINEDALE OUTWASH

FIGURE 31