Geology of the Blue Mountains Region of Oregon, Idaho, and Washington:

Cenozoic Geology of the Blue Mountains Region

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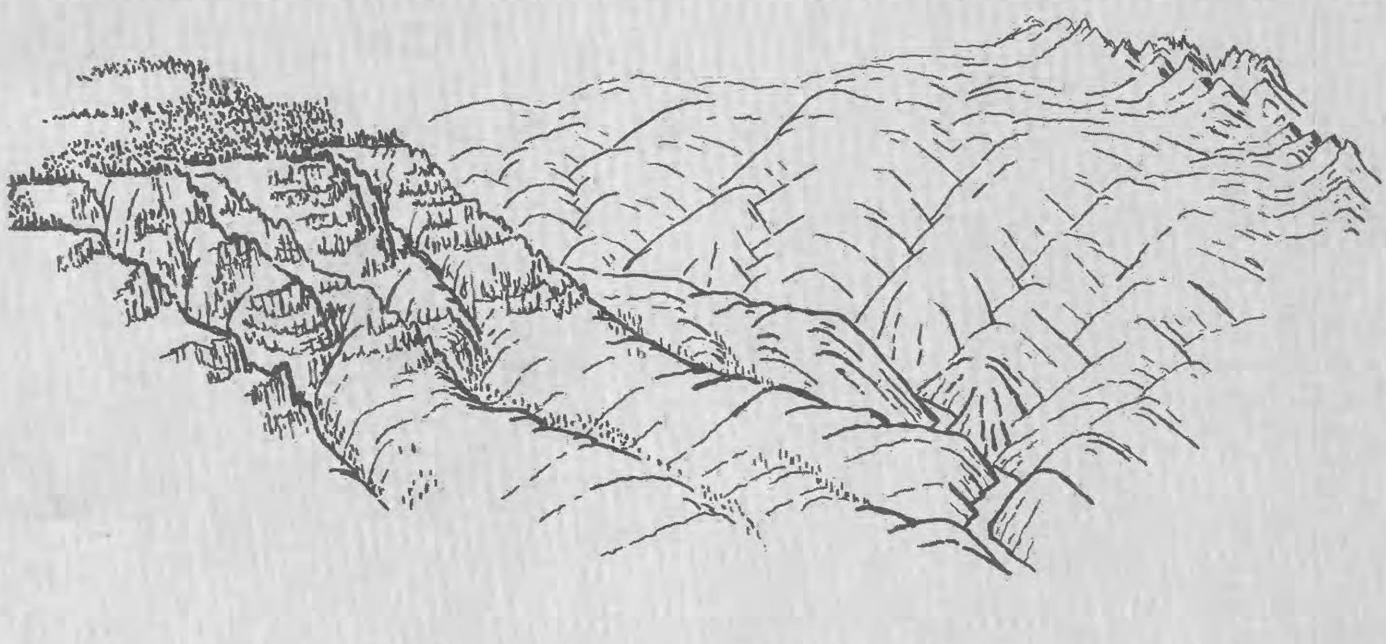

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Geology of the Blue Mountains Region of Oregon, Idaho, and Washington:

Cenozoic Geology of the Blue Mountains Region

GEORGE W. WALKER, *editor*

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PREFACE

This U.S. Geological Survey Professional Paper is one volume of a series that focuses on the geology, paleontology, and mineral resources of eastern Oregon, western Idaho, and southeastern Washington. The purpose of this series is to familiarize readers with the work that has been completed in the Blue Mountains region and to emphasize this region's importance for understanding island-arc processes, the accretion of an allochthonous terrane, and postaccretion magmatism and volcanism. These professional papers provide current interpretations of a complex island-arc terrane that was accreted to ancient North America during the late Mesozoic, of a large batholith that intruded after accretion had occurred, and of overlying Cenozoic volcanic and sedimentary rocks that were subsequently uplifted and partly stripped off the older rocks by erosion.

Most of the Cenozoic volcanogenic rocks were locally derived, although the source vents for some large volumes of rock remain unknown; other rocks formed from material erupted in areas west of the Blue Mountains and were transported eastward into the region. Volcanism within this large region appears to have been more or less continuous throughout the Cenozoic, but the kinds of volcanic products varied both geographically and over time. Earth scientists who have worked on the Cenozoic volcanic rocks of the Blue Mountains and adjacent areas have attributed the volcanism partly to several different volcanic arcs and partly to extensional volcanism in regions behind arc axes.

Most earth scientists who have worked in the Blue Mountains region agree that the pre-Tertiary rocks there form one or more allochthonous terranes. The importance of such terranes in the evolution of circum-Pacific continental margins has been recognized for about a decade, but many complex questions remain. For example, how, when, and where did most of the circum-Pacific allochthonous terranes form? How did they accrete to continents, and how have they affected and been affected by subsequent processes? What are the mechanisms of amalgamation during terrane formation and transport? And, perhaps most importantly, what are the effects of these processes on mineral and hydrocarbon resources? Although the volumes in this series provide some answers, the data and interpretations contained in them will no doubt raise new and equally intriguing questions for future generations of earth scientists.

CONTENTS

[Numbers indicate chapters]

1. OVERVIEW OF THE CENOZOIC GEOLOGY OF THE BLUE MOUNTAINS REGION

By GEORGE W. WALKER

CONTENTS

ABSTRACT

This volume, besides this brief overview, presents five papers that deal with the Cenozoic volcanic and sedimentary rocks of the Blue Mountains region. Most of the descriptive material concerning these rocks is organized into chapters that discuss stratigraphic relations, isotopic and paleontologic ages, and petrographic and petrologic data by age groups. The final chapter discusses some implications of Cenozoic tectonism and volcanism of this large region.

INTRODUCTION

The Cenozoic geologic history of the Blue Mountains, northeastern Oregon and adjacent southeastern Washington, and of contiguous areas in Idaho and central Oregon is dominated by major episodes of volcanism interspersed with periods of sedimentation that took place mostly in local tectonic or volcanotectonic depressions. The episodes of andesitic, rhyolitic to rhyodacitic, and basaltic volcanism, which are related to widely dispersed vent systems of many different kinds, exhibit both temporal and geographic differences; during some periods, diverse chemical types of rock were being erupted in large volumes more or less synchronously.

Regional correlations and extrapolations are only approximate in this area dominated by complex mixtures of intrusive, volcanic, and volcaniclastic rocks and by numerous separate and ephemeral basins of deposition. Early workers relied mainly on general lithology and sparse paleontologic evidence to extrapolate and extend stratigraphic units over large parts of the region, particularly for Miocene and older Cenozoic rocks. Dependence on these lithologic characteristics has resulted in establishing and mapping units of questionable time-stratigraphic significance. Many of the resulting correlations were approximate at best, and some were incorrect. Since about 1965, however, geologic mapping supported by extensive use of radiometric dating, precise major- and minor-element chemistry, magnetic-polarity determinations, reevaluation of old fossil collections, and evaluation of new ones has provided a reasonably consistent regional time-stratigraphic framework suitable for reconstructing the Cenozoic geology of the region.

Although widely recognized and accepted stratigraphic nomenclature has been developed for important parts of the Cenozoic section, the terminology and definition of units for some parts are still in an elementary stage of development. Many problems of terminology and regional correlation remain for Eocene, Oligocene, and lower Miocene volcanic and volcaniclastic rocks of the Blue Mountains region. Difficulties in regional correlation result partly from the mixed and recurring lithologic assemblages, their different degrees of alteration, and, most importantly, from absence of detailed mapping and adequate age data. However, recent intensive study of the Miocene basalts of the region by numerous individuals representing many diverse interests has led to a fairly precise stratigraphic nomenclature for these rocks over major parts of the region, supported by extensive chemical, radiometric age, and magnetostratigraphic data. Miocene and younger volcaniclastic and sedimentary rocks are mostly restricted to separate local basins of deposition and, in general, have not received the same attention as the Miocene basalts.

For the purposes of this volume, the geographic area of coverage is arbitrarily restricted to an area of approximately 140,000 km^2, including parts of southeastern Washington, northeastern Oregon, and west-central Idaho. This area includes the Blue Mountains province, parts of Columbia Basin, the Columbia Intermontane province, parts of the Deschutes-Umatilla Plateau, the Joseph

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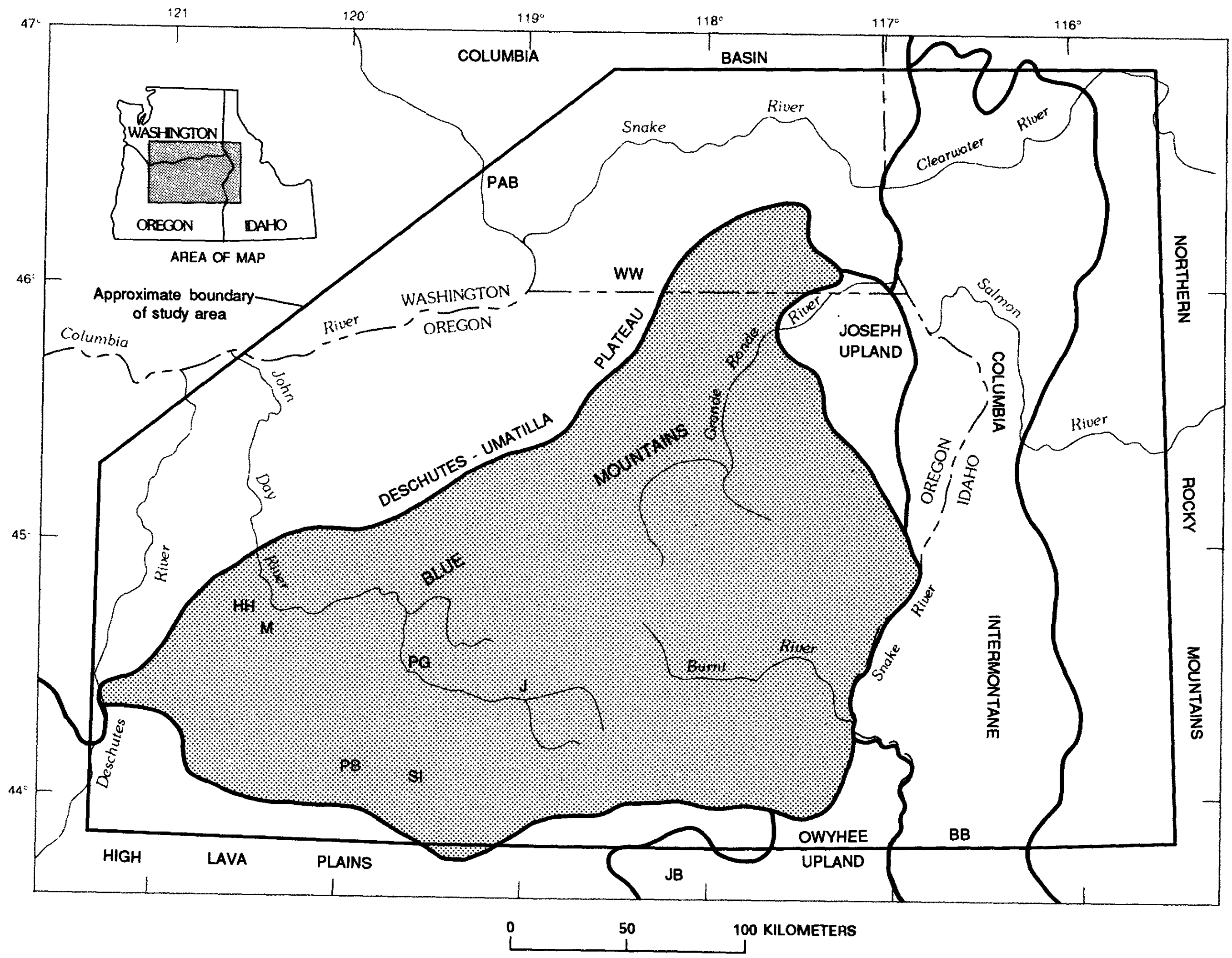
Upland, and small parts of the High Lava Plains and Owyhee Upland provinces of southeastern Oregon (fig. 1.1). Excluded from consideration are areas dominated by volcanic rocks of the Cascade Range, which lie far to the west of the Blue Mountains and also west of the Deschutes River and areas to the south and southeast, within the northern Basin and Range, although these areas contain sequences of volcanic and volcaniclastic rocks correlative, in part, with rocks of the Blue Mountains. Some of those rock units are mentioned, however, in the following pages to obtain a better regional understanding of the temporal and spatial distribution of different kinds of volcanism within the Cenozoic. In addition, some volcanic rocks erupted in these peripheral areas extend long distances into parts of the Blue Mountains province.

A diverse assemblage of Cenozoic terrestrial volcanic and sedimentary rocks occupies most of the surface area within this 140,000-km^2 area; pre-Cenozoic rocks represent perhaps only 10 to 15 percent of the total outcrop area (fig. 1.2). In the Oregon and Washington parts of the region, inliers of Paleozoic and Mesozoic metasedimentary and metavolcanic rocks, partly representing a tectonically mixed assemblage of accreted ophiolite and island-arc terranes, are intruded by several varieties and ages of plutonic rocks. In western Idaho, the Cenozoic cover



INDEX TO LOCALITIES

BB, Boise Basin
HH, Horse Heaven mining district
J, John Day and John Day Basin
JB, Juntura Basin
M, Mitchell
PAB, Pascoe Basin
PB, Paulina Basin
PG, Picture Gorge
SI, Suplee-Izee area
WW, Walla Walla Basin

FIGURE 1.1.—Index map of the Blue Mountains region (screened area), showing major physiographic provinces and localities mentioned in text.

rests on similar Paleozoic and Mesozoic rocks and on high-grade metamorphic rocks of Precambrian age.

Within the Blue Mountains province, the outcrop distribution of pre-Cenozoic and lower Cenozoic rocks is related partly to burial by subsequent volcanism from vents marginal to and within the province and partly to differential erosion of a structurally and physiographically high region. A plexus of folds and tilted fault blocks make up this large, structurally uplifted block, which extends for about 360 km across northeastern Oregon into western Idaho and the southeast corner of Washington. The width of the elevated block is difficult to determine but is generally about 80 km or more. Distribution of Cenozoic units indicates that deformation was initiated during latest Eocene or Oligocene time and has persisted into late Miocene and probably into Quaternary time. The deformation has been inferred by some geologists to be related to northwest-southeast regional compressional stresses and by others to be related in part to deformation of brittle surface and near-surface rocks as a result of tangential stresses from transcurrent movement between crustal blocks. Different workers in the region have labeled this elevated crustal block an anticline, anticlinorium, antiform, or uplift; we prefer the nonspecific term "uplift," inasmuch as the origin of the structure is only poorly known.

Before the early 1970's, investigators assigned stratigraphic units to geologic epochs, the boundary ages of which differed considerably from those acceptable in the 1980's. This practice led to some confusion, particularly where formations originally assigned to the Miocene or Pliocene are now referred to, respectively, as Oligocene or Miocene in age. Throughout this volume, authors of chapters use the epoch-boundary ages of Palmer (1983), which are younger in years than those used in many previous reports on the geology of the Blue Mountains and adjacent areas. Every attempt has been made in the figures and text to integrate these different assigned ages.

Most of the K-Ar ages referred to in this and subsequent chapters are from published reports. Because some of these published ages were originally calculated using decay constants that are no longer acceptable, all ages for which analytical data are available have been recalculated using decay constants acceptable in 1983 (see Fiebelkorn and others, 1983). The recalculated ages using the new constants generally are older by a few percent than the original reported ages that were calculate using the old constants.

ACKNOWLEDGMENTS

Preparation of this series of volumes on the geology of the Blue Mountains region was initiated by T.L. Vallier of the U.S. Geological Survey and H.C. Brooks of the Oregon Department of Geology and Mineral Industries, both of whom have devoted large segments of their careers to the geology of this region. Because they wished to concentrate their efforts on their main interests in the pre-Cenozoic geology of the Blue Mountains, they asked me to oversee the volume on the Cenozoic geology of the region: to identify potential authors of chapters, to assist these authors in establishing appropriate coverage, and to act in an informal editorial capacity.

This summary volume is based on results from the work of nearly all geologists who have worked in the area and on discussions over the years with many of them. Their work is acknowledged mostly through references to their published papers and, in a few places, to either written or oral communications.

The authors thank T.L. Vallier and H.C. Brooks for generating interest in this endeavor, and W.H. Taubeneck, T.L. Robyn, T.P. Thayer, E.M. Taylor, and A.C. Waters for many interesting discussions over the years concerning their investigations of the Cenozoic geology of the region. The critical reviews and helpful suggestions of N.S. MacLeod and T.L. Vallier substantially improved the separate manuscripts that make up this volume.

PREVIOUS WORK

Previous summary reports dealing with aspects of the geology of the Blue Mountains and contiguous areas are mainly topical discussions of selected parts of the stratigraphic column or of special features in restricted geographic areas; several areally extensive geologic maps also have been prepared for different parts of the region.

Early discussions of John Day Basin in Oregon, located mainly at the west end of the Blue Mountains province and including the southwestern part of the Deschutes-Umatilla Plateau, include papers on the general geology (Merriam, 1901), the petrography of the Eocene and Oligocene(?) Clarno Formation (Calkins, 1902), and both fossil floras (Knowlton, 1902; Chaney, 1924) and faunas (Merriam and Sinclair, 1907). Since these early studies, many additional papers have been published on details or reviews of Clarno Formation floras (Arnold, 1952; Scott, 1954; Hegert, 1961; Elemenderf and Fisk, 1978; Manchester, 1981) and Clarno Formation faunas (Stirton, 1944; Hanson, 1973). In addition, many topical papers have been published concerning the distribution, lithology, structure, and alteration of both the Clarno and John Day Formations of central Oregon; foremost among these papers are those by Waters and others (1951), Hay (1962), Hogenson (1964), Fisher (1967), Fisher and Rensberger (1972), and Noblett (1981).

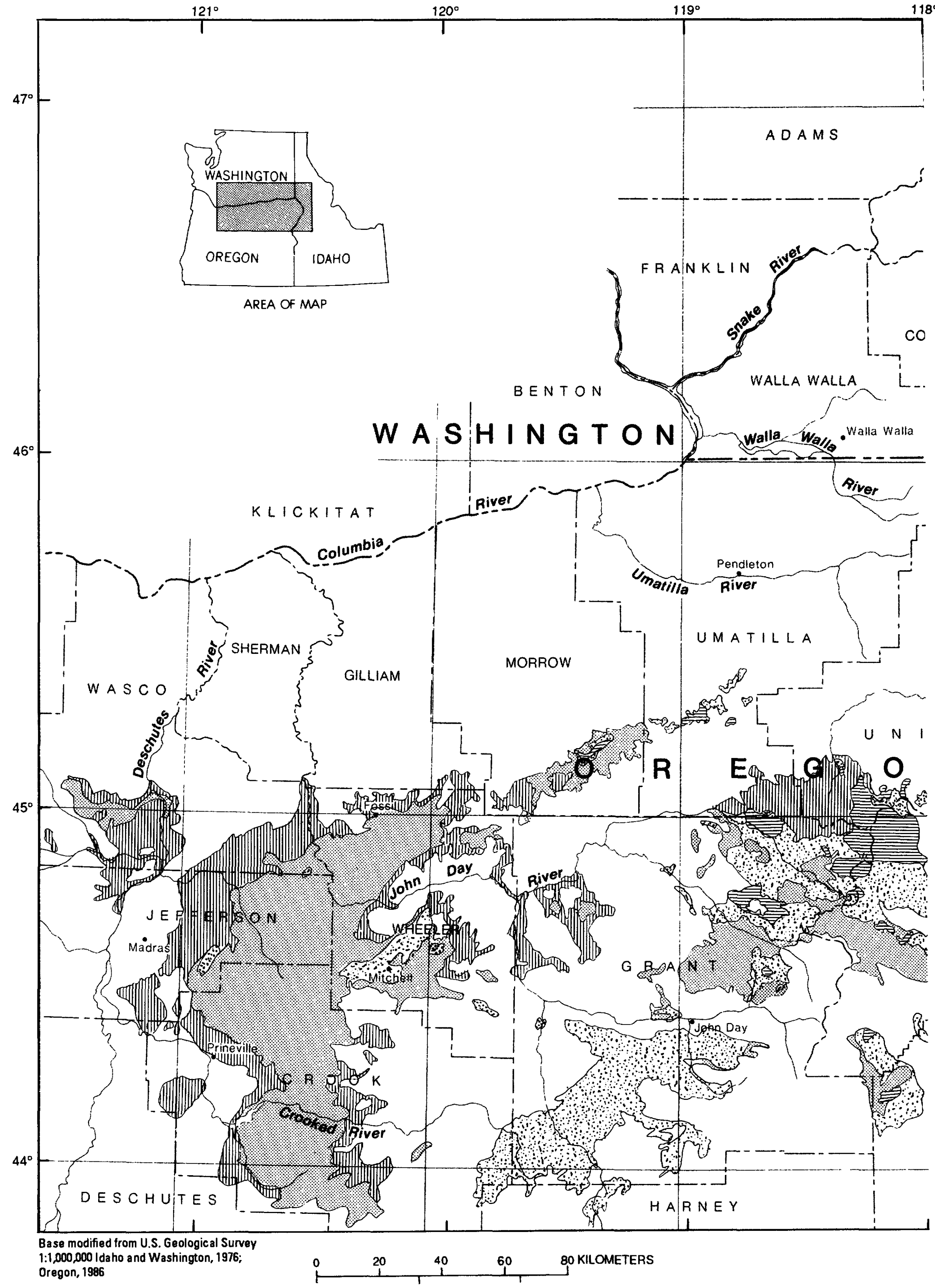
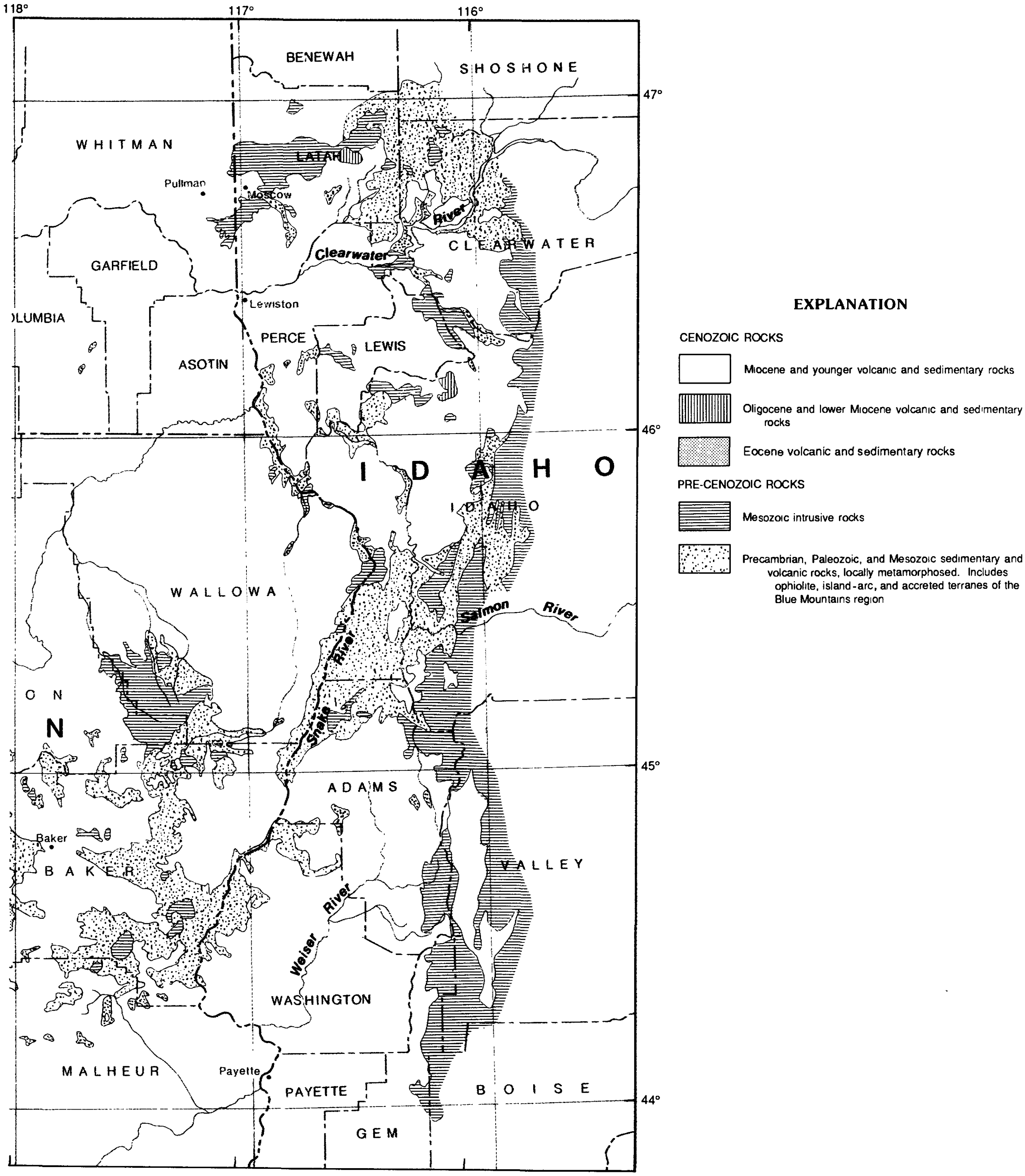


FIGURE 1.2.—Generalized geologic map of northeastern



Geology modified from Wells and Peck (1961), Walker (1977), and Bond (1978)

Oregon and adjacent areas in Washington and Idaho.

Brief mention was made by Anderson (1930, p. 25), Bond (1963) and, more recently, Jones (1982, p. 43–52) of andesitic and latitic volcanic rocks in western Idaho that are probably equivalent to the Challis Volcanics and correlative with the Clarno Formation; some of these volcanic rocks exposed in western Idaho were thought to be as old as Permian(?) (Tullis, 1944, p. 40).

The Miocene basaltic rocks, mostly making up what is now known as the Columbia River Basalt Group but including some petrographically and chemically distinct mafic flow units, have been treated in many publications; earlier ones are by Mackin (1961) and Waters (1961), and more recent ones are by Bond (1963), Thayer and Brown (1966), Schmincke (1967), Wright and others (1973), Hooper (1974), and Nathan and Fruchter (1974). Proliferation of formal and informal names for basaltic flows within this group led to a review by Swanson and others (1979) in which a revised formal stratigraphic nomenclature was established and the petrologic, magnetic polarity, and chemical characteristics were described. That report also summarized the geographic distribution of units and K-Ar ages of selected flows. Other formal publications and numerous dissertations concerning these rocks are referenced throughout this volume.

Several early publications dealt with both lava flows and terrestrial sedimentary rocks of Miocene and younger age in and adjacent to Boise Basin. For example, studies by Lindgren (1898), Russell (1902), Merriam (1918), and Buwalda (1924) established early stratigraphic concepts for many of the units exposed in the basin. Regional stratigraphic relations and nomenclature, as well as precise age assignments, of these diverse rocks recently have been reconsidered and revised, largely through the work of Malde and Powers (1962), Armstrong (1975), Armstrong and others (1975), and other workers.

Many reports are available on the stratigraphy and structure of Pasco Basin in southeastern Washington, most of them related to investigations of ground water and radioactive-waste disposal at and near Hanford. Principal among these reports are those by Strand and Hough (1952), Newcomb (1958), Brown and McConiga (1960), Newcomb and others (1972), and Gustafson (1978). Newcomb (1965) described the geology and water resources of Walla Walla River basin, which lies about 60 km southeast of Pasco Basin.

Regional geologic maps and compilations that have contributed to our understanding of the distribution of Cenozoic volcanic and sedimentary units include principally those by Gaston and Bennett (1979), Mitchell and Bennett (1979), Rember and Bennett (1979), and Swanson and others (1981) for Idaho; Brown and Thayer (1966), Walker and others (1967), Swanson (1969), Greene and others (1972), Walker (1973, 1979), Robinson (1975), and Brooks and others (1976) for northeastern Oregon; and Swanson and others (1980) for southeastern Washington. Smaller-scale maps of Idaho (Bond, 1978), Washington (Huntting and others, 1961), eastern Oregon (Walker, 1977), and Oregon and Washington (Luedke and Smith, 1982) also provide background information on the regional distribution of volcanic and sedimentary sequences. Each of these regional maps provides an index to more detailed work in specific small areas; much of this more detailed work is in unpublished theses.

PURPOSE AND SCOPE

The chapters in this volume describe the volcanic and sedimentary rocks of the Blue Mountains region, their distribution by time-lithologic units, and their relations to vents; the Cenozoic structural development of the Blue Mountains and of several subsidiary depositional basins; and, finally, the interrelations of tectonism and volcanism. The volume will be a companion to several other volumes published (Vallier and Brooks, 1986, 1987) or in preparation that discuss the pre-Cenozoic geology of the same major parts of eastern Oregon, southeastern Washington, and western Idaho, including the Idaho batholith and its wallrocks.

The large size of the area being considered in this report and the diversity of the Cenozoic geology required the talents of several authors whose interests dealt with specific parts of the stratigraphic column or with single basins of deposition or related groups of basins. There is little uniformity in the amount and kind of data available for the different subjects that are considered. Intensive studies of some parts of the stratigraphic column or of small geographic areas have provided much data on thicknesses of section, lithology, distribution of lithologic units, chemistry, isotopic ages, and magnetic-polarity characteristics. Other areas and other parts of the stratigraphic column have been studied by only the broadest reconnaissance methods, and the resulting data are sparse and generally provisional.

A final summary chapter deals with the Cenozoic tectonic and volcanic history of the region. It attempts to bring data presented in the other sections into focus, particularly in relation to structural development and volcanism and to concepts of magma generation and crustal development.

In 1981, I initially provided each author with a general outline of the entire volume, suggesting general topics to be covered in each chapter and giving very general guidelines as to scope and philosophy of approach. Within these guidelines, authors independently prepared their reports dealing with particular parts of the Cenozoic section. Because of different approaches to the subject by

the authors and the vast differences in availability of data, the chapters are not uniform in either style or coverage. Furthermore, several chapters prepared in 1982 and 1983 represent the status of knowledge at that time and have not been substantially updated as new work was completed and published.

In this region, as in most volcanic regions, generalized stratigraphic columns are of limited usefulness, and regional correlations are only approximate. In and marginal to the Blue Mountains, numerous separate ephemeral and mostly structurally controlled depositional basins have received various volcanic, volcaniclastic, and sedimentary materials from many source areas. Before radiometric dating and the development of precise geochemical and magnetostratigraphic techniques, many correlations of lithologic units among the separate basins were somewhat haphazard; only in recent years have time-stratigraphic relations reached a stage where regional correlation charts are meaningful and reasonably consistent.

Correlations that are well supported by numerous radiometric ages, by clearly defined rock chemistry, or by systematic magnetostratigraphy are restricted largely to the Oligocene and Miocene part of the section and concern primarily the John Day Formation of central Oregon and the overlying flows of the Columbia River Basalt Group.

REGIONAL STRATIGRAPHIC RELATIONS

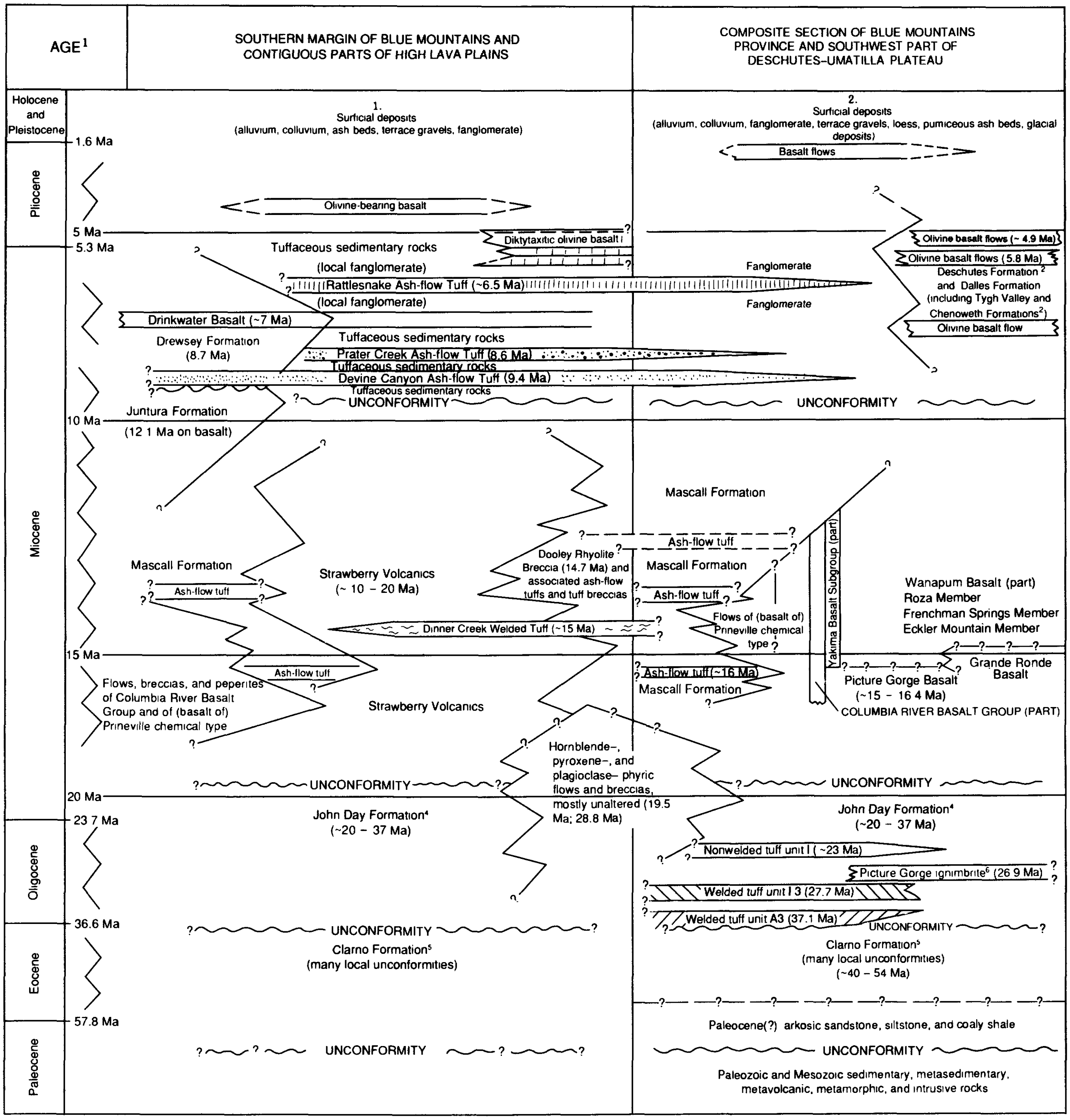
One purpose of this introductory chapter is to provide a regional stratigraphic framework for the individual chapters that deal with specific part of the stratigraphic column. Those chapters present detailed descriptions of the physical and chemical characteristics of the different lithologic units and the relations within the described sequences of rocks.

Composite and highly generalized stratigraphic columns have been prepared for four separate but interrelated parts of the region (fig. 1.3). These four columns represent (1) the Blue Mountains province and southwestern part of the Deschutes-Umatilla Plateau; (2) the southern part of Columbia Basin and adjacent parts of Idaho, including the Clearwater embayment; (3) the southern margin of the Blue Mountains province and contiguous parts of the High Lava Plains; and (4) the northern Owyhee Upland and northwestern part of the King Hill section of the Columbia Intermontane province, forming the northwestern and western part of Boise Basin. The columns are arranged in a correlation chart (fig. 1.3) and referenced to lines representing ages of 20, 15, 10, and 5 Ma. Stratigraphic units common to more than one column are indicated, and age spans of isotopically dated units are shown in parentheses. Gaps in the columns, in part shown as regional unconformities, indicate that volcanism in this region was episodic. However, as new radiometric ages have been determined from various parts of the region, some of the apparent gaps in the volcanic history seem to be diminishing in duration or disappearing. Where interfingering relations are known or inferred, they are shown by jagged lines (fig. 1.3); unconformities, which are present in many parts of the section, are shown only if they are of regional extent or, if of local extent, are important in understanding relations in this complex volcanic area.

The Cenozoic rocks of the region are divisible into several major geologic units, some with wide geographic distribution and others restricted to small ephemeral basins of deposition or to geographically restricted volcanic piles. Even though some of these units are recognizable over large parts of the the region, the Cenozoic column is so segmented by regional and local unconformities and by faults that estimates of composite thickness are almost meaningless. Nowhere are all the diverse units displayed in a continuous sequence, and in most parts of the region only small segments of the column are present. One of the thickest sequences of Eocene and Oligocene(?) rocks, representing the Clarno Formation, is that described by Waters and others (1951) for the Horse Heaven mining district of Oregon, where the total thickness is about 1,800 m. Oles and Enlows (1971) described a section of similar thickness in the Mitchell quadrangle; they divided this section into upper and lower parts and identified the total section as the Clarno Group. The thickest described section of uppermost Eocene(?), Oligocene, and lower Miocene rocks, representing the John Day Formation, is that estimated by Peck (1964) to be about 1,200 m thick near the west limits of the formation. The John Day Formation has a composite thickness of about 820 m near Mitchell (Hay, 1963) and 760 m near Picture Gorge (Fisher and Rensberger, 1972). Miocene basaltic rocks and the partly coeval Strawberry Volcanics vary extremely in thickness. Maximum total thickness of this part of the section was estimated at about 2,450 m by Thayer and Brown (1966), although Swanson and others (1979) indicated that this thick section of basaltic flows may be partly duplicated by faulting. In the deep canyons adjacent to the Snake and Grande Ronde Rivers, the total exposed thickness is about 800 m, and this value represents only a part of the Miocene basalt column. The Mascall Formation of middle Miocene age, which locally overlies the basalt flows and elsewhere appears to interfinger with them, was reported (Thayer and Brown, 1966) to be locally as thick as 1,830 to 2,135 m.

Measured thicknesses of the Miocene silicic volcanic and volcaniclastic rocks and of the upper Miocene and overlying volcaniclastic and sedimentary rocks vary so

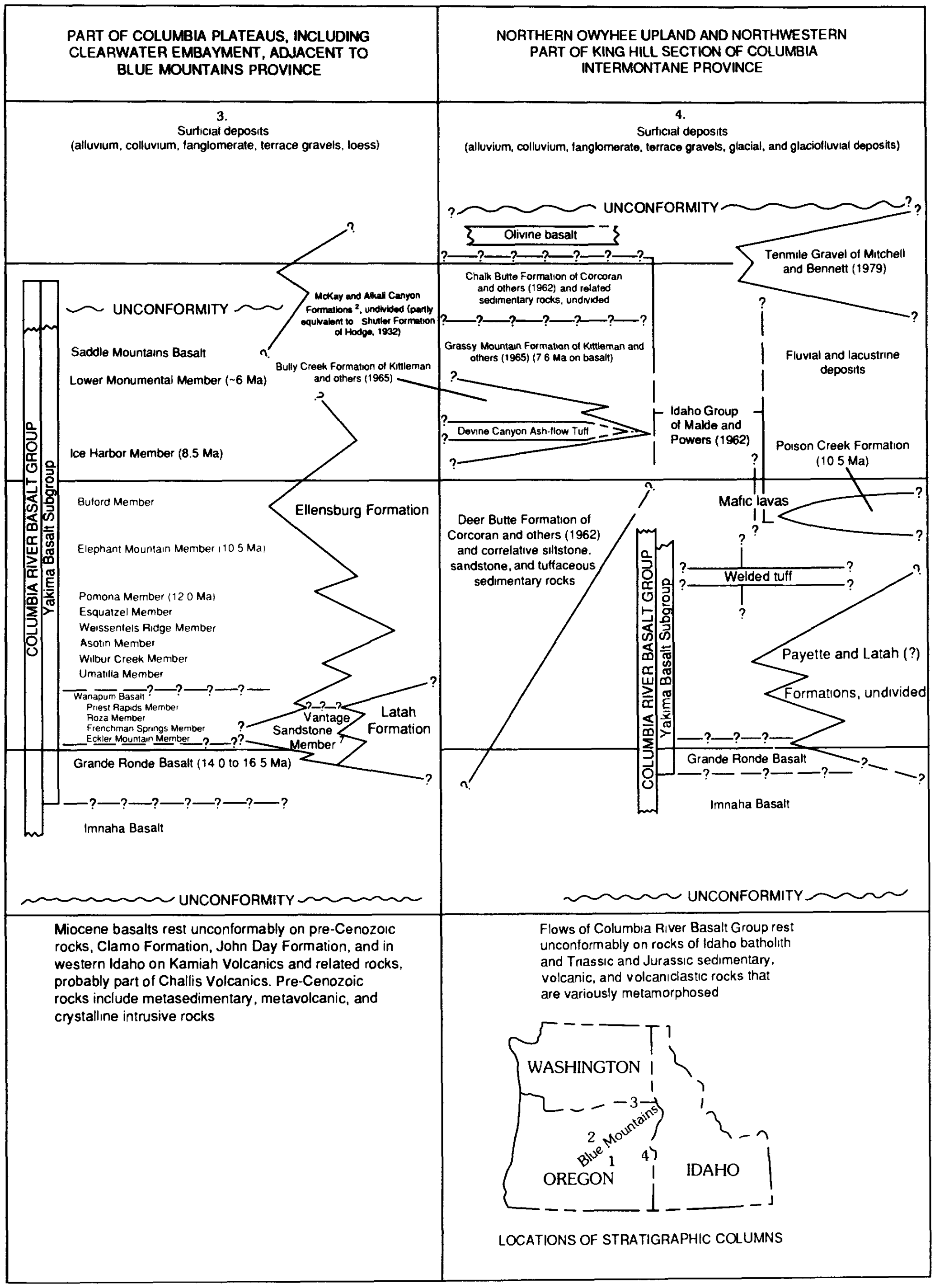
much from one depositional basin to another and within basins that few generalizations are possible. In basins marginal to the Blue Mountains, volcaniclastic and tuffaceous sedimentary deposits of middle Miocene (Bar-



[1] Time scale of Palmer (1983)
[2] Of Farooqui and others (1981)
[3] Peck (1964)
[4] Unit age is latest Eocene (?) to early Miocene
[5] Unit age is Eocene and earliest Oligocene (?)
[6] Informal unit of Fisher (1966)
[7] Of the Ellensburg Formation

FIGURE 1.3.—Generalized stratigraphic columns

stovian) and younger age are locally as much as 650 m thick, but they generally thin rapidly and pinch out on older units. Bowen and others (1963) reported measured thickness of approximately 270 m for middle Miocene



of Blue Mountains region and adjacent areas.

(Barstovian) clastic rocks, about 130 m for overlying upper Miocene (Clarendonian), and 250 m for uppermost Miocene and Pliocene(?) (Hemphillian) rocks, for a total section thickness of nearly 640 m in Juntura Basin. Similar thicknesses of correlative tuff and sedimentary rocks are present in areas to the west of the Blue Mountains, where parts of the sections are exposed in deep canyons of the Deschutes River and its tributaries. Rock sequences several hundred meters thick are present in several basins marginal to the Blue Mountains on the southeast—in the north end of the Owyhee Upland province and in the Columbia Intermontane province. Within the Blue Mountains, local sections of these middle Miocene and younger clastic rocks are generally much thinner; the thickest sections may be those that are about 300 m thick in Paulina Basin (Davenport, 1971) and to the east in the Suplee-Izee area (Dickinson and Vigrass, 1965).

REFERENCES CITED

Anderson, A.L., 1930, The geology and mineral resources of the region about Orofino, Idaho: Idaho Bureau of Mines and Geology Pamphlet 34, 63 p.

Armstrong, R.L., 1975, The geochronometry of Idaho: Isochron/West, no. 14, 50 p.

Armstrong, R.L., Leeman, W.P., and Malde, H.E., 1975, K-Ar dating, Quaternary and Neogene volcanic rocks of the Snake River basin, Idaho: American Journal of Science, v. 275, no. 3, p. 225–251.

Arnold, C.A., 1952, Fossil Osmundaceae from the Eocene of Oregon: Palaeontographica, v. 92, pt. B, p. 63–78.

Bond, J.G., 1963, Geology of the Clearwater embayment: Idaho Bureau of Mines and Geology Pamphlet 128, 83 p.

———1978, Geologic map of Idaho: Moscow, Idaho Bureau of Mines and Geology, scale 1:500,000.

Bowen, R.G., Gray, W.L., and Gregory, D.C., 1963, General geology of the northern Juntura Basin, *in* Shotwell, J.A., The Juntura Basin—Studies in earth history and paleoecology: American Philosophical Society Transactions, new ser., v. 53, pt. 1, p. 22–34.

Brooks, H.C., McIntyre, J.R., and Walker, G.W., 1976, Geology of the Oregon part of the Baker 1° by 2° quadrangle: Oregon Department of Geology and Mineral Industries Geologic Map Series Map GMS–7, scale 1:250,000.

Brown, C.E., and Thayer, T.P., 1966, Geologic map of the Canyon City quadrangle, northeastern Oregon: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–447, scale 1:250,000.

Brown, R.E., and McConiga, M.W., 1960, Some contributions to the stratigraphy and indicated deformation of the Ringold Formation: Northwest Science, v. 34, no. 2, p. 43–54.

Buwalda, J.P., 1924, The age of the Payette Formation and the old erosion surface in Idaho: Science, new ser., v. 60, p. 572–573.

Calkins, F.C., 1902, A contribution to the petrography of the John Day Basin: University of California, Department of Geology Bulletin, v. 3, p. 109–172.

Chaney, R.W., 1924, Fossil floras of the John Day Basin: Carnegie Institution of Washington Year Book 22, p. 349–350.

Corcoran, R.E., Doak, R.A., Porter, P.W., Pritchett, F.I., Jr., and Privrasky, N.C., 1962, Geology of the Mitchell Butte quadrangle, Oregon: Oregon Department of Geology and Mineral Industries Geologic Map Series Map GMS–2, scale 1:125,000.

Davenport, R.E., 1971, Geology of the Rattlesnake and older ignimbrites in the Paulina Basin and adjacent area, central Oregon: Corvallis, Oregon State University, Ph.D. dissertation, 132 p.

Dickinson, W.R., and Vigrass, L.W., 1965, Geology of the Suplee-Izee area, Crook, Grant, and Harney Counties, Oregon: Oregon Department of Geology and Mineral Industries Bulletin 58, 109 p.

Elemendorf, Jean, and Fisk, L.H., 1978, The age of the "Pilot Rock flora," Umatilla County, Oregon [abs.]: Geological Society of America Abstracts with Programs, v. 10, no. 5, p. 215.

Farooqui, S.M., Beaulieu, J.D., Bunker, R.C., Stensland, D.E., and Thoms, R.E., 1981, Dalles Group—Neogene formations overlying the Columbia River Basalt Group in north-central Oregon: Oregon Geology, v. 43, no. 10, p. 131–140.

Fiebelkorn, R.B., Walker, G.W., MacLeod, N.S., McKee, E.H., and Smith, J.G., 1983, Index to K-Ar determinations for the State of Oregon: Isochron/West, no. 37, 60 p.

Fisher, R.V., 1966, The geology of a Miocene ignimbrite layer, John Day Formation, eastern Oregon: University of California Publications in Geological Sciences, v. 67, 73 p.

———1967, Early Tertiary deformation in north-central Oregon: American Association of Petroleum Geologists Bulletin, v. 51, no. 1, p. 111–123.

Fisher, R.V., and Rensberger, J.M., 1972, Physical stratigraphy of the John Day Formation, central Oregon: University of California Publications in Geological Sciences, v. 101, 45 p.

Gaston, M.P., and Bennett, E.H., 1979, Geologic map of the Grangeville quadrangle, Idaho: Idaho Bureau of Mines and Geology Geologic Map Series, scale 1:250,000.

Greene, R.C., Walker, G.W., and Corcoran, R.E., 1972, Geologic map of the Burns quadrangle, Oregon: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–680, scale 1:250,000.

Gustafson, E.P., 1978, The vertebrate faunas of the Pliocene Ringold Formation, south-central Washington: Eugene, University of Oregon, Museum of Natural History Bulletin 23, 62 p.

Hanson, C.B., 1973, Geology and vertebrate faunas in the type area of the Clarno Formation, Oregon [abs.]: Geological Society of America Abstracts with Programs, v. 5, no. 1, p. 50.

Hay, R.L., 1962, Origin and diagenetic alteration of the lower part of the John Day Formation near Mitchell, Oregon, *in* Engel, A.E.J., James, H.L., and Leonard, B.F., eds., Petrologic studies—A volume in honor of A.F. Buddington: New York, Geological Society of America, p. 191–216.

———1963, Stratigraphy and zeolitic diagenesis of the John Day Formation of Oregon: University of California Publications in Geological Sciences, v. 42, no. 5, p. 199–262.

Hergert, H.L., 1961, Plant fossils in the Clarno Formation, Oregon: Ore Bin, v. 23, no. 6, p. 55–62.

Hodge, E.T., 1932, Geologic map of north-central Oregon: University of Oregon Geology Publication Series, v. 1, no. 5, 7 p., scale 1:250,000.

Hogenson, G.M., 1964, Geology and ground water of the Umatilla River basin, Oregon: U.S. Geological Survey Water-Supply Paper 1620, 162 p.

Hooper, P.R., 1974, Petrology and chemistry of the Rock Creek flow, Columbia River Basalt, Idaho: Geological Society of America Bulletin, v. 85, no. 1, p. 15–26.

Huntting, M.T., Bennett, W.A.G., Livingston, V.E., Jr., and Moen, W.S., comps., 1961, Geologic map of Washington: Olympia, Washington Division of Mines and Geology, scale 1:500,000.

Jones, R.W., 1982, Early Tertiary-age Kamiah Volcanics, *in* Bonnichsen, Bill, and Breckenridge, R.M., eds., Cenozoic geology of Idaho: Moscow, Idaho Bureau of Mines and Geology, p. 43–52.

Kittleman, L.R., Green, A.R., Hagood, A.R., Johnson, A.M., McMurray, J.M., Russell, R.G., and Weeden, D.A., 1965, Cenozoic stratigraphy of the Owyhee region, southeastern Oregon: Eugene, University of Oregon, Museum of Natural History Bulletin 1, 45 p.

Knowlton, F.H., 1902, Fossil flora of the John Day Basin, Oregon: U.S. Geological Survey Bulletin 204, 153 p.

Lindgren, Waldemar, 1898, The mining districts of the Idaho Basin and

the Boise Ridge, Idaho: U.S. Geological Survey Annual Report 18, pt. 3, p. 617–719.

Luedke, R.G., and Smith, R.L., 1982, Map showing distribution, composition, and age of late Cenozoic volcanic centers in Oregon and Washington: U.S. Geological Survey Miscellaneous Investigations Series Map I–1091–D, scale 1:1,000,000.

Mackin, J.H., 1961, A stratigraphic section in the Yakima Basalt and Ellensburg Formation in south-central Washington: Washington Division of Mines and Geology Report of Investigations 19, 45 p.

Malde, H.E., and Powers, H.A., 1962, Upper Cenozoic stratigraphy of western Snake River Plain, Idaho: Geological Society of America Bulletin, v. 73, no. 10, p. 1197–1219.

Manchester, S.R., 1981, Fossil plants of the Eocene Clarno Nut Beds: Oregon Geology, v. 43, no. 6, p. 75–81.

Merriam, J.C., 1901, A contribution to the geology of the John Day Basin [Oreg.]: University of California Publications, Department of Geology Bulletin, v. 2, no. 9, p. 269–314.

———1918, New mammalia from the Idaho Formation: University of California Publications, Department of Geology Bulletin, v. 10, no. 26, p. 523–530.

Merriam, J.C., and Sinclair, W.J., 1907, Tertiary faunas of the John Day region: University of California Publications, Department of Geology Bulletin, v. 5, no. 11, p. 171–205.

Mitchell, V.E., and Bennett, E.H., 1979, Geologic map of the Baker quadrangle, Idaho: Idaho Bureau of Mines and Geology Geologic Map Series, scale 1:250,000.

Nathan, Simon, and Fruchter, J.S., 1974, Geochemical and paleomagnetic stratigraphy of the Picture Gorge and Yakima Basalts (Columbia River Group) in central Oregon: Geological Society of America Bulletin, v. 85, no. 1, p. 63–76.

Newcomb, R.C., 1958, Ringold Formation of Pleistocene age in type locality, the White Bluffs, Washington: American Journal of Science, v. 256, no. 5, p. 328–340.

———1965, Geology and ground-water resources of the Walla Walla River basin, Washington-Oregon: Washington Division of Water Resources Water-Supply Bulletin 21, 151 p.

Newcomb, R.C., Strand, J.R., and Frank, F.J., 1972, Geology and ground-water characteristics of the Hanford Reservation of the U.S. Atomic Energy Commission, Washington: U.S. Geological Survey Professional Paper 717, 78 p.

Noblett, J.B., 1981, Subduction-related origin of the volcanic rocks of the Eocene Clarno Formation near Cherry Creek, Oregon: Oregon Geology, v. 43, no. 7, p. 91–99.

Oles, K.F., and Enlows, H.E., 1971, Bedrock geology of the Mitchell quadrangle, Wheeler County, Oregon: Oregon Department of Geology and Mineral Industries Bulletin 72, 62 p.

Palmer, A.R., 1983, The Decade of North American Geology 1983 geologic time scale: Geology, v. 11, no. 9, p. 503–504.

Peck, D.L., 1964, Geologic reconnaissance of the Antelope-Ashwood area, north-central Oregon, with emphasis on the John Day Formation of late Oligocene and early Miocene age: U.S. Geological Survey Bulletin 1161-D, 26 p.

Rember, W.C., and Bennett, E.H., 1979, Geologic map of the Pullman quadrangle, Idaho: Idaho Bureau of Mines and Geology Geologic Map Series, scale 1:250,000.

Robinson, P.T., 1975, Reconnaissance geologic map of the John Day Formation in the southwestern part of the Blue Mountains and adjacent areas, north-central Oregon: U.S. Geological Survey Miscellaneous Investigations Series Map I–872, scale 1:125,000.

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

Schmincke, H.-U., 1967, Stratigraphy and petrography of four upper Yakima Basalt flows in south-central Washington: Geological Society of America Bulletin, v. 78, no. 11, p. 1385–1422.

Scott, R.A., 1954, Fossil fruits and seeds from the Eocene Clarno Formation of Oregon: Palaeontographica, v. 96, pt. B, no. 3–6, p. 66–97.

Stirton, R.A., 1944, A rhinoceros tooth from the Clarno Eocene of Oregon: Journal of Paleontology, v. 18, no. 3, p. 265–267.

Strand, J.R., and Hough, M.J., 1952, Age of the Ringold Formation: Northwest Science, v. 26, no. 4, p. 152–154.

Swanson, D.A., 1969, Reconnaissance geologic map of the east half of the Bend quadrangle, Crook, Wheeler, Jefferson, Wasco, and Deschutes Counties, Oregon: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–568, scale 1:250,000.

Swanson, D.A., Anderson, J.L., Camp, V.E., Hooper, P.R., Taubeneck, W.H., and Wright, T.L., 1981, Reconnaissance geologic map of the Columbia River Basalt Group, northern Oregon and western Idaho: U.S. Geological Survey Open-File Report 81–797, 5 sheets, scale 1:250,000.

Swanson, D.A., Wright, T.L., Camp, V.E., Gardner, J.N., Helz, R.T., Price, S.P., Riedel, S.P., and Ross, M.E., 1980, Reconnaissance geologic map of the Columbia River Basalt Group, Pullman and Walla Walla quadrangles, southeast Washington and adjacent Idaho: U.S. Geological Survey Miscellaneous Investigations Series Map I–1139, scale 1:250,000.

Swanson, D.A., Wright, T.L., Hooper, P.R., and Bentley, R.D., 1979, Revisions in stratigraphic nomenclature of the Columbia River Basalt Group: U.S. Geological Survey Bulletin 1457-G, 59 p.

Thayer, T.P., and Brown, C.E., 1966, Local thickening of basalts and late Tertiary silicic volcanism in the Canyon City quadrangle, northeastern Oregon, *in* Geological Survey research, 1966: U.S. Geological Survey Professional Paper 550–C, p. 73–78.

Tullis, E.L., 1944, Contribution to the geology of Latah County, Idaho: Geological Society of America Bulletin, v. 55, no. 2, p. 131–164.

Vallier, T.L., and Brooks, H.C., eds., 1986, Geology of the Blue Mountains region of Oregon, Idaho, and Washington—Geologic implications of Paleozoic and Mesozoic paleontology and biostratigraphy, Blue Mountains province, Oregon and Idaho: U.S. Geological Survey Professional Paper 1435, 93 p.

———1987, Geology of the Blue Mountains region of Oregon, Idaho, and Washington—The Idaho batholith and its border zone: U.S. Geological Survey Professional Paper 1436, 196 p.

Walker, G.W., 1973, Reconnaissance geologic map of the Pendleton quadrangle, Oregon and Washington: U.S. Geological Survey Miscellaneous Investigations Series Map I–727, scale 1:250,000.

———1977, Geologic map of Oregon east of the 121st meridian: U.S. Geological Survey Miscellaneous Investigations Series Map I–902, scale 1:500,000.

———1979, Reconnaissance geologic map of the Oregon part of the Grangeville quadrangle, Baker, Union, Umatilla, and Wallowa Counties, Oregon: U.S. Geological Survey Miscellaneous Investigations Series Map I–1116, scale 1:250,000.

Walker, G.W., Peterson, N.V., and Greene, R.C., 1967, Reconnaissance geologic map of the east half of the Crescent quadrangle, Lake, Deschutes, and Crook Counties, Oregon: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–493, scale 1:250,000.

Waters, A.C., 1961, Stratigraphic and lithologic variations in the Columbia River Basalt: American Journal of Science, v. 259, no. 8, p. 583–611.

Waters, A.C., Brown, R.E., Compton, R.R., Staples, L.W., Walker, G.W., and Williams, Howel, 1951, Quicksilver deposits of the Horse Heaven mining district, Oregon: U.S. Geological Survey Bulletin 969–E, p. 105–149.

Wells, F.G., and Peck, D.L., 1961, Geologic map of Oregon west of the 121st meridian: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–325, scale 1:500,000.

Wright, T.L., Grolier, M.J., and Swanson, D.A., 1973, Chemical variation related to the stratigraphy of the Columbia River Basalt: Geological Society of America Bulletin, v. 84, no. 2, p. 371–385.

2. PALEOCENE(?), EOCENE, AND OLIGOCENE(?) ROCKS OF THE BLUE MOUNTAINS REGION

By GEORGE W. WALKER and PAUL T. ROBINSON[1]

CONTENTS

ABSTRACT

Lower Cenozoic terrestrial rocks of the Blue Mountains region are separable into a lower, possibly Paleocene, sequence of quartzose and feldspathic sandstone and siltstone overlain by a thick sequence of largely volcanogenic deposits, most commonly referred to the Clarno Formation in Oregon and the Challis Volcanics in western Idaho.

The volcanogenic sequence is dominated by several kinds of andesite flows, mudflows, breccia, and moderate to large volumes of andesitic to rhyolitic tuff and tuffaceous sedimentary rocks. Isotopic ages on volcanogenic rocks shown on published maps as part of the Clarno Formation in the Blue Mountains range in age from about 54 Ma to 19 Ma. Assignment of rock units to the Clarno Formation that have been dated at younger than about 37 to 35 Ma are suspect; reexamination of outcrops of those rocks younger than about 37 Ma indicates that although they are petrographically and, presumably, chemically much like rocks of the Clarno Formation they are invariably less altered. Much evidence indicates that volcanism similar to that of the Clarno Formation extended well into the Oligocene and, possibly, into earliest Miocene time and is, therefore, partly coeval with volcanism that produced the John Day Formation.

INTRODUCTION

A heterogeneous assemblage of terrestrial volcanic, volcaniclastic, and epiclastic rocks primarily of Paleocene(?) and Eocene age crop out in widely separated areas in the Blue Mountains of Oregon and in parts of several adjoining provinces. The assemblage typically comprises a basal sequence of terrestrial quartzose and feldspathic sandstone and siltstone overlain by a thick sequence of largely volcanogenic deposits, dominantly of andesitic composition but including some basaltic and rhyolitic rocks. Although these rocks vary lithologically, many of them are similar in kinds and degree of alteration and in structural deformation. These similarities, along with erratically distributed fossil floras and faunas, have been used by various investigators to correlate detached parts of the lower Cenozoic section over many tens or even hundreds of kilometers, with various degrees of precision.

In northeastern Oregon nearly all of these lower Cenozoic volcanogenic rocks have been assigned to the Clarno Formation by most investigators, none of whom has considered this complex assemblage on a regional basis. Although Mendenhall (1909) and Hogenson (1964) included the epiclastic sedimentary rocks in the Clarno Formation, Pigg (1961) placed them in an older, unnamed unit. Shorey (1976) also considered these arkosic sedimentary rocks to be older than the Clarno Formation and referred them to his Herren unit. Because these rocks are distinct in composition from the bulk of the Clarno Formation and appear to be older (Trauba, 1975; Shorey, 1976; Elemendorf and Fisk, 1978), we also treat them herein as a separate unit.

On the basis of an angular unconformity in the Mitchell area (fig. 2.1), Oles and Enlows (1971) divided the volcanogenic sequence into what they termed the "Upper Clarno" and "Lower Clarno" Formations and identified the aggregate sequence as the Clarno Group. Other investigators (Waters and others, 1951; Noblett, 1981) have also recognized unconformities at various localities within the sequence separating sections with slightly different lithologies, types of alteration, or degrees of deformation. However, these unconformities appear to be local features (Swanson and Robinson, 1968), and there is no evidence of a regional unconformity. Therefore, we disagree with this terminology and instead follow the usage of Merriam (1901) in referring to this volcanogenic sequence as the Clarno Formation.

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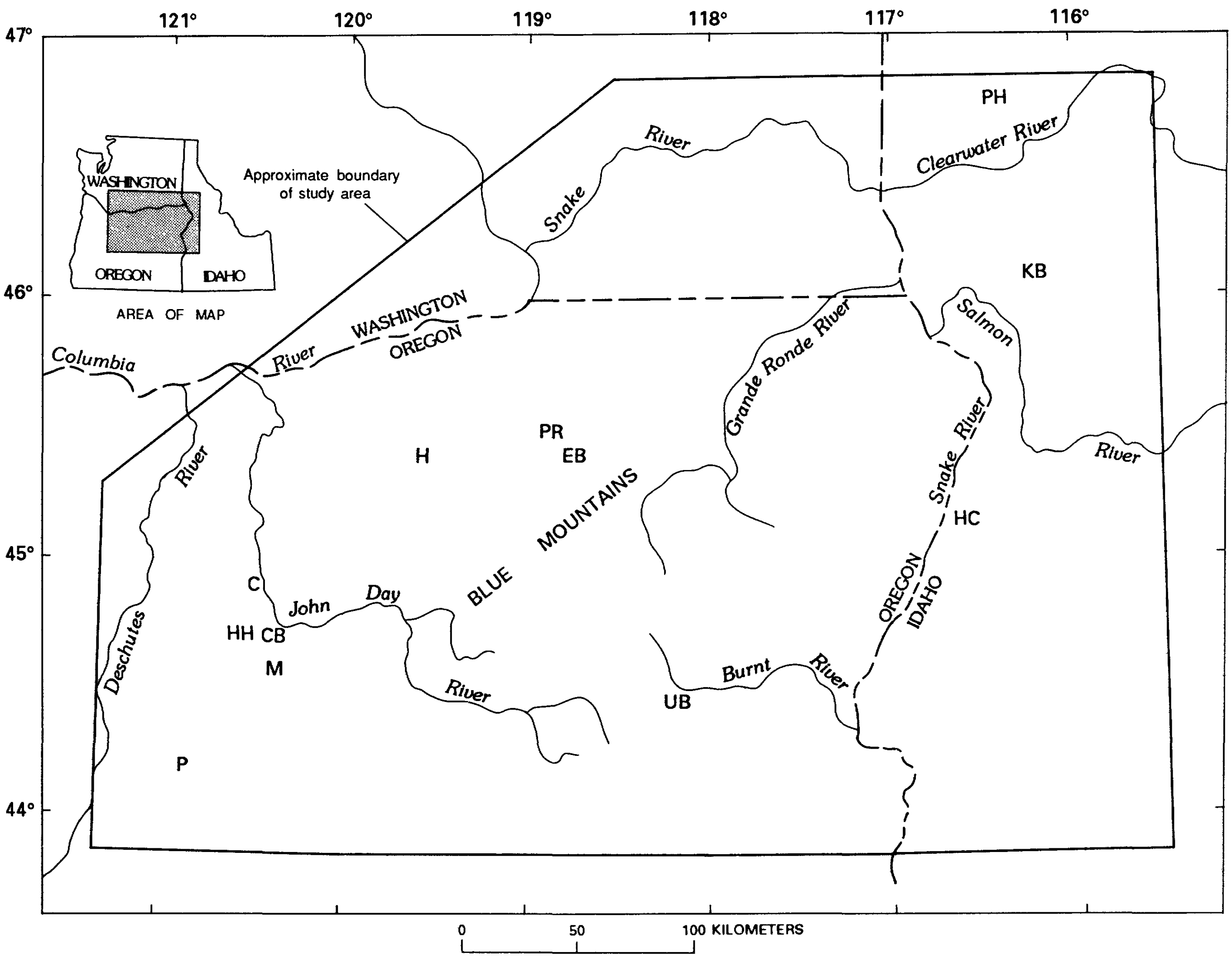
In adjacent parts of western Idaho, two small inliers (kipukas) of volcanic flow, volcaniclastic, and epiclastic(?) rocks surrounded by flows of the Columbia River Basalt Group presumably correlate with parts of the Clarno Formation and, most recently, have been correlated with the Eocene Challis Volcanics of eastern Idaho (Jones, 1982).

Most radiometric ages on samples that have been assigned to the Clarno Formation on the basis of lithology range from 55 to 40 Ma. As discussed below, some ages younger than about 35 Ma suggest possible temporal overlap with parts of the John Day Formation (table 2.1), but many of those ages are suspect. Thus, in this report we consider the Clarno Formation to be chiefly Eocene in age (we actually consider the overall age range to be Eocene and earliest Oligocene?), but if some of these young ages can be substantiated, the unit as presently mapped will have a greater timespan.

A few of these young Clarno-like rocks are less altered and less deformed than those in the type locality of the Clarno Formation, which most workers consider to be near Clarno (or Clarno's Ferry), thus introducing the



INDEX TO LOCALITIES

C	Clarno (Clarno Ferry)	KB	Kamiah Butte
CB	Cherry Creek. Bridge Creek area	M	Mitchell
EB	East Birch Creek	P	Prineville
H	Heppner	PH	Potato Hill
HC	Hells Canyon	PR	Pilot Rock
HH	Horse Heaven mining district	UB	Unity Basin and Reservoir

FIGURE 2.1.—Index map of Blue Mountains region, showing study area and localities mentioned in text.

TABLE 2.1.—*Potassium-argon ages for rocks mapped as the Clarno Formation*

[References: 1, Fiebelkorn and others (1983); 2, Enlows and Parker (1972); 3, Swanson and Robinson (1968); 4, Evernden and others (1964); 5, Walker (1973); 6, Brooks and others (1976); 7, Brown and Thayer (1966). do., ditto]

Sample	General location	Rock type	Material	Age	References
648-628	East of Prineville	Andesite?	Plagioclase	53.7±1.0	1
E-16-77	Mitchell	Andesite	Hornblende	48.9±5.2	1
648-695	North of Mitchell	do.	do.	47.0±9.5	1
W-3-58	South of Mitchell	do.	Hornblende Whole rock	46.1±0.4 41.6±0.9	2
KFO-1702B	Mitchell	Basalt	Whole rock	46.1±0.1	2
KFO-1112	do.	Andesite	Whole rock	44.0±0.6	2
648-657	East of Prineville	?	Hornblende	42.9±5.4	1
DAS-67-80	Horse Heaven	Rhyolite	Sanidine	42.1±0.8	3
DX-1	East of John Day	Basalt	?	41.7±6.0	1
B-1	Northeast of John Day.	Andesite	Whole rock	41.2±0.4	1
KA818	Mitchell	do.	do.	37.5	4
KA824A	North of Mitchell	Bentonite	Sanidine	36.5±0.9	4
M-859	Mitchell	Diabase	Whole rock	34.3±0.9	2
KA1204	Northwest of Mitchell.	Tuff	Plagioclase	34.0	4
KFO-901	Mitchell	Andesite	Whole rock	33.7±3.1	2
10-6-78-2	Smith Rock	Basalt	do.	30.8±0.5	1
EMT-11	Mitchell	Diabase	do.	29.3±0.4	2
SC-1-70	Upper Grande Ronde.	Andesite	Plagioclase	28.8±0.8	5, 1
UB-2-70K	Unity	Dacite (clast)	Hornblende Plagioclase	19.6±8.0 19.5±0.6	6, 7, 1

fundamental problem of what should be included in the formation. Should it include all those rocks assigned by different workers to the Clarno Formation in northeastern Oregon and the questionable age-equivalent rocks in western Idaho assigned to the Challis Volcanics? Should it include only those rocks clearly dated by whatever means as Paleocene and Eocene? Having no direct answers to these questions, we herein arbitrarily include in the Clarno Formation only those bodies of volcanic, volcaniclastic, and sedimentary rocks within the region that meet one or more of the following criteria: (1) They represent rocks of the type locality of the Clarno Formation (Merriam, 1901) or are directly traceable into those rocks(2) they are well established as either Eocene or earliest Oligocene(?) in age, either by diagnostic fossils or by isotopic ages; and (3) they lie either conformably or unconformably beneath rocks clearly identifiable with the uppermost Eocene(?) to lower Miocene John Day Formation and overlie pre-Cenozoic rocks. All other bodies of rock with "typical" Clarno Formation characteristics that do not meet these criteria are looked at with some skepticism; isolated bodies that are near the type locality at and near the hamlet of Clarno (originally Clarno's Ferry; fig. 2.1), evidently are time-stratigraphic correlatives of the Clarno Formation, whereas others farther east and west of the type locality are suspect. Some of these Clarno-like rocks have been radiometrically dated as Oligocene and even early Miocene in age but show relations with the John Day Formation or with other units that make them suspect; others are considerably less altered than typical Clarno Formation rocks (see chap. 3).

DISTRIBUTION

The epiclastic sedimentary rocks of Paleocene(?) and early Eocene age are exposed over an area of approximately 80 km^2 at two localities—south of Heppner, Oreg., and along East Birch Creek southeast of Pilot Rock (fig. 2.2). South of Heppner, the rocks form a northeast-trending body exposed along the crest of the Blue Mountains uplift and are conformably(?) overlain by volcanogenic rocks of the Clarno Formation.

The largest exposures of the Clarno Formation occupy an arcuate area of slightly more than 4,000 km^2 mostly along and to the south of the axis of the Blue Mountains uplift at and near the western margin of the Blue Mountains province in Crook, Jefferson, and Wheeler Counties, Oreg. (fig. 2.2). A smaller northeast-trending body of about 400-km^2 area lies along the axis of the Blue Mountains uplift in southern Morrow County. Smaller inliers of Clarno rocks are surrounded by younger flows and sedimentary rocks in areas that are deeply dissected by the Deschutes River and its tributaries in the southwesternmost tip of the Deschutes-Umatilla Plateau province (see fig. 1.2). Detached bodies of correlative rocks are found sporadically southeast of the axis of the Blue Mountains uplift, particularly in areas of considerable structural relief. The largest of these detached bodies, in central Grant County, covers about 600 km^2, and another northwest-trending elongate body that crosses the Grant-Baker County line covers about 500 km^2. The isolated small inliers (kipukas) at Potato Hill, Latah County, and Kamiah Butte, Idaho County, Idaho, represent erosional highs of volcanic, volcaniclastic, and epiclastic(?) rocks of early Cenozoic(?) age surrounded by flows of the Columbia River Basalt Group. Neither inlier is more than a few tens of square kilometers in maximum extent, and neither has been studied in detail.

Some of these rocks in isolated bodies away from the type localities of the Clarno Formation in Oregon and the Challis Volcanics in Idaho have been assigned ages either by fossils or by isotopic dating, whereas others have not. Isotopic dating of some of these isolated bodies that are considered to be correlative with either the Clarno Formation or the Challis Volcanics indicates that they may be related to later periods of volcanism. Only additional isotopic dating and mapping can resolve these uncertainties.

Distribution of the Paleocene(?) to lowermost Oligocene(?) rocks in the subsurface beneath the very widespread Miocene and younger flows and sedimentary rocks of the region is virtually unknown. Lipman and others (1972, fig. 2a) interpreted the distribution to be extensive and relatively continuous throughout the Pacific Northwest, as inferred from widely separated outcrop areas. However, great variations in thickness and rapid thinning away from volcanic centers indicates that these rocks are probably erratically distributed and may be quite discontinuously distributed on pre-Cenozoic terranes. Rocks that are lithologically similar and appear to be partly the same age as the Clarno Formation are exposed about 100 km south of the Blue Mountains near Paisley, Oreg. (Muntzert, 1969; Walker, 1980, p. 4), and a little farther south in the Warner Range of northeastern California. Also, the older part of the Fisher Formation of the western Cascade volcanic sequence (Peck and others, 1964) shows some lithologic similarities to, as well as differences from, rocks of the Clarno Formation; and K-Ar ages indicate some apparent temporal overlap (Lux, 1982; Fiebelkorn and others, 1983) with younger parts of the Clarno Formation.

STRATIGRAPHY

Paleocene(?) to lowermost Oligocene(?) rock assemblages in northeastern Oregon form a thick sequence of largely volcanogenic deposits mostly of andesitic composition, which locally includes a basal section of terrestrial, epiclastic quartzose and feldspathic fine-grained sedimentary rocks (fig. 2.3). In the Blue Mountains province (see fig. 1.2), this sequence rests unconformably on many different types of rocks, including Cretaceous marine sedimentary rocks, Triassic and Jurassic sedimentary and volcanic rocks, ophiolite, Paleozoic(?) and Mesozoic granitic to gabbroic intrusive rocks, and Paleozoic sedimentary, metasedimentary, and metamorphic rocks. The sequence is overlain by Eocene(?), Oligocene, and Miocene rocks of the John Day Formation or, in some places, by flows of the Columbia River Basalt Group. Contact relations with the overlying John Day Formation are not fully resolved (see chap. 3). In most places the contact is clearly unconformable (Waters and others, 1951; Swanson and Robinson, 1968; Oles and Enlows, 1971; Fisher and Rensberger, 1972; Robinson, 1975), but in a few localities the Clarno and John Day apparently exhibit some temporal overlap, in which the youngest isotopic ages of rocks assigned to the Clarno Formation postdate the oldest ages determined on rocks assigned to the John Day Formation. Misidentification of the isotopically dated units is the most likely explanation for this apparent overlap, inasmuch as no evidence has been found to suggest interfingering of the Clarno and John Day Formations.

The terrestrial, epiclastic sedimentary rocks exposed along the axis of the Blue Mountains uplift south of Heppner are probably the oldest part of the Cenozoic section, as suggested by stratigraphic relations and available age data. However, the absence of regionally extensive marker horizons in the lower part of the section, repetition of some lithologic types in different parts of the section, and considerable relief on the

underlying unconformity surface (Fisher, 1967; Oles and Enlows, 1971, p. 13; Thayer and Brown, 1973) make it nearly impossible to firmly establish which part of the assemblage is the oldest. In the Heppner area, the rocks consist of bedded and folded, well-sorted, fine- to medium-grained, quartzose and feldspathic, micaceous sandstone, as well as micaceous and carbonaceous siltstone and shale. The absence or near-absence of volcanic debris in the fine- to medium-grained, well-sorted sedimentary rocks indicates that they were derived from granitic and, possibly, sedimentary, metasedimentary, or metamorphic source rocks exposed in a terrane with little to moderate relief. South of Heppner they are conformably overlain by andesitic volcanic and volcaniclastic rocks typical of the Clarno Formation as exposed in adjoining areas.

The total thickness of the sedimentary rocks is not known, but it is estimated to range from about 365 to 455 m for the section near Willow Creek 25 to 30 km southeast of Heppner (Pigg, 1961). Shorey (1976) estimated the thickness of these sedimentary rocks, which he identified as his Herren unit, at "slightly less than 2,000 feet" or about 600 m. Whether significant thicknesses of these lithologically distinct rocks are present in areas far removed from Willow Creek is not known. Pigg (1961, p. 58) indicated that pre-Clarno post-Cretaceous sedimentary rocks were encountered in deep oil test holes, one in an area about 50 to 60 km west of the Willow Creek area and another as much as 170 km south of the area. His description of the volcanic and sedimentary rocks penetrated in these test holes indicates that they differ significantly from the quartzose and feldspathic sedimentary rocks south of Heppner and are nearly identical to tuff, claystone, and mudstone typical of the Clarno Formation. None of the sedimentary rocks in these test holes apparently has been dated; thus, if they predate rocks typical of the Clarno Formation and are of Paleocene age, they represent a significantly different facies from the facies represented by the sedimentary rocks southeast of Heppner.

The Clarno Formation is a very thick sequence that comprises extremely diverse volcanic, volcaniclastic, and sedimentary rocks and that apparently overlies the Paleocene(?) and lower Eocene epiclastic sedimentary rocks. The sequence is dominated by several kinds of andesite flows, mudflows, and breccia; it also contains moderate to large volumes of andesitic to rhyolitic volcaniclastic rocks and tuffaceous sedimentary rocks. The aggregate thickness is more than 1,825 m, although in no one area is the total section displayed.

Two thick sections of the Clarno Formation are exposed near the axis of the Blue Mountains uplift, one in the Horse Heaven mining district of Jefferson County, Oreg., and the other a few kilometers to the east between Cherry and Bridge Creeks astride the Jefferson-Wheeler County line. The section near Horse Heaven (Waters and others, 1951) is more than 1,768 m thick. Presumably, the section is stratigraphically above the terrestrial sedimentary rocks exposed south of Heppner, although this relation has not been demonstrated in the field. Waters and others (1951, p. 112–114) subdivided this thick section into four units, as follows (fig. 2.3):

Unit 1.—A basal unit, 183 m thick, of platy andesite flows interbedded with layers of varicolored clay. The base of this unit is not exposed.

Unit 2.—An overlying unit, 411 m thick, of tuff, tuffaceous claystone, volcanic mudflow deposits, and a few thin lava flows, all of andesitic composition.

Unit 3.—About 533 m of tuffaceous claystone containing a few beds of coarse-grained tuff and a few andesite flows.

Unit 4.—White tuff, largely rhyolitic in composition, at least 640 m thick, with some interlayered andesite flows. Quartz-bearing basaltic andesite flows are present in this part of the section and apparently extend eastward into adjoining areas (see unit 5 below).

The section between Cherry Creek and Bridge Creek, which is about 1,646 m thick, was subdivided by Noblett (1981), first, into a lower sequence more than 1,463 m thick and an upper sequence locally from 152 to 182 m thick and, second, into smaller units dominated either by flows or by volcaniclastic and epiclastic deposits. The lower and upper sequences are separated by an angular unconformity and by a saprolite 3 to 6 m thick developed in the lower unit. Noblett's secondary divisions, from bottom to top and somewhat generalized, are:

Unit 1.—A 42-m-thick unit composed of mudflow (lahar) deposits and conglomeratic sandstone containing andesite fragments. The base of this unit is not exposed.

Unit 2.—About 457 m of dominantly plagioclase-, augite-, and hypersthene-phyric andesite flows.

Unit 3.—An overlying 30-m-thick unit of bouldery andesitic mudflow.

Unit 4.—More than 610 m of tuffaceous fluvial and lacustrine sedimentary rocks, locally containing fossil plants, and a few local thin andesite flows.

Unit 5.—A sequence of aphyric olivine-, augite-, and hypersthene-bearing andesite flows with an aggregate thickness of more than 305 m. In the middle part of this unit are distinctive basaltic andesite flows containing quartz xenocrysts probably correlative with those in unit 4 of the Horse Heaven mining district.

Unit 6.—About 100 m of tuffaceous sandstone and plagioclase-, augite-, and hypersthene-phyric andesite and basaltic andesite flows.

These lower units were broadly(?) folded, and a 6-m-thick saprolite was developed before deposition of the overlying units.

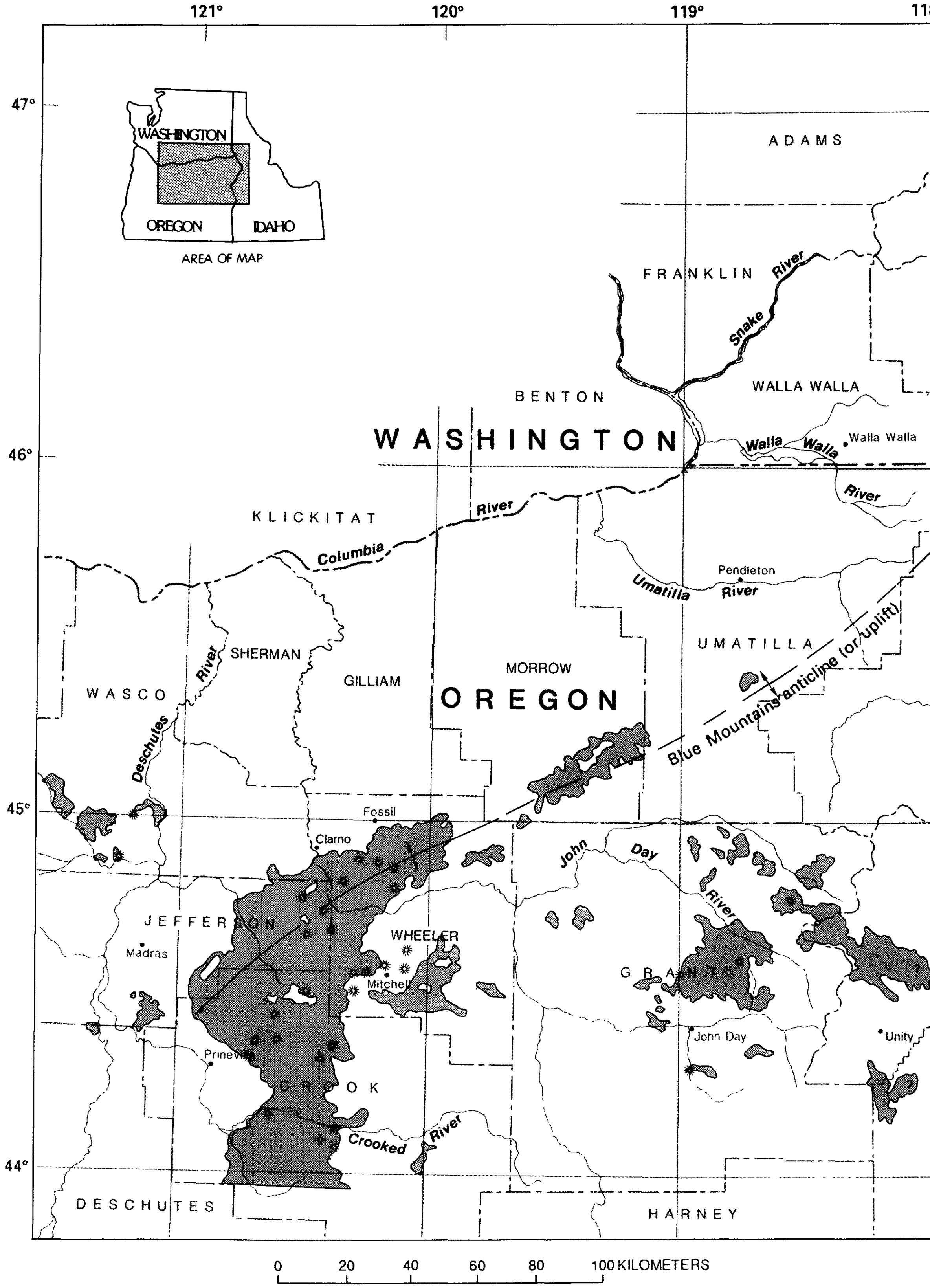
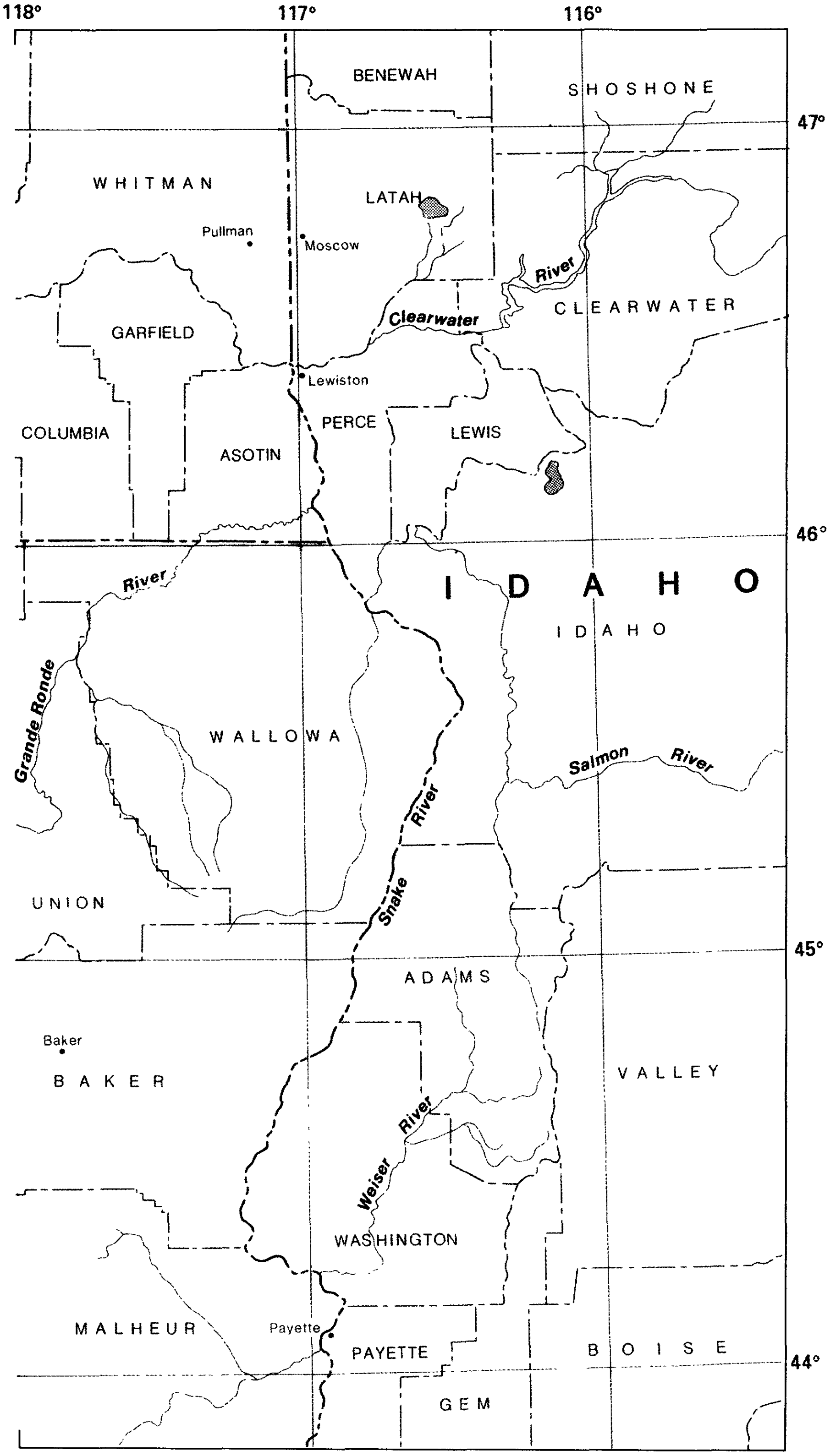


FIGURE 2.2. —Approximate outcrop area of Paleocene(?), Eocene, and



EXPLANATION

Major vents for Eocene volcanic rocks

Outcrop area — Queried where uncertain

Anticline (or uplift) — Dashed where approximately located. Arrow on axis shows direction of plunge

Distribution of outcrops from Brown and Thayer, 1966; Waters, 1968a, b; Swanson, 1969; Walker, 1973; Bond, 1978.

lowermost Oligocene(?) rocks in Blue Mountains region and adjacent areas.

Unit 7.—A fresh pilotaxitic basaltic andesite, 30 m thick, overlying the unconformity.

Unit 8.—A hornblende andesite in the form of a bulbous dome more than 122 m thick but with restricted lateral extent.

Unit 9.—An uppermost aphyric flow, 18 m thick, of basaltic andesite.

Another section, about 1,800 m thick, has been described in the Mitchell area, where Oles and Enlows (1971) subdivided what they called the Clarno Group into a lower unit dominated by thick andesite flows interlayered with varicolored tuffaceous sedimentary rocks and an unconformably overlying unit dominated by andesite flows, mudflows, and lacustrine and fluvial tuffaceous

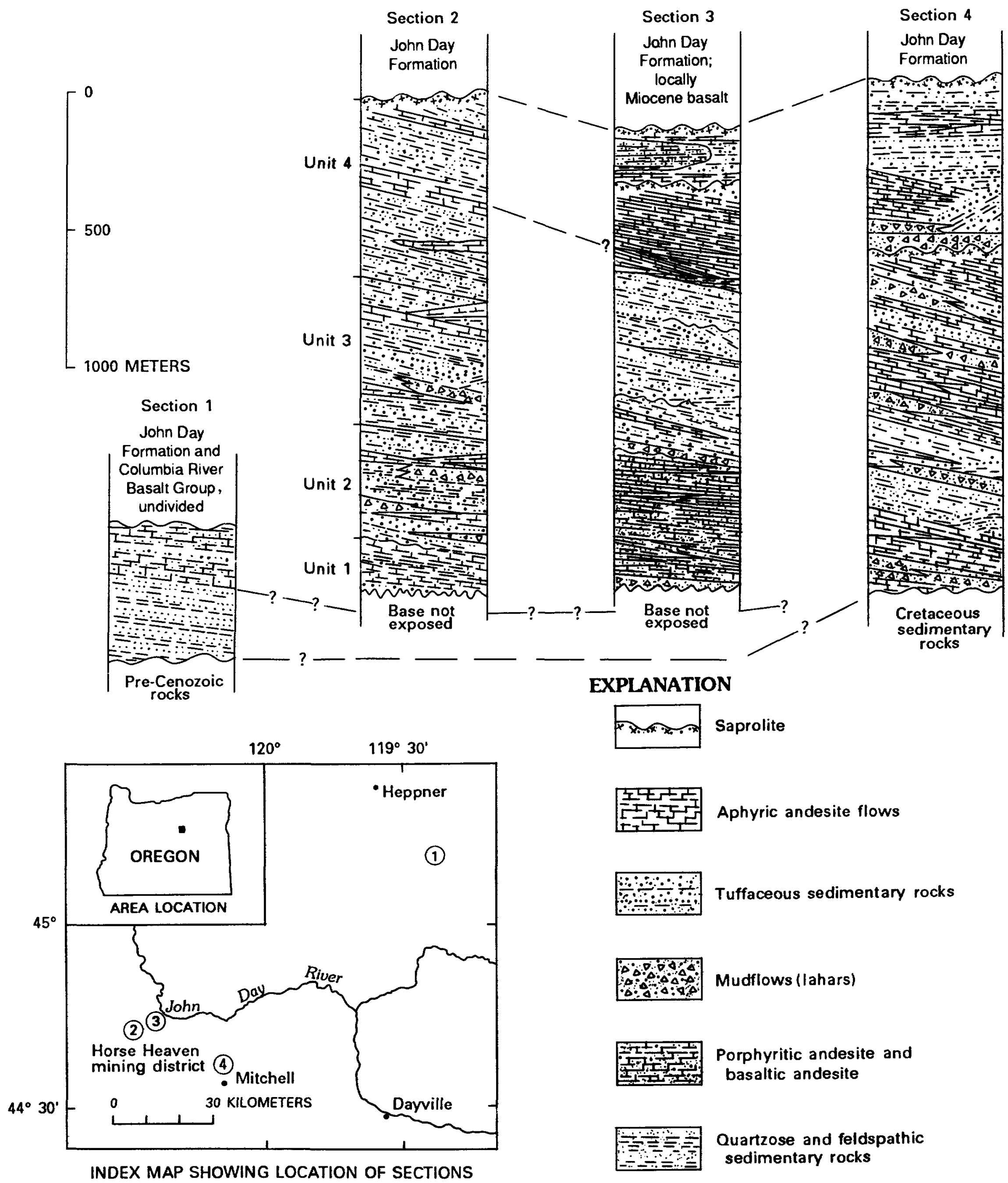


FIGURE 2.3.—Generalized sections of Paleocene(?), Eocene, and lowermost Oligocene(?) rocks. Section 1 from Willow Creek area southeast of Heppner (modified from Pigg, 1961); section 2 from Horse Heaven mining district (Waters and others, 1951); section 3 from Cherry Creek-Bridge Creek area (Noblett, 1981); section 4 from Mitchell area (modified from Taylor, 1960, and Oles and Enlows, 1971).

sedimentary rocks. According to Oles and Enlows (1971, p. 14), rocks of the lower unit were affected by deformation that folded underlying Cretaceous rocks, whereas those of the upper unit were not.

In the northwest corner of the Mitchell quadrangle, near the hamlet of Clarno, Taylor (1960) found the Clarno Formation to be considerably thinner, about 500 to 600 m, and to contain a much higher percentage of volcaniclastic sedimentary rocks. A lower unit there is composed of interlayered andesitic laharic deposits, cliff-forming volcanic conglomerate, and sparse andesitic and basaltic lava; it is conformably overlain by a sequence of tuffaceous sedimentary rocks, air-fall tuff, welded ash-flow tuff, and interlayered basalt and andesite flows. Taylor (1960) recognized basalt and basaltic andesite flows containing quartz xenocrysts in both his lower and upper units, in contrast to the sections at Horse Heaven and in the Cherry Creek-Bridge Creek area, where quartz-bearing mafic flows were reported only in the upper part of the section.

The measured sections—one near Horse Heaven, another between Cherry and Bridge Creeks, and a third near Mitchell—exhibit some of the many lithologic variations in the Clarno Formation, but they are individually and even collectively unrepresentative of the entire sequence. Probably the most characteristic aspect of the volcanogenic part of the formation is its variation from place to place, with coarse-grained, near-vent facies, including vent and flow breccia, volcanic mudflows (lahars), and volcanic conglomerate, grading outward from the many separate volcanic centers into finer grained facies. Intermingled with the andesitic flows and clastic rocks are rhyolite and dacite domes, flows and flow breccia, small intrusions, and beds of coarse-grained near-vent pumiceous ejecta. Most of the rhyolites and dacites are flow-banded and aphyric, but locally they contain phenocrysts of sanidine, quartz, and altered oligoclase. In several widely separated areas, including southwest of Clarno near Current Creek and east of Prineville, Swanson (1969) showed that the Clarno Formation also contains many layers of pumice lapilli tuff, most of which he attributed to an ash-flow origin. The ash-flow tuff layers are dacitic and rhyolitic in composition and contain plagioclase and sanidine in an originally glassy matrix, now mostly devitrified and altered to clay minerals and feldspar.

Complexly intermixed with the volcaniclastic rocks are numerous, commonly local, small-volume flows and flow breccias of plagioclase-, clinopyroxene-, orthopyroxene-, and some olivine-phyric, pilotaxitic to hyalopilitic basaltic andesite. Some hornblende-phyric andesite flows and breccias are locally present.

Many of the flows and breccias are separated by thin, discontinuous layers of saprolite, and more continuous saprolite layers are present locally in the uppermost part of the formation; however, in some places the overlying John Day Formation rests disconformably(?) on unweathered rocks of the Clarno Formation. Many of the different lithologic phases of the Clarno Formation are complexly intermingled and commonly separated by local unconformities. Rarely can individual units be traced for more than 1 or 2 km.

Most of these rocks have been diagenetically and deuterically altered and, locally, subjected to low-grade metamorphism from deep burial in a region of slightly elevated heat flow (Riccio, 1978) and from thermal effects adjacent to the numerous domes, plugs, and small intrusions. The original volcanic glass in these rocks is mostly devitrified and further crystallized into various secondary minerals, and most of the primary minerals are variously altered to smectite, zeolites, calcite, secondary silicic minerals, and, locally, chlorite. Iron-rich hornblende commonly has been completely altered to hematite and an unidentified opaque mineral, and hematite fills fractures and pores in much of the andesite.

Although these alteration features are not ubiquitous, they are characteristic of most of the formation. Where the rocks are unaltered or only slightly altered, they are difficult, if not impossible, to distinguish from similar plagioclase-, pyroxene-, hornblende-, and olivine-phyric flows and breccias of younger age. Therefore, radiometric ages are probably the most reliable means of identifying rocks of the Clarno Formation in areas widely separated from the type locality near Clarno.

Rocks exposed in the inliers at Potato Hill and Kamiah Butte in western Idaho also present problems in regional correlation. The inlier at Kamiah Butte consists of comparatively fresh, commonly trachytic augite and olivine(?) andesite flows and flow breccias, both porphyritic and aphyric, whereas the Potato Hill inlier largely comprises well-indurated and altered breccia and conglomerate(?) deposits that consist of poorly sorted angular fragments of fine-grained sedimentary rocks, gneiss, sheared quartzite, and welded tuff, as well as angular to subangular quartz and feldspar. Thin veinlets of quartz cut the clastic rocks at Potato Hill, and minor epidote and chlorite replace original constituents. The lithology and comparative freshness of the andesite at Kamiah Butte, and its stratigraphic position beneath flows of the Columbia River Basalt Group, certainly permit correlation with similar rocks in the Clarno Formation and the Challis Volcanics. Whether the clastic rocks at Potato Hill are correlative with rocks of either of those formations is questionable because even though they are lapped by flows of the Columbia River Basalt Group, their induration, alteration, and quartz veining are similar to those of the Mesozoic volcaniclastic and sedimentary rocks exposed in Hells Canyon, Idaho, nearly 150 km

to the southeast. However, angular fragments of welded tuff contained in these rocks suggest a Cenozoic rather than Mesozoic age. None of the rocks in either inlier has been dated radiometrically.

AGE

The Paleocene(?) to Oligocene(?) rocks rest unconformably on various Mesozoic or older rocks and are overlain by mainly Oligocene and younger rocks representing several formations. Relations between these lower Cenozoic rocks and overlying rocks, composing mainly the John Day Formation, are not fully resolved, but in most areas the formations are separated by a fairly widespread unconformity and a saprolite zone (regolith) developed on the Clarno Formation (Waters and others, 1951; Waters, 1954; Peck, 1964; Swanson and Robinson, 1968; Oles and Enlows, 1971).

Knowlton (1902) originally assigned a late Eocene age to fossil plants from some of the arkosic sandstone, carbonaceous siltstone, and shale beds south of Heppner and near Pilot Rock on East Birch Creek. However, R.W. Brown (oral commun., 1955; see Hogenson, 1964) considered the fossil plants in these beds to be Eocene or, possibly, Paleocene in age. A reevaluation of the East Birch Creek fossil flora collected from sandstone and shale beds indicated to Elemendorf and Fisk (1978) an age older than typical Clarno Formation floras, that is, Paleocene or early Eocene. Floras collected from the Clarno Formation at localities on Cherry, Current, and Bridge Creeks near and northwest of Mitchell (fig. 2.1), as well as near Clarno, apparently are somewhat younger and have been assigned an Eocene or early Oligocene age (Knowlton, 1902; p. 102–103; Chaney, 1938; Arnold, 1952, p. 68–72; Scott, 1954; Hergert, 1961; McKee, 1970). Apparently, the only isotopic ages on these sedimentary rocks were obtained on white detrital mica separated from the arkosic sedimentary rocks. Two samples from the Herren unit yielded white mica with ages of 85.7 and 84.1 Ma (Heller and others, 1985).

Isotopic ages obtained on rocks of the Clarno Formation range mostly from 55 to 40 Ma, whereas rocks of the John Day Formation range in age from about 37 to 21 Ma. Eight K-Ar ages on rocks that have been assigned to the Clarno Formation by several investigators are younger than 37 Ma (table 2.1), the approximate age of the lowest ash-flow sheet of the John Day Formation[2] (Swanson and Robinson, 1968; Robinson, 1975). In addition, several undated bodies of Clarno-like rocks south and southeast of Prineville (fig. 2.1) appear to postdate the Clarno Formation; these rocks are briefly described in chapter 3. Two of the anomalously young K-Ar ages are on dikes in the Mitchell area that are compositionally similar to basalts of the John Day Formation (Taylor, 1981), and one is so young that the rocks involved clearly have been misidentified as the Clarno (table 2.1). This young age (19.5±6 Ma) is from a comparatively fresh hornblende-porphyritic andesite or mafic dacite exposed in roadcuts just east of the Unity Reservoir (fig. 2.1), Baker County, Oreg. Although this young unit is superficially similar in lithology to some Clarno-age rocks, it shows none of the alteration that has produced secondary clays, zeolites, silica minerals, and calcite in the Clarno Formation. Thus, this unit should not be considered part of the Clarno Formation, nor even correlative with the youngest part of the John Day Formation.

Most of the other young Clarno ages are also suspect because they are on whole-rock specimens of somewhat altered lava (Enlows and Parker, 1972) or on feldspar separates from highly weathered bentonite (Evernden and others, 1964). The effect of such alteration on whole-rock ages is shown by sample W-3–58 (table 2.1). A hornblende separate from this rock yielded a K-Ar age of 46.1±4 Ma, whereas the whole-rock sample yielded an age nearly 5 m.y. younger.

Not only are these ages suspect because of the materials dated, but several conflict with ages on fossil floras and faunas. As pointed out by Manchester (1981, p. 77), a K-Ar age of 34.0 Ma obtained on tuff from the Clarno "Nut Beds" of local usage conflicts with vertebrate fossil dating by Stirton (1944) and Hanson (1973) of about middle Eocene age, or about 48 Ma. According to Hanson (see Manchester, 1981, p. 77), the "Nut Beds" may represent a part of the lower part of the Clarno Formation that was uplifted by folding before deposition of mudflows and tuff of the upper part of the Clarno.

Thus, some and probably all of the ages in the range of 35 to 28 Ma for rocks that have been identified as the Clarno on the basis of lithology should be considered suspect and reevaluated as additional mapping and age data are obtained. It seems reasonable, however, to suppose that Clarno-type volcanism may have persisted locally within the Blue Mountains province until after deposition of the John Day Formation had already begun. Rocks of the Clarno Formation were all erupted locally, whereas those of the John Day Formation were derived mostly from vents west of the present-day outcrops (Robinson and Brem, 1981).

The isotopic ages further imply the possibility of two major hiatuses, one at about 40 to 38 Ma and another, best demonstrated in John Day Formation ages, at about 35 Ma (fig. 2.4). Whether these apparent breaks in the geologic record have any significance in terms of regional

[2] The original published age was 36.4 Ma; the 37.4-Ma age was obtained by using the original analytical data and applying abundance and decay constants acceptable in 1982.

volcanism and resulting stratigraphy is not known. The older break may represent a decline of calc-alkaline volcanic activity near the end of Clarno time, or it may result from insufficient data. The younger break probably reflects incomplete age dating of the lower part of the John Day Formation.

Future investigators would be well advised to question and reevaluate bodies of rocks mapped as the Clarno Formation that are far removed from the type locality and that are not characterized by secondary minerals denoting low-grade metamorphism. Many of the bodies of rock assigned to the Clarno Formation that lie east of U.S. Highway 395 are of particular concern in this regard.

STRUCTURE

Structural and stratigraphic relations among the diverse Paleocene(?) to lowermost Oligocene(?) rock types, as well as distributional relations with overlying Oligocene and Miocene rocks, demonstrate that early Cenozoic structural trends developed before deposition of these younger rocks and that subsequent deformation followed these earlier structural trends.

The distribution of rocks younger than the Clarno Formation indicates that during the early Tertiary a series of northeast-trending structural highs formed essentially along the axis of what is now the Blue Mountains uplift (Rogers, 1966; Fisher, 1967; Taylor, 1977). This early Tertiary uplift (or antiform) includes northeast-trending fold elements that appear to have resulted from northwest-southeast compressional forces (Taylor, 1977, 1981), but it also includes some faults and is marked by many coalescing constructional volcanic edifices. Many of the initial dips from these coalescing constructional highs are to the northwest and southeast, tending to emphasize the antiformal nature of the northeast-trending structural high. How much of the structural high is tectonic and related to arching of the Blue Mountains uplift (or antiform) and how much to the coalescing of numerous volcanic piles along the axis of the arch is not known. However, a significant tectonic uplift in early Oligocene time can be demonstrated from the distribution of and stratigraphic relations within the John Day Formation (Robinson, 1975; Woodburne and Robinson, 1977). A few folds east of Mitchell trend west-northwest at a considerable angle (50°–60°) to the dominant northeastward trend, indicating either local divergence in the regional stress regime or possible differences in the age of development of these divergent structures. The distribution of rock units of different ages associated with both the northeast- and west-northwest-trending folds indicated to Fisher (1967, p. 120) that the folds predate the beginning of John Day deposition—either pre-Oligocene or earliest Oligocene—and probably developed after the Cretaceous. A more precise lower age limit may be indicated by the characteristics of the Paleocene(?) and lower Eocene sedimentary rocks south of Heppner. The fairly uniform and locally thin bedding of these rocks, moderate to good sorting, and presence of carbonaceous shale and coaly lenses, as well as an absence of volcanic debris, indicate a terrain of little physiographic or structural relief. Thus, these structures are apparently of Eocene or earliest Oligocene age.

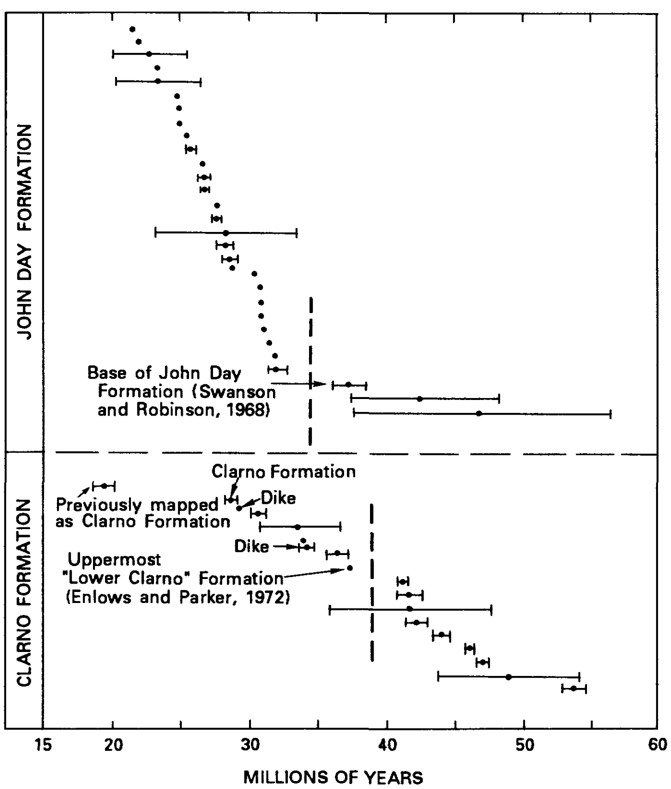


FIGURE 2.4.—K-Ar ages (dots, including analytical errors, where available, shown by horizontal bars) for the Clarno and John Day Formations. Within each formation, dated samples are not necessarily arranged according to stratigraphic position. Age data from Fiebelkorn and others (1983). Vertical heavy-dashed lines indicate apparent age gaps.

COMPOSITION

Considerable chemical data are available on the volcanic and volcaniclastic rocks of the Clarno Formation. Some chemical data also are available on intrusions that represent sources of some of the volcanic rocks. Little is known, however, of the chemistry of the terrestrial quartzose and feldspathic sedimentary rocks, which apparently represent the earliest part of the Cenozoic section.

From the mineralogic composition of their coarse-grained facies, the terrestrial sedimentary rocks appear to represent detritus derived largely from a fairly homogeneous granitic to quartz dioritic or, possibly, metasedimentary or metamorphic source area in which quartz, feldspar, and white mica were major constituents. Minor constituents include apatite, zircon, garnet, sillimanite, staurolite, hornblende, and andalusite (Pigg, 1961; Trauba, 1975). The sandstone, composed mostly of angular to partly subrounded grains, is moderately well indurated and cemented by calcium carbonate. Finer grained beds contain large amounts of carbonaceous material, some altered to lignite and bituminous coal; the coal beds are invariably thin and contain large amounts of clay-size detritus. Sedimentary features indicate that deposition was at least partly in a deltaic environment.

Rocks of the Clarno Formation vary noticeably in composition and form a diverse assemblage that compositionally ranges from basalt to rhyolite (fig. 2.5), with a great preponderance of andesite and basaltic andesite. Some parts of the section are characterized by plagioclase-, augite-, hypersthene-, or hornblende-phyric flows, breccia, lahars, and some tuff, and other parts by abundant aphyric flows. Some of the lava flows are characterized by sparse quartz xenocrysts (Calkins, 1902; Waters and others, 1951; Noblett, 1981). Still other parts of the section are dominated by dacitic, rhyodacitic, and rhyolitic ash-flow tuff, pumice lapilli tuff, and mudstone, siltstone, sandstone, and conglomerate, mostly composed of volcanic debris. A few conglomerates in the eastern part of the Blue Mountains province that have been assigned to the Clarno Formation contain rounded pebbles and cobbles of metasedimentary rocks, mostly quartzite, as well as some metamorphic and igneous rock types.

Inasmuch as the chemistry of these volcanic rocks has been published elsewhere (Calkins, 1902; Oles and Enlows, 1971; Rogers and Novitsky-Evans, 1977; Noblett, 1981) or released in dissertations (for example, Wilson, 1973), only a brief summary will be reported here. Conspicuous chemical characteristics of the Clarno Formation are: (1) Basaltic andesite and andesite (SiO_2, ~54–62 weight percent) are volumetrically predominant and accompanied by lesser amounts of basalt (SiO_2, $<\sim$52 weight percent) and rhyolite or rhyodacite (SiO_2, ~70–76 weight percent); (2) the assemblage is mostly calc-alkaline, although some rocks appear to be transitional to tholeiite (Rogers and Novitsky-Evans, 1977); and (3) there appears to be little relation between variations in rock chemistry and different stratigraphic units within the assemblage, although rhyolitic and rhyodacitic volcaniclastic and tuffaceous sedimentary rocks are more abundant high in the section. A histogram presented by Rogers and Novitsky-Evans (1977) shows that 54 percent of the rocks they analyzed contain from 56 to 63 weight percent SiO_2; this range may indicate actual volumes of rocks with different SiO_2 contents or simply reflect sampling bias.

Trace-element investigations of rocks of the Clarno Formation are of limited extent and pertain principally to abundances of lithophile elements (Rogers and Ragland, 1980; Noblett, 1981) in both porphyritic and nonporphyritic intermediate to mafic flows. Rogers and Ragland (1980) have examined abundances of Rb, Sr, Y, Zr, Nb, Ba, V, Cr, and Ni, and Noblett (1981) examined abundances of Rb, Sr, Y, Zr, Nb, and Ni. Abundances of K, Rb, and Ba in andesite of the Clarno Formation are generally greater than those in andesite from continental margins, and Rogers and Ragland (1980) inferred that these rocks may have formed on a crust that was intermediate or transitional between oceanic and continental. They considered, further, that this assemblage of calc-alkaline rocks of the Clarno Formation most likely formed by partial melting of mantle or modified mantle material. Noblett (1981), however, considered the porphyritic lava to have formed as hydrous melt that rose from a subducting slab and that interacted with the overlying mantle. The nonporphyritic lava presumably formed by partial melting of anhydrous quartz eclogite with little or no subsequent crystal fractionation.

The chemistry and terrestrial character of rocks of the Clarno Formation suggested to Rogers and Novitsky-Evans (1977, p. 63) that the sequence represents a subduction-zone assemblage erupted on a continental margin, rather than in an island arc. On the basis of trace-element chemistry, Rogers and Ragland (1980) inferred that the eruptions took place in an area of thin crust, transitional in character between oceanic and continental. Noblett (1981, p. 995) also considers these rocks to be related to subduction-induced volcanism along the continental margin and to be a late phase of Challis arc volcanism. This interpretation is supported by the abundances of lithophile elements in Clarno lavas relative to those elements in island arcs and continental margins, by the composition of pre-Cenozoic bedrock of the region, in which ophiolite, Paleozoic and Mesozoic island-arc, and continental crustal elements are tectonically intermixed, and by geophysical data (Thiruvathukal and others, 1970) that can be interpreted as indicating a transition zone between continent and ocean basin.

The geochemical characteristics of these rocks, their distribution in the Pacific Northwest, and their geometric relations to postulated crustal plates and their convergence directions suggest that the calc-alkaline volcanogenic assemblage of the Clarno Formation and coeval volcanic rocks near Paisley and Lakeview, Oreg., along the California-Oregon State line, and those in the Washington Cascades are underlain largely by dismem-

bered oceanic crust and accreted terranes and, most likely, are related to a north-south-trending, possibly somewhat sinuous pre-Cascade volcanic arc. These data further suggest that the somewhat similar and partly

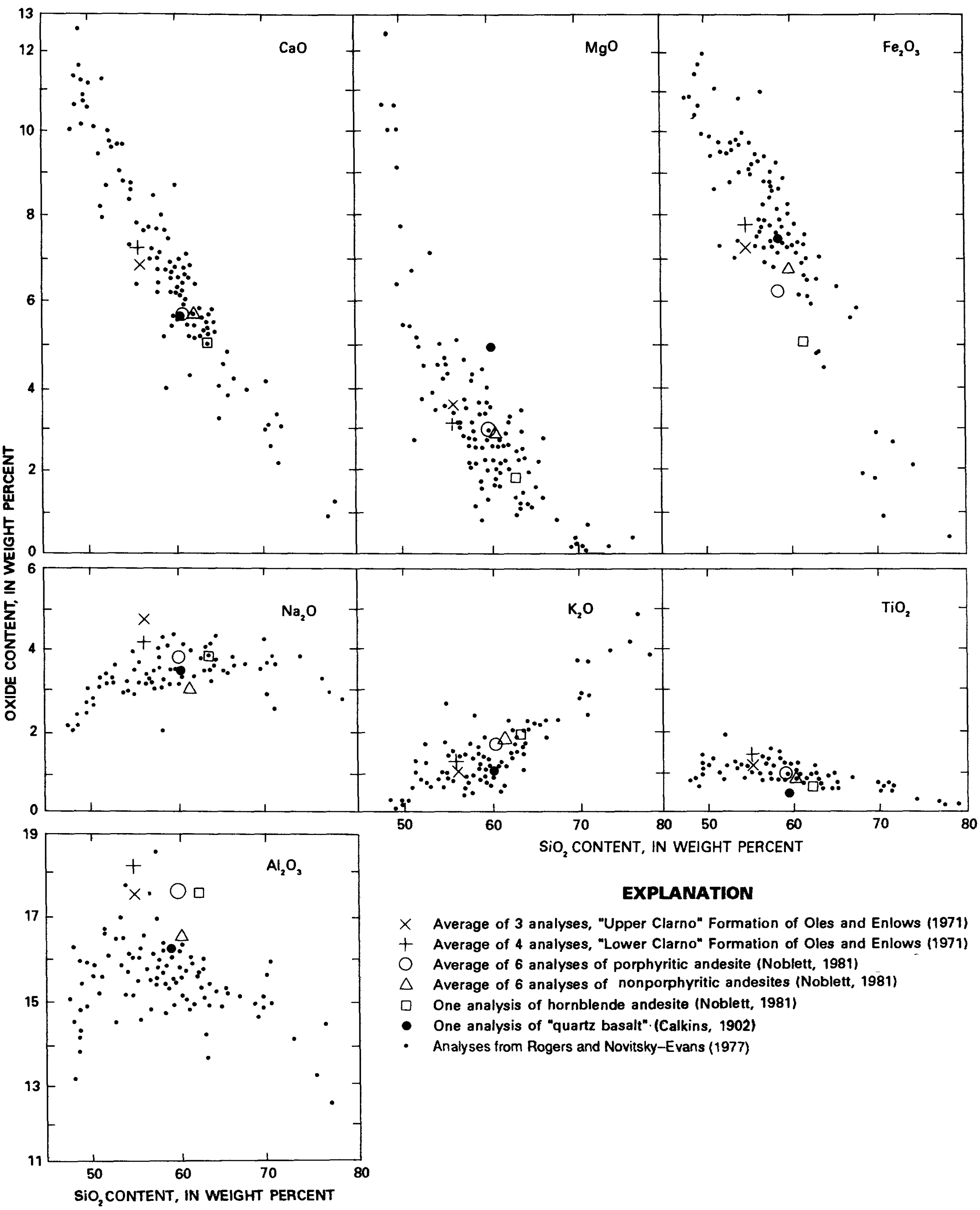


FIGURE 2.5.—Harker diagrams of major oxides in rocks of the Clarno Formation. From Rogers and Novitsky-Evans (1977), with additions from Calkins (1902), Oles and Enlows (1971), and Noblett (1981).

coeval Challis and Absaroka Volcanics, which are underlain by cratonic rocks, are separate and genetically unrelated to this early Cenozoic arc and, thus, to the Clarno Formation. Further discussion of the relations of early Cenozoic volcanism to tectonism is presented in chapter 6.

SOURCE VENTS

Most of the eruptive centers that have been recognized for Clarno Formation rocks are centered on or near the axis of the Blue Mountains uplift (Swanson, 1969) and to the south of this axis within the large Clarno outcrop area in Crook County (fig. 2.2). Some are located in areas to the west near the Deschutes River (Waters, 1968a, b), and others to the southeast of the axis of the Blue Mountains uplift in Grant County (Brown and Thayer, 1966). Taylor (1977, p. 768) indicated that many of the inferred vents are elongate intrusions oriented in a northeast-southwest direction, "coincident with axes of recurrent early Tertiary folding." Many vents are exposed in the area near Mitchell, Oreg., where they are commonly represented by isolated intrusions (fig. 2.6) that penetrate and disrupt Cretaceous sedimentary rocks (Oles and Enlows, 1971). Several vent areas also have been mapped in deep canyons of the Deschutes River and its tributaries (Waters, 1968a, b) and in widely scattered areas southeast of the axis of the Blue Mountains uplift in Grant County (Brown and Thayer, 1966).

Most of the vent complexes are deeply eroded and now consist largely of exhumed dikes, plugs, sills, and irregular intrusions of basaltic andesite or hornblende or pyroxene andesite identical to the lava flows, flow breccias, and lahars that characterize most of the Clarno Formation. A few vent complexes are less eroded and consist of cinders, agglomerate, thin flows, and lahars intruded by thin dikes and sills.

Domal masses of rhyolite and dacite, as well as related near-vent silicic breccias, deposits of pumice blocks, and coarse pumice lapilli tuff, have been recognized in several areas near the axis of the Blue Mountains uplift.

FIGURE 2.6.—Clarno-age conical andesitic plugs (White and Black Buttes on horizon) in rounded and subdued topography eroded from altered volcanic and volcaniclastic rocks of the Clarno Formation. View westward down Bridge Creek toward Mitchell, Wheeler County, Oreg. Elevation of White Butte (left) is 5,665 ft, and of Black Butte (right) 5,080 ft.

REFERENCES CITED

Arnold, C.A., 1952, Fossil Osmudaceae from the Eocene of Oregon: Palaeontographica, v. 92, pt. B, p. 63–78.

Bond, J.G., 1978, Geologic map of Idaho: Moscow, Idaho Bureau of Mines and Geology, scale 1:500,000.

Brooks, H.C., McIntyre, J.R., and Walker, G.W., 1976, Geology of the Oregon part of the Baker 1° by 2° quadrangle: Oregon Department of Geology and Mineral Industries Geologic Map Series Map GMS–7, 25 p., scale 1:250,000.

Brown, C.E., and Thayer, T.P., 1966, Geologic map of the Canyon City quadrangle, northeastern Oregon: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–447, scale 1:250,000.

Calkins, F.C., 1902, A contribution to the petrography of the John Day Basin: University of California, Department of Geology Bulletin, v. 3, p. 109–172.

Chaney, R.W., 1938, Ancient forests of Oregon—A study of earth history in western America, *in* Cooperation in research: Carnegie Institution of Washington Publication 501, p. 631–648.

Elemendorf, Jean, and Fisk, L.H., 1978, The age of the "Pilot Rock flora," Umatilla County, Oregon [abs.]: Geological Society of America Abstracts with Programs, v. 10, no. 5, p. 215.

Enlows, H.E., and Parker, D.J., 1972, Geochronology of the Clarno igneous activity in the Mitchell quadrangle, Wheeler County, Oregon: Ore Bin, v. 34, no. 6, p. 104–110.

Evernden, J.F., Savage, D.E., Curtis, G.H., and James, G.T., 1964, Potassium-argon dates and the Cenozoic mammalian chronology of North America: American Journal of Science, v. 262, no. 2, p. 145–198.

Fiebelkorn, R.B., Walker, G.W., MacLeod, N.S., McKee, E.H., and Smith, J.G., 1983, Index to K-Ar age determinations for the State of Oregon: Isochron/West, no. 37, 60 p.

Fisher, R.V., 1967, Early Tertiary deformation in north-central Oregon: American Association of Petroleum Geologists Bulletin, v. 51, no. 1, p. 111–123.

Fisher, R.V., and Rensberger, J.M., 1972, Physical stratigraphy of the John Day Formation, central Oregon: University of California Publications in Geological Sciences, v. 101, 45 p.

Hanson, C.B., 1973, Geology and vertebrate faunas in the type area of the Clarno Formation, Oregon [abs.]: Geological Society of America Abstracts with Programs, v. 5, no. 1, p. 50.

Heller, P.L., Peterman, Z.E., O'Neil, J.R., and Shafiqullah, Muhammad, 1985, Isotopic provenance of sandstones from the Eocene Tyee Formation, Oregon Coast Range: Geological Society of America Bulletin, v. 96, no. 6, p. 770–780.

Hergert, H.L., 1961, Plant fossils in the Clarno Formation, Oregon: Ore Bin, v. 23, no. 6, p. 55–62.

Hogenson, G.M., 1964, Geology and ground water of the Umatilla River basin, Oregon: U.S. Geological Survey Water-Supply Paper 1620, 162 p.

Jones, R.W., 1982, Early Tertiary-age Kamiah Volcanics, north-central Idaho, *in* Bonnichsen, Bill, and Breckenridge, R.M., eds., Cenozoic geology of Idaho: Idaho Bureau of Mines and Geology Bulletin 26, p. 43–52.

Knowlton, F.H., 1902, Fossil flora of the John Day Basin, Oregon: U.S. Geological Survey Bulletin 204, 153 p.

Lipman, P.W., Prostka, H.J., and Christiansen, R.L., 1972, Cenozoic volcanism and plate tectonic evolution of the Western United States—Pt. I, Early and middle Cenozoic: Royal Society of London Philosophical Transactions, ser. A, v. 271, no. 1213, p. 217–248.

Lux, D.R., 1982, K-Ar and ^{40}Ar-^{39}Ar ages of mid-Tertiary volcanic rocks from the western Cascade Range, Oregon: Isochron/West, no. 33, p. 27–32.

Manchester, S.R., 1981, Fossil plants of the Eocene Clarno Nut Beds: Oregon Geology, v. 43, no. 6, p. 75–81.

McKee, T.M., 1970, Preliminary report on fossil fruits and seeds from the mammal quarry of the Clarno Formation, Oregon: Ore Bin, v. 32, no. 7, p. 117–132.

Mendenhall, W.C., 1909, A coal prospect on Willow Creek, Morrow County, Oregon, *in* Contributions to economic geology, 1907—Pt. II, Coal and lignite: U.S. Geological Survey Bulletin 341, p. 406–408.

Merriam, J.C., 1901, A contribution to the geology of the John Day Basin [Oreg.]: University of California, Department of Geology Bulletin, v. 2, p. 269–314.

Muntzert, J.K., 1969, Geology and mineral deposits of the Brattain district, Lake County, Oregon: Corvallis, Oregon State University, M.S. thesis, 70 p.

Noblett, J.B., 1981, Subduction-related origin of the volcanic rocks of the Eocene Clarno Formation near Cherry Creek, Oregon: Oregon Geology, v. 43, no. 7, p. 91–99.

Oles, K.F., and Enlows, H.E., 1971, Bedrock geology of the Mitchell quadrangle, Wheeler County, Oregon: Oregon Department of Geology and Mineral Industries Bulletin 72, 62 p.

Peck, D.L., 1964, Geologic reconnaissance of the Antelope-Ashwood area, north-central Oregon: U.S. Geological Survey Bulletin 1161–D, 26 p.

Peck, D.L., Griggs, A.B., Schlicker, H.G., Wells, F.G., and Dole, H.M., 1964, Geology of the central and northern parts of the western Cascade Range in Oregon: U.S. Geological Survey Professional Paper 449, 56 p.

Pigg, J.H., Jr., 1961, The lower Tertiary sedimentary rocks in the Pilot Rock and Heppner areas, Oregon: Eugene, University of Oregon, M.S. thesis, 67 p.

Riccio, J.F., 1978, Preliminary geothermal resource map of Oregon: Oregon Department of Geology and Mineral Industries Geologic Map Series Map GMS–11, scale 1:50,000.

Robinson, P.T., 1975, Reconnaissance geologic map of the John Day Formation in the southwestern part of the Blue Mountains and adjacent areas, north-central Oregon: U.S. Geological Survey Miscellaneous Investigations Series Map I–872, scale 1:125,000.

Robinson, P.T., and Brem, G.F., 1981, Guide to geologic field trip between Kimberly and Bend, Oregon, with emphasis on the John Day Formation, *in* Johnston, D.A., and Donnelly-Nolan, Julie, eds., Guides to some volcanic terranes in Washington, Idaho, Oregon, and northern California: U.S. Geological Survey Circular 838, p. 29–40.

Rogers, J.J.W., 1966, Coincidence of structural and topographic highs during post-Clarno time in north-central Oregon: American Association of Petroleum Geologists Bulletin, v. 50, no. 2, p. 390–396.

Rogers, J.J.W., and Novitsky-Evans, J.M., 1977, The Clarno Formation of central Oregon, U.S.A.—Volcanism on a thin continental margin: Earth and Planetary Science Letters, v. 34, no. 1, p. 56–66.

Rogers, J.J.W., and Ragland, P.C., 1980, Trace elements in continental-margin magmatism—Pt. I, Trace elements in the Clarno Formation of central Oregon and the nature of the continental margin on which eruption occurred: Geological Society of America Bulletin, v. 91, pt. 2, no. 4, p. 1217–1292.

Scott, R.A., 1954, Fossil fruits and seeds from the Eocene Clarno Formation of Oregon: Palaeontographica, v. 96, pt. B, no. 3–6, p. 66–97.

Shorey, E.F., 1976, Geology of part of southern Morrow County, northeast Oregon: Corvallis, Oregon State University, M.S. thesis, 131 p.

Stirton, R.A., 1944, A rhinoceros tooth from the Clarno Eocene of Oregon: Journal of Paleontology, v. 18, no. 3, p. 265–267.

Swanson, D.A., 1969, Reconnaissance geologic map of the east half of the Bend quadrangle, Crook, Wheeler, Jefferson, Wasco, and Deschutes Counties, Oregon, *in* Geological Survey Research, 1968: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–568, scale 1:250,000.

Swanson, D.A., and Robinson, P.T., 1968, Base of the John Day Formation in and near the Horse Heaven mining district, north-central Oregon, *in* Geological Survey Research, 1968: U.S. Geological Survey Professional Paper 600–D, p. 154–161.

Taylor, E.M., 1960, Geology of the Clarno Basin, Mitchell quadrangle, Oregon: Corvallis, Oregon State University, M.S. thesis, 173 p.

———1977, The Clarno Formation—A record of early Tertiary volcanism in central Oregon [abs.]: Geological Society of America Abstracts with Programs, v. 9, no. 6, p. 768.

———1981, A mafic-dike system in the vicinity of Mitchell, Oregon, and its bearing on the timing of Clarno-John Day volcanism and early Oligocene deformation in central Oregon: Oregon Geology, v. 43, no. 8, p. 107–112.

Thayer, T.P., and Brown, C.E., 1973, Ironside Mountain, Oregon—A late Tertiary volcanic and structural enigma: Geological Society of America Bulletin, v. 84, no. 2, p. 489–498.

Thiruvathukal, J.V., Berg, J.W., Jr., and Heinrichs, D.F., 1970, Regional gravity of Oregon: Geological Society of America Bulletin, v. 81, no. 3, p. 725–738.

Trauba, W.C., 1975, Petrography of pre-Tertiary rocks of the Blue Mountains, Umatilla County, northeast Oregon: Corvallis, Oregon State University, M.S. thesis, 171 p.

Walker, G.W., 1973, Reconnaissance geologic map of the Pendelton quadrangle, Oregon and Washington: U.S. Geological Survey Miscellaneous Investigations Series Map I–727, scale 1:250,000.

———1980, Preliminary report on the geology of the Lakeview uranium area, Lake County, Oregon: U.S. Geological Survey Open-File Report 80–532, 33 p.

Waters, A.C., 1954, John Day Formation west of its type locality: Geological Society of America Bulletin, v. 65, no. 12, p. 120.

———1968a, Reconnaissance geologic map of the Dufur quadrangle, Hood River, Sherman, and Wasco Counties, Oregon: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–555, scale 1:125,000.

———1968b, Reconnaissance geologic map of the Madras quadrangle, Jefferson and Wasco Counties, Oregon: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–556, scale 1:125,000.

Waters, A.C., Brown, R.E., Compton, R.R., Staples, L.W., Walker, G.W., and Williams, Howel, 1951, Quicksilver deposits of the Horse Heaven mining district, Oregon: U.S. Geological Survey Bulletin 969–E, p. 105–149.

Wilson, P.M., 1973, Petrology of the Cenozoic volcanic rocks of the basal Clarno Formation, central Oregon: Houston, Tex., Rice University, M.A. thesis, 47 p.

Woodburne, M.O., and Robinson, P.T., 1977, A new late Hemingfordian mammal fauna from the John Day Formation, Oregon, and its stratigraphic implications: Journal of Paleontology, v. 51, no. 4, p. 750–757.

3. EOCENE(?), OLIGOCENE, AND LOWER MIOCENE ROCKS OF THE BLUE MOUNTAINS REGION

By PAUL T. ROBINSON[1], GEORGE W. WALKER, and EDWIN H. MCKEE

CONTENTS

ABSTRACT

A distinctive assemblage of tuffaceous sedimentary rocks, air-fall and ash-flow tuffs, and sparse lava flows crops out widely in north-central Oregon. The sequence is mainly Oligocene and early Miocene in age and lies between the Clarno Formation below and the Columbia River Basalt Group above. Most of the rocks in this sequence have been assigned to the John Day Formation, but a few are of uncertain age or affinity.

Three lithologic facies are recognized in the John Day Formation. An eastern facies that crops out east of the Blue Mountains uplift consists of tuffaceous claystone, fine-grained air-fall tuff, and a single ash-flow sheet in the middle part of the section. The western facies, which occurs between the Blue Mountains uplift and the Cascade Range, consists of tuffaceous claystone, coarse-grained tuff, and numerous interlayered ash-flow sheets and lava flows. A southern facies, lithologically similar to the eastern facies, crops out south of the Ochoco Mountains. Correlations between these facies are based on radiometric ages and vertebrate faunas.

The John Day Formation is compositionally heterogeneous. Most of the tuffaceous claystone and air-fall tuffs have pyrogenetic mineralogies indicating an original andesitic or dacitic composition. This air-fall material is believed to have been erupted from vents in what is now the western Cascade Range and thus provides a record of early Cascade volcanism. The interlayered ash-flow sheets are all rhyolitic and are believed to have been erupted from vents east of the present-day Cascade Range, such as those in the Mutton Mountains. Lava flows in the John Day Formation range in composition from alkali olivine basalt to rhyolite and are clearly derived from local vents scattered throughout the outcrop area.

Well-documented radiometric ages for the John Day Formation range from 37.4±1.1 to 22.7±2.7 Ma, in good agreement with the faunal ages.

A few unusual sequences in north-central Oregon have ages equivalent to the John Day Formation but andesitic lithologies more typical of the Clarno Formation, suggesting some overlap in these two different styles of volcanism.

INTRODUCTION

In northern and northeastern Oregon, rocks mainly of Oligocene and early Miocene age form a distinctive sequence lying between the Eocene and lowermost Oligocene(?) Clarno Formation and the lower Miocene and younger Columbia River Basalt Group. In a few places, where flows of the Columbia River Basalt Group pinch out on the mainly Oligocene and lower Miocene rocks, the sequence is overlain unconformably by middle Miocene tuffs and tuffaceous sedimentary rocks. Rocks of the sequence crop out chiefly on the two flanks of the Blue Mountains uplift and in an area south of the Ochoco Mountains. There are major differences in lithology and stratigraphy among these outcrop areas, but the different sections can be correlated on the basis of distinctive fossil flora and fauna and numerous isotopic ages.

The mainly Oligocene and lower Miocene sequence consists chiefly of intermediate and silicic air-fall tuffs

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and tuffaceous sedimentary rocks, locally accompanied by abundant ash-flow tuffs, mafic lava flows, and silicic flows and domes. Most of these rocks have been assigned to the John Day Formation, which is distinguished from the underlying Clarno Formation largely on the basis of lithology, composition, and age (see chap. 2). In addition, the John Day Formation consists of well-layered and laterally continuous units, whereas the Clarno Formation has no throughgoing stratigraphy. However, there is some overlap between the two units in lithology and possibly in age, and the boundary is not everywhere obvious.

Because the bulk of the Clarno Formation is andesitic in composition, isolated rhyolite flows and domes that overlie or intrude Clarno lavas have generally been assigned to the John Day Formation (Waters and others, 1951; Waters and Vaughan, 1968a, b). However, a few isotopic ages suggest that most of these bodies are Eocene in age and should be included in the Clarno Formation. In a few localities, such as the Horse Heaven mining district, these rhyolitic rocks are overlain unconformably by the oldest ash-flow tuffs of the John Day Formation (Swanson and Robinson, 1968); in others, they are commonly mantled with 1 to 2 m of red saprolite, such as is found locally at the top of the Clarno Formation.

More puzzling are isolated sequences of andesitic lava flows of typical Clarno lithology and interlayered rhyolitic volcaniclastic rocks whose isotopic ages are significantly younger than the oldest ages obtained thus far for the John Day Formation. Some of these enigmatic sequences (for example, that at Smith Rock, which has been dated at 30.8±0.5 Ma) may be unconformably overlain by rocks of the upper part of the John Day; others are isolated masses whose structural and stratigraphic relations are unknown. If the young ages can be substantiated, they imply some temporal overlap of Clarno-type volcanism and John Day volcanism. Although physical interfingering of the two units has yet to be demonstrated, such overlap is considered possible because the Clarno Formation was erupted from vents within its outcrop area, whereas the bulk of the John Day Formation was probably derived from vents farther west.

JOHN DAY FORMATION

DEFINITION

The name "John Day Formation" was first used by Marsh (1875) for a sequence of fossiliferous, supposedly lacustrine sedimentary rocks exposed along the banks of the John Day River (Condon, 1871). Merriam (1901) defined the formation as a sequence of varicolored tuffs and tuffaceous sedimentary rocks, approximately 350 m thick, lying between the older Clarno Formation and the younger, then-named Columbia River Basalt. Merriam did not designate a type section, but most of the outcrops he described lie along the John Day River between Kimberly and Picture Gorge, the area generally considered to be the "type" area[2] of the formation (Fisher and Rensberger, 1972).

Rocks of John Day lithology and age are widespread in north-central Oregon, in three principal areas: (1) an eastern area, including the "type" area, that extends along the east flank of the Blue Mountains uplift from the vicinity of Mitchell nearly to La Grande; (2) a western area lying between the Blue Mountains uplift and the Cascade Range; and (3) a southern area lying along the south flank of the Ochoco Mountains, generally between Prineville and Paulina (figs. 3.1, 3.2). The rocks in these three areas can be considered different facies of the formation.

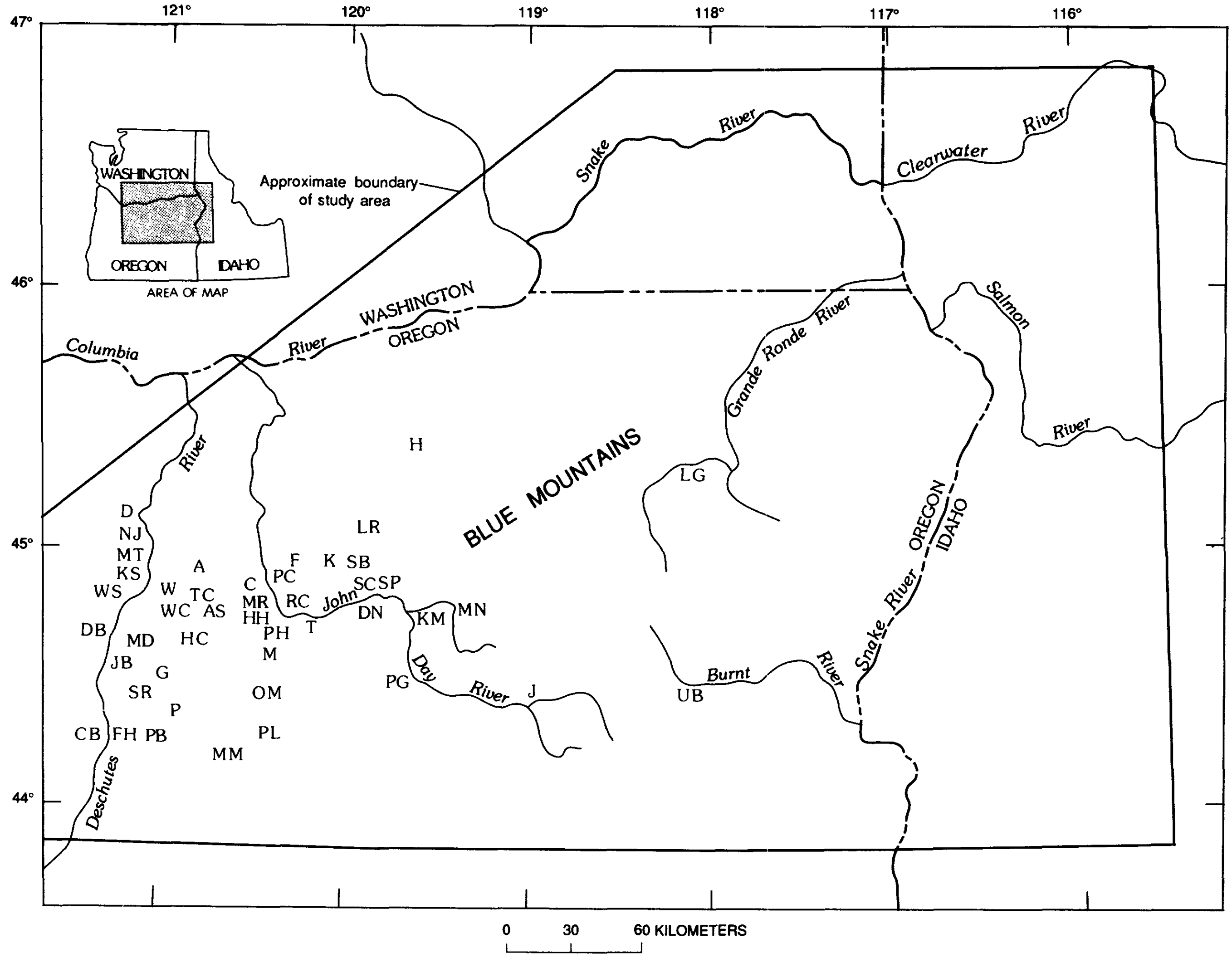
In the eastern area, the formation has a maximum aggregate thickness of approximately 750 m and consists chiefly of fine-grained air-fall tuff and tuffaceous claystone (Calkins, 1902; Coleman, 1949; Fisher and Wilcox, 1960; Hay, 1962a, 1963; Fisher and Rensberger, 1972). A conspicuous rhyolitic ash-flow sheet, the informally designated Picture Gorge ignimbrite of Fisher (1966a), lies near the middle of the formation. Rocks in the southern area are similar to those of the eastern facies and consist chiefly of fine-grained, volcaniclastic sedimentary rocks, with one conspicuous ash-flow sheet near the middle of the sequence. In the area south of Prineville is a second ash-flow sheet, which lies near the base of the formation (Waters and Vaughan, 1968a, b). In the western area, however, the rocks of the formation are much coarser grained and consist chiefly of lapilli tuff and tuffaceous sedimentary rocks, with numerous interbedded ash-flow sheets and lava flows of rhyolitic, basaltic, and trachyandesitic composition. The aggregate thickness in the western area is nearly 1,200 m, but individual sections rarely exceed 500 m in thickness.

BASE

In gross lithology the Clarno and John Day Formations are quite distinct; the Clarno consists chiefly of flows and breccias of pyroxene andesite and basalt (see chap. 2), and the John Day of dacitic to rhyolitic tuffs, tuffaceous sedimentary rocks, and flows. In detail, however, there is considerable overlap in composition, and it is commonly difficult to distinguish the two formations on the basis of

[2]No formal type section or type locality has been established for the John Day Formation. Most investigators generally consider the section exposed between Picture Gorge and Kimberly as typical of the formation.

lithology alone. For example, trachyandesite and alkali olivine basalt flows are common in both the lower part of the John Day Formation (Peck, 1961, 1964; Robinson, 1969, 1975) and the upper part of the Clarno Formation



INDEX TO LOCALITIES

A, Antelope and Antelope Valley
C, Clarno (Clarno Ferry)
CB, Cline Butte
D, Dant
DB, Deschutes Basin
DN, Donnely Basin
F, Fossil
FH, Forked Horn Butte
G, Grizzly
H, Heppner
HC, Hay Creek and Hay Creek Ranch
HH, Horse Heaven mining district
J, John Day
JB, Juniper Butte
K, Kinzua
KM, Kimberly
KS, Kahneeta Hot Springs
LG, La Grande
LR, Lonerock
M, Mitchell
MD, Madras
MM, Maury Mountains
MN, Monument
MR, Muddy Ranch
MT, Mutton Mountains
NJ, North Junction
OM, Ochoco Mountains
P, Prineville and Prineville Reservoir
PB, Powell Buttes
PC, Pine Creek
PG, Picture Gorge
PH, Painted Hills
PL, Paulina
RC, Rowe Creek
SB, Snowboard Ridge
SC, Service Creek
SP, Spray
SR, Smith Rock
T, Twickenham
TC, Trout Creek
UB, Unity Basin and Unity Reservoir
W, Willowdale
WC, Wilson Creek
WS, Warm Springs

FIGURE 3.1.—Index map of Blue Mountains region, showing study area and localities mentioned in text.

(Robinson, 1969). Silicic flows and volcaniclastic rocks similar to those of the John Day Formation are locally abundant in the Clarno Formation, particularly in the Monument quadrangle (Wilcox and Fisher, 1966) and in the Painted Hills area northwest of Mitchell. Tuffs and tuffaceous sedimentary rocks are abundant in the Clarno Formation in the Horse Heaven mining district (Waters and others, 1951) and in the vicinity of Muddy Ranch, and small outcrops of ash-flow tuff also are locally present (Oles and Enlows, 1971; Taylor, 1960).

After deposition of the Clarno Formation, an episode of weathering and erosion resulted in a moderately hilly terrain with as much as 120 m of relief (Waters, 1954). This surface became locally mantled with a red lateritic-soil zone (saprolite) ranging from a few centimeters to as much as 150 m in thickness (Waters and others, 1951; Hay, 1962a, 1963; Fisher, 1964). In many parts of the eastern outcrop area, this soil zone is a distinct lithologic unit that marks the contact between the Clarno and John Day Formations. However, this unit must be interpreted with care because similar soil zones occur at several levels within the Clarno Formation (Waters and others, 1951; Swanson and Robinson, 1968), and in numerous localities the deep-red claystones of the basal part of the John Day Formation rest directly on the post-Clarno soil zone, making it difficult to pinpoint the contact. Hay (1962a) suggested placing the contact in such areas just below the first appearance of pyrogenetic minerals.

In the western facies, the soil zone at the top of the Clarno is thin or missing altogether. There, the base of the John Day Formation is located at the base of an extensive rhyolitic ash-flow sheet that unconformably overlies andesitic and basaltic flows and breccias of typical Clarno lithology (Peck, 1961, 1964; Swanson and Robinson, 1968; Robinson, 1975). Two K-Ar ages of 37.4 and 37.1 Ma indicate that this ash-flow sheet is latest Eocene (Palmer, 1983) in age (table 3.1); the 1-Ma error bars for these ages straddle the Eocene-Oligocene boundary at 36.6 Ma.

The position of the contact between the Clarno and John Day Formations in the Blue Mountains uplift between the eastern and western facies is not so clearly established. Waters and others (1951), working in the Horse Heaven mining district, designated a saprolite-

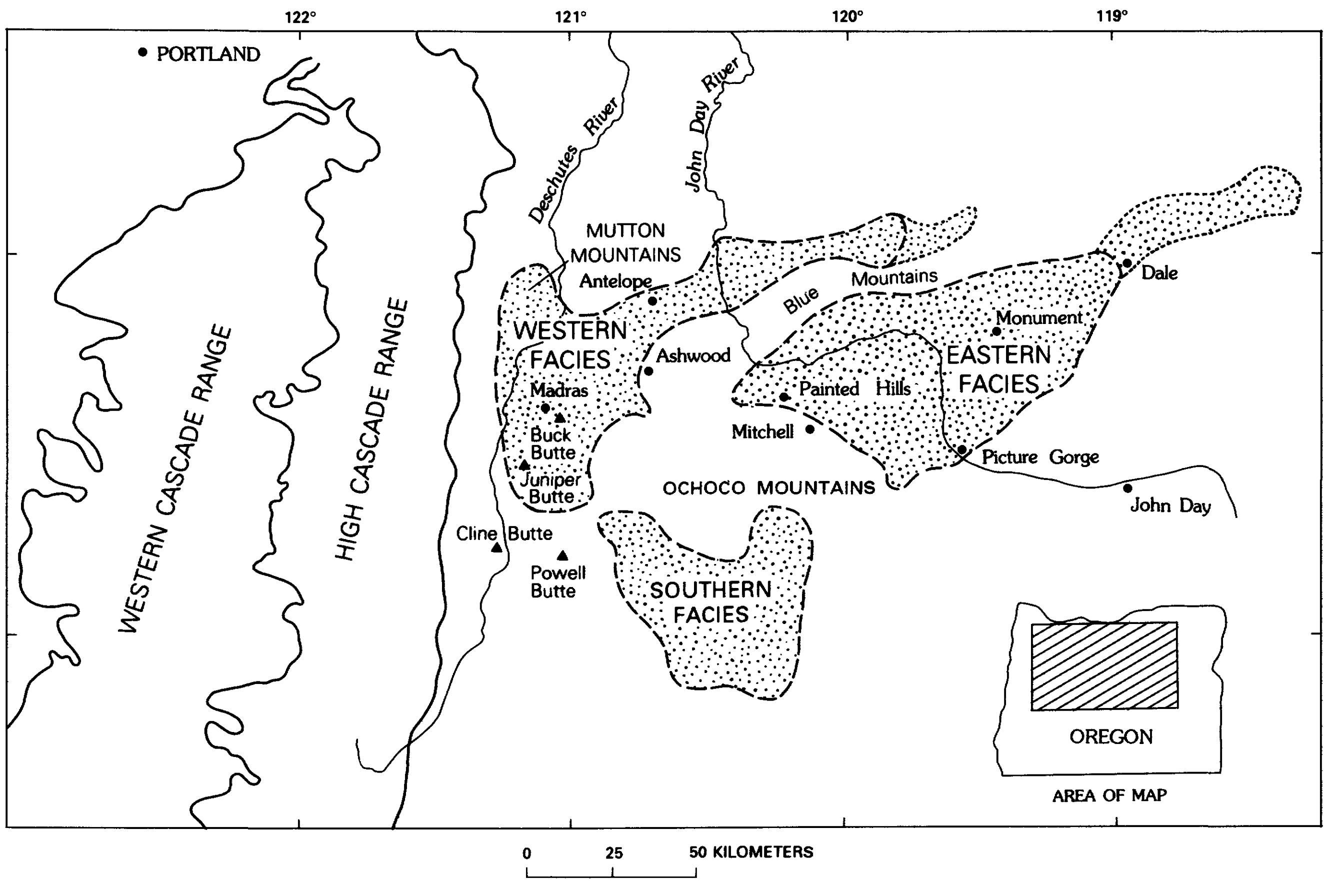


FIGURE 3.2.—Approximate outcrop extent of the John Day Formation in north-central Oregon. Dashed lines enclose areas of known outcrops; dotted lines enclose areas where rocks of probable John Day affiliation are present. Formation also extends an unknown distance to the north beneath the Columbia River Basalt Group.

TABLE 3.1.—*Potassium-argon ages for the John Day Formation, Oregon*

[do. and Do., ditto]

Sample	Unit	Rock type	Material	Age (Ma)	Reference
			Eastern facies		
KA1384	Lower member	Claystone	Adularia	21.6	Evernden and others (1964).
KA845	do.	Basalt	Whole rock	31.5	Do.
KA489	do.	Tuff	Sanidine	31.1	Do.
Unknown	do.	do.	do.	26.7 25.7	Curtis and others (1961).
KA647A	Unknown	do.	do.	23.3	Evernden and others (1964).
648-456	Lower member	do.	do.	26.8±0.4	This chapter.
PTR-71-11	Picture Gorge ignimbrite.	Ash-flow tuff	Plagioclase	26.9±2.3	Do.
KA648	do.	Obsidian	Whole rock	25.5	Evernden and others (1964).
KA649A	Unknown	Unknown	Albite	24.9	Do.
			Southern facies		
648-625A	Lower member	Ash-flow tuff	Plagioclase	32.1±0.7	This report.
			Western facies		
KA1384	Lower member	Bentonite	Sanidine	32.0	Evernden and others (1964).
DAS-66-208	Member A	Ash-flow tuff	do.	37.4±1.1	Swanson and Robinson (1968).
DAS-66-195	do.	do.	do.	37.1±1.0	Fiebelkorn and others (1983).
PTR-71-10	Member G	do.	do.	30.5±0.4	This chapter.
648-584	do.	do.	do.	28.6±0.5	Do.
648-34B	do.	do.	do.	28.1±0.5	Do.
648-552	do.	do.	do.	27.0±0.3	Do.
PTR-71-5A	Member H	do.	Plagioclase	30.9±0.7	Do.
PTR-71-6	Member I	do.	do.	27.7±0.3	Do.
Rob Tuf 3	do.	Tuff	do.	23.4±3.3	Do.
Rob Tuf 1	do.	do.	do.	22.7±2.7	Do.
			Miscellaneous outcrops		
PTR-71-9A	Snowboard	Ash-flow tuff	Plagioclase	57.0	This chapter.
2/UT225	Powell Buttes	Rhyolite	Whole rock	30.1±1.1	Evans and Brown (1981).
PB-5/AH-34	do.	do.	Anorthoclase	28.3±1.0	Do.
648-623B	do.	do.	Sanidine	25.8±0.2	This chapter.
CC-1	Unknown	do.	Anorthoclase	28.1±5.0	Walker (1973).
GWW-56-69	do.	Vitrophyre	Biotite	27.7±0.8	Do.

mantled unconformity as the top of the Clarno Formation and referred to younger andesitic and rhyolitic flows as "post-Clarno," possibly equivalent to John Day rocks elsewhere. However, Swanson and Robinson (1968) found that the basal ash-flow sheet of the western facies of the John Day, which is latest Eocene(?) and earliest Oligocene in age, unconformably overlies the post-Clarno rocks of the Horse Heaven mining district. In addition, one of the post-Clarno rhyolites has a K-Ar age of 42.1±0.8 Ma[3] (Swanson and Robinson, 1968), well within the age limits of the Clarno Formation. Because of these relations, Swanson and Robinson (1968) included the post-Clarno rocks of the Horse Heaven district in the Clarno Formation and suggested that the saprolite zone in this area marks a local unconformity within that unit.

[3]Their original published age was somewhat less than the value given here; this older age was obtained by applying abundance and decay constants acceptable in 1982 (Steiger and Jäger, 1977) to the original analytical data.

Such local unconformities are common in the Clarno Formation, and no means has yet been found to distinguish individual saprolites. Thus, outside the "type" area of the John Day Formation in the eastern facies, the presence or absence of a red saprolite is not a reliable guide to the position of the Clarno-John Day contact.

STRATIGRAPHY

EASTERN FACIES

Merriam (1901) originally divided what he called the John Day Series (or Formation) in the study area into three units based primarily on color: (1) a lower unit consisting chiefly of deep-red clays and tuffs, (2) a middle unit composed of bluish-green tuffs that crop out in rounded hills or pinnacled cliffs, and (3) an upper unit of buff-colored material containing abundant sand and gravel.

Fisher and Rensberger (1972) proposed a somewhat similar fourfold subdivision of the John Day Formation (fig. 3.3), from oldest to youngest: (1) the Big Basin Member, composed of deep-red claystones and approximately comparable to Merriam's lower unit; (2) the Turtle Cove Member, composed chiefly of green, buff, or red zeolitized tuff; (3) the Kimberly Member, consisting of light gray to buff unzeolitized tuff; and (4) the Haystack Valley Member, a sequence of fluviatile, lacustrine, and air-fall gray unzeolitized tuff. Both of these subdivisions are based on diagenetic features, chiefly color and the presence or absence of zeolites, and thus have limited stratigraphic significance. For example, Hay (1962a) showed that the red claystones in the lower part of the John Day Formation were in part derived from the underlying saprolite locally present at the Clarno-John Day contact. Likewise, Hay (1963) and Fisher and Rensberger (1972) showed that the boundary between the zeolitized and fresh-glass facies in the formation is an irregular surface that cuts across stratigraphic boundaries. For these reasons, we prefer the threefold subdivision of Hay (1962a, 1963) for the eastern facies of the John Day Formation, which is based on the presence of a widespread ash-flow sheet (Picture Gorge ignimbrite) near the middle of the formation (fig. 3.4). The lower member, approximately 335 m thick, consists chiefly of tuffaceous claystone and tuff, with sparse, small, interbedded flows of olivine basalt. The (informal) Picture Gorge ignimbrite is a compound cooling unit of rhyolitic tuff, ranging from about 3 to 85 m in thickness. The upper member consists of about 400 to 500 m of yellowish-gray to olive-gray tuffaceous claystone and vitric tuff, with local accumulations of reworked tuff and conglomerate near the top.

Detailed stratigraphic descriptions of the eastern facies were presented by Hay (1962a, 1963) and by Fisher and Rensberger (1972), and so only a brief discussion is given below.

LOWER MEMBER

The lower member of the John Day Formation consists chiefly of thick-bedded, massive tuffaceous claystone, with thin interlayered beds of air-fall tuff and small flows of alkali-olivine basalt. At the base of the formation along Rowe Creek, just north of Twickenham, a 15-m-thick ash-flow tuff overlies a red saprolite developed on andesitic lava flows of Clarno lithology. This white to light-gray, strongly welded eutaxitic tuff contains 5 to 10 percent crystals, chiefly sanidine, quartz, and oligoclase. On the basis of its mineralogy (table 3.2) and stratigraphic position, this tuff was tentatively correlated with the basal ash-flow sheet of member A of the western facies by Hay (1962a, 1963), Swanson and Robinson (1968), and Robinson (1975).

Most claystone in the lower member is pale brown to reddish brown; some is green, yellowish green, or olive gray. Brick-red claystone forms a distinctive basal layer, which sometimes extends through the lower half of the member. Hay (1962a) showed that some red claystone consists of mixtures of weathered John Day ash and red kaolinitic clay derived from the saprolite at the top of the Clarno Formation; Fisher (1964), however, argued that some of the red claystone formed in place by weathering and diagenetic alteration.

In the overlying claystone, brown and yellow varieties consist chiefly of montmorillonite, whereas the green varieties contain abundant celadonite. All of the claystone is tuffaceous, originally containing about 85 to 90 percent vitric material and 3 to 13 percent sand-size crystals and silicic rock fragments (Hay, 1962a, 1963; Fisher, 1968). The abundant andesine and sanidine crystals in the claystone suggest an original dacitic composition for much of the pyroclastic material. The original pumice fragments and shards have been extensively altered to clay minerals and zeolites. Hay (1962a) suggested that most of the lower member consists of vitric material which accumulated slowly at the surface and weathered to montmorillonite before burial. The zeolites and other diagenetic minerals formed after burial.

Air-fall tuff interbedded with the claystone forms light-gray, well-indurated layers, 2 cm to 3 m thick, that crop out as slightly resistant ledges. Most of the tuff is fine to medium grained and vitric and contains sparse pumice lapilli as much as 5 mm across. Sanidine, sodic plagioclase, and clinopyroxene make up 3 to 5 percent of most of the tuff, and quartz is locally present. Silicic rock fragments also are typically present in no more than 5 percent by volume.

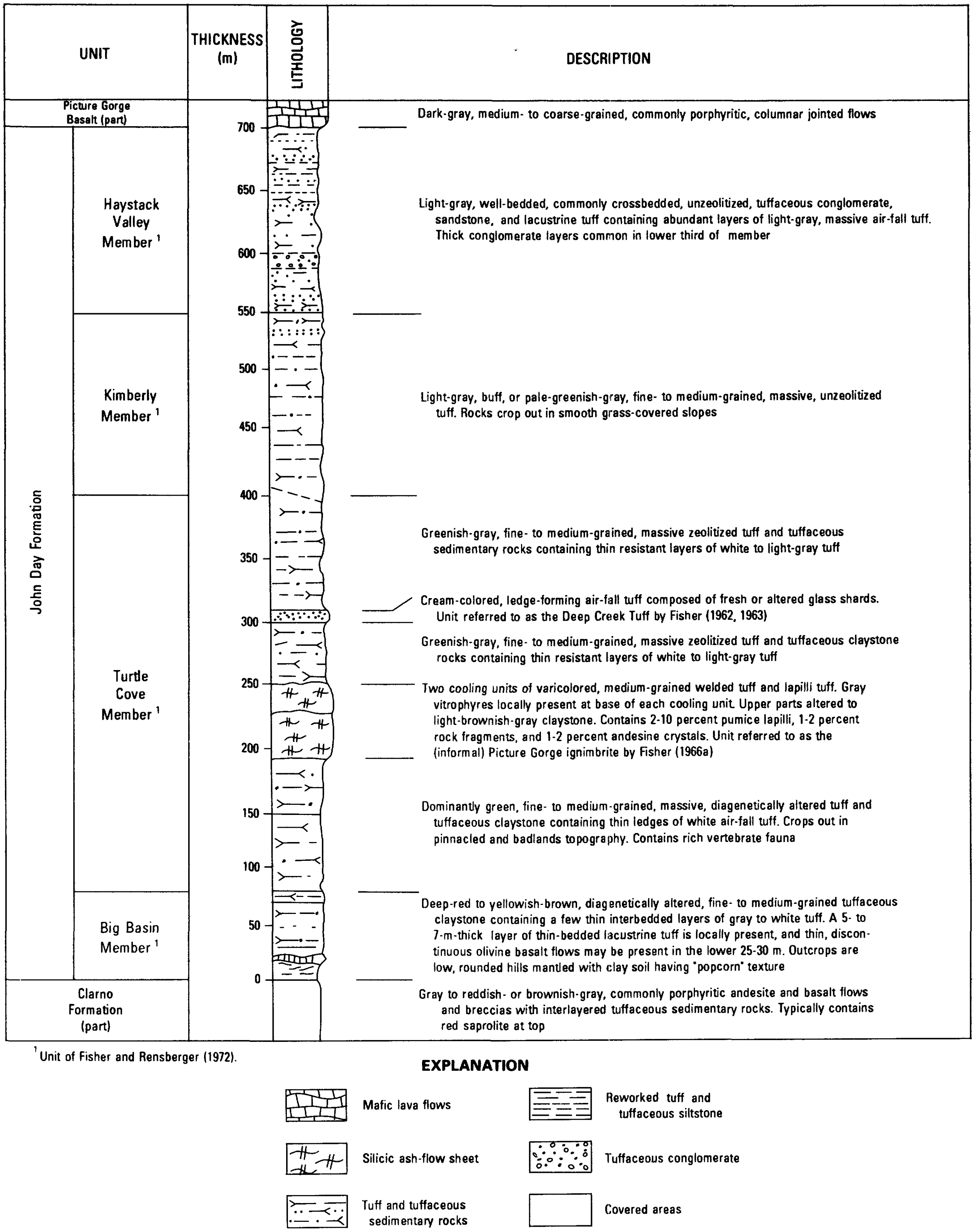


FIGURE 3.3.—Generalized composite stratigraphic column for eastern facies of the John Day Formation in Kimberly, Oreg., area (modified from Fisher and Rensberger, 1972).

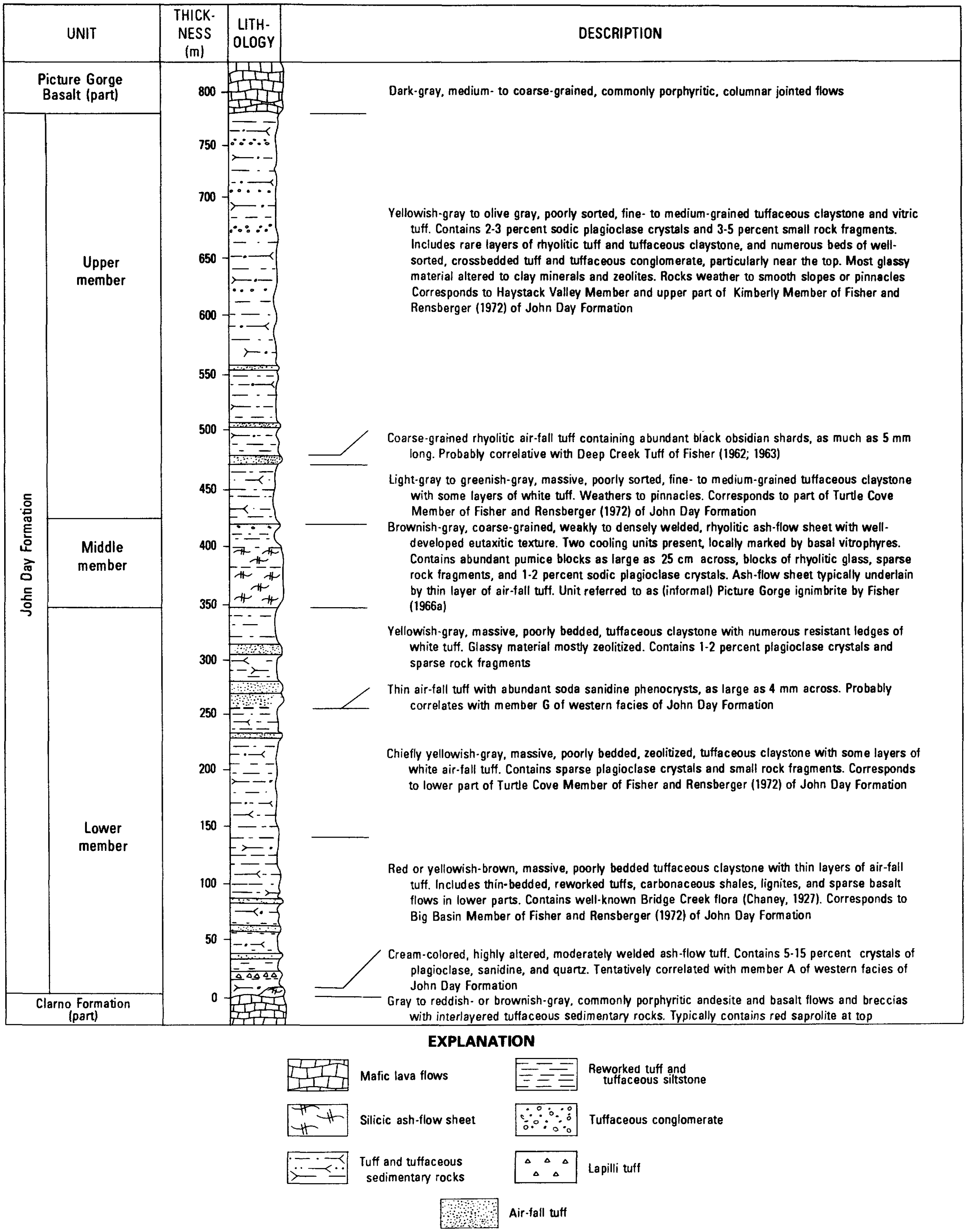


FIGURE 3.4.—Generalized composite stratigraphic column for eastern facies of the John Day Formation in Painted Hills, Oreg., area (modified from Hay, 1962a; 1963).

TABLE 3.2.—*Compositions of feldspars in the lower ash-flow sheet of member A, John Day Formation*

[Oxide values in weight percent; tr, trace; n.d., not determined. Values for end-member components (An, anorthite; Ab, albite; Or, orthoclase; Cn, celsian) in molecular percent]

	Plagioclase				Sanidine			
Sample	648-1A	648-1B	648-1B	71-8	71-8	648-1A	648-1B	71-8
No. of grains	2	1	1	1	2	3	3	4
Major oxides								
SiO_2	63.67	61.22	63.63	61.88	63.21	65.00	65.00	64.43
Al_2O_3	22.37	23.89	22.21	23.51	22.58	19.31	18.68	19.38
CaO	4.12	6.14	3.99	5.45	4.18	.40	.37	.40
Na_2O	8.12	7.52	7.89	6.56	6.74	4.61	4.35	4.35
K_2O	1.54	.90	1.59	1.04	1.54	9.03	9.62	9.15
MgO	tr	n.d.	tr	tr	tr	tr	n.d.	n.d.
FeO	.13	.20	.08	.31	.17	.05	.05	.07
TiO_2	.03	n.d.	.11	.03	.03	n.d.	tr	tr
BaO	.24	.10	.06	.28	.14	2.01	1.45	1.99
Total	100.22	99.98	99.59	99.07	98.59	100.41	99.52	99.77
End-member components								
An	20.4	29.51	9.8	29.3	23.3	2.0	1.9	2.1
Ab	71.1	65.4	70.8	64.0	66.9	42.9	40.2	41.0
Or	8.5	5.1	9.4	6.7	9.8	55.1	57.9	56.9
Cn	.4	.2	.1	.5	.3	3.2	2.6	3.7

A distinctive, coarse-grained, sanidine-rich, air-fall tuff crops out widely in the upper third of the lower member in the Painted Hills area. This unit is generally 0.5 to 1 m thick, but where it has been extensively reworked it may be as much as 10 m thick. It typically lies 150 to 250 m above the base of the formation and 60 to 120 m below the Picture Gorge ignimbrite. The feldspar crystals, which range from 1 to 4 mm in diameter and are commonly mantled with myrmekitic overgrowths, are soda sanidine with an average composition of about $Or_{43}Ab_{54}An_3$ (table 3.3; Hay, 1962b). On the basis of textural and compositional similarities (table 3.3), this tuff is correlated with the basal ash-flow sheet of member G of the western facies (Peck, 1964). This correlation is strengthened by the fact that from southwest to northeast the ash-flow sheet of member G grades from a welded tuff, through an unwelded ash-flow tuff, to an air-fall tuff (Robinson, 1975).

The alkali-olivine basalt flows in the lower member are generally 10 to 15 m thick and rarely extend more than a few hundred meters along strike. Pahoehoe structures on some flow surfaces (Hay, 1962a) and local accumulations of agglomerate and breccia suggest subaerial emplacement.

A 120-m-thick sequence of basalt flows capped by the Picture Gorge ignimbrite is exposed in Donnely Basin southeast of Service Creek. These flows were originally assigned to the Clarno Formation by Lindsley (1961), but Hay (1963) included them in the lower part of the John Day Formation because of the freshness of the flows, the similarity of the tuffs between the flows to the overlying John Day Formation, and the chemical similarity of the basalts to known John Day lavas. Although basalts of similar composition are present in the Clarno Formation elsewhere (Robinson, 1969), they are much more abundant in the John Day Formation.

The basalts are typically fine or medium grained and aphyric, with intersertal to subophitic or, rarely, intergranular textures. They consist chiefly of about equal amounts of labradorite and titaniferous augite, 10 to 15 percent olivine, and 5 to 10 percent ilmenite and magnetite. Common interstitial alkali feldspar and biotite attest

TABLE 3.3.—*Composition of feldspars in the basal ash-flow sheet of member G, western facies, John Day Formation, and its air-fall equivalent in the eastern facies*

[Oxide values in weight percent; tr, trace; n.d., not determined; ---, not found. Values for end-member components (An, anorthite; Ab, albite; Or, orthoclase; Cn, celsian) in molecular percent]

	Ash-flow sheet		Air-fall tuff		
	Sanidine		Sanidine		Plagioclase
Sample	648-34B	648-552	648-456	648-456	648-456
No. of grains	4	3	3	3	2
Major oxides					
SiO_2	67.45	67.14	67.04	65.21	59.01
Al_2O_3	18.94	19.17	18.39	20.12	25.51
CaO	.50	.40	.44	1.92	7.98
Na_2O	6.05	5.92	6.05	7.15	6.24
K_2O	7.3	7.65	7.50	4.47	.26
MgO	tr	n.d.	n.d.	tr	.04
FeO	.15	.18	.16	.18	.37
TiO_2	.07	.04	.04	.04	tr
BaO	.25	.20	.41	1.26	tr
Total	100.73	100.70	100.03	100.35	99.41
End-member components					
An	2.5	2.0	2.2	9.5	40.7
Ab	54.3	53.0	53.8	64.1	57.7
Or	43.2	45.0	44.0	26.4	1.6
Cn	.5	.3	.8	2.2	---

to an alkalic composition. In most specimens, the olivine is partly to completely replaced by clay minerals, and clay minerals are also common in the groundmass. Chemically, these flows are high-titania, alkali olivine basalt similar to widespread flows in the western facies (Lindsley, 1961; Hay, 1962a; Robinson, 1969).

MIDDLE MEMBER

The Picture Gorge ignimbrite, which forms the middle member of the formation, is a compound ash-flow sheet composed of two distinct cooling units and a thin basal layer of air-fall tuff with an aggregate thickness ranging from approximately 3 to 85 m. Fisher (1966a) estimated that the ignimbrite layer originally covered an area of approximately 5,000 km^2, was thickest in the Mitchell-Spray-Service Creek area, and thinned progressively to the northeast. K-Ar ages on the unit are 26.9±3 and 25.5 Ma, or late Oligocene in age (table 3.1).

The two cooling units are texturally and mineralogically similar, with well-defined zonal structures (fig. 3.5; Hay, 1963; Fisher, 1966a). Both contain abundant pumice blocks, as much as 10 cm in diameter, and irregular fragments of black glass, as much as 5 cm across. In densely welded zones the pumice fragments are stretched and as large as 25 cm, imparting a well-developed eutaxitic texture.

Crystals and silicic rock fragments make up about 3 percent of typical specimens (table 3.4). The crystals are chiefly sodic plagioclase (table 3.5), with lesser amounts of sanidine, quartz, and green clinopyroxene. Hay (1963) reported a few pseudomorphs after fayalitic olivine.

UPPER MEMBER

The upper member varies greatly in preserved thickness, depending on the depth of erosion before eruption of the overlying Columbia River Basalt Group. A maxi-

mum thickness between 400 and 520 m was recorded by Hay (1963), but the member is thin or absent where deeply eroded. This member is lithologically similar to the upper part of the lower member and is composed chiefly of tuffaceous claystone and vitric tuff, with small amounts of sandstone and pebble to cobble conglomerate near the top. The principal differences between the two members are diagenetic. The upper member is lighter in color and generally less altered than the lower. The unzeolitized tuff of the upper member corresponds to the Kimberly Member as defined by Fisher and Rensberger (1972).

The claystone is poorly bedded, poorly sorted, and massive and is generally light yellowish gray to light olive gray in color. Crystals make up about 5 percent of typical specimens and consist chiefly of andesine, with trace amounts of clinopyroxene, magnetite, ilmenite, hornblende, and biotite. This mineral assemblage implies a dacitic to rhyodacitic composition for the original pyroclastic material.

Air-fall tuff is commonly interbedded with the claystone, particularly in the lower 175 m of the member. The tuff is typically lighter gray and coarser grained than the claystone and may contain abundant oligoclase. One

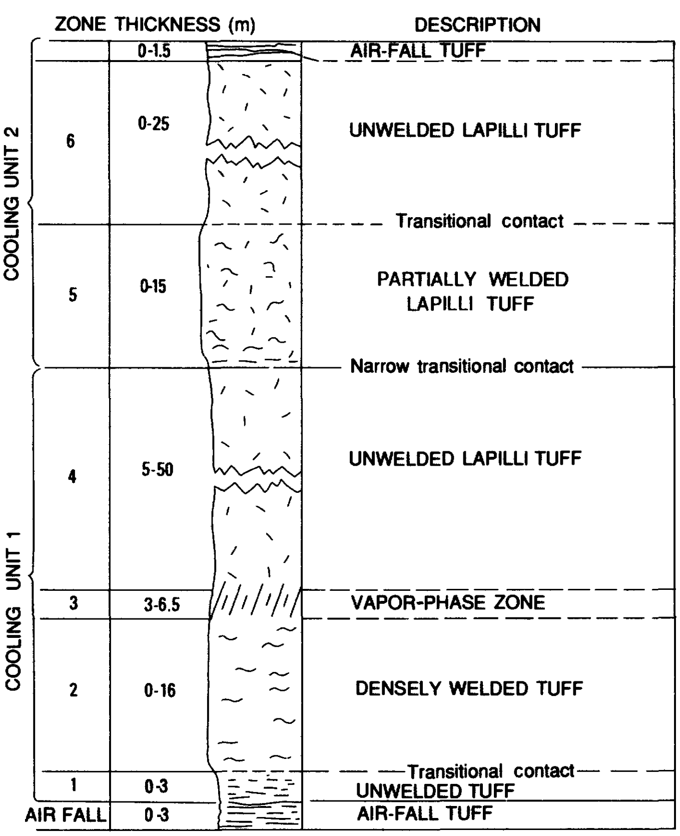


FIGURE 3.5.—Generalized columnar section of (informal) Picture Gorge ignimbrite of Fisher (1966a, fig. 2).

distinctive air-fall layer (Deep Creek Tuff of Fisher, 1962, 1963) in the upper member crops out discontinuously. It lies 3 to 60 m above the Picture Gorge ignimbrite and is as much as 10 m thick. It forms cream-colored outcrops with internal lamination and platy parting and is composed largely of glass shards easily visible in hand specimen. In most outcrops the pyroclastic material is altered to clinoptilolite, but fresh black glass is locally preserved (Fisher and Rensberger, 1972).

Lenticular beds of reworked tuff and conglomerate are interbedded with the claystone at several levels. These layers are typically moderately to well sorted and commonly thin bedded and crossbedded. Conglomerate is most abundant near the top of the formation and correlates approximately with the Haystack Valley Member of Fisher and Rensberger (1972). Clasts are chiefly of tuff eroded from the upper member, but at a few localities clasts of welded tuff and andesite also are present.

SOUTHERN FACIES

Little detailed work has been done in the southern facies, which lies in a belt along the south side of the Ochoco Mountains from Prineville to Paulina. Reconnaissance studies indicate that this section is most like the eastern facies and consists chiefly of fine-grained tuff and tuffaceous claystone (Swinney and others, 1968; Waters, 1968c; Waters and Vaughan, 1968a, b; Fisher and Rensberger, 1972). Two ash-flow sheets are interlayered with the tuff in the vicinity of the Prineville Reservoir, but farther east only one is present. Fossil faunas in the southern facies are similar to those in the eastern facies (Fisher and Rensberger, 1972).

Although the southern facies has not been formally divided into members, the stratigraphy is similar to that of the eastern facies. At the base of the section is a thin, discontinuous saprolite locally overlain by red claystone that corresponds approximately to the Big Basin Member of Fisher and Rensberger (1972). The overlying tuff and tuffaceous claystone can be divided into zeolitized and unzeolitized zones that are similar to the Turtle Cove and Kimberly Members, respectively.

The lower welded tuff south of Prineville lies near the base of the formation and either rests directly on Clarno lavas or is separated from them by 50 to 100 m of altered dacitic tuff. The ash-flow tuff is a light-gray to reddish-gray, coarse-grained, stony rhyolite with a good eutaxitic texture defined by flattened pumice fragments, as long as 10 cm. Crystals make up 1 to 2 percent of the tuff and are chiefly plagioclase with traces of magnetite and altered biotite (table 3.4). The matrix is a microcrystalline mixture of quartz and feldspar in which the original vitroclastic texture has been largely obliterated. Many of the larger pumice fragments have been zeolitized.

TABLE 3.4.—*Average modal compositions of ash-flow tuff sheets in the John Day Formation*

[Component contents in volume percent. An content of plagioclase in molecular percent; tr, trace; —, not found]

Rock unit	1	2	3	4	5	6	7	8	9	10	11
Groundmass	94.9	97.8	96.4	84.4	98.6	96.2	98.4	97.3	93.8	99.1	99.3
Rock fragments	—	1.4	1.6	3.6	.7	2.1	.5	1.7	5.4	.1	.7
Sanidine	1.1	—	tr	10.2	—	—	—	—	—	—	—
Quartz	3.2	.1	.6	1.6	—	—	tr	—	—	—	—
Plagioclase	.8	.7	1.2	—	.7	1.4	1.0	.9	.8	.8	—
Clinopyroxene	—	—	.2	—	—	.2	tr	.1	tr	.1	—
Hornblende	tr	—	—	—	—	—	tr	—	—	—	—
Biotite	—	—	tr	—	—	—	—	—	—	—	—
Opaque minerals	tr	—	tr	.2	tr	.1	.1	tr	tr	tr	tr
Zircon	—	—	tr	tr	tr	tr	tr	—	—	—	—
Apatite	tr	—	—	tr	—	tr	tr	—	—	—	—
No. of samples	22	7	9	9	12	7	3	4	6	1	2

Unit:

1. Basal ash-flow sheet of member A, western facies (Swanson and Robinson, 1968).
2. Upper ash-flow sheet of member A, western facies (Swanson and Robinson, 1968).
3. Ash-flow tuff sheets of member E, western facies.
4. Basal ash-flow sheet of member G, western facies (Robinson and Brem, 1981).
5. Basal ash-flow sheet of member H, western facies.
6. Basal ash-flow sheet of member I, western facies.
7. Picture Gorge ignimbrite, lower cooling unit, eastern facies (Robinson and Brem, 1981).
8. Picture Gorge ignimbrite, upper cooling unit, eastern facies.
9. Ash-flow tuff north of Fossil, Oreg., northern subfacies.
10. Basal ash-flow tuff, southern facies.
11. Upper ash-flow tuff, southern facies.

The upper ash-flow sheet is about 10 m thick and consists of light-gray to reddish-gray, densely welded, fine-grained, generally stony tuff. Small pumice fragments, about 1 cm long, give the rock a crude eutaxitic texture. Crystals are absent, but most specimens contain 1 to 2 percent small andesitic rock fragments (table 3.4).

Neither of these ash-flow sheets can be correlated with welded tuffs in the eastern or western facies; these sheets are believed to have originated from separate vents south or southwest of the Ochoco Mountains.

WESTERN FACIES

The western facies crops out in a broad arc between the Cascade Range and the Blue Mountains uplift (fig. 3.2; see fig. 1.2). Only the upper part of the unit consists of light-colored, massive tuff and tuffaceous sedimentary rocks similar to those exposed between Kimberly and Picture Gorge. The lower half consists chiefly of ash-flow tuff, silicic and basaltic lava flows, and pumice lapilli tuff, all of which were originally included in the Clarno Formation (Hodge, 1932a, b). Peck (1961, 1964) recognized that these lower rocks rested unconformably on andesitic flows of typical Clarno lithology and found a fossil flora near the base of the sequence that indicated an Oligocene age. For this reason, he included the lower sequence in the John Day and subdivided the formation in the Antelope-Ashwood area into nine members (A–H), largely on the basis of the presence of ash-flow sheets and lava flows (fig. 3.6). These members can be recognized throughout the western facies (Robinson, 1975), except in the Mutton Mountains and in the area between Fossil and Lonerock. Outcrops in these areas are separated from those in the Antelope-Ashwood area, and the stratigraphy is quite different.

MEMBER A

Member A consists of an upper and a lower ash-flow sheet separated by about 30 m of tuff and pumice lapilli tuff. The basal ash-flow sheet is areally extensive and well exposed (Swanson and Robinson, 1968), but the upper sheet is thin and discontinuous. Nearly complete sections of member A crop out along the county road about 3 km west of Ashwood, near the Horse Heaven mining district, and near Hay Creek; elsewhere, only the ash-flow sheets are well exposed.

In most outcrops, the basal ash-flow sheet forms a prominent ledge composed of moderately to strongly welded tuff with a well-developed eutaxitic structure.

TABLE 3.5.—*Compositions of feldspars in ash-flow tuff sheets of the John Day Formation*

[Oxide values in weight percent; tr, trace. Values for end-member components (An, anorthite; Ab, albite; Or, orthoclase; Cn, celsian) in molecular percent]

	Picture Gorge Ignimbrite			Member H	Member I	Tuff near Fossil, Oreg.
No. of grains	1	2	1	3	8	6
Major oxides						
SiO_2	66.39	65.61	63.22	67.15	65.94	66.96
Al_2O_3	20.81	21.12	23.02	19.02	20.92	19.58
CaO	2.33	2.65	4.32	.26	2.41	.47
Na_2O	8.96	8.85	8.24	6.07	9.22	6.55
K_2O	1.65	1.42	.62	6.43	1.58	6.13
FeO	.24	.19	.30	.22	.22	.13
TiO_2	tr	tr	tr	.06	.04	.05
BaO	.31	.28	.25	.30	.32	.37
Total	100.69	100.12	99.97	99.51	100.65	100.24
End-member components						
An	11.3	13.1	21.7	1.4	11.5	2.3
Ab	79.1	78.5	74.6	58.2	79.5	60.0
Or	9.6	8.4	3.7	40.4	9.0	37.7
Cn	.6	.5	.5	.6	.6	.7

Along the county road west of Ashwood, two cooling units are clearly recognizable, each with a basal vitrophyre; elsewhere, separate cooling units cannot be distinguished. At one locality southeast of Antelope, the tuff is red to purple and almost completely unwelded, but it still is highly indurated.

The tuff contains about 10 percent phenocrysts, averaging 1 to 2 mm in diameter (table 3.4). Although the modal composition of the tuff varies considerably (Swanson and Robinson, 1968), there are no apparent systematic mineralogic or chemical differences between the two cooling units. Corroded quartz bipyramids, from 0.5 to 3 mm in diameter, are the most common crystals, making up 1 to 6 percent of the rock. Feldspar phenocrysts average about 2 percent and include both sanidine and sodic plagioclase. The sanidine crystals are euhedral to slightly rounded prisms, ranging from 0.5 to 1 mm in diameter. The sanidine is a soda-bearing variety containing 1 to 2 weight percent BaO (table 3.2). Plagioclase occurs as subhedral, moderately zoned crystals of oligoclase that average about 1 mm in diameter.

Overlying the basal ash-flow sheet in many areas is a 1-m-thick layer of red tuff overlain by about 20 m of poorly indurated, poorly bedded, white pumice lapilli tuff, overlain, in turn, by about 15 to 20 m of very poorly exposed red tuff and tuffaceous siltstone. Just below the upper ash-flow sheet is a 1-m-thick layer of white, very thin bedded, fine-grained tuff containing numerous leaf impressions. In a few localities, this tuff is overlain by 2 to 3 m of thick-bedded tuff containing fossilized tree limbs and trunks.

The upper ash-flow sheet is a light-brown to pinkish-gray, fine-grained, densely welded tuff containing 1 to 2 percent plagioclase crystals and 1 to 2 percent small rock fragments (table 3.4). A light-gray vitrophyre, as much as 2 m thick, occurs locally at the base of the sheet, but most of the unit is stony and contains very sparse lithophysae. This ash-flow sheet has a maximum thickness of about 50 m in exposures west of Ashwood. Elsewhere, it rarely exceeds 10 m in thickness, and it is much less continuous in outcrop than the lower sheet.

MEMBER B

A series of dark-gray trachyandesite flows, which are present in several separate outcrop areas, compose

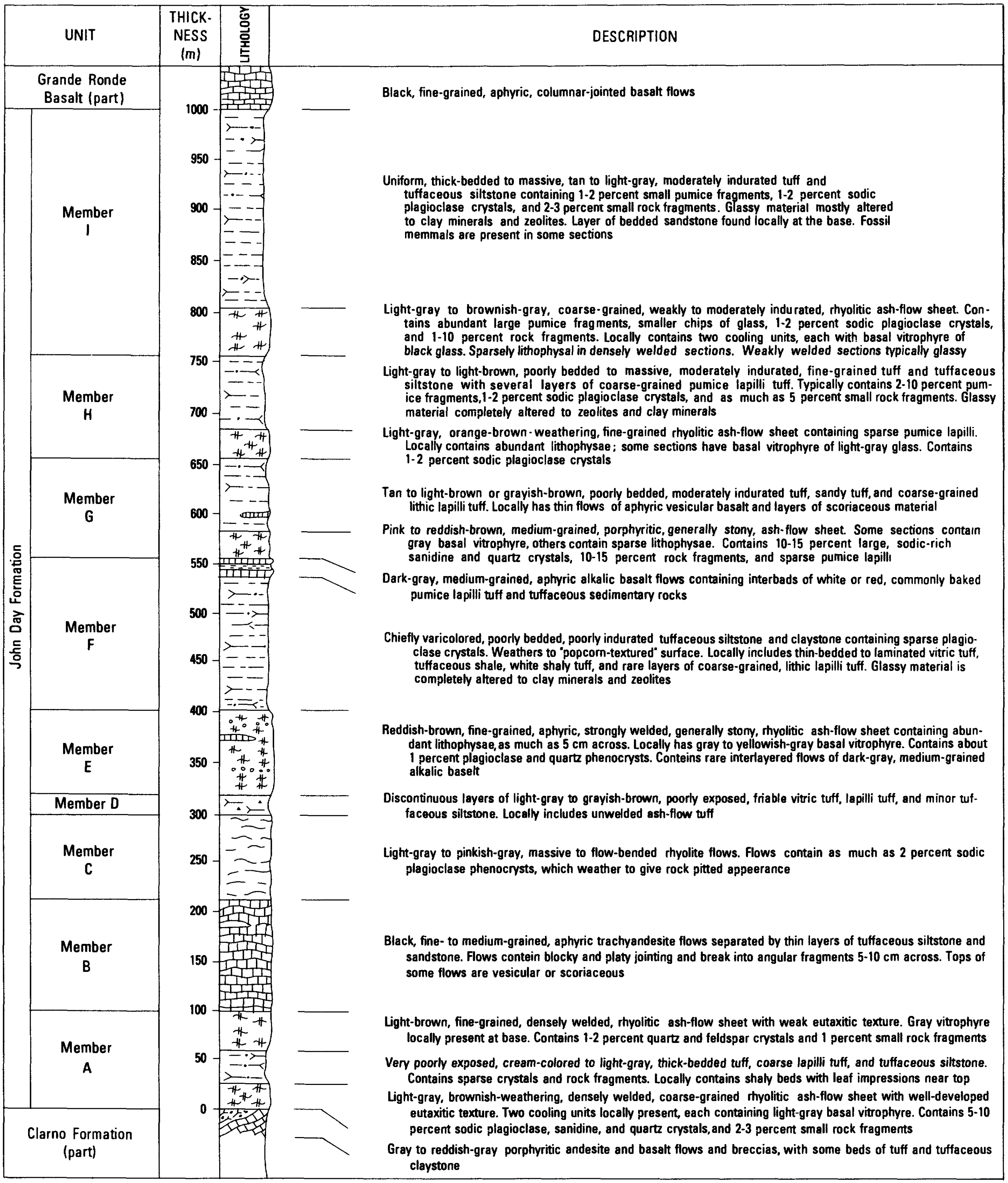


FIGURE 3.6.—Generalized composite stratigraphic column for western facies of the John Day Formation in Antelope-Ashwood-Willowdale, Oreg., area (modified from Peck, 1964, and Robinson, 1975). See figure 3.4 for explanation of lithologic symbols.

TABLE 3.6.—*Average modal compositions of lava types in the John Day Formation*

[Values in volume percent; tr, trace; —, not found]

Rock type	Rhyolite[1]	Quartz latite	Andesite	Trachy-andesite	Basalt
Plagioclase	2.2	35.1	[2]65.6	36.0	49.5
Clinopyroxene	.1	6.1	21.7	22.8	17.4
Olivine	—	3.3	—	1.2	13.4
Opaque minerals	.2	3.2	2.7	8.7	7.7
Glass and interstitial material.	—	52.2	—	29.2	11.1
Apatite	tr	.1	—	.3	.6
Zircon	tr	tr	—	—	—
Other	—	tr	—	1.8	.3
Groundmass	97.5	—	—	—	—
No. of samples	3	3	1	7	22

[1]Member C, western facies

[2]Includes 3.1 volume percent microphenocrysts

member B. The largest mass extends for about 25 km along the western margin of the Blue Mountains uplift from Hay Creek on the south to the Ashwood-Willowdale highway on the north. Smaller outcrops are found in the east end of Antelope Valley and northeast of the hamlet of Clarno. At most localities, the trachyandesite flows rest on member A and are overlain by rocks of members C, D, E, or F.

In the thickest and most complete section, about 4 km southwest of Ashwood, the unit is approximately 500 m thick and consists of at least three flows or sequences of flows separated by thin layers of tuff. The flows form low, rounded outcrops mantled with fist-size fragments that commonly form stone rings and stripes. Typical specimens are dark gray, very fine grained, nearly aphyric, and sparsely vesicular. One flow near Eagle Creek in Antelope Valley is relatively coarse grained, with groundmass crystals averaging 2 to 3 mm across. Sparse microphenocrysts of plagioclase, clinopyroxene, and rare olivine are set in a felted matrix of plagioclase laths, granular clinopyroxene, iron oxides, and glassy interstitial material (table 3.6; Peck, 1964; Robinson, 1969). The groundmass plagioclase averages about An_{45}, but some specimens contain sparse, highly corroded xenocrysts of calcic labradorite. In most specimens the clinopyroxene is light-brown pigeonite accompanied by minor low-calcium augite.

No source vents were located for the trachyandesite flows, but the outcrop pattern and thickness variations indicate that they were erupted locally. A possible vent identified by Peck (1964) is here considered to be a flow remnant.

The interlayered sedimentary rocks, which range from about 0.5 to 5 m in thickness, are light-brown, poorly bedded, sandy pumice lapilli tuff, tuffaceous siltstone, and pebble conglomerate. These rocks are friable and poorly exposed, and the only known outcrops are along the county road between Ashwood and Madras about 8 km southwest of Ashwood.

MEMBER C

A thick sequence of rhyolite flows and associated domes that crops out in the Antelope-Ashwood area makes up member C. The flows are best exposed along the Trout Creek and Wilson Creek drainages, where they have an aggregate thickness of about 125 m. Thinner and less extensive outcrops are also present along Antelope Creek farther north. At most localities the flows rest with structural conformity on tuffs of member A and occupy the same stratigraphic position as the trachyandesite flows of member B. Locally, however, the rhyolite flows clearly rest on the trachyandesite flows of member B.

The flows crop out in low, rounded hills and consist of light-gray to reddish-gray, massive or flow-banded rhyolite. The flows grade into a domelike mass of rhyolite that cuts the Clarno and lower part of the John Day Formation about 6 km north of Ashwood. Here, the rhyolite is exposed in crudely columnar-jointed cliffs that

rise more than 400 m above the surrounding valley. A breccia composed of rhyolite fragments, as much as 1 m in diameter, in a tuffaceous matrix is present along the southwest side of the domal complex. Inasmuch as the contacts are not well exposed, this rhyolite mass could be a very thick flow that ponded in an old topographic low. However, the form of the body indicates that it crosscuts the older rocks, and it is tentatively interpreted as an exogenous dome (Peck, 1964). The flows surround this mass and thin rapidly to the northeast and southwest.

The flows and the probable dome are compositionally similar, consisting of fine-grained, very sparsely porphyritic rhyolite. Phenocrysts are sodic andesine and trace amounts of altered pyroxene(?) and opaque minerals (table 3.6). Weathering of the phenocrysts commonly gives the rock a pitted appearance. The groundmass is a mosaic of very fine grained anhedral quartz and alkali feldspar.

Member D

A thin sequence of white to yellowish-gray tuff and tuffaceous sedimentary rocks makes up member D. The member crops out discontinuously from Wilson Creek on the south to Antelope Valley on the north, where it separates the rhyolite flows of member C from the ash-flow tuffs of member E. At most localities it is less than 5 m thick, but along Trout Creek, between Ashwood and Willowdale, it is more than 30 m thick. The unit consists chiefly of poorly bedded, poorly indurated tuff and minor lapilli tuff, some of which contains small blebs of black glass. Locally, it appears to grade upward into the weakly welded base of member E. Outcrops north and northeast of Trout Creek also contain a light-gray, unwelded lapilli ash-flow tuff, as much as 3 m thick. At one locality the unit consists of angular blocks of gray to brownish-gray glass in a white tuffaceous matrix.

This unit is interpreted as an accumulation of air-fall and minor ash-flow material erupted before the major explosions that produced the thick welded tuffs of member E. The tuff of member D is generally aphyric, but a few layers contain trace amounts of megascopic quartz and feldspar.

Member E

Member E is a light-brown, aphyric ash-flow sheet characterized by large and abundant lithophysae. Known outcrops extend in a broad arc from Hay Creek on the south to the east end of Antelope Valley on the north, a distance of at least 40 km. The unit is missing from sections east of the John Day River and in the Mutton Mountains. It attains a maximum thickness of about 120 m between Willowdale and Ashwood, where it consists of a series of ledges, each 2 to 5 m thick, that may be individual flow units. From this locality the unit thins rapidly to the north, south, and east; to the west it is overlain by higher members of the formation. At many localities the ash-flow sheet is overlain by tuffs and tuffaceous sedimentary rocks of member F; elsewhere, these friable rocks have been stripped by erosion, leaving a series of tablelike outcrops.

The base of the unit is commonly marked by a light-gray to yellowish-gray vitrophyre, as much as 4 m thick, which is overlain by light-pink to reddish-gray, densely welded, stony tuff containing layers of well-developed lithophysae and locally exhibiting crude columnar jointing. In a few localities, vitrophyres also are present within the sequence. The lithophysae are open, elongate cavities, as large as 10 cm in maximum dimension. Most are in distinct layers, a few centimeters to 2 m thick, separated by zones of massive material. Such alternations of lithophysae-rich and lithophysae-free zones are particularly pronounced in exposures along the highway between Ashwood and Willowdale. These zones may define separate flow units within the sheet, but there are no other lithologic breaks between layers, and in some localities the lithophysae are concentrated in pipelike masses that cut across the layers. At several localities (for example, the Priday Agate deposit near Willowdale) this ash-flow sheet contains abundant silica-filled spherulites or thunder eggs (Ross, 1941; Renton, 1951; Brown, 1957) that appear to have formed by secondary silica filling cavities within spherulites or lithophysae.

The ash-flow tuff is very sparsely porphyritic and contains a few crystals of calcic oligoclase, 1 to 3 mm long, and traces of quartz and altered pyroxene; a few small dark rock fragments also are present (table 3.4). The groundmass in glassy specimens consists of densely welded glass shards, averaging about 0.3 mm in length, accompanied by rare, flattened pumice fragments. In the stony layers it consists of microcrystalline quartz and feldspar.

Member F

A heterogeneous sequence, composed chiefly of diagenetically altered tuff and tuffaceous sedimentary rocks, lies between ash-flow sheets of members E and G. Thin basalt flows are locally interlayered with the sedimentary rocks in the uppermost parts of the unit. Rocks of member F can be traced in outcrop from just north of Grizzly on the southwest to just west of Fossil on the northeast, a distance of nearly 75 km. Because the unit consists chiefly of nonresistant tuff and sedimentary rocks, it is poorly exposed, commonly mantled with talus

from overlying ash-flow tuffs and basalts or completely covered by landslide material. Where exposed, outcrops typically form low, rounded hills with a crackly or popcorn-textured surface.

The most complete sections are found in a triangular area, bounded by Hay Creek, Willowdale, and Ashwood, in which the member is more than 150 m thick. From this area the unit thins to the northeast and southwest, eventually pinching out as it overlaps older units and rests directly on the underlying Clarno Formation. In the vicinity of Hay Creek, the base of the member is marked by about 5 m of white, medium-bedded, coarse-grained pumice lapilli tuff. At other localities the lowest beds consist of light-gray to greenish-gray, poorly bedded, moderately indurated vitric tuff, commonly containing 3 to 4 percent pumice lapilli. The rest of the unit consists largely of grayish-brown or grayish-green, poorly bedded, friable tuffaceous claystone and siltstone. Beds are lenticular, however, and the unit varies widely in thickness from place to place. For example, in many areas, particularly where members B, C, D, and E are missing and member F rests directly on member A, the lower 40 to 50 m of the unit are brick red in color. Elsewhere, layers of red or pink claystone and tuff are interbedded with more typical gray to greenish-gray rocks. Other local variations include layers of coarse-grained pumice lapilli tuff containing fragments of perlitic glass, well-bedded and crossbedded sandstone and siltstone, white platy shale with abundant leaf impressions and petrified wood, and coarse-grained tuffaceous conglomerate. In the area between Clarno and Fossil, member F consists chiefly of green and buff tuffaceous claystone, with interbeds of green sandy tuff and white vitric tuff. A thin layer near the top of this sequence contains fossil snails and fish-bone fragments. As elsewhere in the formation, these color variations appear to reflect different degrees of diagenetic alteration. The green claystone is rich in montmorillonite, celadonite, and clinoptilolite, whereas the buff and gray sequences contain chiefly montmorillonite and minor amounts of fresh glass.

In Antelope Valley and in the area north and northeast of Clarno, several extensive flows of olivine basalt are interlayered with the uppermost tuffaceous sedimentary rocks of member F. These flows extend discontinuously from Trout Creek near Willowdale on the west nearly to Fossil on the east, a distance of approximately 50 km. As many as three separate flows, each 10 to 15 m thick, are present in some outcrops; elsewhere, multiple flows have not been recognized. Locally, the flows are associated with beds of brownish-gray, moderately bedded, basaltic lapilli tuff and agglomerate. Some of the tuffs immediately overlying the lava flows contain fragments of vesicular basalt, as large as 10 cm across, indicating erosion before burial.

The lava flows are dark-gray, aphyric, medium- to coarse-grained alkali olivine basalt (table 3.7) (Robinson, 1969). A few exhibit poorly developed columnar jointing, and many have vesicular zones in the upper few meters. Most are highly altered and weathered, but some fresh, even glassy, specimens are present. As seen in thin section, the basalt is a medium-grained rock with intersertal to subophitic textures composed of plagioclase, titaniferous clinopyroxene, ilmenite, and 15 to 20 percent olivine, most of which is altered to orange smectite. A few outcrops contain pegmatitic segregations composed of ilmenite, titanaugite, and plagioclase crystals, as much as 3 cm across.

Member G

A distinctive, crystal-rich ash-flow sheet and an overlying thick sequence of air-fall tuff and tuffaceous sedimentary rocks make up member G. It is one of the most widespread and easily recognized units in the western facies, extending from the Mutton Mountains on the west to beyond the John Day River on the east and from Smith Rock northward to Antelope Valley.

The basal ash-flow sheet consists of reddish- to yellowish-gray, densely welded, crystal-rich tuff, commonly with abundant lithophysae in its upper few meters. It varies systematically in thickness, degree of welding, and crystallization over its outcrop area. In the vicinity of Hay Creek Ranch, east of Madras, the tuff is as much as 30 m thick and is densely welded, completely devitrified, and moderately lithophysal in the upper few meters. In an outcrop about 20 km north of Hay Creek Ranch along Oregon Highway 206 between Willowdale and Antelope, the tuff is 5 to 6 m thick, only weakly welded, and completely glassy. Still farther to the northeast, between Antelope and Clarno, an air-fall tuff containing crystals of sanidine similar to those in the welded tuff is present in the same stratigraphic position. This air-fall tuff averages about 1 m thick, but where it is extensively reworked it is 5 m or more thick. This tuff has been correlated on the basis of mineralogy with a similar air-fall tuff in the eastern facies in the Painted Hills area north of Mitchell (Hay, 1963; Woodburne and Robinson, 1977; Robinson and Brem, 1981).

Where it is devitrified, the ash-flow sheet is typically dark reddish brown, less commonly pink or purple. In the Mutton Mountains it is light yellowish to reddish brown and is characterized by pronounced color banding. Locally, a light-gray vitrophyre as much as 3 m thick is present at the base of the sheet.

Crystals average about 12 percent of this tuff and consist chiefly of sanidine and quartz accompanied by trace amounts of iron oxide and rare altered pyroxene (table 3.4). The sanidine crystals are as large as 4 mm

TABLE 3.7.—*Average chemical compositions of basalt and trachyandesite units in the John Day Formation*

[Values in weight percent; n.d., not determined]

Rock unit	1	2	3	4	5	6
SiO_2	44.98	45.57	45.76	46.17	55.00	51.17
Al_2O_3	15.47	14.56	14.66	16.20	13.55	14.75
Fe_2O_3	6.75	4.50	5.44	7.68	2.90	1.15
FeO	7.98	10.84	9.68	5.42	9.70	3.76
MgO	4.97	4.97	5.20	4.75	2.09	2.63
CaO	7.61	8.42	8.37	9.66	6.14	6.45
Na_2O	3.42	3.12	2.88	3.04	3.25	3.60
K_2O	1.23	.84	.82	.55	1.72	1.64
H_2O^-	1.76	1.76	1.63	1.83	.93	1.60
H_2O^+	1.74	.86	1.28	1.19	1.27	1.17
TiO_2	3.25	3.82	3.51	2.72	2.10	2.93
P_2O_5	.62	.65	.50	.49	.59	.10
MnO	.17	.19	.19	.23	.21	.09
CO_2	.12	n.d.	n.d.	n.d.	.73	n.d.
Total	100.07	100.10	99.92	99.93	100.18	100.04
No. of samples	4	2	7	2	3	1

Unit:

1. Alkalic basalt of lower member, eastern facies (data from Lindsley, 1961; Hay, 1962a; Robinson, 1969).
2. Alkalic basalt of member E, western facies (Robinson, 1969).
3. Alkalic basalt of member F, western facies (Robinson, 1969).
4. Alkalic basalt of sequence north of Fossil, Oreg., northern subfacies (Robinson, 1969).
5. Trachyandesite of member B, western facies (Peck, 1964; Robinson, 1969).
6. Trachyandesite of sequence northeast of Fossil, Oreg., northern subfacies (Robinson, 1969).

across, typically euhedral, and commonly rimmed with myrmekitic intergrowths of quartz and feldspar. These crystals have a distinctive soda-rich composition averaging about $Or_{44}Ab_{54}An_2$ (table 3.3). This unique composition, along with the textural characteristics of the grains, has been used to correlate the air-fall tuff with the ash-flow sheet. The quartz crystals are corroded bipyramids as large as 1 mm across, and the opaque minerals are small laths and irregular grains, generally less than 0.2 mm in diameter. Most specimens also contain 2 to 5 percent silicic to andesitic rock fragments. The groundmass consists chiefly of devitrified shards, commonly with axiolitic textures, and some microcrystalline quartz. In the basal vitrophyre the shards consist of light-brown glass, some of which is bleached and corroded.

The tuffs and tuffaceous sedimentary rocks above the basal ash-flow sheet are typically light gray to light brownish gray, poorly bedded, and moderately to weakly indurated. In the Willowdale-Antelope-Ashwood area these rocks are chiefly lapilli tuff and tuffaceous siltstone, but they become finer grained to the east and grade into green to brownish-gray tuff and tuffaceous claystone.

A fairly well exposed section, near Wilson Creek, consists of about 7 m of gray pumice-lapilli tuff overlain by 35 m of light-brownish-gray, fine-grained tuff and tuffaceous siltstone, 4 m of light-gray, coarse-grained lapilli tuff, and, finally, 4 m of orangish-brown, massive, fine-grained tuff containing 10 to 15 percent pumice lapilli. These tuffs and tuffaceous sedimentary rocks typically contain 1 to 2 percent crystals, chiefly plagioclase, and sparse silicic rock fragments (table 3.8).

Thin, discontinuous basalt flows are interbedded with the tuffs near Wilson Creek and in Antelope Valley between Willowdale and Antelope. In the Wilson Creek area the thickest flow is about 4 m thick and consists of green to black, highly altered amygdaloidal basalt. In Antelope Valley the flows are thin and consist of highly porous, altered basalt with associated tuff, agglomerate,

TABLE 3.8.—*Modal compositions of tuffaceous sedimentary rocks in the John Day Formation*

[Values in volume percent; tr, trace amount; —, not found]

Unit	Member G		Member H			Member I
Sample	S7-6	S1-15	207	S2-9	S2-11	S2-14A
Groundmass	97.6	92.6	93.1	93.8	91.3	95.8
Rock fragments	.7	3.1	4.7	2.5	4.8	2.8
Sanidine	—	—	—	.1	—	—
Quartz	.3	—	—	tr	.1	—
Plagioclase	1.4	3.9	2.1	3.1	1.8	.8
Biotite	—	—	tr	—	—	—
Opaque minerals	—	.2	.1	.5	—	.5
Zircon	—	tr	—	tr	—	—
Other	—	tr	—	—	—	—

and breccia. The agglomerate and breccia are coarse-grained, tan to light-gray or reddish-gray, moderately well bedded rocks containing scoria and basalt fragments, as much as 40 cm across, which crop out in resistant ledges extending for 130 to 150 m along strike. Several accumulations of cinders and scoria as much as 35 m across with rather steep, quaquaversal dips are interpreted as near-vent deposits.

MEMBER H

A basal, fine-grained ash-flow sheet and about 35 m of overlying fine- to medium-grained tuff and tuffaceous sedimentary rocks make up member H. This unit can be traced in nearly continuous outcrop from the vicinity of Smith Rock on the south almost to Fossil on the northeast. The tuffs and tuffaceous sedimentary rocks of member H are similar to those of the overlying and underlying members and cannot be distinguished except where the ash-flow sheet is present. For this reason, the unit cannot be recognized in outcrops west of Willowdale.

The ash-flow sheet ranges from about 2 to 30 m in thickness and consists of a single cooling unit of fine-grained, moderately to densely welded tuff containing sparse lithophysae. In outcrops east of Smith Rock, the ash-flow sheet rests on about 1 m of white, medium- to fine-grained, thin-bedded tuff that appears to be an air-fall precursor of the ash-flow sheet. Commonly present at the base of the sheet is a layer, 0.5 to 2 m thick, of light-gray to black, fine-grained vitrophyre with platy jointing. This layer is overlain by 3 to 30 m of light-gray or brownish-gray stony tuff, which weathers a distinctive orangish brown. In a few outcrops, such as those near Willowdale, the stony tuff is brick red in color and is quarried for building stone. Many outcrops have well-developed platy jointing; others have crude columnar joints. In outcrops east of Madras, where the stony section is thickest, lithophysae as large as 3 cm across are common, particularly in the upper parts of the sheet.

The ash-flow sheet is composed chiefly of shards, and only in a few places does it contain small pumice lapilli or rock fragments. Crystals of zoned oligoclase (table 3.5) average 1 to 2 percent of the rock and are commonly accompanied by traces of quartz (table 3.4).

The overlying tuffs and tuffaceous sedimentary rocks are poorly exposed in low, talus-mantled outcrops. They consist chiefly of poorly bedded, fine- to medium-grained tuff and minor cross-bedded, tuffaceous siltstone and sandstone. Thin layers of light-gray air-fall tuff, 10 to 50 cm thick, form resistant ledges in many sections. Most of the tuffaceous sedimentary rocks have a speckled appearance due to the presence of 5 to 10 percent white pumice lapilli, as much as 5 mm across. Crystals of sodic andesine make up 1 to 2 percent of most of the tuffs, and small lithic fragments are locally present (table 3.8). Some of the tuffs also contain small fragments of black glass, but most glassy material in these rocks has been diagenetically altered to clay minerals and zeolites, chiefly clinoptilolite.

MEMBER I

A distinctive, coarse-grained ash-flow sheet marks the base of member I. This member is overlain by about 5 to 10 m of thin-bedded tuffaceous sedimentary rock, followed by as much as 120 m of white to light-gray massive

tuff and tuffaceous claystone. This member crops out intermittently in a triangular area extending from Smith Rock on the south to Willowdale on the northwest and nearly to Fossil on the northeast. Light-colored massive tuffs and tuffaceous sedimentary rocks similar to those in the upper parts of member I also crop out extensively in the Mutton Mountains and in the area north and northeast of Fossil. In the absence of the basal ash-flow sheet, the exact stratigraphic position of these rocks is uncertain, but they occur near the top of the formation and probably correlate with tuffs of both members H and I.

Member I is thickest in the Willowdale area and generally thins to the northeast and south. In Antelope Valley, part or all of the member was eroded before eruption of the Columbia River Basalt Group, and so locally the basalt flows of that group rest directly on rocks of member H or older units. Between Madras and Smith Rock the Columbia River Basalt Group is absent, and member I is unconformably overlain by the Madras Formation (Deschutes Formation of Farooqui and others, 1981).

The basal ash-flow sheet is easily distinguished from all other tuffs of the John Day Formation by its very coarse grain size. It contains 5 to 30 percent pumice fragments, as large as 10 cm across, and 5 to 10 percent rock fragments. Densely welded sections have a pronounced eutaxitic texture, whereas less welded sections are open and porous. Both the thickness and degree of welding in the sheet decrease from the southwest to the northeast over the outcrop area. South of Hay Creek Ranch the sheet is a compound cooling unit, 35 to 40 m thick, composed of two thick vitrophyres, each overlain by 10 to 25 m of densely to moderately welded, stony tuff containing sparse lithophysae. About 25 km to the north, in the vicinity of Willowdale, the sheet is approximately 35 m thick and consists of a single cooling unit of unwelded, glassy, pumice lapilli tuff. Between Clarno and Fossil, about 50 km northeast of Willowdale, the sheet is 15 to 20 m thick and consists of unwelded, fine-grained vitric tuff (fig. 3.7).

The tuff consists of angular to flattened pumice fragments, rock fragments, and sparse crystals in a matrix of glass shards and dust. In addition, most specimens contain 2 to 5 percent structureless glass fragments, which may be either fresh or devitrified. Rock fragments make up 1 to 10 percent of the tuff and are as much as 1 cm in diameter. Most are composed of silicic, crystalline material, but a few are fine-grained, welded vitric tuff. Crystals make up 1 to 2 percent of the rock and consist chiefly of sodic andesine (table 3.5) accompanied by trace amounts of iron oxides, altered pyroxene, and zircon (table 3.4).

The overlying tuffs and tuffaceous sedimentary rocks are light gray to yellowish or brownish gray, poorly bedded to massive, and medium to coarse grained. Outcrops have steep, even slopes, typically mantled with talus from the overlying Columbia River Basalt Group. Large landslide masses composed of basalt fragments in a tuff matrix are common along the major river canyons.

Near the base of the sequence, the rocks are typically thin-bedded and crossbedded tuffaceous sandstone and siltstone composed chiefly of pumice and small lithic fragments. They also contain 1 to 2 percent andesine crystals, sparse lithic fragments, and 2 to 5 percent small, white, pumice lapilli that give the rocks a speckled appearance. The matrix consists of montmorillonite and zeolites, chiefly clinoptilolite, derived from altered glass. Some of the least altered rocks contain small fragments of black vesicular glass, 1 to 3 mm across.

A rich and well-preserved vertebrate fauna is present in the upper parts of member I, particularly in the Warm Springs area (Woodburne and Robinson, 1977; Dingus, 1979).

NORTHERN SUBFACIES

The ash-flow tuffs and lava flows that are used to define members of the John Day Formation in the western facies thin from southwest to northeast and pinch out against an old Clarno high between Clarno and Fossil (fig. 3.8). Northeast of this high, in the area between Fossil and Lonerock, the stratigraphy of the formation differs markedly from that of the normal western facies. Several ash-flow tuffs and lava flows are present, but these cannot be correlated directly with those of the Antelope-Ashwood sequence.

This lithologically different sequence is thickest and best exposed in the area immediately north of Fossil. Here, the top of the Clarno Formation is marked by a red saprolite, as much as 3 m thick, developed on a series of porphyritic andesite flows. The overlying John Day Formation is about 500 m thick and consists of a lower unwelded ash-flow sheet and an upper sequence of tuff and tuffaceous sedimentary rocks, with several interlayered lava flows.

The basal ash-flow sheet is about 245 m thick and consists of two cooling units separated by a lava flow of quartz latite. The sheet can be traced as far west as the head of Pine Creek east of Clarno, where it appears to interfinger with the upper part of member G of the western facies. To the northeast, it can be traced to the vicinity of Lonerock, where it is considerably thinner and better bedded than in the Fossil area. Except in the vicinity of Fossil, exposures are poor, and outcrops are commonly mantled with talus from the Columbia River Basalt Group or from lava flows within the John Day Formation.

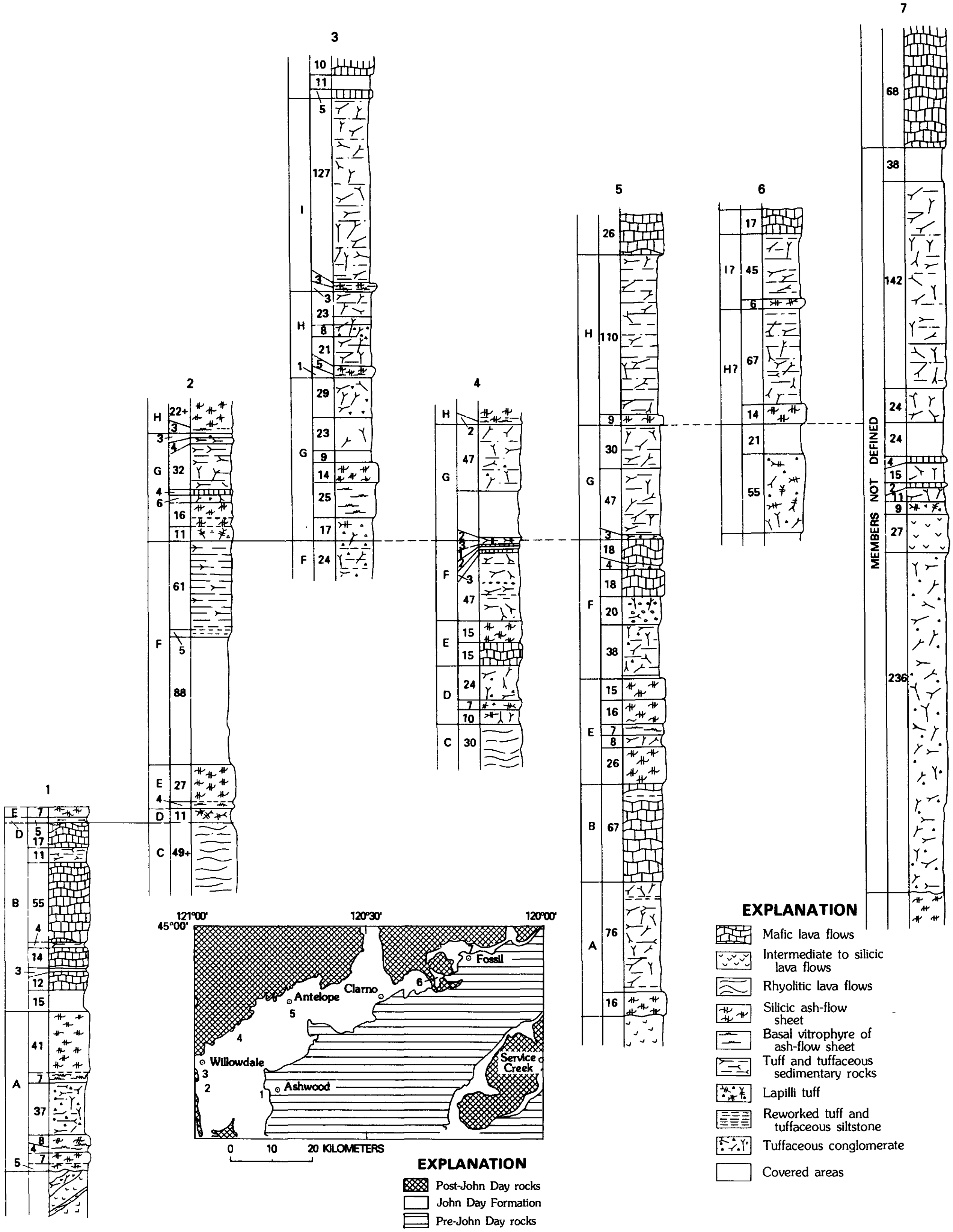


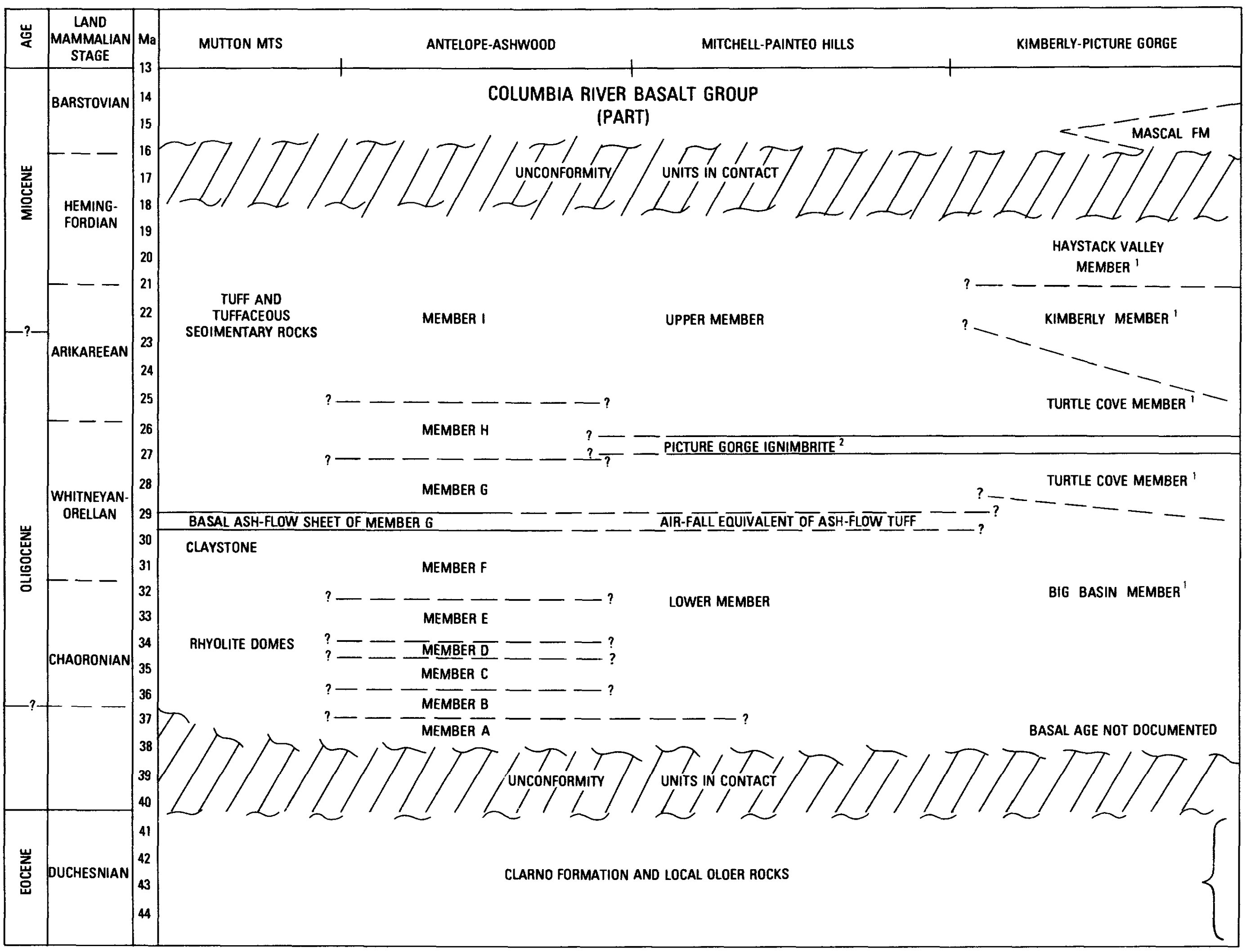
FIGURE 3.7.—Measured sections showing stratigraphic variations in western facies of the John Day Formation in Ashwood-Willowdale-Fossil, Oreg., area. Letters indicate members of formation. Sections measured in meters. Index map shows locations of sections. Horizontal lines connecting sections represent basal ash-flow sheet of indicated members.

In the Fossil area the lower cooling unit is about 235 m thick. At the base it consists of coarse-grained, poorly sorted tuff containing as much as 15 percent angular pumice lapilli and 10 percent dark rock fragments. The unit grades upward into moderately well sorted vitric tuff containing very sparse pumice and rock fragments. The upper cooling unit is 9 to 10 m thick and consists of coarse-grained tuff similar to that in the lower unit. Both cooling units are well indurated but completely unwelded.

Crystals average 1 to 2 percent of each cooling unit and are mostly small prisms of anorthoclase (table 3.5), accompanied by trace amounts of plagioclase, quartz, green pyroxene(?), and zircon (table 3.4). The sparse rock fragments are chiefly andesitic and silicic volcanic material. The original glass has been partly devitrified and partly altered to clinoptilolite.

The ash-flow sheet is overlain by about 25 m of poorly exposed, white to red, medium-bedded, fine- to medium-grained tuff. Locally, a thin amygdaloidal basalt flow is interbedded with the tuff. Overlying the white tuff is a 230-m-thick section of uniform, light-brown to gray, massive to thick-bedded, moderately indurated tuff containing 1 to 2 percent small pumice fragments, which give the rock a slightly speckled appearance. Crystals of andesine feldspar also make up 1 to 2 percent and are accompanied by trace amounts of iron oxide and altered pyroxene (table 3.4). Most of the glass is zeolitized, but angular fragments of fresh black glass occur in the uppermost parts of the section.

Interlayered lava flows are present near the middle of the unit and range from about 2 to 25 m in thickness. The thickest flow is a dark-reddish-brown, flow-banded dacite or quartz latite that crops out discontinuously in a



[1] Unit of Fisher and Rensberger (1972).
[2] Informal unit of Fisher (1966a).

FIGURE 3.8.—Chronostratigraphic relations within the John Day Formation, based on interpretations of existing age data (modified from Woodburne and Robinson, 1977). Time scale from Palmer (1983).

12-km-long band north and west of Fossil (Robinson, 1975). It is a fine-grained, hypocrystalline rock containing sparse microphenocrysts of plagioclase, clinopyroxene, and altered olivine in a hyalopilitic groundmass of plagioclase laths, clinopyroxene granules, magnetite, and abundant brown glass (table 3.6).

Farther east, north of Kinzua, the quartz latite is missing, but a basalt flow occupies approximately the same stratigraphic position (Robinson, 1975). This flow is generally less than 10 m thick and extends for about 8 km along strike, pinching out both to the east and to the west. It consists of light-gray to brick-red, fine- to medium-grained, commonly vesicular and amygdaloidal, alkali olivine basalt, similar to that of member F of the western facies (Robinson, 1969). Locally, the basalt contains megacrysts of alkali feldspar, as large as 2 cm across, probably xenocrysts derived from the pre-Clarno basement.

A thin flow of andesite locally overlies the basalt. This flow is a dark-gray, fine-grained rock containing 2 to 3 percent small plagioclase and altered olivine microphenocrysts in a pilotaxitic groundmass of plagioclase, clinopyroxene, iron oxides, and sparse olivine (table 3.6).

In the area between Fossil and Lonerock, several small, isolated masses of welded tuff crop out in the sequence. These are reddish- to yellowish-gray, densely welded rocks with poorly developed eutaxitic textures and platy jointing. A light-gray basal vitrophyre, about 5 m thick, is preserved in one outcrop. Crystals and rock fragments are sparse, and these rocks appear to be most similar to parts of the basal ash-flow sheet of member H in the western facies. However, because the outcrops are isolated and the tuff has no distinctive features, a definite correlation is not possible.

MUTTON MOUNTAINS SUBFACIES

Rocks of the John Day Formation crop out over an area of about 700 km^2 in the Mutton Mountains area. Most ash-flow sheets of the western facies in the Antelope-Ashwood-Willowdale area are missing in the Mutton Mountains, and the sequence consists chiefly of tuffs and tuffaceous sedimentary rocks, with numerous domes and flows of rhyolite. The Mutton Mountains themselves consist of a series of large rhyolite domes that locally interfinger with the tuffs. Although outcrops in the Mutton Mountains area are not continuous with those of the western facies, correlations are possible on the basis of age and lithology.

In the vicinity of Kahneeta Hot Springs and along the Deschutes River approximately between North Junction and Dant, the John Day Formation rests unconformably on andesitic and dacitic lavas and breccias assigned to the Clarno Formation on the basis of lithology (Waters, 1968a, b). Near Kahneeta Hot Springs, the contact between the two units is marked by a bright-red saprolite, some of which was reworked into the lower part of the John Day Formation. Farther north, along the Deschutes River, no saprolite is preserved, but there is an angular discordance between the two formations.

On the north, east, and southeast flanks of the Mutton Mountains, the John Day Formation is unconformably overlain by flows of the Columbia River Basalt Group. These basalt flows thin and pinch out toward the core of the Mutton Mountains, indicating that the area was a topographic high in middle Miocene time. On the west, between the Mutton Mountains and the Cascade Range, the John Day Formation is overlain directly by the Miocene and lower Pliocene Madras Formation (Deschutes Formation of Farooqui and others, 1981) and younger basalt flows.

At the base of the John Day Formation in the Mutton Mountains is a 20- to 30-m-thick sequence of red tuffaceous claystone, which overlies Clarno lavas near Kahneeta Hot Springs. This sequence is overlain by about 30 m of white to light-brownish-gray, thin- to medium-bedded tuff and lapilli tuff. The thin-bedded zones consist of coarse-grained ash, whereas the thicker bedded zones contain abundant pumice lapilli and lithic fragments, as much as 2 cm across. These weakly to moderately indurated tuffs are completely argillized and zeolitized, and some contain narrow bands and small spots of red material, probably hematite.

The lower tuffs and tuffaceous claystones are overlain by a distinctive, crystal-rich ash-flow tuff that is correlated with the basal ash-flow sheet of member G of the western facies. Locally, the underlying tuffs and claystones pinch out, and the ash-flow sheet rests directly on flow-banded rhyolite of the Mutton Mountains.

The ash-flow tuff, which is about 5 to 10 m thick, is yellowish to reddish brown, moderately welded, and moderately indurated. It contains 7 to 8 percent crystals and 1 to 2 percent rock fragments in a fine-grained matrix composed of shards and glass dust. Locally, sparse pumice fragments impart a poorly developed eutaxitic texture to the rock. The crystals are mostly 1 to 3 mm across and consist chiefly of sanidine and quartz. The sanidine is compositionally and texturally similar to that in the basal ash-flow sheet of member G near Willowdale (table 3.3). A single K-Ar analysis from the tuff in the Mutton Mountains yielded an age of 27.0$\pm$3 Ma (table 3.1), or about the same age as member G.

Overlying the ash-flow tuff is about 100 to 200 m of white to light-gray, poorly bedded tuff and lapilli tuff similar to those of member I near Willowdale. These rocks typically contain 1 to 2 percent plagioclase crystals and sparse lithic fragments. Pumice fragments are common in the coarser grained layers and are as much as

about 5 mm across. The original glassy material is almost completely replaced by montmorillonite and clinoptilolite. Locally, these beds contain well-preserved vertebrate fossils, particularly near the top of the sequence (Woodburne and Robinson, 1977; Dingus, 1979).

The main mass of the Mutton Mountains consists of light-gray to reddish-gray, fine-grained, generally aphyric, flow-banded rhyolite. Thicknesses are estimated to be in excess of 500 m, and the flow banding is commonly steep, possibly indicating a domal structure. Along the eastern margin of the Mutton Mountains, individual flows can be recognized where they are interbedded with layers of white pumice-lapilli tuff. The exact stratigraphic position of these flows and domes is uncertain, but because they underlie the welded tuff dated at 27 Ma and unconformably overlie rocks of Clarno lithology, they are presumed to be equivalent in age to the middle and lower parts of the John Day Formation.

A separate sheetlike mass of rhyolite that overlies John Day tuffs southwest of Kahneeta Hot Springs appears to be a lava flow, locally as much as 200 m thick. The pluglike mass of rhyolite that forms Eagle Butte, about 4 km south of Kahneeta Hot Springs, is believed to mark the vent from which this flow erupted.

ISOLATED RHYOLITE DOMES AND FLOWS

Many isolated masses of flow-banded lava have been mapped as part of the John Day Formation by various workers. Some of these rocks, such as the ones in the Antelope-Ashwood area and the Mutton Mountains, are interlayered with John Day tuffs, and their age assignments are undoubtedly correct. Others along the crest of the Blue Mountains uplift overlie or cut andesitic rocks of typical Clarno lithology but have no field relation to younger rocks. Where age relations have been established for these bodies, they appear to be primarily of Clarno age (Swanson and Robinson, 1968). Still others, such as Cline Buttes, Powell Buttes, Forked Horn Butte, and several unnamed bodies, are completely surrounded by basalt flows and sedimentary rocks of the Madras Formation (Williams, 1957). Only Powell Buttes has been dated by the K-Ar method, and it has a definite John Day age (table 3.1). It is a porphyritic rhyolite (table 3.9)

TABLE 3.9.—*Average chemical compositions of silicic lava flows and domes of the John Day Formation*

[Oxide values in weight percent; * total iron as Fe_2O_3; n.d., not determined; LOI, loss on ignition]

Rock unit	1	2	3	4	5	6
SiO_2	76.3	70.5	81.8	78.7	76.9	75.9
Al_2O_3	12.6	13.8	9.7	11.7	12.4	12.3
Fe_2O_3*	2.2	6.1	1.5	1.3	2.1	1.6
MgO	.06	.10	.01	.01	.02	.14
CaO	.5	1.6	.15	.20	.26	.90
Na_2O	3.2	3.8	2.1	3.5	3.6	3.1
K_2O	5.1	3.6	4.3	4.4	4.4	5.2
TiO_2	.29	.53	.14	.16	.27	.15
P_2O_5	.02	.07	.01	n.d.	.02	n.d.
MnO	.06	.03	.02	.01	.01	.03
Total	100.3	100.1	99.7	100.0	100.0	99.3
LOI	1.1	1.0	.9	.6	.9	1.1
No. of samples	2	2	3	1	4	2

Unit:

1. Rhyolite flows of member C, western facies.
2. Quartz latite flows north of Fossil, Oreg., northern subfacies (Robinson, 1969).
3. Rhyolite of Juniper Butte, western facies.
4. Rhyolite of Powell Buttes, western facies.
5. Rhyolite flows of Smith Rock, western facies.
6. Rhyolite flows and domes of the Mutton Mountains.

containing about 10 percent sanidine, quartz, and sparse amphibole phenocrysts in a microcrystalline groundmass. Several other bodies, including Cline Buttes, consist of pyroxene andesite and are most likely part of the Clarno Formation.

COMPOSITION

The John Day Formation consists of three main components: air-fall tuff and tuffaceous sedimentary rocks, which form the bulk of the unit, welded and unwelded ash-flow tuff, which is abundant in the western facies, and lava flows and domes.

Little direct compositional information is available for the air-fall tuffs and tuffaceous sedimentary rocks because most of the original glass has been replaced by clay minerals and zeolites, either during weathering before burial or by later diagenetic alteration. Chemical analyses of altered rocks indicate extreme hydration accompanied by significant changes in SiO_2, MgO, CaO, Na_2O, and K_2O contents (Hay, 1963). Even in the fresh-glass facies in the upper part of the formation the glass is bleached, etched, and commonly partly replaced by clay minerals. Thus, the initial composition of these rocks can only be deduced from their pyrogenetic mineralogy.

Crystals rarely make up more than 2 to 3 percent of the tuffaceous sedimentary rocks and are chiefly andesine feldspar, accompanied by trace amounts of labradorite, oligoclase, clinopyroxene, iron oxides, biotite, and rare quartz (table 3.8; Hay, 1962a; Fisher, 1968). Amphibole and sanidine are rare accessory minerals. Lithic components in these rocks are chiefly andesitic and dacitic in composition, more rarely rhyolitic. Glass shards have refractive indices ranging from 1.496 to 1.502 (Hay, 1963), implying compositions in the range of about 70 to 74 weight percent SiO_2. These data indicate a dacitic to rhyodacitic composition for most of the original pyroclastic material in the sedimentary rocks.

The interbedded vitric tuffs are commonly somewhat coarser grained than the sedimentary rocks, and many contain small pumice lapilli. Sanidine is present in many of these rocks and equals or exceeds plagioclase in about half of the specimens (Hay, 1963). On average, the plagioclase is somewhat more sodic than in the claystones, being chiefly oligoclase and sodic andesine (Hay, 1962a). Quartz is common in a few tuff layers, particularly those rich in sanidine. Mafic minerals are sparse and consist chiefly of clinopyroxene and iron oxides. On the basis of their mineralogy, the quartz- and sanidine-rich tuffs are inferred to have been originally rhyolitic in composition, whereas the others are inferred to have been dacitic or rhyodacitic.

Both primary mineralogy and whole-rock chemical compositions indicate that the ash-flow tuffs are rhyolitic (table 3.10). Phenocrysts are primarily oligoclase, sanidine, anorthoclase, and quartz, with only traces of such mafic minerals as hornblende, clinopyroxene, biotite, and iron oxides (table 3.4). Quartz and sanidine dominate in the tuffs of members A and G and are present in small amounts in the tuff of member E (table 3.4). Plagioclase is present in all of the ash-flow tuffs except that in member G, but it varies considerably in composition (tables 3.2, 3.5).

Although the tuffs are all rhyolitic in composition, whole-rock compositional data show wide variations even within the same unit (Swanson and Robinson, 1968; Hay, 1963). These variations are believed to be due to differences in devitrification, vapor-phase crystallization, and diagenetic alteration rather than to primary compositional differences. Densely welded, devitrified zones with low porosity are believed to represent the least altered parts of the sheets (Robinson and Brem, 1981). Within these zones, analyses are consistent, K_2O and Na_2O contents are subequal, water content is low, and there is no petrographic evidence of diagenesis or vapor-phase crystallization. Porous glassy material from near the base of individual sheets has been largely altered to clay minerals and zeolites, whereas less porous vitrophyres have been hydrated. Compositions in the upper parts of the sheets appear to have been affected by diagenetic alteration rather than vapor-phase crystallization. The soft, unconsolidated tuff at the top of the Picture Gorge ignimbrite has been completely altered to claystone (Hay, 1963).

Ash-flow tuffs containing abundant quartz and sanidine phenocrysts are undoubtedly rhyolitic, whereas those containing only plagioclase as a phenocryst phase may be either rhyolitic or rhyodacitic. Minor chemical variations must have existed originally within individual ash-flow sheets, but these variations have been obscured by later alteration and hydration. Some original compositional differences between sheets are implied by different phenocryst assemblages. In general, the ash-flow tuffs are slightly alkalic rhyolite containing about 75 to 76 weight percent SiO_2 (table 3.10).

Mafic lavas of the formation range in composition from alkali olivine basalt to trachyandesite. Most basalts contain 15 to 20 percent modal olivine and are undersaturated with respect to silica if the Fe_2O_3/FeO ratios are adjusted to take into account the effects of deuteric alteration and oxidation (Robinson, 1969). The trachyandesites are fine-grained, fresh lavas that contain sparse microphenocrysts of olivine, clinopyroxene, and plagioclase. They are all silica saturated but, like the basalts, are fairly rich in TiO_2 and alkalis (table 3.7). These flows are closely associated in space and time to the basalt flows, but the chemical relation between the two, if any, is unclear (fig. 3.9).

TABLE 3.10.—*Average chemical compositions of ash-flow tuff sheets in the John Day Formation*

[* Total iron as Fe_2O_3; n.d., not determined; LOI, loss on ignition]

Rock unit	1	2	3	4	5	6	7	8	9	10	11
SiO_2	76.2	73.0	76.2	78.9	76.4	77.2	77.3	76.4	74.9	82.2	76.4
Al_2O_3	12.5	12.8	12.6	10.7	12.0	11.6	12.3	12.4	12.9	9.2	13.5
Fe_2O_3*	1.6	3.0	1.5	2.0	2.1	2.0	1.5	2.3	2.6	1.0	1.8
MgO	12	.20	.07	.05	.11	.10	.10	.24	.17	.04	1.0
CaO	7	1.0	.7	.5	1.0	.4	.4	.7	1.0	.6	2.9
Na_2O	2.6	2.0	2.5	2.6	2.3	2.8	3.7	2.8	3.6	2.4	.8
K_2O	5.6	6.6	6.2	4.6	5.1	5.5	4.1	4.7	4.5	4.0	3.9
TiO_2	32	.73	.19	.20	.21	.29	.19	.27	.30	.24	.13
P_2O_5	.06	.23	.01	0	.02	.05	.04	.04	.11	n.d	0
MnO	.05	.14	.06	.02	.05	.02	.03	.05	.03	.03	0
Total	99.89	9.7	100.0	99.6	99.3	100.0	99.7	99.9	100.1	99.7	100.4
LOI	7	1.4	1.5	1.5	3.8	.8	.9	2.3	n.d.	.9	5.7
No. of samples	10	4	5	4	4	10	9	6	7	3	2

Unit:

1. Basal ash-flow sheet of member A, lower cooling unit, western facies.
2. Basal ash-flow sheet of member A, upper cooling unit, western facies.
3. Upper ash-flow sheet of member A, western facies.
4. Lithoidal welded tuff of member E, western facies.
5. Vitrophyric welded tuff of member E, western facies.
6. Basal ash-flow sheet of member G, western facies.
7. Basal ash-flow sheet of member H, western facies.
8. Basal ash-flow sheet of member I, western facies.
9. Picture Gorge ignimbrite, lower cooling unit, eastern facies.
10. Picture Gorge ignimbrite, upper cooling unit, eastern facies.
11. Basal ash-flow sheet north of Fossil, Oreg., northern subfacies.

The silicic lava flows are primarily rhyolite, but one flow of quartz latite is present in the area north of Fossil. The rhyolites are microcrystalline rocks that show little evidence of alteration except for dissolution of rare feldspar phenocrysts. Most specimens are compositionally similar to the ash-flow tuffs, consisting of slightly alkalic rhyolite containing 76 to 82 weight percent SiO_2 and with an Na_2O/K_2O ratio ranging from 0.60 to 0.82. The most siliceous specimens are from domes containing vapor-phase tridymite. The quartz latite flow is significantly less silicic than the rhyolite flows and has an Na_2O/K_2O ratio greater than 1.

AGE

From studies of stratigraphic relations and extensive faunal collections, the John Day Formation was clearly identified as Oligocene and Miocene in age many decades ago. Recent radiometric dating has substantiated this general age assignment, has added greater precision to the age of selected units within the formation, and has provided accurate lower and upper age limits to the entire formation. All of the available K-Ar ages for the formation are listed in table 3.1. Several of the ages listed by Fiebelkorn and others (1983) and identified as from the John Day Formation are repetitions of ages published both by Hay (1962a) and by Evernden and others (1964), and two ages (samples 648–657, 648–695) identified as being from the John Day Formation should be ascribed to the Clarno Formation. The interpreted age relations in the John Day Formation are summarized in figure 3.8.

In the western facies, Peck (1961, 1964), Swanson and Robinson (1968), and Robinson (1975) designated the base of the formation as the base of a distinctive ash-flow sheet that rests unconformably on rocks of Clarno

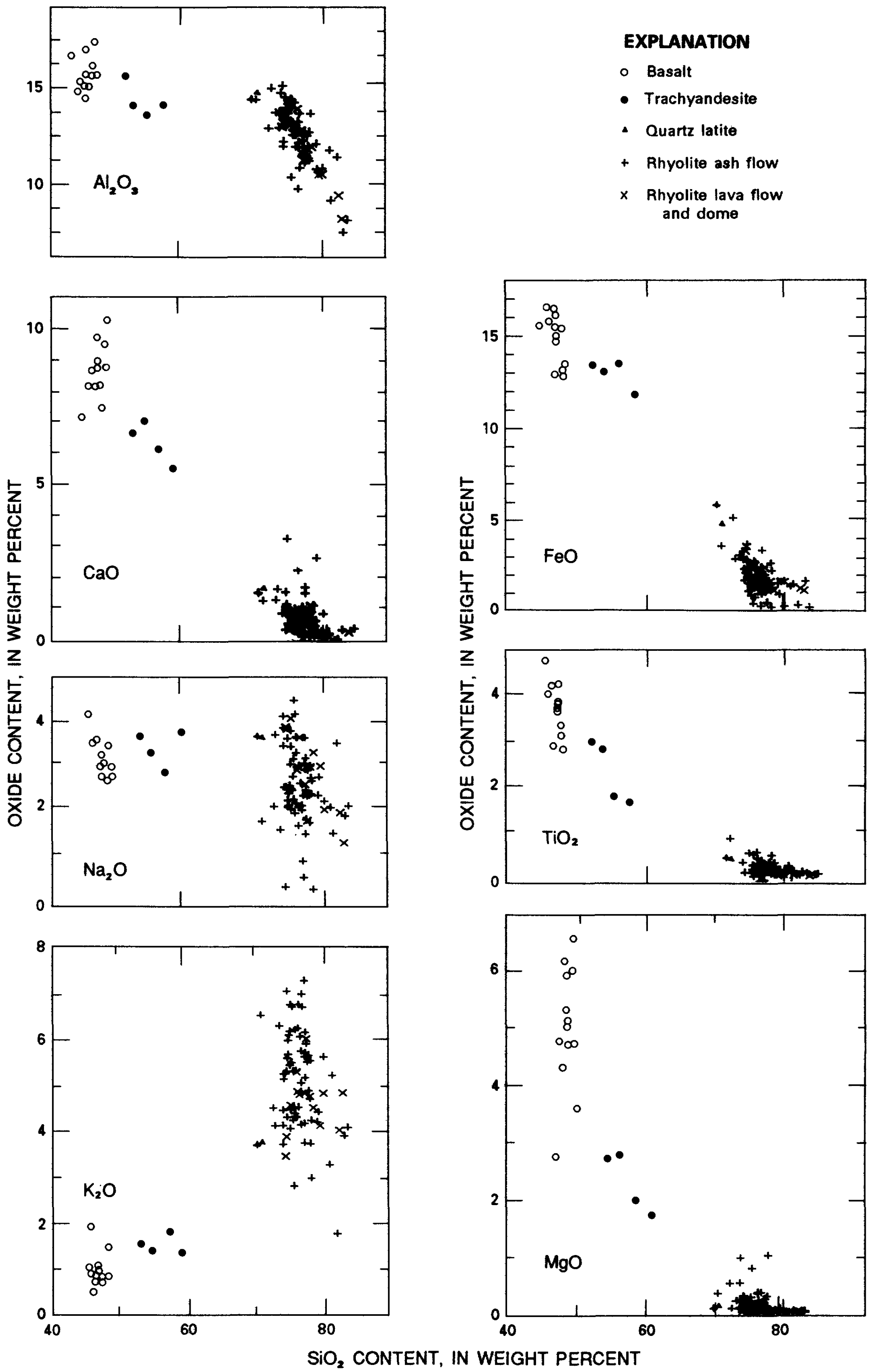


FIGURE 3.9.—Oxide versus SiO_2 content, showing compositional ranges of major rock types within the John Day Formation.

lithology and is separated from them locally by a thin layer of red saprolite. The appearance of this ash-flow sheet marks a change in the style of volcanism from local emplacement of predominately andesitic lavas flows and breccias to regional deposition of predominately silicic pyroclastic material. Two radiometrically dated samples from this sheet (DAS–66–208, DAS–66–195) yielded ages of 37.4 and 37.1 Ma, respectively, indicating an age near the Eocene-Oligocene boundary. This ash-flow sheet can be traced eastward from the Antelope-Ashwood area to the Horse Heaven mining district on the crest of the Blue Mountains uplift (Swanson and Robinson, 1968) and probably as far as Rowe Creek, where an isolated outcrop is found at the base of the eastern facies (Hay, 1962a; Robinson, 1975; Woodburne and Robinson, 1977; Robinson and Brem, 1981). If this correlation is correct, it implies that the base of the formation in the eastern facies is also about 37 Ma.

Several ages have been reported from lavas of the upper part of the Clarno Formation in the Mitchell area that are younger than 37 Ma (Evernden and others, 1964; Enlows and Parker, 1972). However, most of these are whole-rock ages of somewhat altered lava and dike rocks, and the reported ages are suspect. Others are on feldspar separates from bentonite that had been subjected to intense weathering during alteration of the enclosing ash. The feldspars could have lost argon during this process, thus giving rise to younger apparent ages. Taylor (1981) concluded that the anomalously young ages for the Clarno Formation in the Mitchell area are incorrect on geologic grounds and that the age of the base of the John Day Formation is approximately 37 Ma.

The next younger radiometrically dated unit in the western section is the basal ash-flow sheet of member G. Four ages on sanidine separates range from 30.5±0.4 to 27.0±0.3 Ma and average about 28.5 Ma. The range in ages may be due to incorporation of small but varying amounts of accidental or accessory feldspar in the ash-flow sheet, or to slight alteration. On the basis of its unusual feldspar composition, this sheet has been correlated with an air-fall tuff in the eastern facies that lies about 150 to 250 m above the base of the formation (Hay, 1962a, 1963; Woodburne and Robinson, 1977). An early determination (sample KA647A) by Evernden and others (1964) indicated a probable age of 23.3 Ma for this tuff, but a more recent determination (sample 648–456) yielded an age of 26.8±0.4 Ma (table 3.1), which is within the range of ages on the ash-flow sheet of member G.

Several other determinations from the lower member of the eastern facies yielded ages ranging from 31.5 to 26.7 Ma (table 3.1). One age (sample KA1384) of 21.6 Ma from a claystone near the base of the formation was obtained from authigenic feldspar and reflects the age of diagenesis rather than deposition (Hay, 1962a).

K-Ar ages from the basal ash-flow sheets of members H and I of the western facies are 30.9±0.7 and 27.7±0.3 Ma, respectively. Both of these ages are slightly older than the Picture Gorge ignimbrite of the eastern facies, for which two dates are available, 26.9±2.3 and 25.5 Ma (table 3.1). Because the ash-flow tuff of member H is stratigraphically higher than that of member G, its reported age of 30.9 Ma is too old. This tuff is very sparsely porphyritic in plagioclase and may have been contaminated with some slightly older material. The age from the basal ash-flow sheet of member I is probably somewhat in error for the same reason.

The youngest rocks in the formation dated by the K-Ar method are two air-fall tuff beds near the top of the formation in the Mutton Mountains area. These tuff samples have ages of 23.4±3.3 and 22.7±2.7 Ma, and they are associated with a fossil fauna believed to be Hemingfordian in age, approximately 20 to 18 Ma (Woodburne and Robinson, 1977; Dingus, 1979).

Several ages are also available from isolated outcrops of silicic rocks believed to be correlative with the John Day Formation. The oldest of these samples (PTR-71–9A) is from a welded tuff that rests unconformably on rocks of Clarno lithology on Snowboard Ridge about 7 km southeast of Kinzua. This outcrop was tentatively correlated with the basal ash-flow tuff of member A of the western facies (Robinson, 1975), but its reported age of 57.0 Ma is much too old (table 3.1). Either the correlation is wrong and this is a Clarno tuff, or the date is incorrect, possibly because the unit has been contaminated with older material or altered in some way. Powell Buttes, a rhyolite domal complex in Crook County, has three age determinations ranging from 30.1±1.1 to 25.8±0.2 Ma, indicating a definite John Day age. Thus, in summary, in this report we consider the John Day Formation to range in age from latest Eocene(?) to early Miocene.

STRUCTURE

Rocks of the John Day Formation are generally little deformed and typically dip less than 15°. Most structures within the formation are subparallel to the major tectonic features of north-central Oregon (fig. 3.10). Small northeast-trending folds are present on both flanks of the Blue Mountains uplift, and a few east-trending structures parallel the Ochoco uplift (Swanson, 1969). Several small, tight folds are present in the western facies adjacent to faults or small intrusions (Peck, 1964; Robinson, 1975). Northeast-trending faults are common on the west flank of the Blue Mountains uplift, where they offset all rocks older than the Madras Formation of Miocene and early Pliocene age. These are chiefly normal faults, with throws in the range 2 to 100 m, but a few exhibit minor strike-slip displacement. Along one rather large fault

system, between Willowdale and Grizzly, the John Day rocks are exposed in a west-tilted block with dips between 15° and 30°.

The distribution and stratigraphy of the Tertiary units in north-central Oregon indicate that a topographic high has persisted along the Blue Mountains axis from at least early Oligocene time (Rogers, 1966; Fisher, 1967; Robinson, 1975; Taylor, 1977). In Eocene time, numerous Clarno vents were active along this trend (Swanson, 1969), indicating that it was a volcanotectonic feature. By earliest John Day time, however, erosion had produced a terrain of little relief, and the earliest ash-flow sheet of

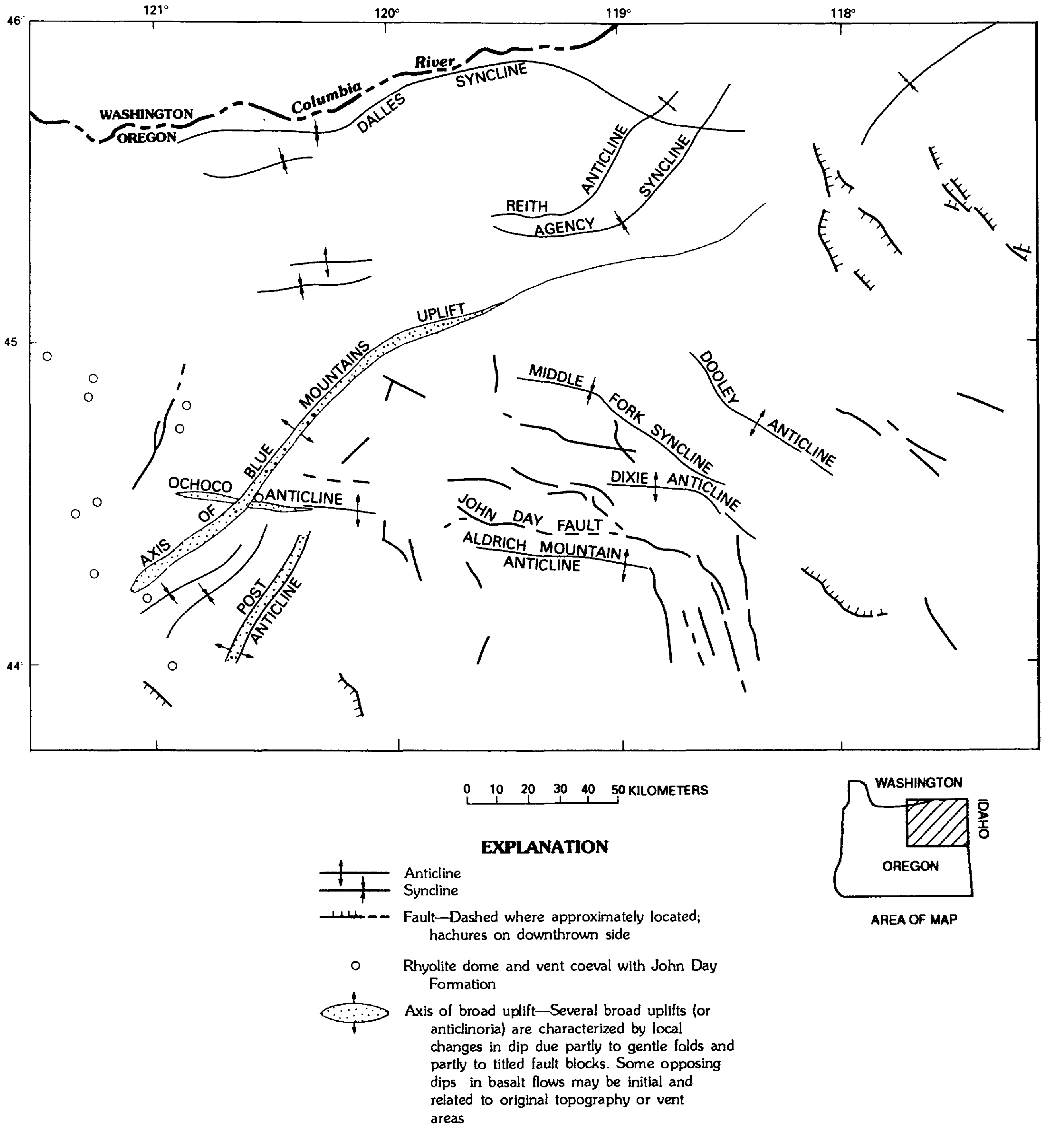


FIGURE 3.10.—North-central Oregon, showing major tectonic features (modified from Walker, 1977).

the John Day Formation (member A of the western facies) was deposited across this axis (Swanson and Robinson, 1968). After deposition of member A, uplift along a northeastward trend produced a topographic high that acted as a barrier between the eastern and western facies. The ash-flow sheets of members E, G, H, and I of the western facies thin to the east and pinch out against this Clarno high, as do the intervening tuffs and tuffaceous sedimentary rocks (Robinson, 1975; Woodburne and Robinson, 1977). The Picture Gorge ignimbrite, which was erupted east of the Blue Mountains uplift, did not extend into the western facies. Air-fall tuff, mostly derived from vents in the present-day Cascade Range, was deposited across the entire area.

Continued uplift along the Blue Mountains axis is indicated by the angular discordance between John Day tuffs and lava flows of the overlying Columbia River Basalt Group. In the area northeast of Kinzua, the John Day Formation pinches out entirely, and the Columbia River Basalt Group there rests directly on rocks of the Clarno Formation. Deformation since the middle Miocene is indicated by broad, open folds and local faults in the Columbia River Basalt Group. However, most tectonic activity ceased before deposition of younger parts of the Madras Formation (Deschutes Formation of Farooqui and others, 1981), probably about 10 to 8 Ma. Rocks of this unit are nearly flat lying and only rarely cut by faults.

ORIGIN

The John Day Formation is a compositionally heterogeneous unit that consists of air fall and ash-flow tuff, tuffaceous sedimentary rocks, and various lava flows and domes. The thickness of the unit varies extremely, ranging as high as about 750 m in a given section. The maximum age span of the formation is about 37 to 19 Ma. The bulk of the unit in the eastern, southern, and western facies consists of tuffaceous sedimentary rocks, chiefly claystones and siltstones with relict shard textures. Most of these rocks are believed to have formed from material deposited as thin ash falls on the land surface (Hay, 1962a, 1963; Fisher, 1966b). Such structures as root cavities and animal burrows in the sedimentary rocks and pahoehoe surfaces on some of the lava flows provide direct evidence of subaerial deposition. Reworking of the ash by roots, burrowing organisms, and sheetwash eliminated the initial stratification and produced the massive bedding that is so characteristic of these rocks (Hay, 1962a, 1963). Some reworked tuffs in the lower part of the unit forms very thin bedded lacustrine deposits, whereas those in the upper part of the unit are fluviatile.

On the basis of age data then available, Hay (1963) calculated an average sedimentation rate for the lower member of the eastern facies of about 36 mm per 1,000 yr. More recent K-Ar ages indicate that the 335-m-thick lower member spans a period of about 12 m.y. rather than the 7 m.y. estimated by Hay, implying an average sedimentation rate of about 28 mm per 1,000 yr. The upper member, which is about 550 m thick, was deposited over a period of about 5 m.y., indicating an average rate of accumulation of about 110 mm per 1,000 yr.

The tuffaceous sedimentary rocks become progressively thicker and coarser grained from east to west (Waters, 1954; Peck, 1964; Robinson, 1975), indicating a probable source for these materials west of the present-day outcrops. The pyrogenetic mineralogy of these rocks implies a generally dacitic composition similar to that of the pyroclastic rocks of the western Cascades (Peck and others, 1964). K-Ar ages from western Cascade rocks indicate extensive volcanism coeval with deposition of the John Day Formation in latest Eocene(?), Oligocene, and early Miocene time (Fiebelkorn and others, 1983). The sorting characteristics of the pyroclastic material in the eastern facies are compatible with transportation for distances of 50 to 150 km (Hay, 1963; Fisher, 1966b). These relations indicate that that the bulk of the pyroclastic material in the John Day Formation was derived from volcanos in the western Cascades. Hay (1962a, 1963) suggested that the claystone in the lower part of the formation formed by weathering at the land surface, whereas that of the Picture Gorge ignimbrite presumably formed after burial. Probably both weathering and diagenesis were responsible for formation of the claystone in the upper member.

The air-fall tuffs interlayered with the tuffaceous claystones and siltstones differ from the sedimentary rocks in being commonly coarser grained and in having somewhat more silicic compositions. They may have been formed by more violent eruptions from the same volcanos that produced the pyroclastic material in the sedimentary rocks or may, in part, have come from vents east of the Cascades, such as those now marked by rhyolite domes in the western facies. The fact that the tuffs retain their original bedding characteristics implies burial before extensive weathering and reworking had taken place.

Most of the ash-flow sheets in the western facies become thicker, coarser grained, more intensely welded, and richer in lithophysae from east and northeast to west and southwest, implying sources in that direction. These sheets cannot be traced to any known source and presumably were erupted from vents west of the present-day outcrops. Because the tuffs are coarse grained and apparently more silicic than coeval deposits in the western Cascades, they are believed to have been derived from vents west of long 121° W. but east of the Cascade Range. A separate source for these rocks is also indicated by their rare-earth geochemistry (I.G. Gibson

and P.T. Robinson, unpub. data, 1985), which differs significantly from that of Cascade rocks (Lux, 1981). The ash-flow tuff sheets of member E in the western facies lie in an old topographic depression between Ashwood and Willowdale and pinch out in all directions from this area. This unit is closely related in space, time, and composition to the rhyolite flows of member C and may have been derived from the same vent, believed to be marked by a rhyolite dome about 6 km north of Ashwood.

The ash-flow tuff sheets vary sufficiently in age and composition to indicate that they did not originate from a single source. Most have volumes in the range 10 to 20 km^3, larger than expected for the many small vents in the western facies marked by rhyolite domes, such as Juniper Butte or Powell Buttes. However, eruption of these volumes of material would probably not produce collapse calderas. The large rhyolite domes in the Mutton Mountains were originally considered as possible sources for some of the ash-flow sheets, but only the tuff of member G can be traced into that area, and it is thinner, finer grained, and less densely welded there than in outcrops to the east and south. In summary, most of the ash-flow sheets appear to have originated from vents west of long 121° W. but east of the Cascades. This area is now largely covered by the Madras (or Deschutes) Formation, and only a few isolated outcrops of older rock are present.

The Picture Gorge ignimbrite crops out only in the eastern facies. It was originally thought to correlate with member H of the western facies (Robinson, 1966), but its lithology and distribution indicate that it is probably a separate unit (Fisher, 1966a). It has an estimated volume of about 40 km^3, suggesting a fairly large vent system. From the variations in thickness, crystal content, and size of lithic fragments, Fisher (1966a) inferred a possible source somewhere in the Ochoco Mountains southwest of Mitchell; however, no vent has yet been found in that area.

Both the mafic and silicic lava flows in the John Day Formation were erupted from local vents, many of which are still preserved. Feeder dikes for the mafic flows have been found in the vicinity of Antelope, Mitchell, and Service Creek (Lindsley, 1961; Hay, 1963; Robinson, 1975), and deposits of basaltic agglomerate are common locally. Peck (1964) suggested that the voluminous trachyandesite flows of member B of the western facies were erupted from a vent in sec. 14, T. 11 S., R. 16 E., Foley Butte (7.5-minute) quadrangle, but this suggestion has not been confirmed. However, even where known vents are missing, the distribution and thickness variations of individual flows leave no doubt that their origin was local.

Most of the silicic lava flows within the formation can be related to vents now marked by rhyolite domes or plugs. The largest rhyolite flow (member C of the western facies) is believed to have been erupted from a vent about 6 km north of Ashwood. Other small flows are associated with Juniper Butte and with domes in the Mutton Mountains. A source has not been located for the quartz latite flow north of Fossil, but it probably lies north of the present-day outcrops and in an area buried beneath flows of the Columbia River Basalt Group.

UNUSUAL ROCKS OF JOHN DAY AGE

Within the Blue Mountains province, isolated sequences of comparatively fresh lava flows and pyroclastic rocks have been recognized that are lithologically similar to rocks of the Clarno Formation but whose stratigraphic relations and age indicate that they are at least in part coeval with the John Day Formation. Little is known of these basaltic to silicic volcanic rocks in terms of their regional distribution, volume, petrology, and origin; almost invariably, they have been mapped either as part of the Clarno Formation or, in a few places, as anomalous parts of the John Day Formation. Examples of these enigmatic sequences are found at Smith Rock in the northeast corner of Deschutes County (Peterson and others, 1976; Robinson and Stensland, 1979), southwest of the Maury Mountains in southern Crook County (Walker and others, 1967), in the headwaters of the Grande Ronde River in southern Union County (Walker, 1973), and in several areas marginal to Unity Basin in western Baker County (Brown and Thayer, 1966; Brooks and others, 1976). Other sequences of lithologically similar but undated rocks east of the main outcrop area of the Miocene and Pliocene Strawberry Volcanics (Brown and Thayer, 1966) are suspect and may be coeval with either the upper part of the John Day Formation or the lower part of the Strawberry Volcanics, which have been radiometrically dated at about 20 Ma (Robyn, 1977).

Although in the field most of these rocks look superficially like rocks of the Clarno Formation and consist commonly of hornblende-, pyroxene-, and plagioclase-porphyritic andesite and dacite lava and breccia, aphyric platy andesite and basaltic andesite, and rhyolitic to dacitic tuff and lapilli tuff, they generally differ in degree and type of alteration from most rocks of the Clarno Formation. Rocks of the Clarno Formation typically show some evidence of deuteric and low-temperature hydrothermal alteration to montmorillonite, secondary silica and carbonate minerals, zeolites, hematite, and, locally, chlorite. Areas of most intense alteration are commonly adjacent to small intrusions of andesite, dacite, or rhyolite. The lithologically similar but younger rocks are typically much less altered, although in a few places reddish-brown iron oxide, probably hematite, occurs on fractures and in vesicles. Hornblende is commonly replaced by magnetite and other unidentified

alteration products of late deuteric or diagenetic processes.

At Smith Rock a sequence of andesitic and basaltic lavas, laharic breccias, and tuffs, at least 1,000 m thick, is overlain conformably by a succession of rhyolite flows and tuffs, which, in turn, are succeeded by white tuffs of typical John Day lithology. The tuffs of typical John Day lithology are capped by flows of the Columbia River Basalt Group (Robinson and Stensland, 1979). The lower andesitic sequence is similar lithologically to rocks of the Clarno Formation and locally is overlain unconformably by ash-flow tuffs of the upper part of the John Day Formation (Robinson and Stensland, 1979). Williams (1957) originally included the entire sequence in the John Day Formation, but Robinson and Stensland (1979) mapped the lower andesitic sequence as part of the Clarno Formation. However, a single K-Ar analysis recently obtained from the upper part of this sequence indicates an age of 30.8±0.5 Ma, well within the recognized age span of the John Day Formation (Fiebelkorn and others, 1983). More recent work suggests that most of this sequence is equivalent in age to the John Day Formation or even younger (Walter Obermiller, oral commun., 1987).

Southwest of the Maury Mountains, largely unaltered, platy, plagioclase-, pyroxene-, and hornblende-pyroxene-porphyritic andesite flows rest with discordance on deeply weathered and altered flows and breccias typical of the Clarno Formation; neither sequence has been isotopically dated, but apparent interfingering of the overlying, unaltered unit with tuffaceous sedimentary rocks that in nearby areas contain a vertebrate fauna typical of the John Day Formation indicates that it may be coeval with some part of that formation.

In the upper drainage basin of the Grande Ronde River, unaltered to moderately altered platy, aphyric, commonly trachytic andesite flows and associated plagioclase-, pyroxene-, and hornblende-porphyritic andesite flows and monolithologic flow breccias rest unconformably on pre-Cenozoic rocks and are overlain (discordantly?) by dacitic to rhyolitic volcaniclastic rocks, flows, perlite, and coarse breccias, which, in turn, are overlain by flows of the Columbia River Basalt Group. The andesitic flows and breccias were originally mapped as part of the Clarno Formation, but more recently they have been considered coeval with part of the John Day Formation. Hornblende in the andesite flows commonly is baked and largely converted to opaque iron oxides and other unidentified alteration products, and fractures in a few places contain zeolites presumably derived from groundmass feldspar. A K-Ar analysis on fresh plagioclase phenocrysts from a porphyritic hornblende andesite (Walker, 1973) indicates an age of 28.8±0.8 Ma (Fiebelkorn and others, 1983), approximately in the midrange of radiometric ages on the John Day Formation. A K-Ar age on biotite from the (discordantly?) overlying silicic assemblage is 27.7±0.8 Ma. Whether all of the andesitic rocks in this area are of late Oligocene age is not known.

In the area southwest of Lonerock and along Oregon Highway 207 about 15 km to the east, fairly fresh basalt crops out along the crest of the Blue Mountains uplift. These rocks underlie sequences of white tuffaceous claystone of typical John Day lithology and appear to rest on more altered lava flows of the Clarno Formation. Exposures are poor, however, and the structural relations of these lavas are unclear. On the basis of their stratigraphic position, Robinson (1975) included these rocks in the Clarno Formation, whereas Walker (1977) mapped them as part of the John Day Formation because they are fresher than most Clarno lavas. They are medium-grained aphyric rocks with pilotaxitic to subophitic textures composed of plagioclase laths, clinopyroxene, iron oxides, and about 5 percent olivine. Petrographically, they are more similar to the alkali-olivine basalt flows of the John Day Formation than to the calc-alkalic lavas of the Clarno Formation. However, alkalic rocks are present in both units (Robinson, 1969), and composition is not a reliable guide to stratigraphic position. Resolution of this problem will have to await more detailed studies, including radiometric-age determinations.

Exposed in areas mostly northeast, east, and southeast of the Unity Reservoir are rocks that have been mapped as the Clarno Formation. These rocks consist of a diverse assemblage of hornblende- and plagioclase-porphyritic andesitic to dacitic flows and flow breccias, tuffs, a few thin basalt or basaltic andesite flows, and intrusions related to local eruptive centers. In the lava flows containing both plagioclase and quartz crystals, some of the plagioclase and all of the quartz show resorption features implying that they are xenocrysts. These rocks are mostly unaltered, and none shows evidence of the type of alteration that characterizes the Clarno Formation. K-Ar ages on hornblende and plagioclase from a monolithologic andesitic to dacitic breccia originally mapped as the Clarno Formation are 19.6±0.8 and 19.5±0.6 Ma, respectively. The breccia is approximately coeval with the youngest part of the John Day Formation, as represented by rhyolitic tuffs and tuffaceous sedimentary rocks of late Hemingfordian (early Miocene) age exposed near Warm Springs along the Deschutes River (Woodburne and Robinson, 1977), and, possibly, also with the oldest part of the Strawberry Volcanics.

REFERENCES CITED

Brooks, H.C., McIntyre, J.R., and Walker, G.W., 1976, Geology of the Oregon part of the Baker 1° by 2° quadrangle: Oregon Department

of Geology and Mineral Industries Geologic Map Series Map GMS–7, scale 1:250,000.

Brown, C.E., and Thayer, T.P., 1966, Geologic map of the Canyon City quadrangle, northeastern Oregon: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–447, scale 1:250,000.

Brown, R.W., 1957, Plantlike features in thunder eggs and geodes: Smithsonian Institution Annual Report, 1956, p. 329–339.

Calkins, F.C., 1902, A contribution to the petrography of the John Day Basin [Oreg.]: University of California, Department of Geology Bulletin, v. 3, p. 109–172.

Chaney, R.W., 1927, Geology and paleontology of the Crooked River basin, with special reference to the Bridge Creek flora, *in* Additions to the paleontology of the Pacific coast and Great Basin regions of North America: Carnegie Institution of Washington Publication 346, p. 45–138.

Coleman, R.G., 1949, The John Day Formation in the Picture Gorge quadrangle, Oregon: Corvallis, Oregon State University, M.S. thesis, 211 p.

Condon, Thomas, 1871, The rocks of the John Day Valley: Overland Monthly, May issue, p. 393.

Curtis, G.H., Savage, D.E., and Evernden, J.F., 1961, Critical points in the Cenozoic: New York Academy of Science Annals, v. 91, art. 2, p. 342–351.

Dingus, L.W., 1979, The Warm Springs fauna (mammalia Hemingfordian) from the western facies of the John Day Formation, Oregon: Riverside, University of California, M.S. thesis, 98 p.

Enlows, H.E., and Parker, D.J., 1972, Geochronology of the Clarno igneous activity in the Mitchell quadrangle, Wheeler County, Oregon: Ore Bin, v. 34, no. 6, p. 104–110.

Evans, S.H., and Brown, F.H., 1981, Summary of potassium-argon dating—1981: Washington, U.S. Department of Energy, Division of Geothermal Energy Report ID–12079–45, 29 p.

Evernden, J.F., Savage, D.E., Curtis, G.H., and James, G.T., 1964, Potassium-argon dates and the Cenozoic mammalian chronology of North America: American Journal of Science, v. 262, no. 2, p. 145–198.

Farooqui, S.M., Beaulieu, J.D., Bunker, R.C., Stensland, D.E., and Thomas, R.E., 1981, Dalles Group—Neogene formations overlying the Columbia River Basalt Group in north-central Oregon: Oregon Geology, v. 43, no. 10, p. 131–140.

Fiebelkorn, R.B., Walker, G.W., MacLeod, N.S., McKee, E.H., and Smith, J.G., 1983, Index to K-Ar age determinations for the State of Oregon: Isochron/West, no. 37, 60 p.

Fisher, R.V., 1962, Clinoptilolite tuff from the John Day Formation, eastern Oregon: Ore Bin, v. 24, no. 12, p. 197–203.

———1963, Zeolite-rich beds of the John Day Formation, Grant and Wheeler Counties, Oregon: Ore Bin, v. 25, no. 11, p. 185–197.

———1964, Resurrected Oligocene hills, eastern Oregon: American Journal of Science, v. 262, no. 6, p. 713–725.

———1966a, Geology of a Miocene ignimbrite layer, John Day Formation, eastern Oregon: University of California Publications in Geological Sciences, v. 67, 73 p.

———1966b, Textural comparison of John Day volcanic siltstone with loess and volcanic ash: Journal of Sedimentary Petrology, v. 36, no. 3, p. 706–718.

———1967, Early Tertiary deformation in north-central Oregon: American Association of Petroleum Geologists Bulletin, v. 51, no. 1, p. 111–123.

———1968, Pyrogenic-mineral stability, lower member of the John Day Formation: University of California Publications in Geological Sciences, v. 75, 36 p.

Fisher, R.V., and Rensberger, J.M., 1972, Physical stratigraphy of the John Day Formation, central Oregon: University of California Publications in Geological Sciences, v. 101, 45 p.

Fisher, R.V., and Wilcox, R.E., 1960, The John Day Formation in the Monument quadrangle, Oregon, *in* Short papers in the geological sciences: U.S. Geological Survey Professional Paper 400-B, p. B302–B304.

Hay, R.L., 1962a, Origin and diagenetic alteration of the lower part of the John Day Formation near Mitchell, Oregon, *in* Engle, A.E.J., James, H.L., and Leonard, B.F., eds., Petrologic studies—A volume in honor of A.F. Buddington: New York, Geological Society of America, p. 191–216.

———1962b, Soda-rich sanidine of pyroclastic origin from the John Day Formation of Oregon: American Mineralogist, v. 47, no. 7–8, p. 968–971.

———1963, Stratigraphy and zeolite diagenesis of the John Day Formation of Oregon: University of California Publications in Geological Sciences, v. 42, no. 5, p. 199–262.

Hodge, E.T., 1932a, Geologic map of north-central Oregon: Oregon State College Studies in Geology 1, scale 1:250,000.

———1932b, New evidence of the age of the John Day Formation: Geological Society of America Bulletin, v. 43, no. 3, p. 695–702.

Lindsley, D.H., 1961, Geology of the Spray quadrangle, Oregon: Baltimore, Johns Hopkins University, Ph.D. dissertation, 236 p.

Lux, D.R., 1981, Geochronology, geochemistry, and petrogenesis of basaltic rocks from the western Cascades, Oregon: Columbus, Ohio State University, Ph.D. dissertation, 171 p.

Marsh, O.C., 1875, Ancient lake basins of the Rocky Mountain region: American Journal of Science, ser. 3, v. 9, p. 49–52.

Merriam, J.C., 1901, A contribution to the geology of the John Day Basin [Oreg.]: University of California Publications, Department of Geology Bulletin, v. 2, p. 269–314.

Oles, K.F., and Enlows, H.E., 1971, Bedrock geology of the Mitchell quadrangle, Wheeler County, Oregon: Oregon Department of Geology and Mineral Industries Bulletin 72, 62 p.

Palmer, A.R., 1983, The Decade of North American Geology 1983 geologic time scale: Geology, v. 11, no. 9, p. 503–504.

Peck, D.L., 1961, John Day Formation near Ashwood, in north-central Oregon, art. 343 *of* Short papers in the geologic and hydrologic sciences: U.S. Geological Survey Professional Paper 424–D, p. D153–D156.

———1964, Geological reconnaissance of the Antelope-Ashwood area, north-central Oregon, with emphasis on the John Day Formation of late Oligocene and early Miocene age: U.S. Geological Survey Bulletin 1161–D, 26 p.

Peck, D.L., Griggs, A.B., Schlicker, H.G., Wells, F.G., and Dole, H.M., 1964, Geology of the central and northern parts of the western Cascade Range in Oregon: U.S. Geological Survey Professional Paper 449, 56 p.

Peterson, N.V., Groh, E.A., Taylor, E.M., and Stensland, D.E., 1976, Geology and mineral resources of Deschutes County, Oregon: Oregon Department of Geology and Mineral Industries Bulletin 89, 66 p.

Renton, J.L., 1951, Some notes on thunder eggs: Mineralogist, v. 19, no. 4, p. 171–177.

Robinson, P.T., 1966, Facies changes in John Day Formation, Oregon [abs.]: Geological Society of America, Rocky Mountain Section Annual Meeting, Las Vegas, Nev., 1966, Program, p. 49.

———1969, High-titania alkali-olivine basalts of north-central Oregon, U.S.A.: Contributions to Mineralogy and Petrology, v. 22, no. 4, p. 349–360.

———1975, Reconnaissance geologic map of the John Day Formation in the southwestern part of the Blue Mountains and adjacent areas, north-central Oregon: U.S. Geological Survey Miscellaneous Investigations Series Map I–872, scale 1:125,000.

Robinson, P.T., and Brem, G.F., 1981, Guide to geologic field trip between Kimberly and Bend, Oregon, with emphasis on the John Day Formation, *in* Johnston, D.A., and Donnelly-Nolan, Julie, eds., Guides to some volcanic terranes in Washington, Idaho,

Oregon, and northern California: U.S. Geological Survey Circular 838, p. 29–40.

Robinson, P.T., and Stensland, D.H., 1979, Geologic map of the Smith Rock area, Jefferson, Deschutes, and Crook Counties, Oregon: U.S. Geological Survey Miscellaneous Investigations Series Map I–1142, scale 1:48,000.

Robyn, T.L., 1977, Geology and petrology of the Strawberry Volcanics, NE Oregon: Eugene, University of Oregon, Ph.D. dissertation, 224 p.

Rogers, J.J.W., 1966, Coincidence of structural and topographic highs during post-Clarno time in north-central Oregon: American Association of Petroleum Geologists Bulletin, v. 50, no. 2, p. 390–396.

Ross, C.S., 1941, Origin and geometric form of chalcedony-filled spherulites from Oregon: American Mineralogist, v. 26, no. 12, p. 727–732.

Steiger, R.H., and Jäger, Emilie, 1977, Subcommission on geochronology—Convention on the use of decay constants in geo- and cosmochronology: Earth and Planetary Science Letters, v. 36, no. 3, p. 359–362.

Swanson, D.A., 1969, Reconnaissance geologic map of the east half of Bend quadrangle, Crook, Wheeler, Jefferson, Wasco, and Deschutes Counties, Oregon, *in* Geological Survey Research, 1968: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–568, scale 1:250,000.

Swanson, D.A., and Robinson, P.T., 1968, Base of the John Day Formation in and near the Horse Heaven mining district, north-central Oregon, *in* Geological Survey Research, 1968: U.S. Geological Survey Professional Paper 600–D, p. 154–161.

Swinney, C.M., Waters, A.C., and Miller, C.P., 1968, Reconnaissance geologic map of the Lookout Mountain quadrangle, Crook and Wheeler Counties, Oregon: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–543, scale 1:62,500.

Taylor, E.M., 1960, Geology of the Clarno Basin, Mitchell quadrangle, Oregon: Corvallis, Oregon State University, M.S. thesis, 173 p.

———1977, The Clarno Formation—A record of early Tertiary volcanism in central Oregon [abs.]: Geological Society of America Abstracts with Programs, v. 9, no. 6, p. 768.

———1981, A mafic-dike system in the vicinity of Mitchell, Oregon, and its bearing on the timing of Clarno-John Day volcanism and early Oligocene deformation in central Oregon: Oregon Geology, v. 43, no. 8, p. 107–112.

Walker, G.W., 1973, Reconnaissance geologic map of the Pendleton quadrangle, Oregon and Washington: U.S. Geological Survey Miscellaneous Investigations Series Map I–727, scale 1:250,000.

———1977, Geologic map of Oregon east of the 121st meridian: U.S. Geological Survey Miscellaneous Investigations Series Map I–902, scale 1:500,000.

Walker, G.W., Peterson, N.V., and Greene, R.C., 1967, Reconnaissance geologic map of the east half of the Crescent quadrangle, Lake, Deschutes, and Crook Counties, Oregon: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–493, scale 1:250,000.

Waters, A.C., 1954, John Day Formation west of its type locality [abs.]: Geological Society of America Bulletin, v. 65, no. 12, pt. 2, p. 1320.

———1968a, Reconnaissance geologic map of the Dufur quadrangle, Hood River, Sherman, and Wasco Counties, Oregon: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–555, scale 1:125,000.

———1968b, Reconnaissance geologic map of the Madras quadrangle, Jefferson and Wasco Counties, Oregon: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–556, scale 1:125,000.

———1968c, Reconnaissance geologic map of the Post quadrangle, Crook County, Oregon: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–542, scale 1:62,500.

Waters, A.C., and Vaughan, R.H., 1968a, Reconnaissance geologic map of the Eagle quadrangle, Crook County, Oregon: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–540, scale 1:62,500.

———1968b, Reconnaissance geologic map of the Ochoco Reservoir quadrangle, Crook County, Oregon: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–541, scale 1:62,500.

Waters, A.C., Brown, R.E., Compton, R.R., Staples, L.W., Walker, G.W., and Williams, Howel, 1951, Quicksilver deposits of the Horse Heaven mining district, Oregon: U.S. Geological Survey Bulletin 969–E, p. 105–149.

Wilcox, R.E., and Fisher, R.V., 1966, Geologic map of the Monument quadrangle, Grant County, Oregon: U.S. Geological Survey Geologic Quadrangle Map GQ–541, scale 1:62,500.

Williams, Howel, 1957, A geologic map of the Bend quadrangle, Oregon, and a reconnaissance geologic map of the central portion of the High Cascade Mountains: Portland, Oregon Department of Geology and Mineral Industries, scales 1:125,000 and 1:250,000.

Woodburne, M.O., and Robinson, P.T., 1977, A new late Hemingfordian mammal fauna from the John Day Formation, Oregon, and its stratigraphic implications: Journal of Paleontology, v. 51, no. 4, p. 750–757.

4. THE COLUMBIA RIVER BASALT GROUP AND ASSOCIATED VOLCANIC ROCKS OF THE BLUE MOUNTAINS PROVINCE

By PETER R. HOOPER[1] and DONALD A. SWANSON

CONTENTS

ABSTRACT

Between about 17.5 and 6.0 Ma, flows of the Columbia River Basalt Group were erupted from north-northwest-oriented fissures concentrated in the accreted terrane(s) of the Blue Mountains province. The flows spread northward over the cratonic crust of eastern Washington and westward into Pasco Basin, where they accumulated as a pile as much as 4.5 km thick.

In this chapter, the stratigraphic succession is briefly described and the characteristic features of the flows within each unit are given. Feeder systems and the areal extents of each unit are illustrated.

The Columbia River Basalt Group and associated volcanic rocks are subdivided on the basis of their isotopic compositions. Each of the six subdivisions is distinct in its time-space relations, and most possess a unique range of composition. The first subdivision, the main series of the Columbia River Basalt Group consisting of the Imnaha, Grande Ronde, and Wanapum Basalts, contains over 90 percent (by volume) of the group and was erupted from the Chief Joseph dike swarm in the eastern part of the Blue Mountains province. This subdivision displays a small stepped increase in the $^{87}Sr/^{86}Sr$ ratio (and a corresponding decrease in the $^{143}Nd/^{144}Nd$ ratio) over time and may record varying degrees of mixing between at least two source compositions. The subdivision represented by the Saddle Mountains Basalt was also erupted from the Chief Joseph dike swarm and represents the last minor episodes of Columbia River Basalt Group volcanism. The Saddle Mountains flows have varying compositions and high but varying values of the ratio $^{87}Sr/^{86}Sr$; they lie unconformably on earlier flows, and there is evidence of prolonged periods of deformation and erosion between eruptions.

The Picture Gorge Basalt and the other three remaining subdivisions are restricted to the southern margin of the Columbia Plateau, each with its own feeder system. The Picture Gorge was erupted from the Monument dike swarm in John Day Basin and is coeval with part of the eruption of the Grande Ronde Basalt farther northeast. The isotopic signature of the Picture Gorge Basalt implies an origin from a depleted mantle source. The basalt of Prineville was erupted from unknown vents close to the east edge of the Cascade Range. It is remarkably homogeneous and was erupted in late Grande Ronde time. The basalts of Powder River and Weiser were erupted during Wanapum and Saddle Mountains time; compositionally, they resemble the Slide Creek Member and other lower units of the Strawberry Volcanics.

Deformation of the Columbia River Basalt Group was apparently continuous throughout the approximately 11 m.y. of eruption and continues today. Structural elements include: (1) southeastward to northwestward tilting due to isostatic rise of the Idaho batholith and other plutons to the east, together with depression of Pasco Basin to the west; (2) older structural discontinuities, including the suture zone between the thicker cratonic crust and the thinner accreted terranes, as well as a pre-Miocene northwest-southeast and northeast-southwest fracture pattern; and (3) a regional stress system, including a horizontal north-northwest/south-southeast direction of maximum stress and a horizontal west-southwest/east-northeast direction of minimum stress, which resulted in potential north-northwest/south-southeast extensional fissures at depth that became feeders for the basalt flows, approximately northwest-southeast (right-lateral) and northeast-southwest (left-lateral) strike-slip faults at more moderate depths, and approximately east-west anticlinal folds associated with reverse faults near the surface where the vertical direction became the direction of minimum stress. In addition, clockwise rotation of individual blocks has occurred, and preliminary work indicates that such rotation may be concentrated adjacent to the suture zone and the southeastward extension of the Olympic-Wallowa lineament.

Genesis of the Columbia River Basalt Group and related volcanic rocks was complex. Multiple sources are required by isotopic data, and crystal fractionation of plagioclase, olivine, and some pyroxene of as

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much as 50 percent may have occurred between flows within the same isotopic sequence. Magma mixing has also occurred and may have been widespread in the Grande Ronde Basalt. Of the many unanswered questions, some of the most profound are why the Columbia River Basalt Group erupted when and where it did, what were the depth and compositions of the various mantle sources, and what was the relative significance of crustal contamination in their genesis. It is particularly difficult to distinguish among crustal contamination of a mantle source, a mantle-enrichment process, and contamination by crustal assimilation as the magma made its way to the surface.

INTRODUCTION

The Blue Mountains province of northeastern Oregon and adjacent parts of Washington and Idaho is defined by pre-Tertiary rocks that form a complex of one or more allochthonous terranes accreted to the west side of the North American craton during the Mesozoic. These accreted terranes are separated from the craton by a suture zone running northward along the west side of the Idaho batholith and eastward through southeastern Washington, defining a 90° reentrant angle (fig. 4.1; Hamilton, 1962; Armstrong and others, 1977; Fleck and Criss, 1985; Mohl and Thiessen, in press).

The northwestern margin of the Blue Mountains province is marked by the Klamath-Blue Mountains lineament (KBML; Riddihough and others, 1986), a sharp geophysical discontinuity corresponding to the northwestward limit of accreted-terrane rocks and trending parallel to their regional foliation. The west edge of the cratonic crust under the Columbia River Basalt Group in eastern Washington is not well defined but almost certainly trends northward along the east side of Pasco Basin (fig. 4.1). The composition of the crust beneath the basalt in the triangular area between the Cascades in the west, the KBML, and the cratonic margin east of Pasco Basin is not known. Available geophysical data are consistent with 4.5 km of the Columbia River Basalt Group overlying a wedge of sedimentary rocks resting directly on oceanic crust.

In contrast to the thick Precambrian and lower Paleozoic continental crust of the craton, the accreted terrane(s) is of oceanic origin—a complex of Devonian to Jurassic island-arc volcanic rocks, oceanic lithosphere, and fore-arc graywacke turbidites, capped by carbonate and associated metasedimentary units and welded together by Jurassic and younger granitoids (Brooks and Vallier, 1978; Dickinson and Thayer, 1978; Vallier, 1985). The units have a pronounced northeastward structural trend at all scales and are wedged into the craton's reentrant angle as the southern limb of a structural bend that has been called the Columbia arc (Carey, 1958; Wise, 1963).

The accreted terrane is largely covered by Eocene to Pliocene volcanic rocks. Of these rocks, the lower to upper Miocene Columbia River Basalt Group is dominant in the northern part of the province, and lava flows spread northward and westward from there to form the Columbia Plateau (fig. 4.1).

The Columbia River Basalt Group was erupted between about 17.5 and 6.0 Ma (Watkins and Baksi, 1974; McKee and others, 1977, 1981; Swanson, Wright, and others, 1979; Long and Duncan, 1982) from north-northwest-trending fissures (Taubeneck, 1970; Swanson and others, 1975; Swanson, Wright, and others, 1979) concentrated in the southeastern part of the Columbia Plateau and almost wholly within the Blue Mountains province. Most eruptions took place within the province, although much lava flowed long distances downslope and now is found north and west of the province. A few dikes, however, crossed the east-west suture zone to feed eruptions on the craton in southeastern Washington (fig. 4.1; also, see figs. 4.4, 4.8), and the feeder dikes of the Ice Harbor Member of the Saddle Mountains Basalt (Swanson and others, 1975; Swanson, Wright, and others, 1979) lie entirely outside the Blue Mountains province. Also, Taubeneck (1970) reported dikes within the Idaho batholith that resemble dikes of the Columbia River Basalt Group.

A clear consensus on the origin and evolution of the Columbia River Basalt Group has not yet been reached, but field, petrographic, chemical, and isotopic data clearly indicate that different flow sequences were derived from different sources by a wide variety of processes (Carlson, 1984; Hooper, 1984a, 1985). The probability of multiple sources was recognized in early studies of the Columbia River Basalt Group (Waters, 1961; Wright and others, 1973).

With these various sources and processes in mind, we divide the Columbia River Basalt Group and associated volcanic rocks into six sequences: (1) the main series of the Columbia River Basalt Group (Imnaha, Grande Ronde, and Wanapum Basalts), (2) the Saddle Mountains Basalt of the Columbia River Basalt Group, (3) the Picture Gorge Basalt of the Columbia River Basalt Group, (4) the basalt of Prineville, (5) the basalt of Powder River and associated volcanic rocks in Oregon, and (6) the basalt of Weiser and associated volcanic rocks in Idaho.

The Strawberry Volcanics, of middle to late Miocene age (Robyn, 1977, 1979), are discussed in chapter 5. As noted below, the Slide Creek Member, the basal unit of the Strawberry Volcanics, is probably correlative with both the basalts of Powder River and Weiser, and the flows of these two sequences are thus best considered part of the Strawberry Volcanics. The basalt of Bear Creek occurs farther west and may be of similar age; this unit was described by Goles (1986) and is not reviewed here.

The rocks of all six sequences have been included in the Columbia River Basalt Group by various authors (Waters, 1961; Uppuluri, 1974; Swanson, Wright, and others, 1979; Carlson and others, 1981; Swanson and others, 1981; Fitzgerald, 1984). Their relative ages, where known, are shown in figure 4.2. Their approximate extents are shown in figures 4.1, 4.4, 4.5, and 4.8 to 4.10. Each sequence is distinctive in its time-space relations and, with the possible exceptions of the Powder River and Weiser rocks, can be recognized by a distinctive range of chemical and isotopic compositions.

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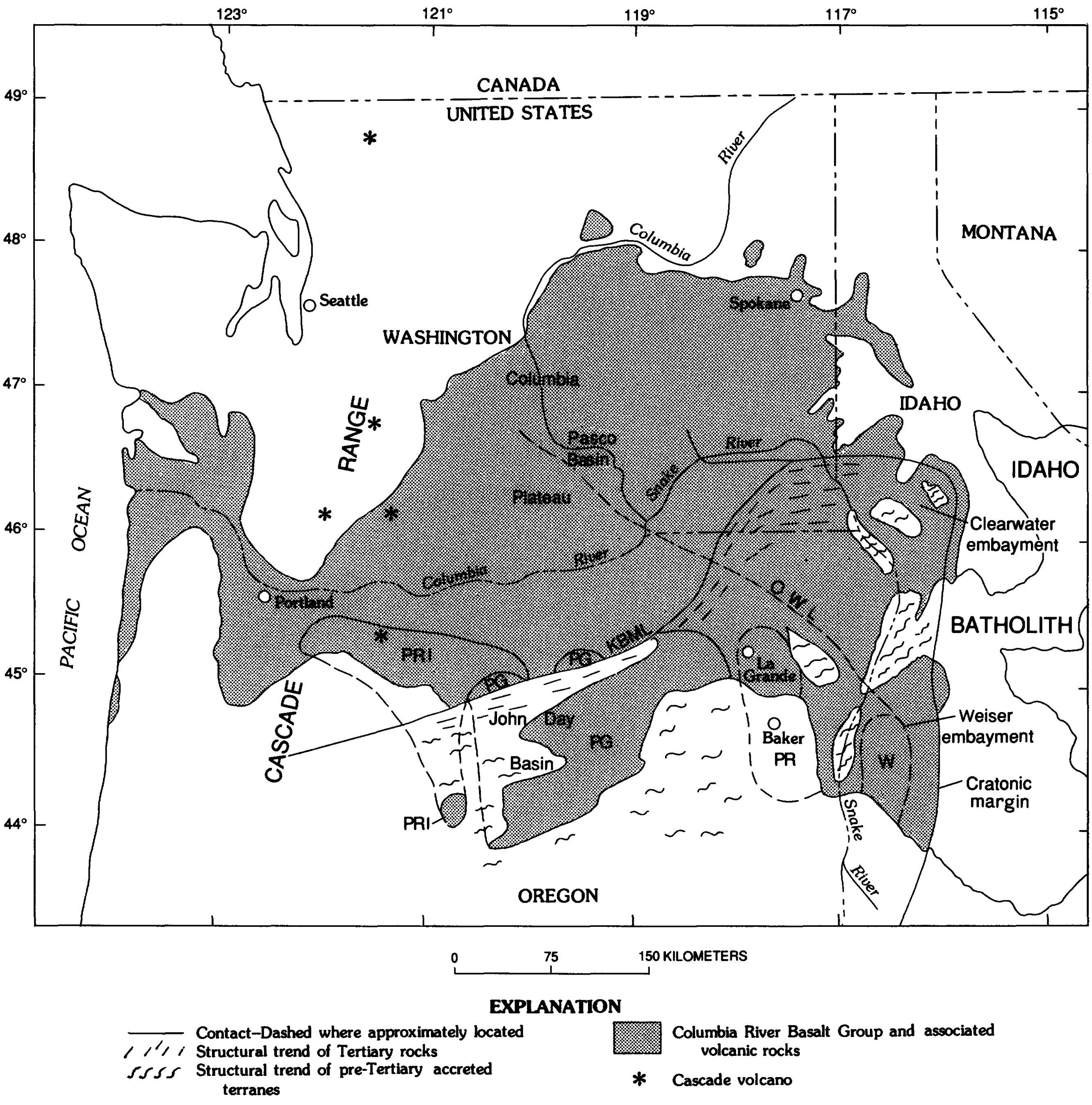


FIGURE 4.1.—Northwestern United States, showing distribution of the Columbia River Basalt Group and associated volcanic rocks. Blue Mountains province lies between cratonic margin on east and Klamath-Blue Mountains line (KBML) in west. OWL, Olympic-Wallowa lineament; PG, Picture Gorge Basalt; PR, basalt of Powder River; PRI, basalt of Prineville; W, basalt of Weiser.

basalt of Prineville and Chris Hawkesworth for permission to use his $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ values, and we are indebted to G.G. Goles for his careful review of the manuscript.

MAIN SERIES

The Imnaha, Grande Ronde, and Wanapum Basalt constitute at least 90 percent (probably more than 95 percent) of the volume of the whole Columbia River Basalt Group, including the associated volcanic sequences shown in figure 4.2. These three units were all erupted from the broad Chief Joseph dike swarm in the southeastern part of the Columbia Plateau (Taubeneck, 1970). Four petrographically distinct units (Imnaha Basalt, Grande Ronde Basalt, Eckler Mountain Member of the Wanapum Basalt, and main part of the Wanapum Basalt) can be recognized on the basis of chemical and isotopic composition (fig. 4.3; Waters, 1961; Swanson, Wright, and others, 1979; Carlson, 1984; Hooper, 1984a). The differences among these units indicate that they could not have originated from a single primary magma derived by partial melting of a homogeneous source.

IMNAHA BASALT

Early eruptions of the Columbia River Basalt Group produced coarsely porphyritic (plagioclase+olivine+

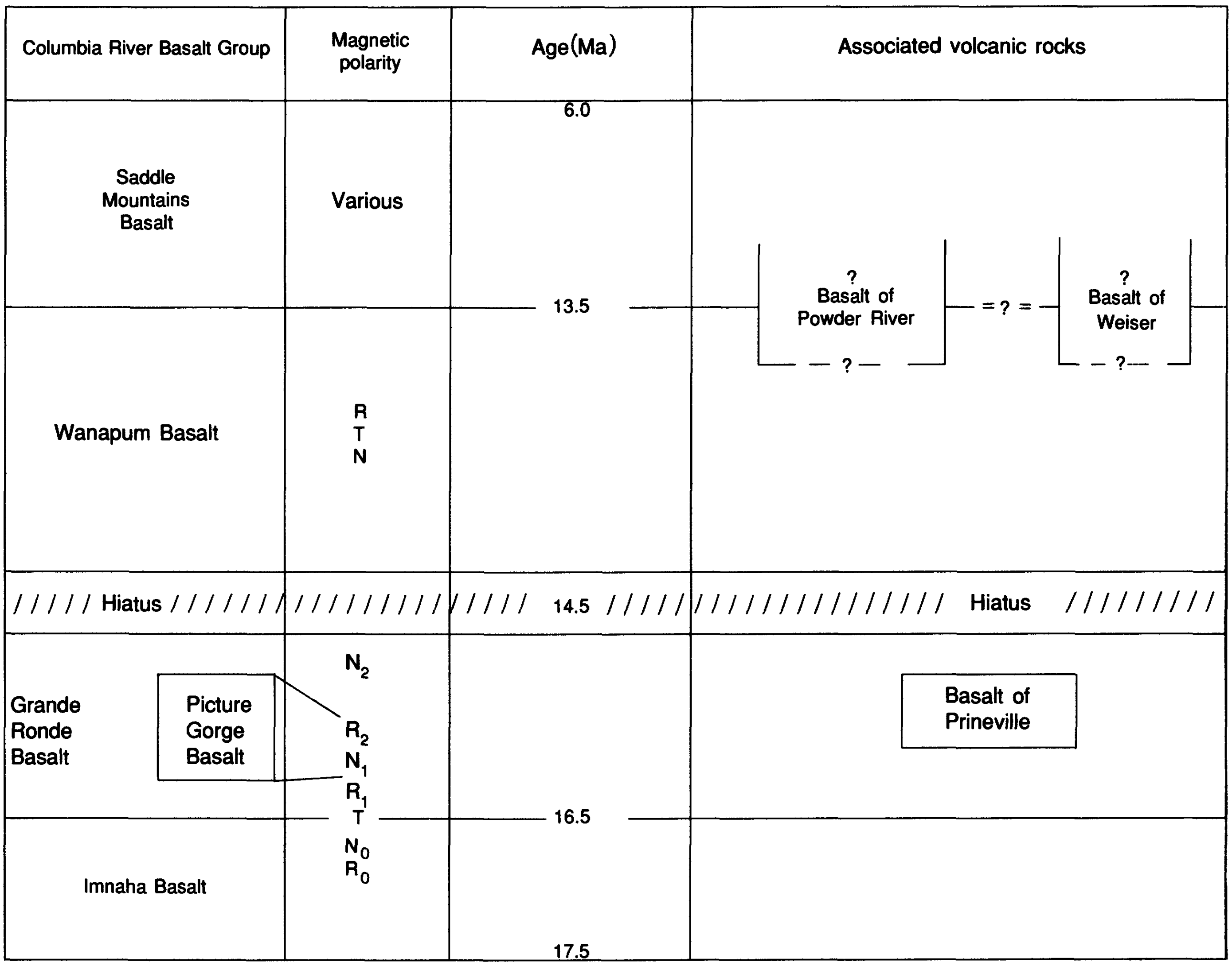


FIGURE 4.2.—K-Ar ages (Watkins and Baksi, 1974; McKee and others, 1977, 1981). Age of the Vantage Member of the Ellensburg Formation, between the Grande Ronde and Wanapum Basalts, is 15.6 Ma, determined by $^{40}Ar/^{39}Ar$ methods (Long and Duncan, 1982). Magnetic polarities: N, normal; R, reversed; T, transitional. Polarity intervals numbered from oldest to youngest (for example, N_0 to N_3) where sequential.

augite) flows about 17.5 to 16.5 Ma (McKee and others, 1981; Long and Duncan, 1982). These flows make up the Imnaha Basalt and cover at least 30,000 km^2 in western Idaho and adjacent parts of Oregon and Washington (fig. 4.4; Camp and Hooper, 1981; Fitzgerald, 1984; Hooper and others, 1984). Their north-northwest-trending feeder dikes and vent complexes are located in Oregon close to the Idaho-Oregon State line, largely within Waters' (1961) Cornucopia dike swarm (fig. 4.4). The fissures crossed the major east-west topographic divide defined by the Wallowa and Seven Devils Mountains. Flows of similar composition filled deep canyons north and south of

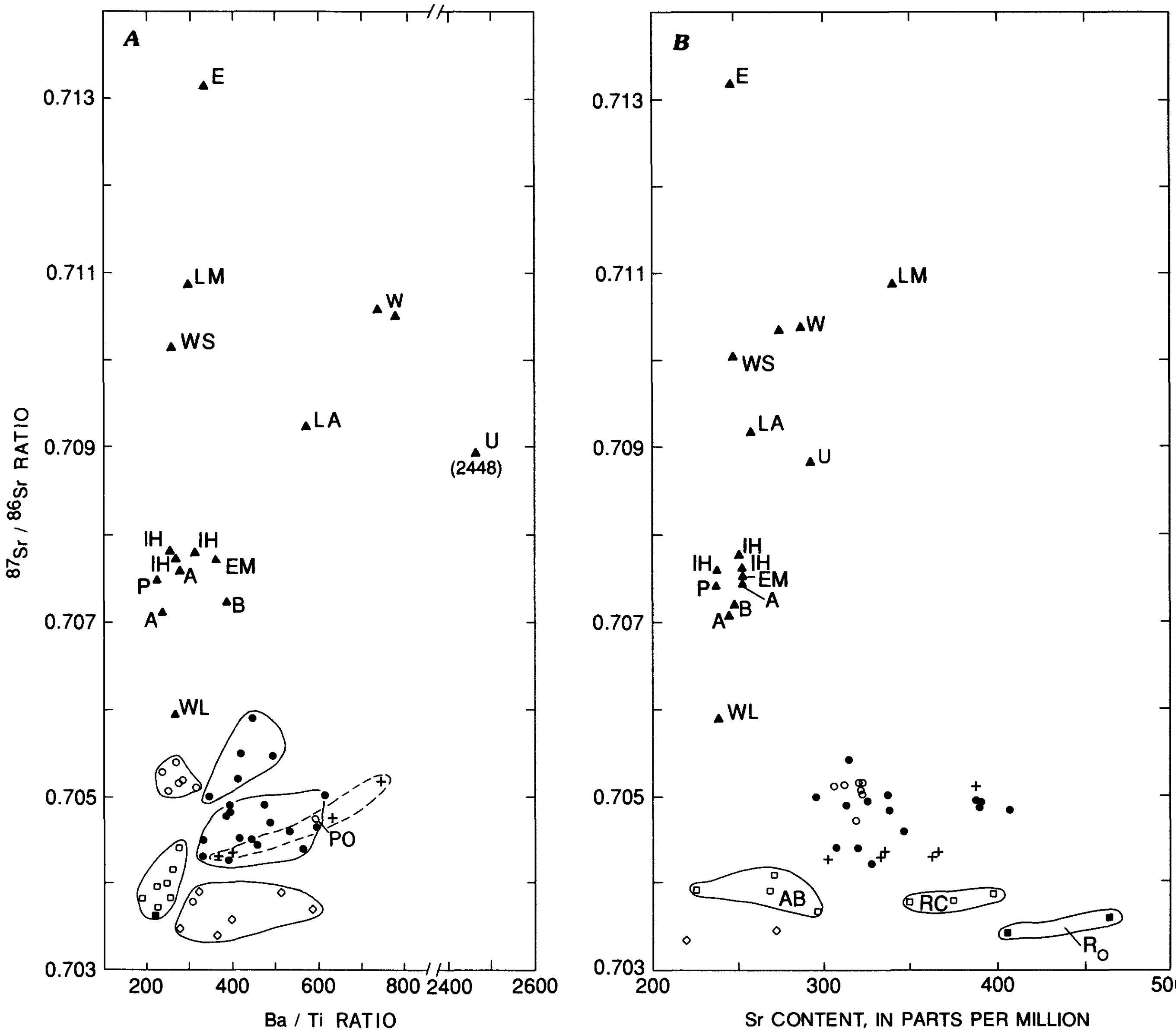


FIGURE 4.3.—Plots of $^{87}Sr/^{86}Sr$ ratios against Ba/Ti ratios and Sr contents for flows of the Columbia River Basalt Group. □, Imnaha Basalt (N_0 flows); ■, Imnaha Basalt (R_0 flows); ◇, Picture Gorge Basalt; ●, Grande Ronde Basalt; +, Eckler Mountain Member (of the Wanapum Basalt); ○, main part of the Wanapum Basalt (PO, Powatka flow); ▲, Saddle Mountains Basalt (A, Asotin; B, Buford; E, Esquatzel; EM, Elephant Mountain; IH, Ice Harbor; LA, Lapwai; LM, Lower Monumental; P, Pomona; U, Umatilla; W, Wilbur Creek; WL, Lewiston Orchard flow; WS, Slippery Creek flows). Data from McDougall (1976), Carlson and others (1981), Carlson (1984), and C.J. Hawkesworth (unpub. data, 1984). Magnetic polarities: N, normal; R, reversed. Polarity intervals numbered from oldest to youngest (for example, N_0 to N_3) where sequential. *A*, $^{87}Sr/^{86}Sr$ ratio against Ba/Ti ratio. *B*, $^{87}Sr/^{86}Sr$ ratio against Sr content. AB and RC, American Bar and Rock Creek subunits, respectively, of the Imnaha Basalt with reverse magnetic polarity. Analyses from McDougall (1976) are omitted because they are consistently biased to lower Sr contents relative to those of Carlson and others (1981) and analyses in this study (table 4.2) by approximately 40 ppm.

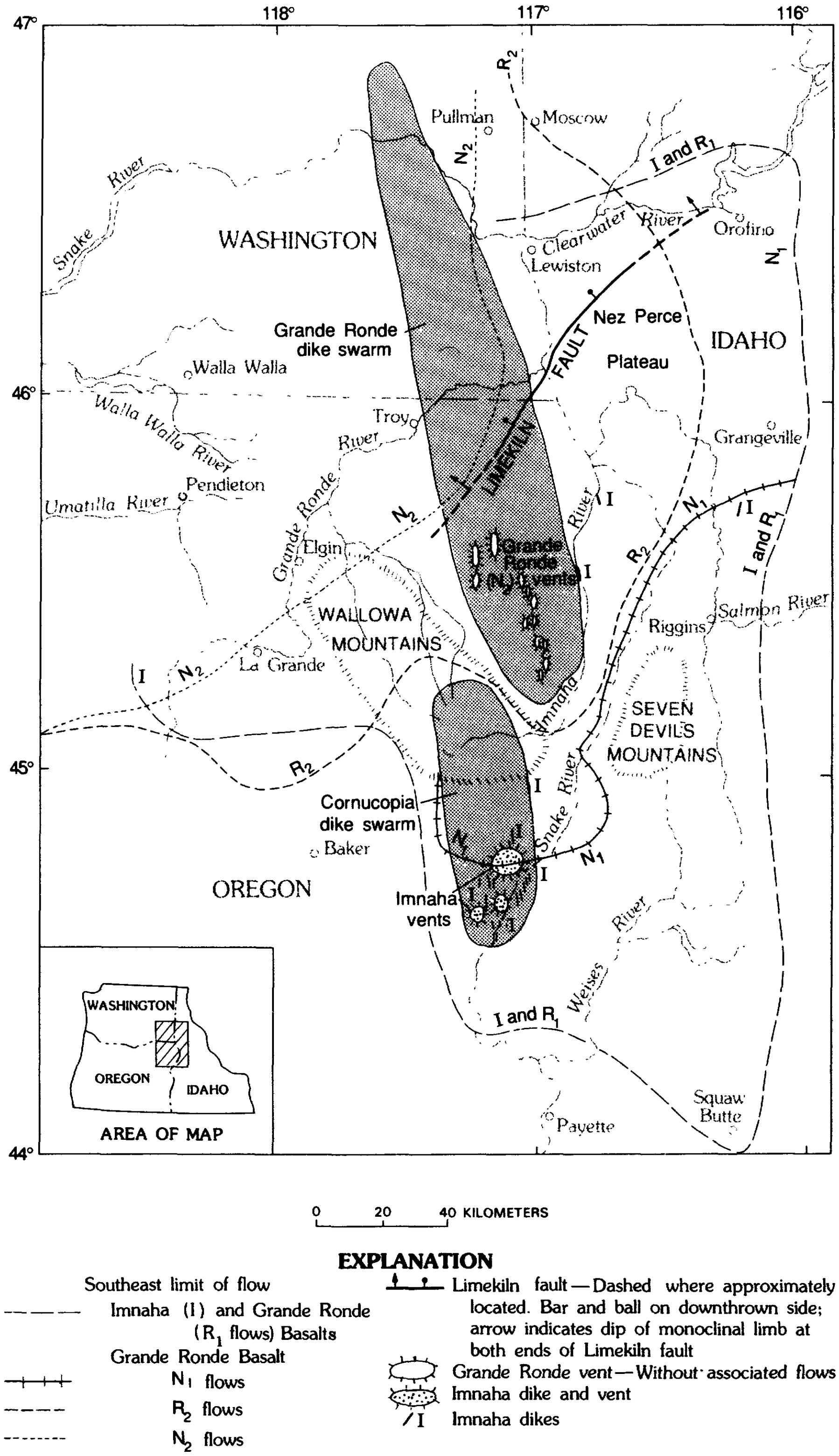


FIGURE 4.4.—Blue Mountains region, showing distribution of flows, dikes, and vents of the Imnaha and Grande Ronde Basalts. Grande Ronde dikes and vents are concentrated in two main swarms (shaded), the Grande Ronde dike swarm in the north and the Cornucopia dike swarm in the south. Known dikes of the Imnaha Basalt are marked separately. Magnetic polarities: N, normal; R, reversed. Polarity intervals numbered from oldest to youngest (for example, N_1 to N_2) where sequential.

the divide and covered all but the highest peaks of the two ranges. As the canyons filled and the basalt plain expanded laterally, however, successive flows became thinner though covering larger areas. This change permitted the continuing gradual uplift of the predominantly granitic mountain blocks to overtake the rising tide of basalt. As a result, subsequent flows of the Grande Ronde Basalt sequentially offlap the rising mountains (Camp and Hooper, 1981), even though eruptions were more frequent and possibly more voluminous than during Imnaha time (Shaw and Swanson, 1970; Hooper, 1981, fig. 3).

In the Imnaha Basalt, the lowest flows exposed at Squaw Butte (fig. 4.4) and on China Cap Ridge on the southwest flank of the Wallowa Mountains have reversed magnetic polarity (fig. 4.2; Swanson, Wright, and others, 1979; Fitzgerald, 1984; Martin, 1984), but nearly all flows have normal magnetic polarity, with transitional to reversed polarity in the youngest flows (Hooper and others, 1979).

Hooper and others (1984) provided stratigraphic, petrologic, and chemical data for the Imnaha Basalt north of the Wallowa-Seven Devils divide. Two basalt types are distinguished within the Imnaha Basalt (Kleck, 1976): The American Bar (AB) type dominates the lower part of the stratigraphic section, and the Rock Creek (RC) type dominates the upper part, although the two types interfinger throughout the section. The AB type is recognized in the field by its fine-grained groundmass, plagioclase+olivine+augite phenocryst assemblage, and minor amounts of olivine. The RC type has a coarser grained groundmass, much more abundant and conspicuous olivine, and no augite phenocrysts.

Representative major- and trace-element analyses and isotopic data for the Imnaha Basalt are presented in tables 4.1 and 4.2. The two Imnaha types show a considerable overlap in the concentrations of most elements, but AB flows tend to have higher Al_2O_3 and MgO contents and lower SiO_2 contents than RC flows. The two flow types are clearly distinguished by their different concentrations of Sc, V, Ni, and Sr. Fractionation dominated by plagioclase appears to have been a major process relating individual flows within each type. Major-element modeling demonstrates that the fractionated assemblage is similar to the observed phenocryst assemblage for each type (Hooper, 1984a; Hooper and others, 1984). However, large-ion-lithophile (LIL)-element contents increase at a greater rate than do high-field-strength (HFS)-incompatible elements. Such decoupling might result from an open magma-chamber system, in which crystal fractionation was accompanied by crustal assimilation and possible recharge by primitive magma (DePaolo, 1981; O'Hara and Mathews, 1981). As yet, however, no quantitative model has been devised that is able to account for the chemical variations within each type and the differences between them. In particular, such simple models are quite unable to account for the differences between the two flow types (AB and RC). It is significant that the low values of isotopic ratio $^{87}Sr/^{86}Sr$ for the Imnaha Basalt provide little evidence of crustal contamination (fig. 4.3) and, instead, indicate magmas derived from different mantle sources (Carlson, 1984).

GRANDE RONDE BASALT

The mostly aphyric flows of Grande Ronde Basalt may form as much as 85 percent of the Columbia River Basalt Group (Reidel and others, 1982). These flows conformably overlie the porphyritic flows of the Imnaha Basalt wherever the latter are present in the southeastern part of the Columbia Plateau, but they spread much farther to the north and west (fig. 4.4). Some of the younger flows reached the Pacific Ocean (Anderson and others, 1987). All known feeder dikes are within the Blue Mountains province and are concentrated in the Grande Ronde and Cornucopia dike swarms (fig. 4.4).

By detailed geochemical studies, previous workers have succeeded in distinguishing individual flows or limited sets of flows in the Grande Ronde Basalt over substantial parts of the Columbia Plateau (Reidel, 1983; Mangan and others, 1986), but the fine-grained aphyric texture and comparatively homogeneous chemical composition of the flows have inhibited the mapping of individual flows except in limited areas. However, the Grande Ronde Basalt has been subdivided and mapped regionally as four informal, successively younger magnetostratigraphic units: R_1, N_1, R_2, and N_2 (fig. 4.2; Swanson, Wright, and others, 1979).

As much as 200 m of unit R_1 of the Grande Ronde Basalt overlies the Imnaha Basalt in the Weiser River embayment (fig. 4.4; Fitzgerald, 1984); this thickness increases northwestward and reaches a maximum of approximately 400 m in the Imnaha River valley (Camp and Hooper, 1981). Flows of unit N_1 form a thin veneer over R_1 flows in the northern part of the Weiser River embayment, and unit N_1 thickens progressively into southeastern Washington. Unit R_2 thickens northwestward but pinches out southeastward against the northwest flanks of the Wallowa and Seven Devils Mountains and is not found southeast of the divide. Plateau-forming flows of unit N_2 are limited to the northwest side of the Limekiln fault (fig. 4.4), with only local exceptions (Camp and Hooper, 1981).

The apparent northwestward trend (fig. 4.4) of progressively younger stratigraphic units has been ascribed to continued regional tilting of the ground surface, the southeastern part of the area rising relative to the northwestern part during Grande Ronde time. The

TABLE 4.1.—*Representative chemical analyses of samples of the Columbia River Basalt Group and associated volcanic rocks*

[Flow types (see figs. 4.2, 4.6): RC and Log Cr., Rock Creek subtype; AB, American Bar subtype; Rb. Mt., Robinette Mountain flow; Sh. Cr., Shumaker Creek flow; FSP, Frenchman Springs Member; Pr., Priest Rapids Member; Tus Umat., Umatilla, sillusi flow; Tu Umat., Umatilla flow; Powat., Powatka flow; Wilbur, Wilbur Creek flow; Slipp. Cr., Slippery Creek flow; Lew. Or., Lewiston Orchards flow; Wc, Weissenfels Ridge Member, Cloverdale flow; Wt, Weissenfels Ridge Member, Tenmile Creek flow; Pom., Pomona Member; El. Mt., Elephant Mountain Member; L. Mon., Lower Monumental Member; Tamm., Tammany Creek flow. XRF, X-ray fluorescence; ICP, inductively coupled plasma spectrometry. XRF analyses performed at Washington State University's Geoanalytical Laboratory by methods of Hooper and Atkins (1969) and Hooper and others (1976). Major elements are normalized on a volatile-free basis, with iron expressed as Fe_2O_3. Rare-earth elements, including yttrium, were determined by P.R. Hooper, Kings College, London, by ICP methods of Thompson and Walsh (1983); numbers in parentheses are rare-earth element analyses run on trace-element program. ---, no data. Samples of basalt of Weiser courtesy of J.F. Fitzgerald; samples of basalt of Prineville courtesy of G.A. Smith]

	Imnaha Basalt						Grande Ronde Basalt		Wauapum Basalt							Saddle Mouutains Basalt	
Flow type—	RC-0	AB-2	AB-4	AB-7	RC-1	Log Cr.	—	—	Rb. Mt.	Rb. Mt.	Dodge	Sh. Cr.	FSP	Roza	Pr.	Tus Umat.	Tu Umat.
Sample—	SQB-100	BUK-1	BUK-5	W-25	BUK-16	BUK-21	BUK-22	I-68	EP-2	D-5	EP-1	HAS-40	RS-2	K-9040	HCO-85	HCO-86	HAS-45
Major elements, in weight percent, by XRF																	
SiO_2	48.48	50.94	50.73	51.07	49.72	49.42	52.76	54.86	50.28	50.34	51.67	54.55	51.00	49.72	49.39	54.32	53.56
Al_2O_3	17.20	15.36	14.30	12.90	15.07	15.32	13.24	13.71	15.42	14.96	15.02	13.33	12.64	13.15	13.25	13.33	13.41
TiO_2	1.871	1.579	1.883	3.033	2.481	2.123	2.325	2.017	1.119	1.346	1.396	2.564	3.043	3.122	3.095	2.764	3.241
Fe_2O_3	13.96	12.53	13.86	15.71	13.69	13.05	14.58	13.11	11.83	12.96	11.34	13.27	16.11	15.87	15.25	13.27	13.53
MnO	.197	.190	.225	.236	.170	.179	.219	.231	.198	.215	.168	.256	.227	.239	.223	.213	.203
CaO	9.69	10.73	10.23	8.37	9.09	9.62	8.18	7.51	10.73	10.46	10.68	7.08	8.32	9.03	9.17	6.49	7.24
MgO	4.99	5.44	5.30	4.27	5.59	6.68	4.15	4.02	7.46	6.42	6.03	2.92	4.16	4.66	5.40	2.58	2.64
K_2O	.52	.43	.51	1.07	.95	.60	1.28	1.36	.35	.45	.62	1.82	1.37	1.01	.95	2.84	2.30
Na_2O	2.65	2.44	2.52	2.72	2.72	2.58	2.65	2.70	2.26	2.41	2.61	2.98	2.32	2.36	2.40	2.75	2.60
P_2O_5	.283	.214	.282	.440	.353	.269	.440	.299	.211	.257	.298	.987	.616	.642	.704	.977	.846
Trace elements, in parts per million, by XRF																	
Ba	254	220	260	459	395	234	503	620	232	327	343	1,204	563	550	487	3,532	3,047
Rb	9	11	15	37	29	16	33	32	8	9	15	54	36	24	23	51	45
Cr	47	92	86	31	113	112	22	16	187	106	178	17	19	172	101	0	0
Cu	99	131	160	88	103	92	85	33	84	89	54	16	27	11	39	13	0
Nb	6	7	7	15	12	9	13	11	5	5	7	18	13	12	13	21	21
Ni	121	34	24	12	77	95	19	3	65	35	38	0	4	77	37	0	0
Sc	28	38	41	38	32	34	39	31	39	44	40	35	39	41	42	27	30
Sr	509	296	244	271	384	388	314	368	338	376	389	408	327	329	297	300	308
V	295	336	368	415	276	323	402	356	306	339	363	189	468	440	381	185	276
Zn	100	102	106	143	126	106	133	123	89	106	97	176	154	142	142	136	134
Zr	130	127	145	231	191	150	210	185	92	108	117	258	192	184	171	464	416
Trace elements, in parts per million, by ICP																	
La	10.97	10.52	12.66	21.67	17.10	13.05	21.80	20.82	10.00	(12)	14.42	38.71	26.18	25.20	24.15	45.42	(41)
Ce	24.18	22.59	27.48	48.26	38.42	29.55	47.34	42.63	20.95	(36)	30.62	81.95	54.00	52.40	51.24	90.30	(82)
Pr	3.88	3.58	4.28	7.24	5.89	4.64	6.96	6.24	3.18	---	4.72	11.50	7.79	7.32	7.49	11.83	---
Nd	17.85	16.12	19.45	32.72	26.65	21.41	30.51	26.23	14.04	---	20.64	50.29	33.73	32.54	32.78	50.11	---
Sm	4.46	4.20	5.08	8.12	6.51	5.48	7.51	6.10	3.40	---	4.78	11.62	7.66	7.43	7.70	10.20	---
Eu	1.67	1.46	1.71	2.47	2.09	1.87	2.34	1.86	1.18	---	1.56	3.89	2.42	2.42	2.63	4.48	---
Gd	4.95	4.99	6.05	9.03	7.04	6.04	8.56	6.51	3.90	---	5.15	12.43	8.18	8.07	8.45	10.55	---
Dy	4.71	5.30	6.42	8.83	6.59	5.78	8.60	6.15	4.03	---	5.31	11.20	7.75	7.45	7.70	9.06	---
Ho	.89	1.04	1.26	1.72	1.24	1.10	1.71	1.22	.82	---	1.07	2.20	1.50	1.46	1.49	1.75	---
Er	2.60	3.16	3.74	5.02	3.57	3.18	5.03	3.42	2.51	---	3.20	6.22	4.33	4.17	4.20	4.91	---
Yb	2.18	2.82	3.44	4.44	3.01	2.70	4.66	3.10	2.36	---	3.02	5.63	3.82	3.63	3.67	4.34	---
Lu	.35	.46	.54	.73	.48	.44	.76	.49	.40	---	.48	.95	.61	.59	.60	.72	---
Y	27	32	39	53	38	33	53	36	26	28	33	69	46	45	46	54	47

Flow type	Saddle Mountains Basalt—Continued													Picture Gorge Basalt			Basalt of Prineville
Flow type	Powst.	Wilbur	Lapwai	Asotin	Slipp. Cr.	Lew. Or.	Wc	Wt	Pom.	El. Mt.	Buford	L. Mon.	Tamm.	---	---	---	---
Sample	RS-1a	HSI-51	HAS-56	HAS-57a	HAS-74	HAS-51	HAN-03	HAS-34	CLA-42b	HAS-32	HWE-03	HAS-25	HAS-64	BBJ-309	BBJ-311	BBJ-314	PD-1
Major elements, in weight percent, by XRF—Continued																	
SiO_2	53.37	53.76	51.66	49.76	51.19	48.72	52.69	49.61	51.16	50.78	54.39	50.40	53.22	50.70	48.87	51.20	51.33
Al_2O_3	12.88	14.31	14.86	16.08	13.61	13.85	14.80	12.92	14.27	12.44	14.03	13.50	13.12	13.67	16.40	13.90	13.52
TiO_2	2.663	1.921	1.549	1.404	2.500	2.517	1.752	3.012	1.647	3.543	2.192	2.962	2.747	1.904	1.297	1.934	2.690
Fe_2O_3	14.89	11.81	11.76	9.93	13.54	13.24	11.50	15.19	11.81	16.32	11.55	14.82	14.27	15.00	11.55	14.64	13.47
MnO	.252	.181	.165	.153	.199	.198	.156	.225	.186	.212	.169	.221	.196	.228	.176	.224	.237
CaO	7.00	8.70	9.73	11.99	9.61	11.18	9.71	9.72	10.85	8.59	8.83	8.93	7.88	9.36	11.32	9.43	8.06
MgO	2.96	4.41	6.41	8.01	5.46	6.90	5.31	5.29	7.07	3.95	4.70	4.50	3.90	5.09	7.33	4.80	4.44
K_2O	1.76	1.74	1.19	.38	.97	.49	1.19	.94	.56	1.25	1.09	1.43	1.71	.83	.31	.64	1.76
Na_2O	2.92	2.44	2.15	1.99	2.31	2.14	2.28	2.22	2.08	2.18	2.55	2.40	2.39	2.75	2.37	2.76	2.71
P_2O_5	1.179	.514	.348	.165	.451	.578	.396	.664	.240	.551	.321	.661	.382	.280	.234	.301	1.443
Trace elements, in parts per million, by XRF—Continued																	
Ba	953	913	560	202	416	436	758	694	227	512	546	571	601	378	223	401	2,126
Rb	46	41	27	9	26	9	31	17	16	34	45	36	52	16	4	12	42
Cr	0	30	139	296	36	234	133	75	116	37	42	22	29	62	182	44	14
Cu	22	21	55	86	37	52	38	32	52	19	36	0	15	348	119	247	21
Nb	18	16	11	9	17	19	14	25	11	24	20	26	19	6	3	4	6
Ni	0	37	72	132	20	60	67	40	44	23	19	8	9	31	108	27	18
Sc	34	29	29	33	35	42	29	39	37	35	34	28	31	44	36	42	37
Sr	342	294	274	261	265	249	306	268	237	246	265	371	270	198	239	234	393
V	181	284	281	273	344	344	276	391	290	431	310	366	355	489	283	481	339
Zn	181	129	103	82	125	121	116	146	93	159	120	138	124	153	99	123	124
Zr	248	239	175	107	175	185	200	264	134	251	212	186	216	122	70	111	151
Trace elements, in parts per million, by ICP—Continued																	
La	35.68	37.59	(28)	11.88	27.68	30.76	(30)	(40)	17.23	34.17	30.12	35.75	(34)	11.54	---	---	---
Ce	76.59	72.98	(58)	23.96	54.79	63.07	(62)	(82)	34.24	69.23	58.94	67.77	(70)	24.16	---	---	---
Pr	11.00	9.58	---	3.49	7.50	8.88	---	---	4.90	9.48	7.93	8.65	---	3.59	---	---	---
Nd	48.92	38.79	---	15.29	31.11	37.54	---	---	20.82	40.04	32.36	35.22	---	17.20	---	---	---
Sm	11.23	7.70	---	3.63	6.85	8.12	---	---	4.79	9.00	7.09	7.30	---	4.80	---	---	---
Eu	3.68	2.06	---	1.28	2.19	2.55	---	---	1.59	2.79	2.00	2.42	---	1.66	---	---	---
Gd	12.05	7.68	---	4.04	7.26	8.54	---	---	5.25	9.67	7.53	7.60	---	6.03	---	---	---
Dy	10.87	7.40	---	4.11	6.94	7.84	---	---	5.31	9.19	7.34	6.69	---	6.96	---	---	---
Ho	2.11	1.47	---	.77	1.33	1.52	---	---	1.04	1.77	1.41	1.27	---	1.47	---	---	---
Er	5.96	4.30	---	2.30	3.82	4.35	---	---	3.01	5.07	4.12	3.61	---	4.33	---	---	---
Yb	5.24	3.88	---	2.01	3.23	3.78	---	---	2.62	4.43	3.60	3.08	---	4.08	---	---	---
Lu	.86	.65	---	.32	.53	.61	---	---	.41	.70	.58	.49	---	.65	---	---	---
Y	65	43	35	23	41	46	37	51	31	54	43	39	40	43	25	39	49

TABLE 4.1.—*Representative chemical analyses of samples of the Columbia River Basalt Group and associated volcanic rocks*—Continued

	Basalt of Prineville—Continued					Basalt of Powder River and associated volcanic rocks					Basalt of Weiser						
Flow type---	—	—	—	—	—	—	—	—	—	—	Star Butte		Sugar Loaf	Cambridge		Cuddy Mountain	Associated welded tuff
Sample-------	PD-5	PD-10	PD-12	LS-1	LS-2	PH80-124	PH80-126	PH80-092	PH80-113	PH80-133	BX6043	BX6110	BX6123	BX6000	BX6005	X20	X21
						Major elements, in weight percent, by XRF—Continued											
SiO_2-------	51.41	51.07	50.94	51.02	54.76	51.40	50.75	60.64	53.30	54.56	50.00	48.61	54.50	48.18	51.93	43.25	70.97
Al_2O_3------	13.57	13.56	13.55	13.80	14.05	15.68	15.52	16.31	17.46	17.08	16.25	16.16	17.76	16.39	16.13	12.85	15.31
TiO_2-------	2.644	2.639	2.671	2.745	2.428	1.370	1.388	1.031	1.498	1.193	1.622	2.169	1.284	1.747	1.674	3.567	.435
Fe_2O_3------	13.25	13.61	13.60	13.08	10.79	9.90	10.47	7.17	7.49	8.90	10.92	11.23	8.05	11.58	10.49	13.37	2.81
MnO-------	.242	.239	.241	.249	.224	.148	.149	.126	.103	.130	.175	.162	.132	.164	.194	.161	.022
CaO-------	8.02	7.95	8.05	8.30	6.25	9.76	9.40	6.06	10.68	8.51	9.59	10.49	9.86	11.27	7.61	11.71	1.76
MgO-------	4.53	4.58	4.52	4.33	3.34	8.08	9.07	3.06	5.47	5.13	7.60	7.01	3.59	7.68	5.92	11.46	.19
K_2O--------	1.82	1.78	1.79	1.64	3.26	.69	.47	1.83	.71	1.03	.80	.66	1.32	.32	1.60	.63	4.61
Na_2O-------	2.76	2.83	2.86	2.99	3.20	2.41	2.29	3.11	2.70	2.91	2.44	2.53	2.82	1.96	3.43	2.35	3.48
P_2O_5-------	1.421	1.421	1.436	1.479	1.383	.362	.310	.463	.395	.373	.436	.758	.467	.555	.775	.403	.090
						Trace elements, in parts per million, by XRF—Contiuued											
Ba---------	2,068	2,040	2,110	2,314	2,141	345	307	742	390	544	390	472	658	304	778	371	2,181
Rb---------	43	41	43	34	50	8	5	22	11	16	10	7	19	3	21	38	145
Cr---------	31	28	29	34	36	353	335	91	351	101	176	202	101	200	128	283	5
Cu---------	30	40	15	15	19	35	22	41	31	3	55	60	93	59	43	80	1
Nb---------	6	7	6	6	7	8	8	15	10	7	6	12	8	7	20	33	28
Ni---------	20	21	20	27	23	151	152	52	109	71	104	136	116	109	95	138	21
Sc---------	37	38	38	38	34	29	29	16	31	22	35	31	30	41	24	36	2
Sr---------	393	389	396	411	303	461	421	530	521	537	479	555	558	361	817	775	336
V----------	339	313	341	351	233	266	261	114	271	199	272	289	268	293	225	384	29
Zn---------	136	137	125	129	118	90	88	92	78	90	89	121	90	83	113	97	57
Zr---------	150	151	152	154	143	114	112	202	126	124	105	142	138	90	165	186	433
						Trace elements, in parts per million, by ICP—Continued											
La---------	---	---	---	---	---	13.39	11.78	28.07	14.71	17.51	---	---	---	---	---	---	---
Ce---------	---	---	---	---	---	28.36	25.25	52.37	30.35	34.78	---	---	---	---	---	---	---
Pr---------	---	---	---	---	---	4.16	3.85	6.84	4.36	4.88	---	---	---	---	---	---	---
Nd---------	---	---	---	---	---	18.52	16.85	29.64	19.92	21.53	---	---	---	---	---	---	---
Sm---------	---	---	---	---	---	4.11	3.77	5.67	4.41	4.49	---	---	---	---	---	---	---
Eu---------	---	---	---	---	---	1.36	1.29	1.61	1.47	1.36	---	---	---	---	---	---	---
Gd---------	---	---	---	---	---	4.09	3.87	5.11	4.46	4.39	---	---	---	---	---	---	---
Dy---------	---	---	---	---	---	3.97	3.80	4.55	4.31	4.24	---	---	---	---	---	---	---
Ho---------	---	---	---	---	---	.84	.83	.94	.90	.89	---	---	---	---	---	---	---
Er---------	---	---	---	---	---	2.42	2.20	2.63	2.52	2.49	---	---	---	---	---	---	---
Yb---------	---	---	---	---	---	2.02	1.95	2.42	2.27	2.28	---	---	---	---	---	---	---
Lu---------	---	---	---	---	---	.32	.29	.38	.35	.35	---	---	---	---	---	---	---
Y----------	48	49	49	52	50	25	25	29	28	27	---	---	---	---	---	---	---

TABLE 4.2.—*Selected isotopic data for samples of the Columbia River Basalt Group and associated volcanic rocks*

[Data sources: H (samples listed in table 4.1), Sr and Nd isotope analyses by C.J. Hawkesworth, unpub. data (Open University, England); LT, $\delta^{18}O$ values, P.B. Larson and H.P. Taylor, unpub. data; C, Carlson and others (1981) and Carlson (1984); MC, Sr from McDougall (1976) and Pb from Church (1985). Many of the $^{87}Sr/^{86}Sr$ values are from McDougall (1976) and Pb ratios, from Church (1985). Flow types as for table 4.1 with these additions: symbols N and R refer to normal and reversed magnetic polarity (fig. 4.2); Huntz., Huntzinger flow; Ind. Mt., basalt of Indian Mountain, which, along with basalts of Basin City, Martindale, and Goose Island (Goose Is.), is part of the Ice Harbor Member. ?, uncertain; —, no data; do., ditto]

Sample	Data sources	Flow type	$^{87}Sr/^{86}Sr$	$^{143}Nd/^{144}Nd$	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pbb$	$\delta^{18}O$
Imnaha Basalt								
SQB100	H,LT	RC-0	0.70363	0.51290	—	—	—	6.50
C1	C	RC-0	.70349	.51302	18.769	15.511	38.28	5.60
BUK-1	H,LT	AB-2	.70415	.51281	—	—	—	6.40
BUK-5	H,LT	AB-4	.70397	.51284	—	—	—	6.00
W-25	H,LT	AB-7	.70398	.51286	—	—	—	5.90
BUK-16	H,LT	RC-1	.70383	.51285	—	—	—	—
BUK-21	H,LT	Log Cr.	.70383	.51289	—	—	—	6.10
M-1	C	?	.70373	.512938	19.086	15.646	38.73	6.09
M-13	C	?	.70395	.512995	19.040	15.610	38.68	6.48
Picture Gorge Basalt								
PG-1	C	—	0.70340	0.513033	18.837	15.555	38.38	5.64
PG-14	C	—	.70351	.513017	19.088	15.646	38.73	5.64
Grande Ronde Basalt								
SW73-355	C	—	0.70500	0.51265	—	—	—	7.13
SW75-219	C	—	.70489	.51268	18.810	15.564	38.66	6.62
SW77-127	C	N_1	.70444	.51284	18.978	15.640	38.81	6.01
SW78-258	C	R_2	.70507	.51269	18.785	15.601	38.76	6.77
SW78-221	C	N_2	.70502	.512736	—	—	—	—
L-11	C	R_1	.70428	.51284	18.969	15.603	38.67	6.27
L-41	C	N_1?	.70462	.512808	18.918	15.607	38.74	6.91
YB-1	C	R_2	.70445	.51274	18.925	15.611	38.74	6.33
BCR-1	C	—	.70493	.512654	18.801	15.609	38.66	—
Wanapum Basalt								
SW73-322	C	Rb. Mt.	0.70438	0.51277	19.009	15.584	38.58	6.60
EP-2	H	do.	.70432	.51273	—	—	—	—
SW73-296	C	Dodge	.70435	.51274	18.818	15.588	38.54	6.69
SW73-152	C	Sh. Cr.	.70481	.51267	18.881	15.612	38.75	6.82
SW78-408	C	FSP	.70517	.51263	—	—	—	6.60
CR-19	MC	FSP	.7053	—	18.84	15.59	38.67	—
K-9040	H	Roza	.70517	—	—	—	—	—
SW71-42	C	Lolo	.70506	.51259	18.729	15.558	38.70	6.47
SW77-321	C	Lookingglass	.70546	.51266	18.773	15.625	38.77	6.36
RS-1A	H	Powatka	.70475	.51266	—	—	—	—
Saddle Mountains Basalt								
WU-1	C	Umatilla	0.70892	0.51227	18.148	15.629	39.36	6.85
SW72-213	C	Wilbur	.71047	.51177	17.736	15.589	38.81	8.03
HS1-51	H	do.	.71055	.51173	—	—	—	—
HAS-56	H	Lapwai	.70922	—	—	—	—	—
SW72-31	C	Asotin	.70755	.51234	18.085	15.627	39.17	6.37
20-6.63.1	C	Huntz.	.70884	.51199	17.819	15.587	38.89	6.73
HAS-51	H	Lew. Or.	.70593	—	—	—	—	—
HAS-74	H	Slipp. Cr.	.71013	—	—	—	—	—
WSM-2	C	Esquatzel	.71317	.51206	19.751	15.952	40.13	6.09
WPOM-1	C	Pomona	.70750	.51245	18.367	15.656	39.74	6.09
WEM-2	C	El. Mt.	.70762	.51243	19.109	15.777	39.91	6.34
SW75-46	C	Basin City	.70782	.51231	18.626	15.653	39.45	—
SW75-114	C	Martindale	.70764	.51233	18.656	15.652	39.44	5.81
SW75-35	C	Goose Is.	—	.51234	—	—	—	—
SW75-36	C	Ind. Mt.	.70780	.51233	—	—	—	—
HWE-03	H	Buford	.70723	—	—	—	—	—
SW73-361	C	L. Mon.	.71087	.51213	18.464	15.654	38.76	7.41
Basalt of Powder River								
F-2	C	Olivine basalt	0.70353	0.512931	—	—	—	—
D-6	C	Andesite	.70354	.512932	—	—	—	—
7-50	C	Nepheline basalt	.70343	.513008	—	—	—	—

tilting is attributed to the continued, possibly isostatic, rise of the Idaho batholith and other associated granite-cored mountain blocks (Hooper and Camp, 1981). The tilting was augmented by normal faulting, such as that displayed by the northeast-trending Limekiln fault at the end of unit R_2 (Grande Ronde) time (fig. 4.4).

The fissure system and, presumably, the magma source may have remained in much the same position while the magma erupted progressively farther north, where the fissures intersected a lower ground surface. This apparent stability is supported by the presence of low shields of unit N_2 of the Grande Ronde Basalt (Joseph Volcanics unit of Kleck, 1976; see Walker, 1979) directly along the axis of the Grande Ronde dike swarm on the uplifted side of the Limekiln fault (fig. 4.4). The shields demonstrate the presence of unit N_2 feeder dikes farther south than the main eruption site of unit N_2, yet they form only isolated units resting on the unit R_2 plateau surface. The large volumes of unit N_2 magma, which covered the greater part of the Columbia Plateau and reached the Pacific Ocean, were erupted from those parts of the fissures that intersected the lower topographic surface on the northwestern, downthrown side of the Limekiln fault. A similar relation can be observed for the Roza Member (of the Wanapum Basalt) across the same fault.

Detailed descriptions of the stratigraphic and petrologic features and the chemical composition of the Grande Ronde Basalt in the Blue Mountains province were provided by Reidel (1983), and isotopic data were given by McDougall (1976), Carlson and others (1981), Nelson (1983), and Carlson (1984). Major-element and trace-element analyses and isotopic data for two representative samples of the Grande Ronde Basalt are listed in tables 4.1 and 4.2.

The Grande Ronde Basalt is consistently richer in SiO_2, K_2O, and LIL-incompatible trace elements (such as Ba and Rb) than is the Imnaha Basalt (Hooper and others, 1984), but the two formations contain similar concentrations of most other major and trace elements. Gradual fluctuations of incompatible elements through the Grande Ronde stratigraphy can be partly explained by crystal fractionation, in which removal of plagioclase and pyroxene played a dominant role (Reidel, 1983). The scarcity of well-developed phenocrysts of either of these phases is anomalous. However, rare orthopyroxene phenocrysts mantled by clinopyroxene are present (Reidel, 1983).

A small stepped-upward increase in the $^{87}Sr/^{86}Sr$ ratio (with an upward decrease in the $^{143}Nd/^{144}Nd$ ratio) is evident in the Grande Ronde Basalt (Hooper, 1984a, fig. 2). These changes, together with Pb isotopic data, led Carlson (1984) to conclude that progressive crustal contamination, in addition to crystal fractionation, played a role in the evolution of the magmas. Hooper (1984a) suggested, however, that the absence of physical and chemical evidence for crustal assimilation points more obviously to mixing of two mantle sources.

ECKLER MOUNTAIN MEMBER OF THE WANAPUM BASALT

A break in volcanic activity separates the eruptions of the Grande Ronde Basalt from those of the Wanapum Basalt across the whole Columbia Plateau (fig. 4.2; Beeson and others, 1985). During this break, sediment was deposited in many areas of the western plateau, giving rise to the Vantage Member of the Ellensburg Formation.

In the Blue Mountains area a residual soil developed, and sediment was trapped in local basins. The oldest flows of the Wanapum Basalt are of relatively small volume, diverse chemical compositions, and normal magnetic polarity and consist of the basalts of Robinette Mountain, Dodge, Shumaker Creek, and Lookingglass, which together form the Eckler Mountain Member (Swanson and others, 1979b, 1981). Despite their chemical diversity, these flows have similar isotopic signatures, in that the $^{87}Sr/^{86}Sr$ ratio increases with the Ba/Ti ratio (fig. 4.3*A*), a relation suggesting mixing between two sources. These sources are clearly different from those that fed the large eruptions of the main part of the Wanapum Basalt which followed. A comparison of erupted-magma volume with approximate K-Ar ages implies a substantial decrease in the rate of magma supply relative to the Grande Ronde (Swanson and Wright, 1981; Hooper, 1981, 1982). The areal extents of individual flows of the Eckler Mountain Member and of their feeder dikes are shown in figure 4.5. Major-element and trace-element analyses with isotopic data are listed in tables 4.1 and 4.2.

Five chemically distinguishable flows or sets of flows can be recognized in the Eckler Mountain Member (Swanson, Bentley, and others, 1979; Swanson, Wright, and others, 1979; Swanson and Wright, 1983). The oldest of these, the Robinette Mountain flow, is present in an area elongate north-northwest area flanking its feeder dike, which crosses the Blue Mountains uplift (fig. 4.5). It is a coarse-grained diktytaxitic basalt containing olivine phenocrysts that is chemically the most primitive flow in the Columbia River Basalt Group. Two or more Dodge flows (Swanson, Wright, and others, 1979) have been reported to lie above the Robinette Mountain flow (Ross, 1978; Shubat, 1979; Wright and others, 1982). Southwest of Troy Basin, north of Elgin (fig. 4.5), the lowest of these flows is compositionally similar, but not identical, to the Robinette Mountain flow (figs. 4.6, 4.7; Shubat, 1979). This flow (Robinette Mountain II) is coarse grained and aphyric. Elsewhere, however, as on the northwest side of Troy Basin in the Wenaha Tucannon Wilderness (Swanson and Wright, 1983), as many as three plagioclase-

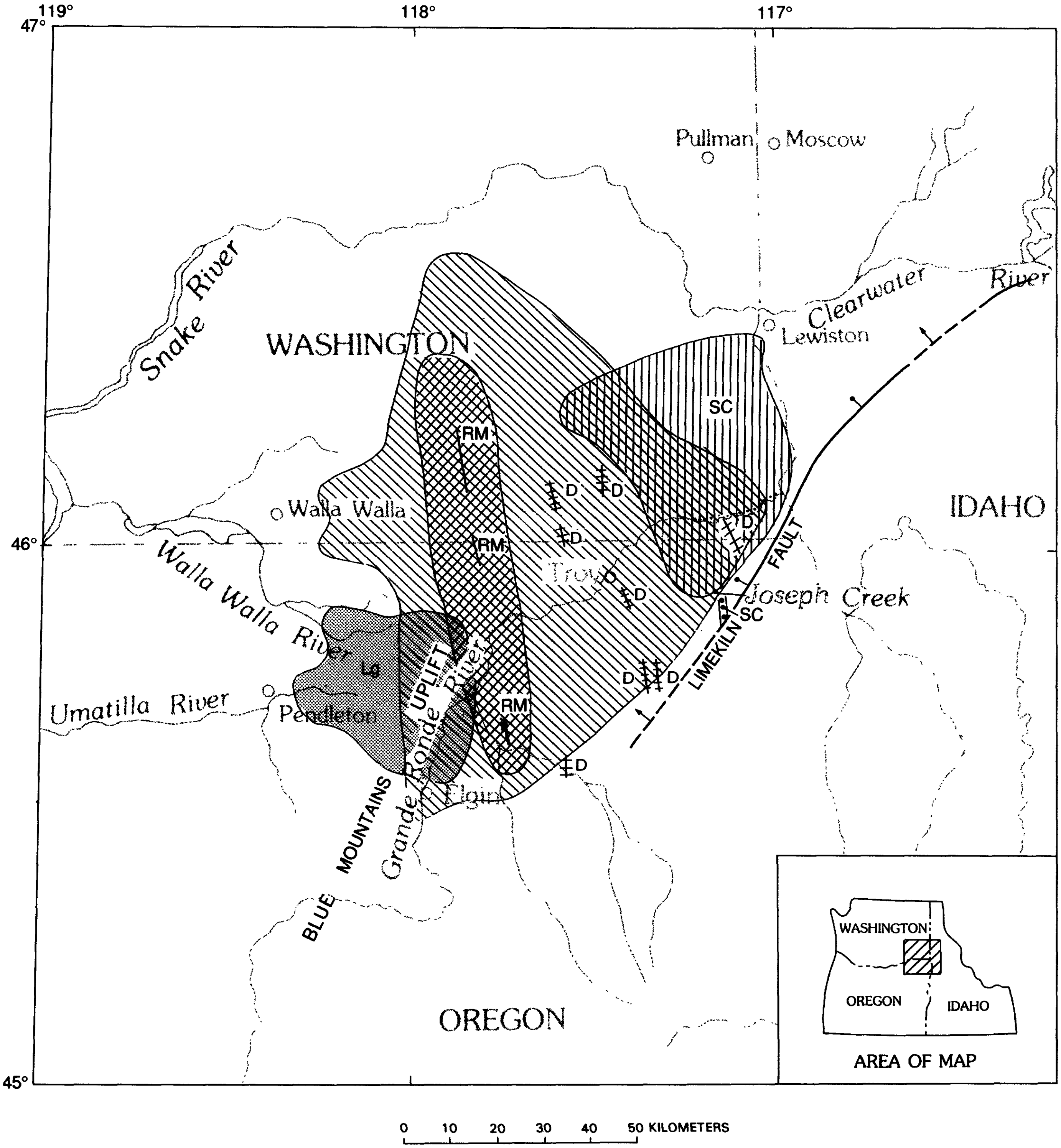


FIGURE 4.5.—Blue Mountains region, showing distribution of flows of the Eckler Mountain Member (of the Wanapum Basalt). Diagonal northeast-southwest line pattern, Robinette Mountain flow; RM, associated dike. Diagonal northwest-southeast line pattern, Dodge flow; D, associated dikes. Shaded area (no dikes), Lookingglass flow (Lg). Vertical line pattern, Shumaker Creek flow; SC, associated dike. Limekiln fault, dashed where approximately located, bar and ball on downthrown side; arrow indicates dip of monoclinal limb at both ends of fault.

porphyritic Dodge flows are present in individual sections, with only the Robinette Mountain I flow at the base.

The undisputed Dodge has a much wider distribution than do the two Robinette Mountain flows discussed above (fig. 4.5). The Dodge is a coarse-grained basalt containing plagioclase and olivine phenocrysts and has a characteristically rounded, yellow- to brown-weathering, commonly grusy outcrop appearance that makes it an excellent marker unit. Two main feeder-dike systems were mapped by Ross (1978) and Swanson and Wright (1983), but other dikes of similar composition are found to the east (fig. 4.5). The wide distribution of Dodge feeder dikes is unusual. Another uncommon feature of the Dodge is the presence of a second, somewhat less porphyritic flow of otherwise identical texture and mineralogic and chemical composition and of indistinguishable age, far to the northeast in the St. Maries River drainage of northern Idaho (Swanson, Bentley, and others, 1979).

The Shumaker Creek and Lookingglass flows both lie directly above the Dodge and below flows of the main part of the Wanapum. Their relative ages are not known because of their geographic separation (fig. 4.5). A dike of Shumaker Creek composition is present in upper Joseph Creek (fig. 4.5). Both flows have relatively evolved chemical compositions and are aphyric, with chemical compositions, including P_2O_5/TiO_2 ratios and incompatible-element concentrations, similar to but less extreme than those of the Umatilla Member (of the Saddle Mountains Basalt). Their Ba contents, only one-third that of the Umatilla, are greater than that of any other flow in the Columbia River Basalt Group and are equivalent in this respect to the basalt of Prineville.

The similar P_2O_5/TiO_2 ratios of the more primitive flows of the Eckler Mountain Member (see fig. 4.7), together with the increase in such ratios as Ba/Ti coupled with an equivalent increase in the ratio $^{87}Sr/^{86}Sr$ (fig. 4.3), suggest crystal fractionation accompanied by crustal assimilation in an open reservoir system (O'Hara and Mathews, 1981).

MAIN PART OF THE WANAPUM BASALT

Three voluminous multiple-flow members (Frenchman Springs, Roza, and Priest Rapids Members) and the single less extensive Powatka flow constitute the main

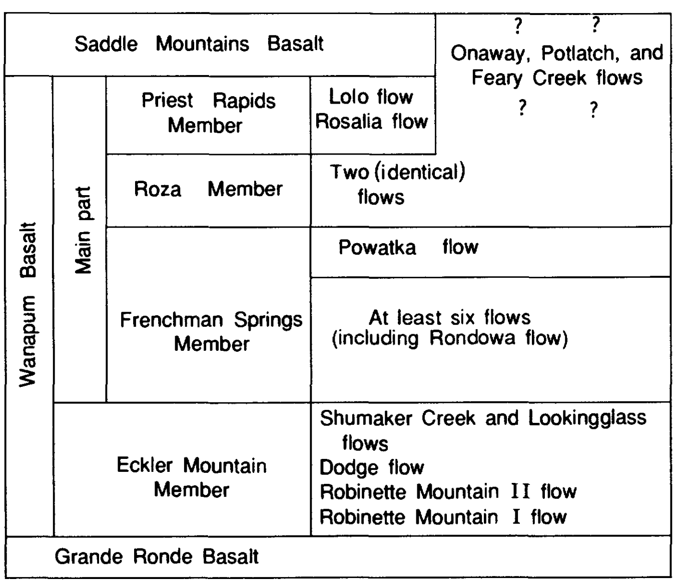


FIGURE 4.6.—Stratigraphic succession of the Wanapum Basalt. Queried where relation uncertain.

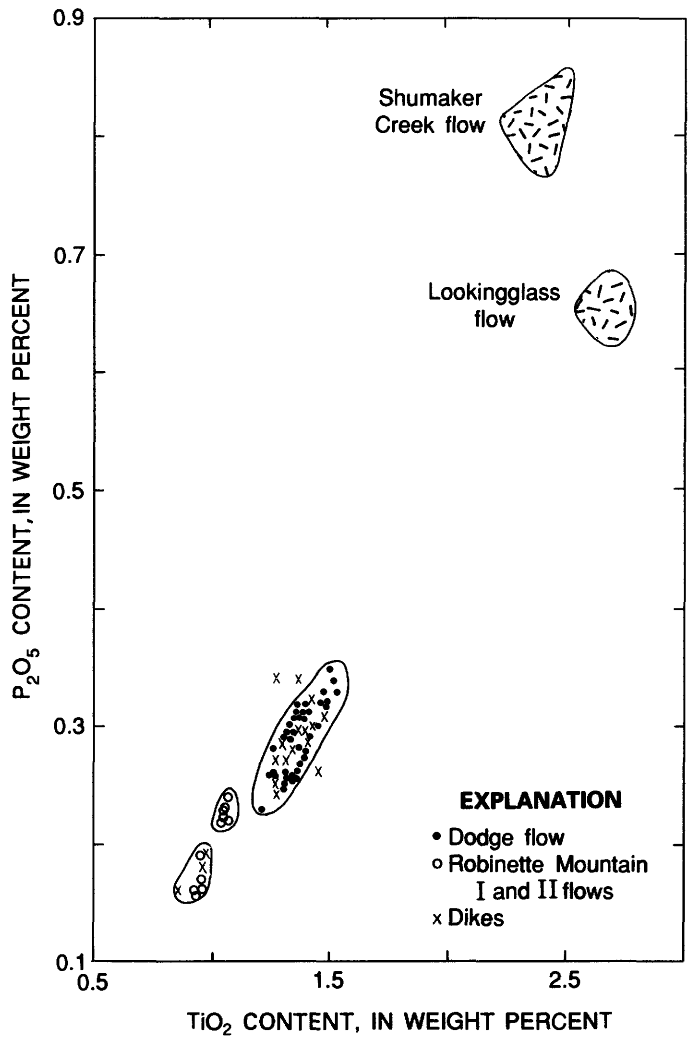


FIGURE 4.7.—TiO_2 versus P_2O_5 content for flows of the Eckler Mountain Member (of the Wanapum Basalt). Encircled patterned areas enclose plots of the Shumaker Creek and Lookingglass flows. Also encircled are plots of Robinette Mountain flow I (lowest P_2O_5 and TiO_2 contents), Robinette Mountain flow II, and Dodge flow.

part of the Wanapum Basalt. The three members and, to some degree, the individual flows within each member are best distinguished in the field by the different sizes and proportions of plagioclase and olivine phenocrysts and by their magnetic polarity (normal for the Frenchman Springs, transitional for the Roza, reverse for the Priest Rapids). Chemical abundances overlap and are difficult to use as discriminants within the Roza and Frenchman Springs (see Beeson and others, 1985), but they are distinctive for the Lolo and Rosalia flows of the Priest Rapids Member.

Six units of the Frenchman Springs Member, some containing several flows, have been recognized over considerable distances on the north side of the Blue Mountains uplift (Beeson and others, 1985). All have normal polarity, but they show subtle chemical differences. Feeder dikes are found in a wide zone across the crest and north flank of the uplift (fig. 4.8), but they have not yet been correlated with individual flows. One Frenchman Springs flow is present near Rondowa (Shubat, 1979) and locally in the Wenaha Tucannon Wilderness (Swanson and Wright, 1983), but other flows of this member appear to have flowed northward and westward, a pattern suggesting that the northeastward extension of the Blue Mountains uplift had become active by that time.

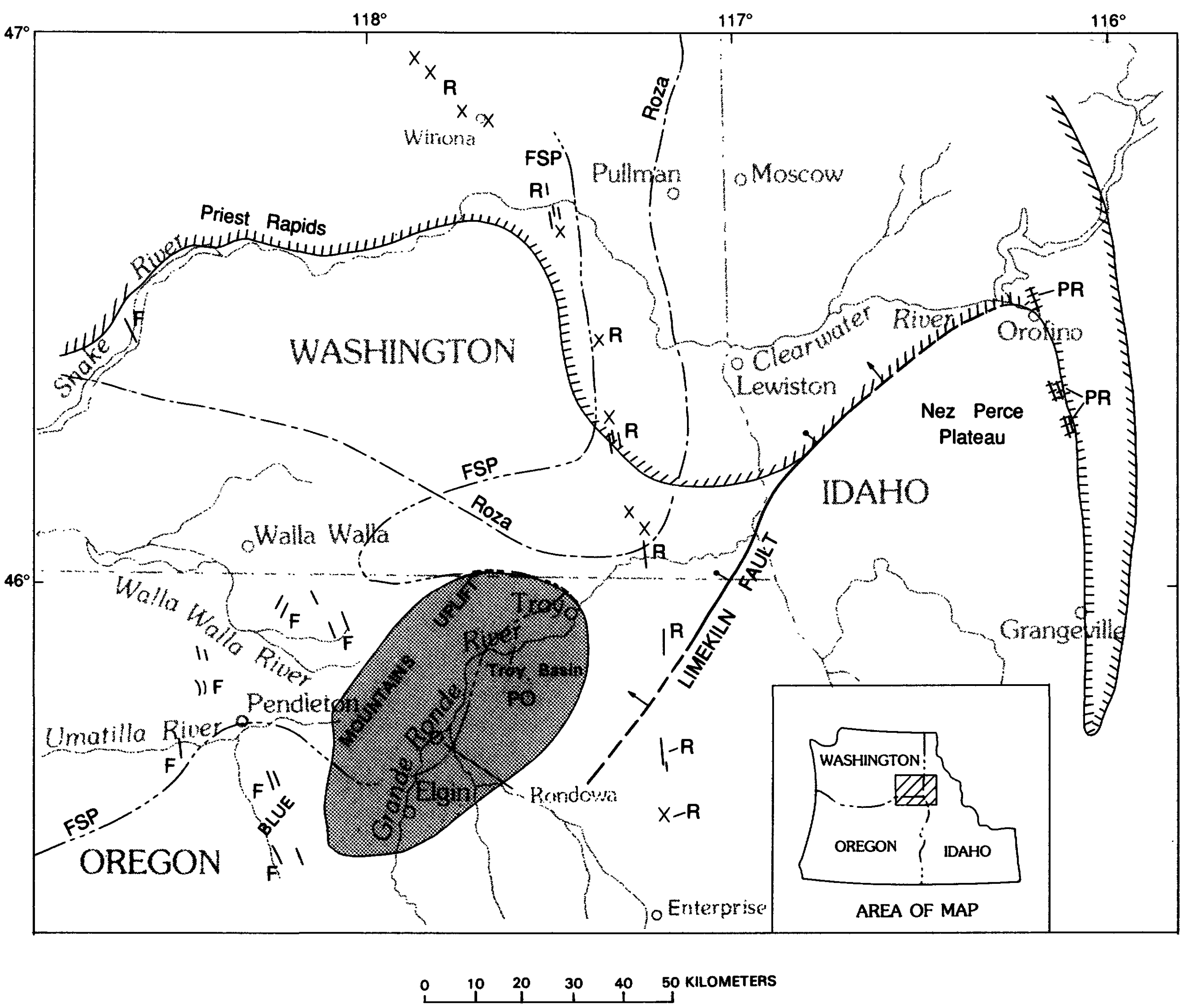


FIGURE 4.8.—Blue Mountains region, showing distribution of flows of main part of the Wanapum Basalt. Broken line with three dashes, southeast limit of Frenchman Springs (FSP) flows; F, associated dikes. Shaded area (no dikes), Powatka flow (PO). Broken line with single dash, southeast limit of Roza flow; R, associated dikes and vents (X). Solid line with hachures, Priest Rapids flows; PR, associated dikes. Limekiln fault, dashed where approximately located, bar and ball on downthrown side; arrow indicates dip of monoclinal limb at both ends of fault.

The Roza may be the best documented member on the Columbia Plateau (Shaw and Swanson, 1970; Swanson and others, 1975; Swanson, Wright, and others, 1979). Two chemically identical flows are commonly recognized, but more flows occur locally near the vent system. At least 20 vents and many dikes (fig. 4.8) define a linear vent system along the axis of the Chief Joseph dike swarm and extend from 30 km north of Enterprise, Oreg., to 10 km northwest of Winona, Wash. Calculations of the rate of discharge from this 150- by 5-km fissure-vent system (Shaw and Swanson, 1970; Swanson and others, 1975) indicate that a flood of basaltic lava covered a large part of the Columbia Plateau in a matter of days. The flows produced by two major eruptions, each with a volume of about 700 km^3, are mineralogically and chemically indistinguishable; a typical chemical analysis of the Roza Member is listed in table 4.1.

The length of this and other fissure systems, and the homogeneity of such large volumes of magma, have profound implications for the magma plumbing system. The magma reservoir must have had a volume many, possibly 10, times that of the magma erupted. The reservoir for the Roza magma must therefore have been very large (approx. 14,000 km^3) and must have been located at least as deep as the crust-mantle boundary, which is approximately 30 km (Hill, 1978) below the surface of the accreted terrane(s). The combination of magma homogeneity and high eruption rate make significant assimilation of crustal rock during passage between reservoir and surface unlikely. To the extent that crustal assimilation took place, it must have occurred around the top of the magma reservoir before eruption and must have been accompanied by very efficient magma mixing within the reservoir.

The outcrop area of the Roza Member (fig. 4.8) reflects the continued rise of the southeast quadrant of the Columbia Plateau relative to Pasco Basin area and the initiation of significant uplift of the north end of the Blue Mountains.

The Priest Rapids Member includes two compositionally and stratigraphically distinct sets of flows, the Rosalia and the younger Lolo (fig. 4.6). Both sets appear to have been erupted across the northeast corner of the Blue Mountains province (fig. 4.8). Feeder dikes to the Priest Rapids flows are present near Orofino along the eastern margin of the Columbia Plateau. The Priest Rapids fissure system, like that of the Roza, must have extended north-northwestward into the craton (fig. 4.1; Camp, 1981). The rising south and east sides of the Columbia Plateau and the northeast end of the Blue Mountains uplift confined the flows to areas north of the Blue Mountains province once they had escaped from the embayment of the Clearwater River drainage (fig. 4.8).

The Powatka flow (Ross, 1978; Hooper, 1981; Swanson and others, 1981) poses a problem in classification. In field appearance (medium to fine grained aphyric) and magnetic polarity it resembles the upper (N_2) flows of the Grande Ronde Basalt. Its exceptionally high P_2O_5 content, however, makes its composition distinctive (table 4.1). It overlies the Rondowa flow north of Elgin (fig. 4.8), which has a Frenchman Springs composition (Swanson and others, 1981), and so its stratigraphic position appears to be within the main part of the Wanapum Basalt. Its isotopic signature, however, more closely resembles those of the Grande Ronde Basalt and Eckler Mountain Member than that of the main part of the Wanapum Basalt (fig. 4.3). The Powatka flow was the first to be confined by the topography of Troy Basin (fig. 4.8).

The similarity in chemical and isotopic composition of the three voluminous members of the Wanapum Basalt is noteworthy. The Frenchman Springs was erupted through the accreted oceanic terrane of the Blue Mountains province or through the apparently thinner crust northwest of the KBML (fig. 4.1). Half of the Roza fissure-vent system lies within the Blue Mountains province, and half within the old craton farther north. The Priest Rapids also appears to have been erupted through both cratonic and accreted oceanic rocks (fig. 4.8).

The crustal rocks of the craton and of the accreted terranes have diverse chemical and isotopic compositions. These differences are known to have profoundly affected the composition and isotopic signature of the granitic rocks that intruded or were generated within them (Armstrong and others, 1977; Fleck and Criss, 1985). The flows of the main part of the Wanapum show no variation in the ratio $^{87}Sr/^{86}Sr$ comparable to that in the different types of crustal rocks through which they passed. Nor do they show significant variation in such trace-element ratios as Ba/Ti (except for the anomalous Powatka flow) that might be expected from varying degrees of crustal contamination. Again, we must conclude that little crustal assimilation occurred between magma reservoir and surface and that a mantle source of varying composition, created either by subduction of sediment and ocean crust or by metasomatic processes in the mantle, would better explain the rather high $^{87}Sr/^{86}Sr$ ratios of these flows than does assimilation of crust.

SADDLE MOUNTAINS BASALT

Flows of the Saddle Mountains Basalt (figs. 4.9, 4.10) are the youngest on the main part of the Columbia Plateau (Swanson, Wright, and others, 1979). They rest unconformably on the Wanapum Basalt and on each other. They probably represent less than 1 percent of the

total volume of the Columbia River Basalt Group but were erupted over an extended period (approx. 13.5–6.0 Ma) of waning volcanic activity. Like those in the Eckler Mountain Member, flows in the Saddle Mountains Basalt vary as widely in chemical composition as the entire Columbia River Basalt Group. They are distinguished isotopically from all other flows on the Columbia Plateau by their consistently large, if varying, $^{87}Sr/^{86}Sr$ ratio (fig. 4.3; table 4.2), which is apparently unrelated to their age or their chemical composition (Hooper, 1984a).

The unconformities between most flows in the Saddle Mountains Basalt result directly from the long periods

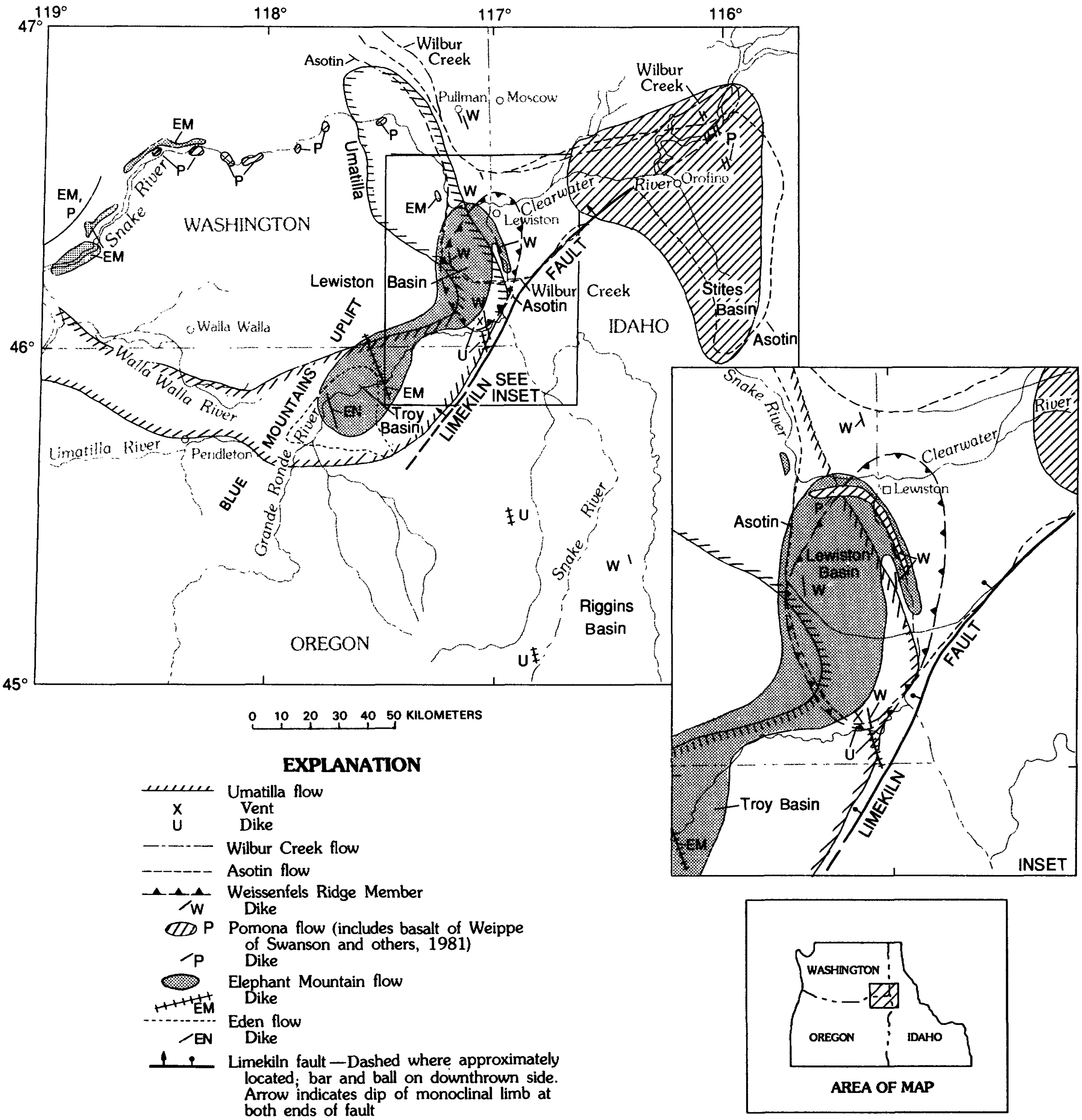


FIGURE 4.9.—Blue Mountains region, showing distribution of larger flows of the Saddle Mountains Basalt.

between eruptions. The cumulative evidence points to the Columbia Plateau deforming continuously throughout Columbia River Basalt Group time (Hooper and Camp, 1981; Reidel, 1984). During Imnaha, Grande Ronde, and, to a lesser extent, Wanapum time (17.5–13.5 Ma), major eruptions occurred much more frequently, averaging one every 10,000 to 20,000 years (Swanson and others, 1975; Hooper and others, 1984), than during later Saddle Mountains time. This frequency was great enough to allow little erosion or deformation between eruptions; only such cumulative changes as regional tilting (Hooper and Camp, 1981) and relatively sudden movement along particular faults (such as along the Limekiln fault during the interval between units R_2 and N_2 of the Grande Ronde Basalt) can be documented. In contrast, during Saddle Mountains time some successive flows differ in age by 1 Ma or more, and so evidence of deformation and erosion between eruptions becomes obvious. Individual flows in the Saddle Mountains Basalt fill and clearly define developing structural basins and drainage systems. Detailed mapping of these flows provides a fascinating view of the structural and morphologic evolution of the Columbia Plateau between about 13.5 and 6.0 Ma.

In addition to the large Pasco Basin that was developing throughout the eruptions of the Columbia River Basalt Group in east-central Washington, four smaller structural basins containing flows of the Saddle Mountains Basalt developed within the Blue Mountains province. Two of these, Stites Basin (Camp, 1981) and Riggins Basin (Hooper, 1984b), lie along the ancient but still active suture zone forming the eastern margin of the Blue Mountains province. Throughout Columbia River Basalt Group time, this broad zone formed a downwarp, elongate north-south, that was locally filled with the Imnaha and lower part (unit R_1) of the Grande Ronde Basalt until the whole region was uplifted enough to prevent younger

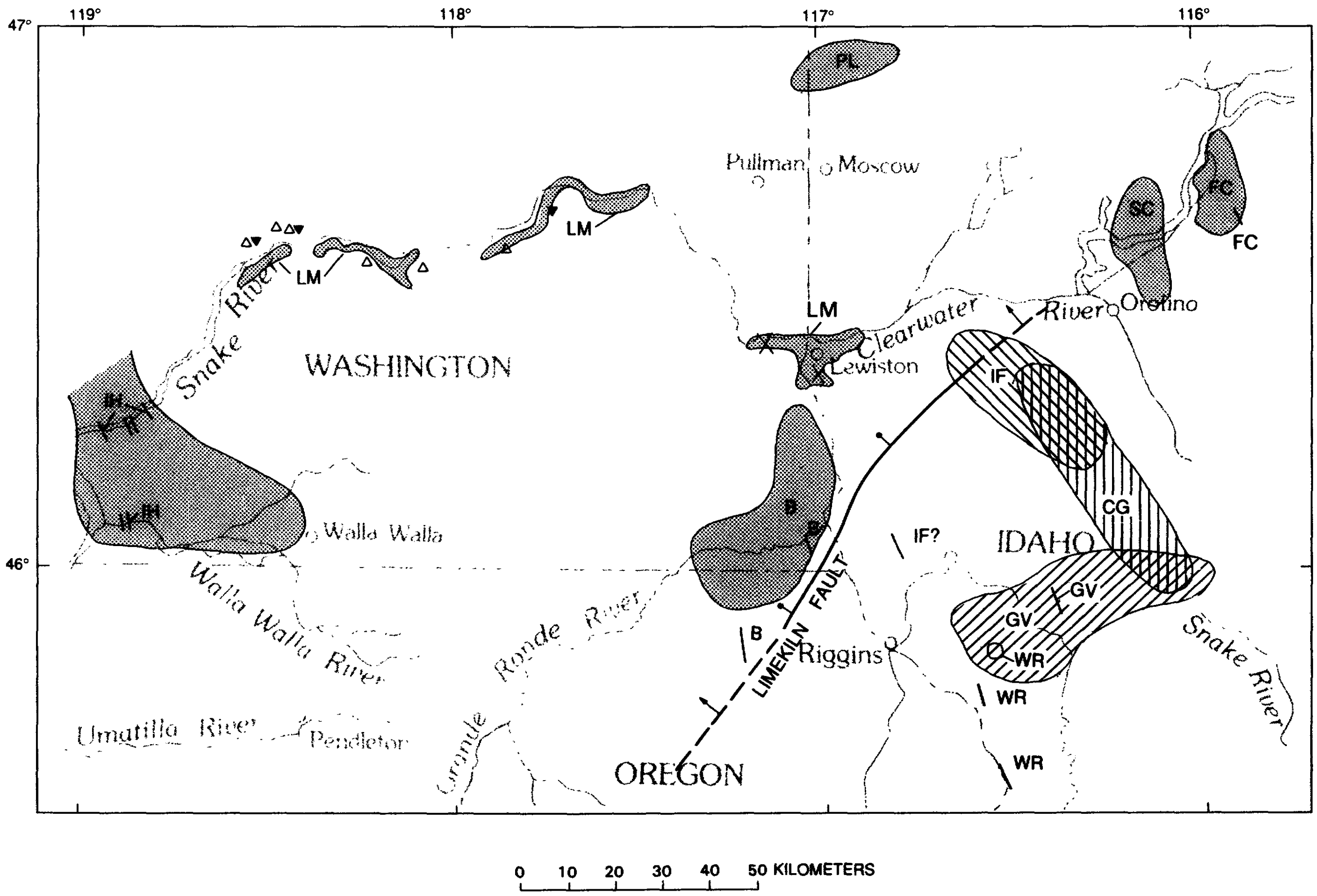


FIGURE 4.10.—Blue Mountains region, showing distribution of smaller flows of the Saddle Mountains Basalt: B, Buford (with possible dikes); CG, Craigmont; FC, Feary Creek (with dike); GV, Grangeville (with dike); IF, Icicle Flat (with possible dike); IH, Ice Harbor (with dikes); LM, Lower Monumental; PL, Potlatch; SC, Swamp Creek; WR, Windy Ridge (with dikes). Feary Creek and Potlatch flows may be equivalent in age to either the Wanapum or Saddle Mountains Basalt (Camp, 1981). △, isolated intracanyon outcrops of Esquatzel flow. ▼, isolated intracanyon outcrops of an unnamed flow. X, isolated outcrops of Tammany Creek intracanyon flow. Limekiln fault: dashed where approximately located, bar and ball on downthrown side; arrow indicates dip of monoclinal limb at both ends of fault.

flows entering from the west. The Wanapum and Saddle Mountains Basalts occur in these two basins only where feeder dikes intersected them (figs. 4.8–4.10; Camp and others, 1984).

The other two basins, Troy and Lewiston Basins, developed northwest of the Limekiln fault (fig. 4.9). These basins, synclines flanking the rising easterly extension of the Blue Mountains uplift, first became apparent during Wanapum time.

LEWISTON BASIN

No evidence exists of deformation in the Lewiston area from Imnaha through Grande Ronde (unit R_2) time. Flows of these formations, which are now folded and faulted, appear to retain their thickness across the area, and sedimentary interbeds are absent. Flows of unit N_2 of the Grande Ronde Basalt pinch out on the west edge of Lewiston Basin (Swanson and others, 1980) against a northwesterly slope formed by drag along the recently developed Limekiln fault (fig. 4.4). The Dodge and Shumaker Creek flows (of the Eckler Mountain Member), which have local feeder systems, behave similarly and indicate no deformation other than the surface rising toward the Limekiln fault (fig. 4.5).

Flows of the Frenchman Springs Member, erupted from fissures farther west, also pinch out against the northwest slope before reaching the Lewiston area (fig. 4.8). The Roza Member vented along the western margin of the present Lewiston Basin yet never reached the center of the basin, although it flowed westward for hundreds of kilometers. This behavior implies a northwesterly sloping surface between the vent system and the Limekiln fault (fig. 4.8). However, there is still no evidence of a Lewiston Basin at that time.

The first evidence of such a basin, albeit a very shallow one, comes from the Lolo flow of the Priest Rapids Member (fig. 4.8). The base of this flow is a pillow-palagonite complex over a wide area in the center of the present Lewiston Basin indicating that a shallow lake was present when the flow entered the area from the east (Camp, 1981). However, ponding of the drainage could also have resulted from construction of the Roza vent system to form a low topographic barrier across the southeast-to-northwest regional slope; thus, the ponding does not provide unambiguous evidence of deformation.

Younger flows and interbeds, however, indicate the presence of a syncline. The next flow (Umatilla Member of the Saddle Mountains Basalt) invaded a thick sedimentary deposit confined to the basin. Subsequent flows of the Saddle Mountains Basalt, from the Wilbur Creek Member to the four flows of the Weissenfels Ridge Member (figs. 4.9, 4.11), either thicken in or are entirely limited to Lewiston Basin (Swanson and others, 1980; Hooper and others, 1985), and many flows are separated by thick interbeds. These features could only result from contemporaneous subsidence, and they define an east-west synclinal basin approximately 40 km long.

Two major events modified this slowly developing syncline after the eruption of the Weissenfels Ridge Member. The Pomona Member (12 Ma; McKee and others, 1977) partly filled a deep canyon above and below Lewiston similar in size and shape to the current Snake River canyon. This deep canyon contrasts with the earlier, shallow drainage system that led northward and northwestward from the basin and was filled by flows of the Wilbur Creek and Asotin Members (of the Saddle Mountains Basalt). Clearly, a striking change in the drainage pattern occurred at this time.

The most probable explanation for this change in drainage pattern is the capture by the poorly developed Lewiston Basin drainage of the much larger ancestral Salmon River, which had previously flowed westward up the present Grande Ronde valley to cross the Blue Mountains uplift east of Walla Walla. This old channel is now defined by the presence of gravel below the Umatilla Member (fig. 4.9; Swanson and Wright, 1983), which apparently dammed the westerly flow to form a deep lake that eventually spilled northward over a low ridge into Lewiston Basin. Once the overflow began, the north-flowing stream would have rapidly cut down along the Limekiln fault, and the whole ancestral Salmon River system (which probably included the ancestral Clearwater River) was diverted through Lewiston Basin in a short period of time. The increased volume of water rapidly cut deep canyons along the lower Snake River, probably guided by an older, shallower drainage pattern. The new canyons provided channels for the Pomona and Elephant Mountain Members, and many subsequent intracanyon flows of the Saddle Mountains Basalt (figs. 4.9, 4.10).

The second major change in the form of Lewiston Basin occurred when the northern limb of the gentle syncline broke and was thrust northward and upward to fashion the Lewiston structure (Camp, 1976; Camp and Hooper, 1981; Hooper and others, 1985). This complex brittle structure includes a broken east-west anticlinal block thrust upward along steep reverse faults. The break probably occurred when steady north-northwest/south-southeast compression, the cause of the original syncline, produced a limb angle so steep that further compression led to brittle failure.

The result was a much deeper Lewiston Basin. The main faulting displaces the Elephant Mountain Member (10.5 Ma). Both the Tammany Creek flow and the Lower Monumental Member (6.0 Ma) of the Saddle Mountains Basalt were apparently erupted into a broad basin similar in shape to the present basin, and the main development

of the Lewiston structure probably preceded their eruption, that is, between 10.5 and 6.0 Ma. Reverse faulting associated with the structure continued, however, and has offset Pleistocene gravel (Kuhns, 1980; Webster and others, 1984).

TROY BASIN

Like Lewiston Basin, Troy Basin developed on the downthrown side of the Limekiln fault and its southwesterly monoclinal extension (figs. 4.8, 4.9). Unit N_2 flows of the Grande Ronde pinch out against the northwest slope of these structures, as does the Dodge flow, the most extensive flow of the Eckler Mountain Member (figs. 4.4, 4.5). Of the other flows in the Eckler Mountain, the oldest—the Robinette Mountain flow—was erupted west of the basin, and its easterly limit apparently reflects its limited volume. The younger Shumaker Creek and Lookingglass flows were erupted on the slope and flowed northwestward, with no obvious restriction in that direction except that imposed by their own limited volumes (fig. 4.5).

In the main part of the Wanapum Basalt, flows of the Frenchman Springs Member crossed the Blue Mountains uplift east of Walla Walla, but too little is known of the areas covered by individual flows to know whether any uplift had earlier resulted in a basin between it and the Limekiln monocline to the southeast. The Roza flowed only northwestward from its vent system (fig. 4.8), suggesting that significant movement on the northeasterly end of the Blue Mountains uplift had occurred by that time. More clearly, the Lolo flow (of the Priest Rapids Member; see fig. 4.2) was held north of high ground developed along the Blue Mountains anticlinal uplift. The Powatka flow, of Wanapum age, was the first flow clearly confined to Troy Basin (fig. 4.8).

The two main flows of the Umatilla Member were erupted near the divide between Lewiston and Troy Basins, entering both basins and invading basin sediment (fig. 4.9; Ross, 1978; Hooper and others, 1985). Their greatest volume, however, filled still-shallow Troy Basin and advanced westward across the now-rising Blue Mountains uplift, probably following the course of the

Lower Snake River Canyon	Lewiston Basin	Troy Basin	Cleerwater and Stites Basins	West-central Idaho		
Lower Monumental Mbr.	Lower Monumental Member —?—?—?—?—		Swamp Creek flow	Craigmont flow		Craigmont flow
Ice Harbor Member: Baselt of Goose Island; Basalt of Martindale; Basalt of Basin City	Tammany Creek flow —?—?—?—?—	—?—?—?—?—				
	Buford Member	Buford Member		Grangeville flow	Grangeville flow	Icicle Flats flow
Elephant Mountain Member	Elephant Mountain Member	Elephant Mountain Member			—?—?—?—?—	
Pomona Member	Pomona Member	Sedimentary rocks	Pomona Member (Weippe flow)	Pomona Member (Weippe flow)	Windy Ridge flow	
Unnemed flow (Tic)	Unnamed flow (Tic)					
Esquatzel Member	Sedimentary rocks					
	Weissenfels Ridge Member: Slippery Creek flow; Tenmile Creek flow; Lewiston Orchards flow; Cloverland flow	Eden flow		—?—?—?—	—?—?—	—?—?—
	Sedimentary rocks	Sedimentary rocks				
	Asotin Member		Asotin Member			
	Lapwai flow		Lepwei flow			
	Wilbur Creek Member		Wilbur Creek Member			
	Umatilla Member: Sillusi flow	Sillusi flow				
	Umatilla Member: Umatilla flow	Umatilla flow				
	Sedimentary rocks	Sedimentary rocks		? ? Onaway, Potlatch, and Feary Creek flows ? ?		
Priest Rapids Member	Priest Rapids Member	Powatka flow	Priest Rapids Member			
Wanapum Basalt						

FIGURE 4.11.—Stratigraphic succession of the Saddle Mountains Basalt. Relative ages of the Buford, Tammany Creek, and Lower Monumental flows are not clearly established. Relative ages of an unnamed intracanyon flow (unit Tic of Swanson and others, 1980) and the Esquatzel flow are not known; both are intracanyon and thus post-Weissenfels Ridge Member in age. Boundaries dashed and queried where uncertain.

ancestral Salmon and Grande Ronde Rivers (Swanson and Wright, 1983; Hooper and Swanson, 1987). These flows may have blocked the drainage for an extended period because thick epiclastic sedimentary rocks containing significant lignite beds overlie the Umatilla throughout Troy Basin (Ross, 1978; Stoffel, 1984). The next younger flow—the basalt of Eden (Ross, 1978)—was erupted from a large dike within the basin and invaded the loose sediment.

The dike that supplied the eruption of the Elephant Mountain Member also cuts across the center of Troy Basin (fig. 4.9). By the time of that eruption, the blocked drainage had apparently found an outlet northward across the low divide into Lewiston Basin and had cut rapidly down the Limekiln fault zone to form the deep lower Snake River canyon. The Elephant Mountain filled Troy Basin (already almost full of sediment), spilled over its northern divide into Lewiston Basin (Swanson and others, 1980), and then moved down the new canyon to reach beyond Pasco Basin. The younger Buford Member of the Saddle Mountains Basalt (fig. 4.10) provides no evidence of a basin in the Troy area, and so the present topographic basin must have developed since that time.

COMPOSITION OF THE SADDLE MOUNTAINS BASALT IN LEWISTON AND TROY BASINS

The flows of the Saddle Mountains Basalt present in Lewiston and Troy Basins are shown in figure 4.11. A thin ash-flow tuff above the Powatka flow in the southwestern part of Troy Basin is overlain by the Cricket Flat flow. This is one of several olivine-rich basalt flows mapped southward into the La Grande-Baker graben by W.H. Taubeneck (oral commun., 1980; Wright and others, 1980) and farther south to Baker (Swanson and others, 1981). Here these flows are regarded as belonging to the basalt of Powder River (Hooper and Swanson, 1987).

The Umatilla Member (Swanson, Bentley, and others, 1979) consists of two flows, the Umatilla flow and the overlying Sillusi flow, both of which are very fine grained and characterized by exceptionally high LIL-element contents (for example, more than 3,000 ppm Ba; fig. 4.3; table 4.1). Although their $^{87}Sr/^{86}Sr$ ratio (0.709) is average for the Saddle Mountains Basalt, their large Ba/Ti ratio and the presence of silicic xenoliths (S.P. Reidel, oral commun., 1985) are good evidence of upper-crustal contamination.

The Eden flow (Ross, 1978; Swanson and others, 1981) in Troy Basin is a conspicuous black glassy basalt containing clear-green olivine phenocrysts. It was fed by a single big dike (fig. 4.9). The unusually large P_2O_5 content of the Eden flow (1.2–1.4 weight percent; table 4.1) is matched only by the basalt of Prineville (Uppuluri, 1974). Samples from a small outcrop directly above the dike have even higher P_2O_5 and TiO_2 contents and probably represent a more evolved low-volume variant of the same magma near its vent.

In Lewiston Basin, the Umatilla Member is overlain by the Wilbur Creek Member, Lapwai flow, and Asotin Member. The Wilbur Creek is a dense, aphyric, commonly glassy flow with the highest recorded $\delta^{18}O$ value for the Columbia Plateau (table 4.2; Carlson, 1984). Like the Umatilla Member below, the Wilbur Creek typically invades the thick sediment of the basin. The Asotin is a dense, fine-grained flow containing abundant small, fresh olivine phenocrysts; it is one of the more primitive flows on the plateau. These two flows have similar distributions (fig. 4.9; Hooper, 1985), forming thick intracanyon flows in the North Fork of the Clearwater River, filling Lewiston Basin, and following a shallow ancestral drainage northward from the basin down the present Union Flat-Palouse River drainage (fig. 4.9) and thence westward to Pasco Basin north of the Saddle Mountains (Swanson, Bentley, and others, 1979; Swanson and others, 1980; Anderson and others, 1987). Locally, the Lapwai flow lies between the Wilbur Creek and Asotin. This flow is intermediate in composition and appears to be the result of simple mixing of the magmas forming the two larger flows (Hooper, 1985).

The Elephant Mountain Member, present in both Troy and Lewiston Basins (Swanson and others, 1980), has one of the largest TiO_2 contents in the Columbia River Basalt Group, but its other chemical and isotopic characteristics resemble those of many flows of the Saddle Mountains Basalt (fig. 4.3; tables 4.1, 4.2).

In Lewiston Basin the Elephant Mountain is underlain by the intracanyon Pomona Member. The Pomona traveled farther than any other known flow, from its feeder dike on the east edge of the Clearwater River embayment, Idaho (fig. 4.9), to the Pacific Ocean, 600 km to the west (Beeson and others, 1979; Swanson, Wright, and others, 1979; Camp, 1981; Hooper, 1982; Magill and others, 1982). Despite its rather primitive composition, the Pomona flow contains small phenocrysts of olivine, plagioclase, and clinopyroxene.

OTHER FLOWS OF THE SADDLE MOUNTAINS BASALT

Numerous other small flows of Saddle Mountains age have been recorded on the Columbia Plateau (fig. 4.10). Most of these flows are of limited areal extent, and so stratigraphic relations are imprecisely known (fig. 4.11). Descriptions of these flows may be found in Swanson, Wright, and others (1979), Camp (1981), Swanson and others (1981), and Hooper and others (1985). The flows of the Ice Harbor Member (Swanson and others, 1975) are particularly significant because they were fed by a

fissure-vent system oriented north-northwest close to Pasco Basin, apparently entirely outside the Blue Mountains oceanic-crustal terranes. Their wide chemical diversity, in combination with restricted $^{87}Sr/^{86}Sr$ values similar to those of many other flows in the Saddle Mountains Basalt (fig. 4.3), make them of special interest in discussion of possible crustal contamination.

Most other local flows of the Saddle Mountains Basalt are found either in west-central Idaho or in the Snake River canyon as intracanyon flows (figs. 4.10, 4.11).

The Buford Member (Walker, 1973) overlies the Elephant Mountain flow on the divide between Lewiston and Troy Basins. Erupted locally, the Buford is of limited volume and has a texture, mineralogy, and major-element composition indistinguishable from those of the Grande Ronde Basalt, though with a different proportion of Sr and a different $^{87}Sr/^{86}Sr$ ratio (fig. 4.3).

The Lower Monumental Member, probably the youngest flow in the Columbia River Basalt Group (6.0 Ma; McKee and others, 1977), is entirely confined to the Snake River canyon. Another nearby flow, the Tammany Creek flow (Hooper and others, 1985), is of similar age but contains augite in addition to olivine phenocrysts and is of more local distribution. Compositions of these flows are listed in tables 4.1 and 4.2 and shown in figures 4.12 to 4.16.

A dike, a plug, and a small capping flow in the center of a small exposure west of Riggins, Idaho, have a composition similar to that of Lewiston Orchards flow (of the Weissenfels Ridge Member) in Lewiston Basin (fig. 4.10; Hooper, 1984b). Small outcrops of a basalt of similar composition are present above stream sediment near the towns of Colton (between Pullman and Lewiston) and Moscow, 30 mi north of Lewiston (fig. 4.10). The basalt of Sprague Lake (Swanson, Wright, and others, 1979; Wright and others, 1980), found in a small area southwest of Spokane (fig. 4.1), has a similar composition (table 4.3). All of these outcrops lie along a north-northwest line and may represent small eruptions from a single fissure system almost 300 km long.

ORIGIN OF THE SADDLE MOUNTAINS BASALT

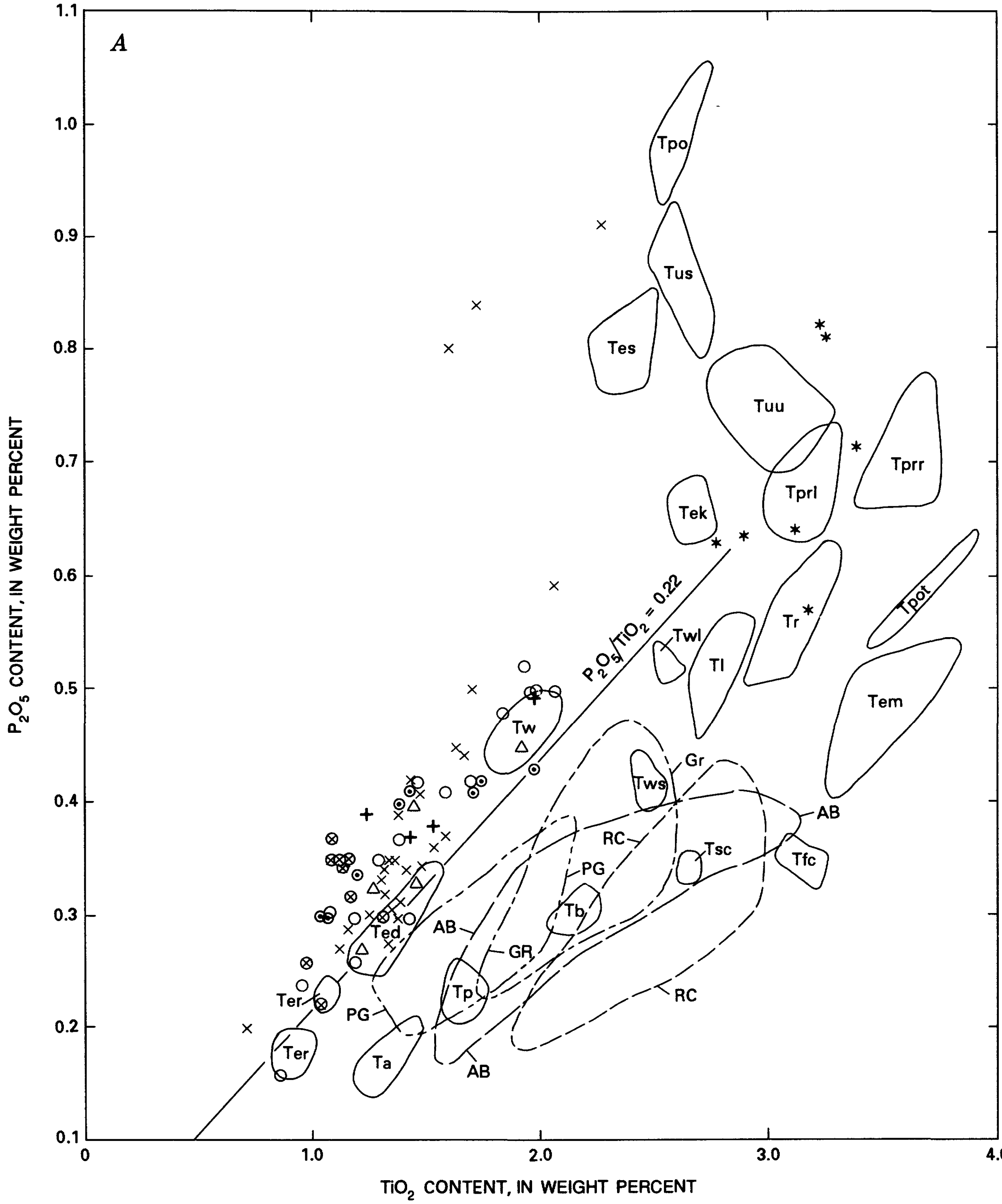
The Saddle Mountains Basalt displays as much mineralogic and chemical diversity as the entire Columbia River Basalt Group. Some flows are aphyric, and others are porphyritic, containing phenocrysts of plagioclase, olivine, and, more rarely, pyroxene. No correlation exists between either abundance or type of phenocryst phase and bulk chemical composition. For example, the presence of pyroxene as a third phenocryst phase does not correlate with a more evolved chemical composition as reflected in the abundance of incompatible trace elements; the Pomona Member, containing plagioclase, olivine, and clinopyroxene microphenocrysts, has smaller abundances of incompatible trace elements than do most other flows (fig. 4.12). Simple crystal fractionation and varying degrees of partial melting are therefore inadequate in themselves to explain the variations between flows in the Saddle Mountains Basalt.

The high but varying values of the ratio $^{87}Sr/^{86}Sr$ (and corresponding small $^{143}Nd/^{144}Nd$ ratios) of these flows suggest crustal contamination, although their $\delta^{18}O$ values are only moderate (table 4.2). Multiple sources are certainly required (Carlson, 1984), and contamination of a mantle-derived magma by one or more crustal components may have occurred. This hypothesis is the more attractive considering the small volume of most of the flows and the long intervals between many eruptions, which could imply long residence times within the crust.

However, strong arguments remain against crustal contamination as an essential process in the evolution of the Saddle Mountains Basalt. The $\delta^{18}O$ value is relatively small in most flows and does not correlate well with the ratio $^{87}Sr/^{86}Sr$ (Carlson and others, 1981). We have already remarked on the evident absence of correlation between the isotopic signature of a flow and the composition of the crust penetrated by its feeder system (Hooper, 1984a). More critical is the fact that all the Sr isotopic data available for rocks from the accreted ter-

FIGURE 4.12.—Plots of P_2O_5 versus TiO_2 contents for the Columbia River Basalt Group and associated volcanic rocks. *A*, Flows of the Columbia River Basalt Group and basalt of Powder River and associated volcanic rocks. Flows of the Columbia River Basalt Group: AB, Imnaha (AB subgroup); GR, Grande Ronde field; PG, Picture Gorge field; RC, Imnaha (RC subgroup); Ta, Asotin; Tb, Buford; Ted, Dodge; Tek, Lookingglass; Tem, Elephant Mountain; Ter, Robinette Mountain; Tes, Shumaker Creek; Tfc, Feary Creek; Tl, Lower Monumental; Tp, Pomona ; Tpo, Powatka; Tpot, Potlatch; Tprl, Lolo; Tprr, Rosalia; Tr, Roza; Tsc, Swamp Creek; Tus, Sillusi; Tuu, Umatilla; Tw, Wilbur Creek; Twl, Lewiston Orchards; Tws, Slippery Creek. Modified from Hooper (1982). Basalt of Powder River and associated volcanic rocks (less than 55 weight percent SiO_2): ⦿, Cricket Flat area; ⊗, olivine basalt in La Grande Basin (W.H. Taubeneck, *in* Wright and others, 1980); ○, diktytaxitic olivine basalt from La Grande Basin (W.H. Taubeneck, *in* Wright and others, 1980); ×, east side of Baker Valley; △, Medicine Springs area; +, west and south sides of La Grande Basin (Barrash and others, 1980); ✳, nepheline basalt from La Grande Basin (Wright and others, 1980). *B*, Basalt and basaltic andesite of the basalt of Weiser (equivalent to the Weiser Basalt of Fitzgerald, 1984). ○, Sugarloaf member; ×, Star Butte member; △, Cambridge member; ✳, basalt of Cuddy Mountain (alkali basalt) member (informal members as designated by Fitzgerald, 1984). *C*, Andesites (more than 55 weight percent SiO_2) of the basalt of Powder River and associated volcanic rocks (×) relative to andesitic rocks of the Strawberry Volcanics (●) and the Slide Creek (basaltic) member (⦿) of the Strawberry Volcanics (Robyn, 1977).

ranes (Armstrong and others, 1977; Fleck and Criss, 1985) indicate $^{87}Sr/^{86}Sr$ ratios lower than those of the Saddle Mountains Basalt. These are the crustal rocks through which almost all of the Saddle Mountains Basalt magmas passed on their way to the surface. Clearly, contamination by such crustal material could not have been responsible for the high isotopic ratios (more than 0.706) observed in these flows. This observation appears to leave no alternative but to ascribe the high $^{87}Sr/^{86}Sr$ ratios of these basalts to an enriched mantle source.



PICTURE GORGE BASALT

The Picture Gorge Basalt (Waters, 1961; Swanson, Wright, and others, 1979) is confined almost entirely to John Day Basin south of the Blue Mountains uplift (fig. 4.1). Flows of the Picture Gorge cross low saddles in the uplift and locally interfinger with the Grande Ronde Basalt. They cover an area of more than 1,200 km^2

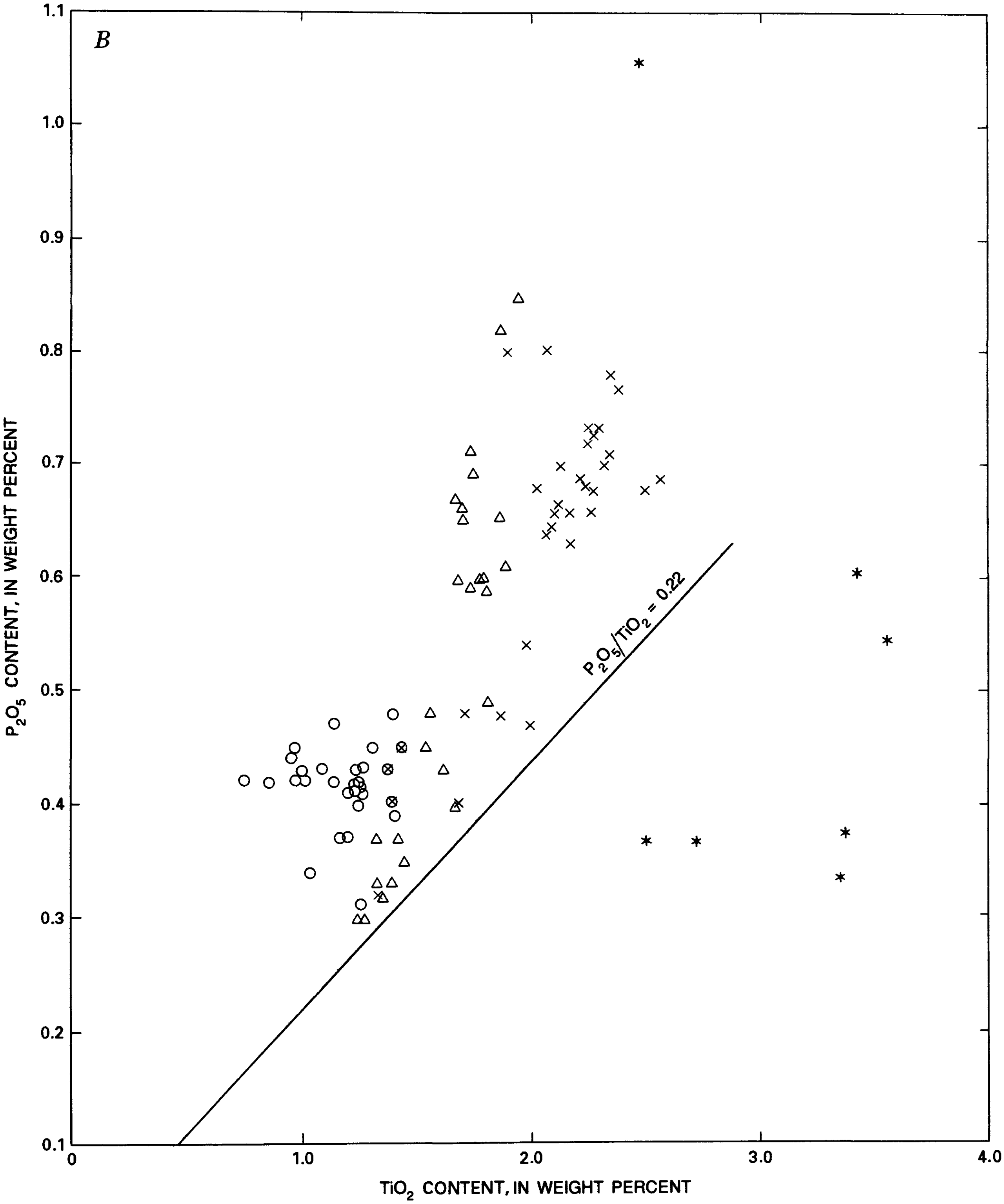


FIGURE 4.12.—Continued

(Bailey, 1985) and were fed by the local Monument dike swarm (Fruchter and Baldwin, 1975). Some stratigraphic sections are as thick as 800 m (Swanson, Bentley, and others, 1979). Many flows near the base and top of the formation contain plagioclase phenocrysts. Porphyritic flows in the northwest corner of John Day Basin are capped by the Grande Ronde Basalt. This relation led Waters (1961) to correlate the Picture Gorge with the equally porphyritic and chemically similar Imnaha Basalt. Watkins and Baksi (1974) showed that most flows in the Picture Gorge had normal magnetic polarity and that only a short sequence of flows at the top were reversed. They regarded this as confirmation of Waters' correlation.

More recent work has questioned this correlation. Nathan and Fruchter (1974) and McDougall (1976) argued that the Imnaha and Picture Gorge Basalts have distinguishable compositions. Swanson and others (1981) traced the Grande Ronde magnetostratigraphic units from east to west and showed that the upper, magnetically reversed flows in the Picture Gorge interfinger with unit R_2 flows in the Grande Ronde. This relation implies that the underlying normal flows of Picture Gorge are time equivalents of units N_1 of the Grande Ronde Basalt and not of unit N_0 of the Imnaha Basalt (fig. 4.2). More detailed comparisons of the Picture Gorge and Imnaha Basalts (Osawa and Goles, 1970; Bailey, 1986) make clear that the Picture Gorge is more primitive than the Imnaha Basalt and can be clearly separated by some trace elements, such as Sc versus Zr (fig. 4.17). The most primitive flow in the Imnaha (AB-2; see Hooper and others, 1984) comes closest to the Picture Gorge in composition but is distinguished in a plot of Sc versus Zr content (fig. 4.17); flow AB-2 is coarsely porphyritic, but the flows of the Picture Gorge Basalt with the most nearly equivalent chemical compositions are aphyric (M.M. Bailey, oral commun., 1985).

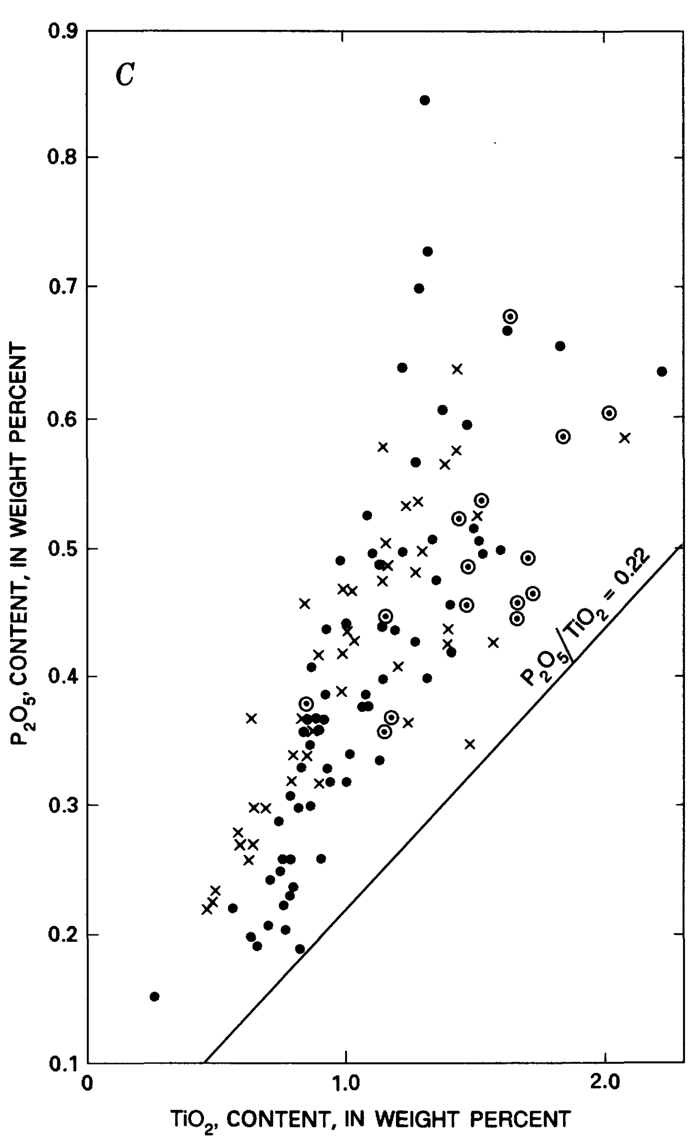


FIGURE 4.12.—Continued

A further distinguishing feature of the Picture Gorge Basalt relative to all other units in the Columbia River Basalt Group is the rare-earth-element (REE)-abundance pattern (fig. 4.18). The Picture Gorge Basalt has a consistently flatter pattern, less fractionated between light REE's and heavy REE's, that tends to cross those of the more primitive flows of the main series of the Columbia River Basalt Group. This REE pattern implies a mantle source for the Picture Gorge that was more depleted than that for those flows supplied by the Chief Joseph dike swarm of the main series.

Bentley and Cockerham (1973) divided the Picture Gorge into three stratigraphic units: a coarsely porphyritic lower unit (basalt of Twickenham of Swanson, Wright, and others, 1979), an aphyric middle unit (basalt of Monument Mountain of Swanson, Wright, and others, 1979), and a porphyritic upper unit (basalt of Dayville of Swanson, Wright, and others, 1979). Using major- and trace-element analyses and magnetic polarity, Bailey (1985, 1986) further subdivided the formation into 15 map units, each composed of 1 to as many as 9 separate flows. The Twickenham flows are very thick, fill deep canyons, and have restricted areal extent. They contain more than 6.0 weight percent MgO, are magnetically normal, and may include very coarse grained segregation veins (Lindsley and others, 1971). The Monument Mountain flows are thinner but cover a larger area than the Twickenham flows; they also contain more than 6.0 weight percent MgO and have normal magnetic polarity, but they are aphyric. The Dayville flows form a much more diverse greup of porphyritic flows; their magnetic polarity is normal at the base and reversed at the top of the unit, and six of the eight subunits contain less than 6.0 weight

percent MgO. The chemical variation within the Picture Gorge is controlled, at least in part, by gabbroic (plagioclase+olivine+clinopyroxene) fractionation (Bailey, 1985, 1986; Goles, 1986).

BASALT OF PRINEVILLE

Uppuluri (1974) described 13 flows of a distinct chemical composition in north-central Oregon (fig. 4.1). Nathan and Fruchter (1974) found flows of this composition interfingered with unit N_2 flows of the Grande Ronde Basalt on the north side of John Day Basin (fig. 4.1), and other similar flows have been reported in the western Cascade Range (Anderson, 1978; Swanson, Wright, and others, 1979).

No feeder dikes have been reported, but the restricted extent of these flows requires several local sources (Smith and Cushing, 1985), whether of fissure or central type, far west of any known Columbia River Basalt Group feeder dike. On almost any plot of elemental compositions (see figs. 4.13, 4.14), these flows show remarkably little variation and possess a distinctive combination of high Ba, Sr, and P contents and moderate TiO_2 and Zr contents. Although these flows are clearly contemporaneous with the upper part of the Grande Ronde Basalt, their relative geographic isolation, the absence of evidence for a fissure eruptive mechanism, and their distinctive composition all indicate an origin for the Prineville-type flows different from that of the Columbia River Basalt Group. New chemical analyses for Prineville flows are listed in table 4.1.

BASALTS OF POWDER RIVER AND WEISER AND ASSOCIATED VOLCANIC ROCKS

POWDER RIVER

Diverse, relatively local flows in the La Grande-Baker Valley of northeastern Oregon (fig. 4.1) are considered together under this heading. The older flows lie unconformably on units R_1 and N_1 of the Grande Ronde Basalt and consist of olivine-rich basalt. Overlying these flows

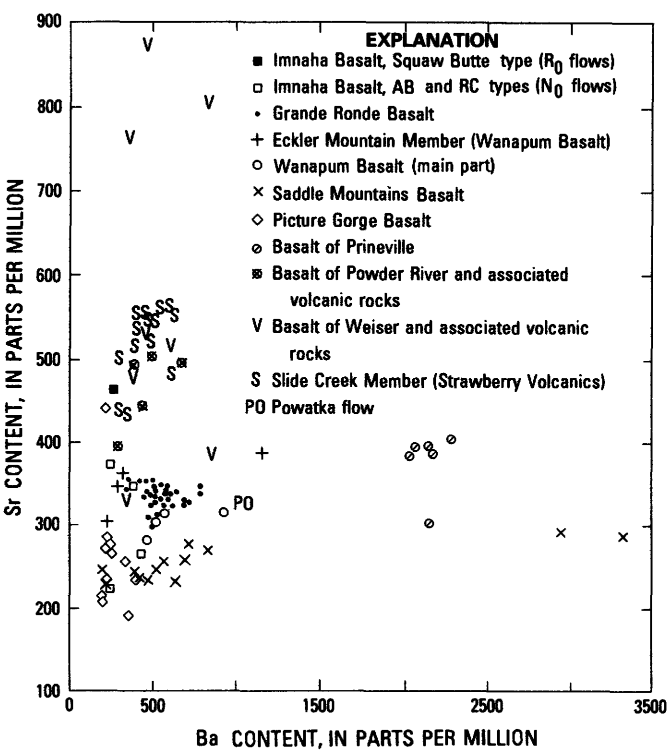


FIGURE 4.13.—Sr versus Ba contents for the Columbia River Basalt Group and associated volcanic rocks. Analyses in table 4.1 are plotted with analyses of additional samples from the Grande Ronde Basalt (X-ray-fluorescence analyses from Reidel, 1983) and analyses of samples from the Slide Creek Member of the Strawberry Volcanics (X-ray-fluorescence and instrumental neutron-activation analyses from Robyn, 1977). Magnetic polarities: N, normal; R, reversed. Polarity intervals numbered from oldest to youngest where sequential.

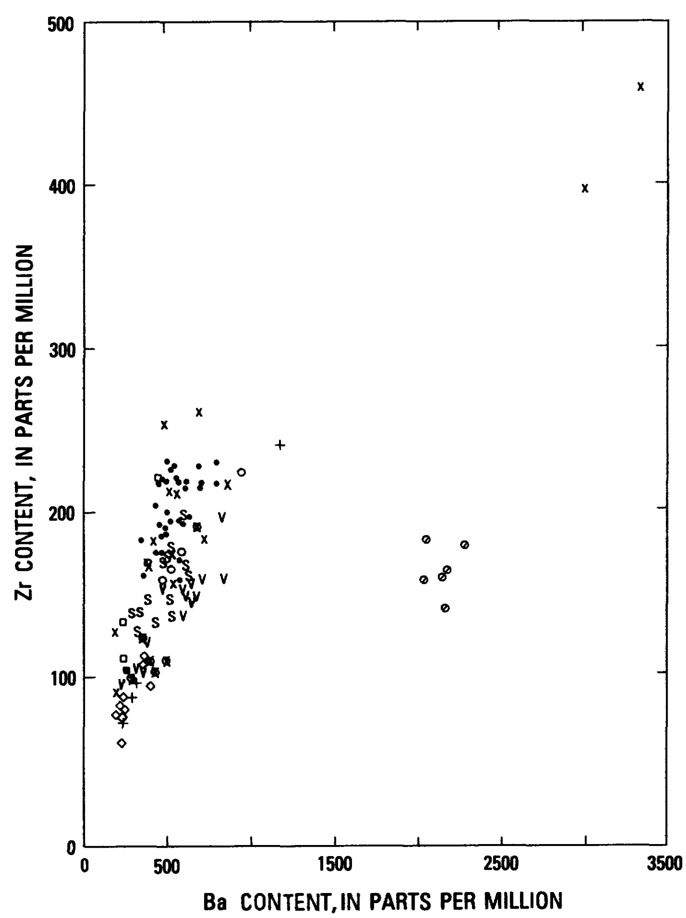


FIGURE 4.14.—Zr versus Ba contents for the Columbia River Basalt Group and associated volcanic rocks. Analyses and symbols same as in figure 4.13.

are basaltic andesite, trachybasalt, andesite, and at least one flow of nepheline basalt or basanite (W.H. Taubeneck, written communs., 1978, 1979; Barrash and others, 1980; Wright and others, 1980; Swanson and other, 1981; Hooper, 1984a; Hooper and Swanson, 1987). At the north end of the La Grande graben, one olivine basalt flow, the Cricket Flat, rests on the Powatka flow and is overlain by the Umatilla Member (figs. 4.2, 4.6, 4.11), and it is therefore a late Wanapum age. The Cricket Flat flow is locally underlain by a silicic ash-flow tuff. From La Grande southward to Baker (fig. 4.1), thick andesite flows cap the olivine basalt and spread northward over unit N_2 of the Grande Ronde Basalt (Swanson and others, 1981). Measured ages for these andesite flows range from 7.0 Ma northwest of La Grande (Kienle and others, 1979) to 12 to 9 Ma east of La Grande (E.H. McKee and W.H. Taubeneck, written commun., 1979). South of Baker, olivine basalt flows are interleaved with and overlain by flows of basaltic andesite and andesite that are indistinguishable from the Strawberry Volcanics described by Robyn (1977, 1979).

Unlike most flows of the Columbia River Basalt Group, those of the basalt of Powder River are of small extent, lie primarily along the faulted margins of La Grande and Baker Basins (fig. 4.19), and form small low-angle cones. Both basins are grabens in which the predominant fault trend is northwest-southeastward (Walker, 1977). The grabens form a series of major basins—Elgin, La Grande, Baker, and Bowen—that lie in a north-south line (fig. 4.19). Reconnaissance mapping suggests a direct correlation between the development of the grabens and the eruption of olivine basalt along their margins. Some of the overlying andesite domes and flows have well-developed craters (Patterson, 1969; Brooks and others, 1977). No feeder dikes have been identified.

The basalt flows are commonly diktytaxitic, containing small but conspicuous olivine phenocrysts in a groundmass in which isolated crystals of ophitic augite are

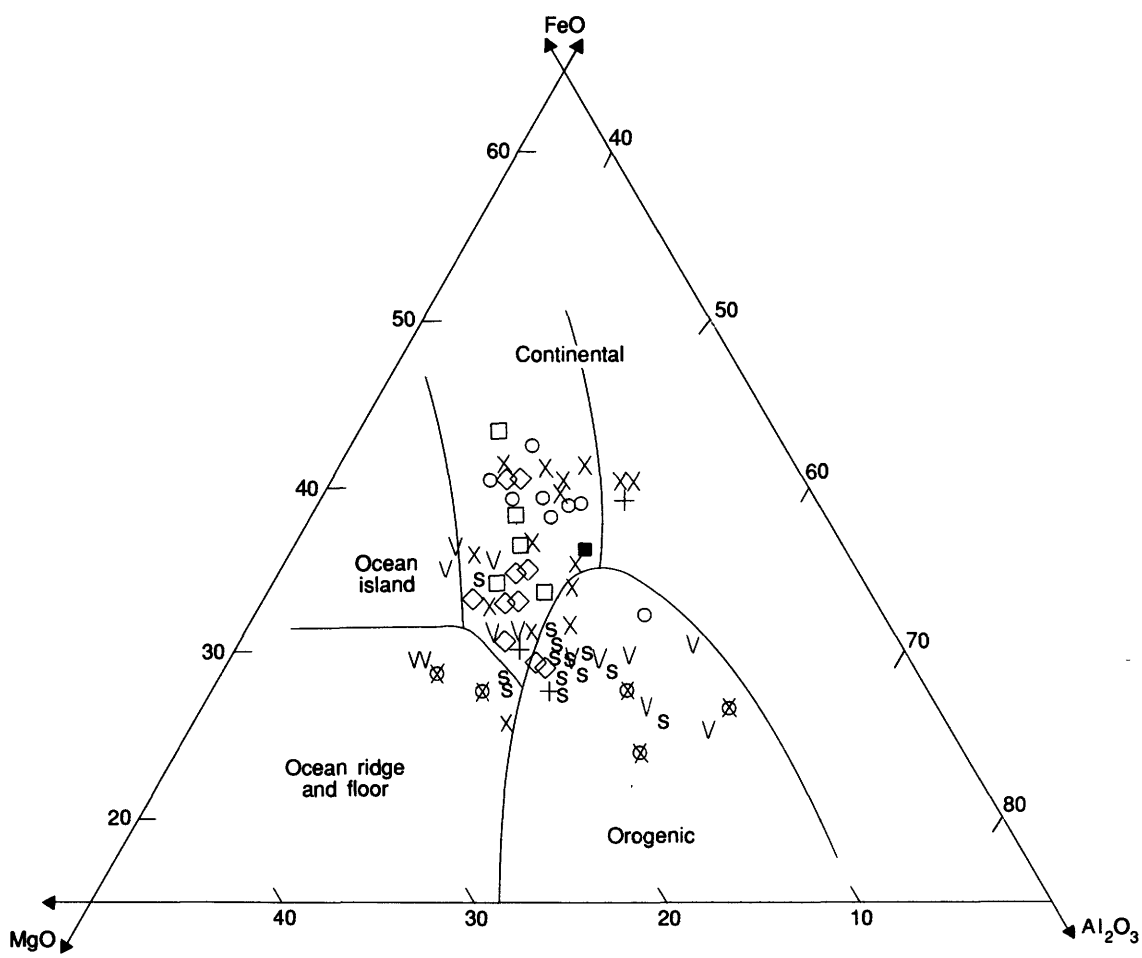


FIGURE 4.15.—Ternary plot of FeO, MgO, and Al_2O_3 contents for the Columbia River Basalt Group and associated volcanic rocks and for the Slide Creek Member (S) of the Strawberry Volcanics (Robyn, 1977). Symbols same as in figure 4.13. Tectonic fields after Pearce and others (1977). Arrows point to pure end members.

separated by areas of devitrified glass between plagioclase laths. In many flows the olivine is largely altered. Pyroxene phenocrysts are rare. Some andesite flows are glassy with a dense trachytic texture; others contain phenocrysts that may include plagioclase, quartz, hornblende, and, rarely, relics of olivine, which are typically rounded by resorption.

Chemically, the Powder River rocks are characterized by a higher P_2O_5/TiO_2 ratio (0.22) than those of flows of the Columbia River Basalt Group (fig. 4.12*A*). Many Powder River flows also have higher Al_2O_3, lower Fe, and much higher Sr contents than Columbia River Basalt Group flows with the same MgO content (figs. 4.13, 4.20). Qualifications to each of these statements are required. First, the nepheline basalt does not fit this pattern (fig. 4.12*A*). Second, some of the most evolved flows of the Columbia River Basalt Group (Eden, Powatka, Umatilla, Lookingglass, and Wilbur Creek flows) have P_2O_5/TiO_2 ratios close to or greater than 0.22. Each of these flows, however, is clearly distinguishable from the Powder River flows in both mineralogy and chemical composition.

The only Columbia River Basalt Group flows with chemical compositions that could be confused with the Powder River olivine basalt flows are the Robinette Mountain and Dodge flows of the Eckler Mountain Member (of the Wanapum Basalt). Of these flows, the Dodge has a consistently lower Al_2O_3 content than Powder River basalt of equivalent MgO content (fig. 4.20). Both the Dodge and the Robinette Mountain flows also have lower K_2O and lower Sr contents than otherwise-similar Powder River flows (figs. 4.13, 4.20). The Columbia River Basalt Group flow most similar to the Powder River rocks, the Robinette Mountain flow, is of broadly similar age and was erupted along a north-northwest fissure close to the northerly extension of the Powder River-area graben structures.

WEISER

Fitzgerald (1984) described a sequence of basalt flows (which he called the Weiser Basalt) resting on the lower part of the Grande Ronde Basalt in the Weiser embayment of west-central Idaho (fig. 4.1). The flows are of diverse composition, ranging from olivine alkali basalt (basalt of Cuddy Mountain), through various types of basalt and basaltic andesite, to typical calc-alkaline andesite, dacite, and silicic ash-flow tuff (table 4.1). The flows are of limited volume and associated with northwest to north-northwest grabens and horsts.

Comparison of chemical compositions (figs. 4.12–4.16) shows that these flows have the same gross chemical characteristics as the Powder River flows.

DISCUSSION

Few data are available for the Powder River flows, and even fewer for the Weiser flows. Nonetheless, the similar tectonic setting, mode of occurrence, and range of gross chemical characteristics show that the two units have

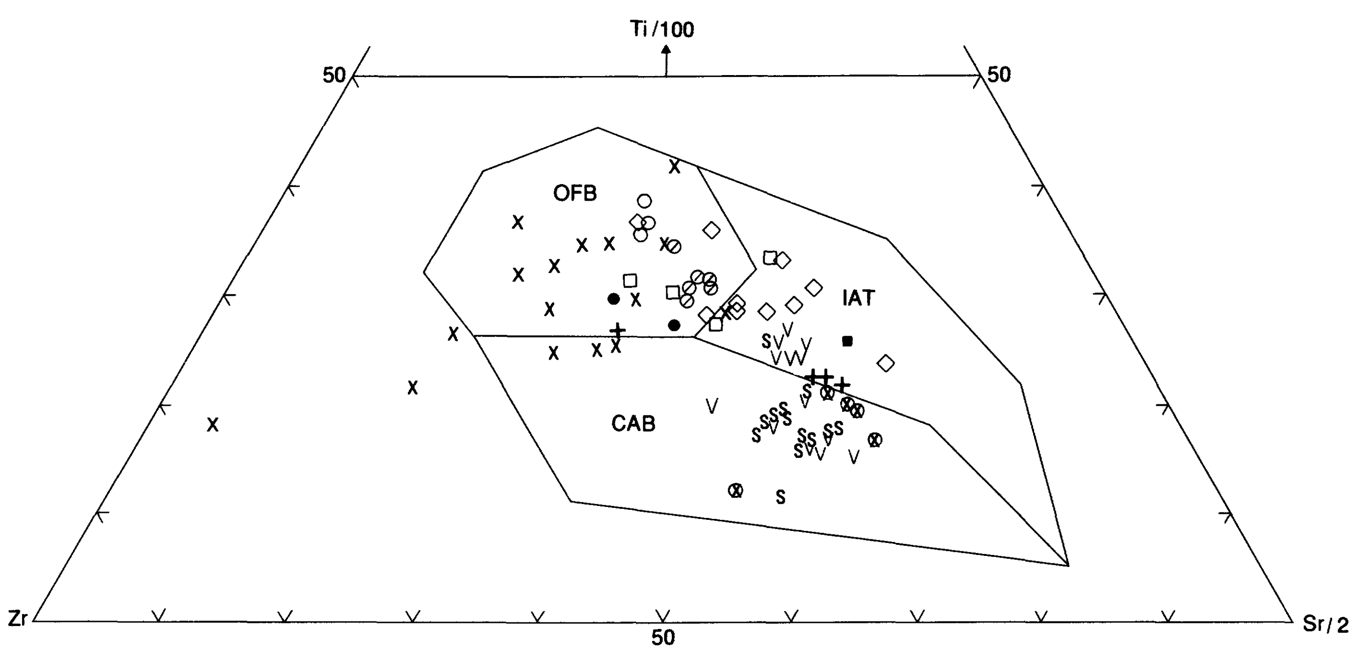


FIGURE 4.16.—Ternary plot of Ti/100, Zr, and Sr/2 contents for the Columbia River Basalt Group and associated volcanic rocks and for the Slide Creek Member (S) of the Strawberry Volcanics (Robyn, 1977). Symbols same as in figure 4.13. Tectonic fields after Pearce and Cann (1973). OFB, ocean-floor basalt; IAT, island-arc (low-K) tholeiite; CAB, calc-alkaline basalt.

TABLE 4.3.—*Comparative compositions of flows and dikes of Lewiston Orchards (Weissenfels Ridge Member) chemical type from localities in Washington and Idaho*

[Analyses determined at Washington State University's Geoanalytical Laboratory by X-ray-fluorescence analysis. Analyses normalized on a volatile-free basis, with iron (Fe*) expressed as 2.00 weight percent Fe_2O_3 added to remainder as FeO]

Sample	1	2	3	4	5	6	7
SiO_2	49.24	49.83	50.30	49.30	51.44	48.86	49.98
Al_2O_3	14.40	15.12	15.63	15.23	15.42	14.61	14.96
TiO_2	2.75	2.56	2.61	2.52	2.77	2.59	2.37
Fe*	13.48	12.52	11.53	11.74	12.07	12.26	12.20
MnO	.23	.20	.22	.20	.19	.20	.20
CaO	10.05	10.60	11.13	11.06	9.14	10.95	10.31
MgO	6.12	6.54	5.93	6.69	5.03	6.94	6.67
K_2O	.75	.46	.31	.41	.99	.62	.56
Na_2O	2.44	1.66	1.82	2.34	2.41	2.41	2.19
P_2O_5	.58	.51	.52	.52	.54	.55	.48

Sample:
1. Average of four analyses of samples of the basalt of Sprague Lake, Wash. (Wright and others, 1980).
2. Sample HUT-8, flow above interbed near Colton, Wash.
3. Sample HVI-1, thin surface flow northwest of Moscow, Idaho
4. Average of 16 analyses of samples of the Lewiston Orchards flow, Lewiston Basin, Wash. and Idaho (Hooper and others, 1985).
5. Average of two analyses of separate dikes (Weissenfels Ridge Member) in the Lewiston Basin, Wash. and Idaho (Hooper and others, 1985).
6. Squaw Butte vent, Riggins Basin, Idaho (Hooper, 1984b).
7. Average of two analyses of dike feeding Squaw Butte vent, Riggins Basin, Idaho (Hooper, 1984b).

much in common, despite their geographic separation. Original descriptions placed both within the Columbia River Basalt Group (Carlson and others, 1981; Swanson and Wright, 1981; Fitzgerald, 1984). Because of their different tectonic settings and compositions, however, this assignment is questionable (Hooper, 1984a), and for the purposes of this report we treat them as separate informal units (referring to them herein as the basalts of Powder River and Weiser, respectively).

Both units correlate more convincingly with the Strawberry Volcanics described by Robyn (1977, 1979; Goles, 1986) than with the Columbia River Basalt Group. Robyn (1977) described a sequence of basalt flows (Slide Creek Member) forming the basal part of the Strawberry Volcanics that he distinguished from the Picture Gorge Basalt by their higher Al_2O_3, K_2O, P_2O_5, and Sr contents and lower Fe contents. The Powder River and Weiser rocks resemble the Slide Creek in these and other chemical characteristics (figs. 4.12–4.16). Like the Slide Creek, the basalts of Powder River and Weiser are closely associated with basaltic andesite, andesite, dacite, and silicic ash-flow tuff of typical calc-alkaline affinity. The low Fe content (relative to Mg and Al) and the high Sr content (relative to Ti and Zr)—two characteristics that distinguish the Slide Creek, Powder River,

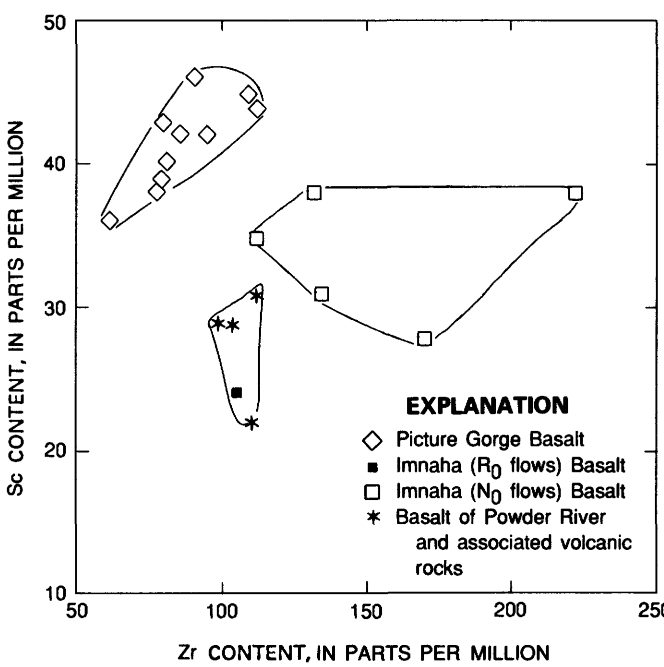


FIGURE 4.17.—Sc versus Zr contents for the Picture Gorge Basalt, the Imnaha Basalt, and basalt of Powder River and associated volcanic rocks. Magnetic polarities: N, normal; R, reversed. Polarity intervals numbered from oldest to youngest (for example, N_0 to N_3) where sequential.

and Weiser basalts from the Columbia River Basalt Group (figs. 4.15, 4.16)—are typical of subduction-related volcanic rocks (Pearce and Cann, 1973; Pearce and others, 1977). In addition, both the Powder River and Weiser units include flows of obvious alkaline affinity, such as the nepheline basalt near La Grande (Wright and others, 1980) and the Cuddy Mountain flows of the Weiser unit (Fitzgerald, 1984).

Detailed studies of the Powder River and Weiser units have not been made. Data on the $^{87}Sr/^{86}Sr$ ratio for three samples of the Powder River unit (an olivine basalt, an andesite, and a nepheline basalt) were presented by Carlson and others (1981). In contrast to the younger Wanapum and Saddle Mountains Basalts of the Columbia River Basalt Group north of La Grande, all three Powder River samples retain a so-called primitive Sr isotopic ratio (0.7035) similar to that of the older Picture Gorge and Imnaha Basalts (Hooper, 1984a).

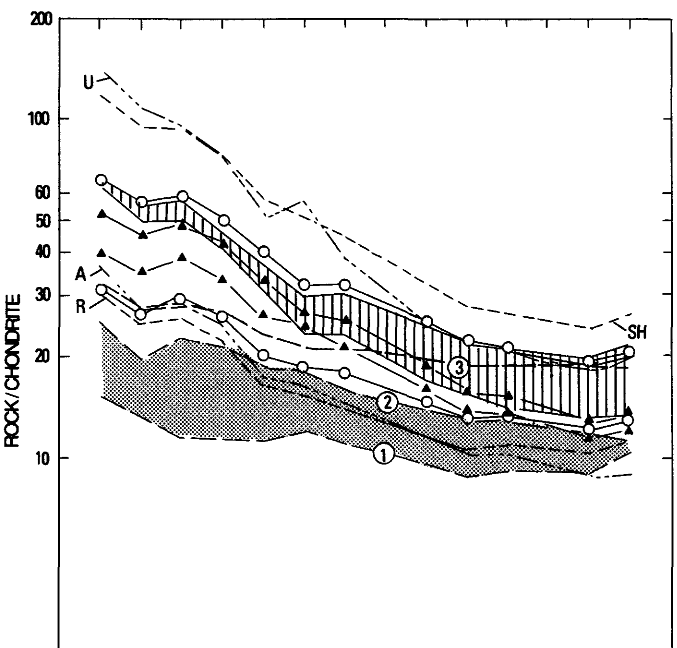


FIGURE 4.18.—Chondrite versus normalized rare-earth element abundance pattern for the Picture Gorge Basalt and other units of the Columbia River Basalt Group (table 4.1). Long-dashed lines 1, 2, and 3, Picture Gorge Basalt (1 and 2 represent extreme values for all but one flow; 3 is the more evolved exception). Circles, extreme values for subgroup AB (of the Imnaha Basalt); triangles, extreme values for subgroup RC (of the Imnaha Basalt). Vertical-line pattern represents two flows of the Grande Ronde Basalt. Short-dashed lines represent extreme values for the Eckler Mountain Member (R, Robinette Mountain flow; SH, Shumaker Creek flow of the Wanapum Basalt). Dashed lines with three short dashes represent extreme values of the Saddle Mountains Basalt (A, Asotin flow; U, Umatilla flow).

TECTONIC SETTING OF THE BASALTIC VOLCANISM

Delineation of the stratigraphic details of the basaltic pile (made possible by rapid analytical and paleomagnetic techniques in combination with geologic mapping) provides an unusually clear view of the structural and geomorphic evolution of the Blue Mountains province during the early to late Miocene. The regional stress regime of horizontal compression oriented north-northwest/south-southeast and horizontal extension oriented east-northeast/west-southwest remained in effect throughout the period of volcanism. However, tectonic activity toward the end of Grande Ronde time along a broad zone corresponding approximately to the Olympic-Wallowa Lineament (OWL) resulted in a different struc-

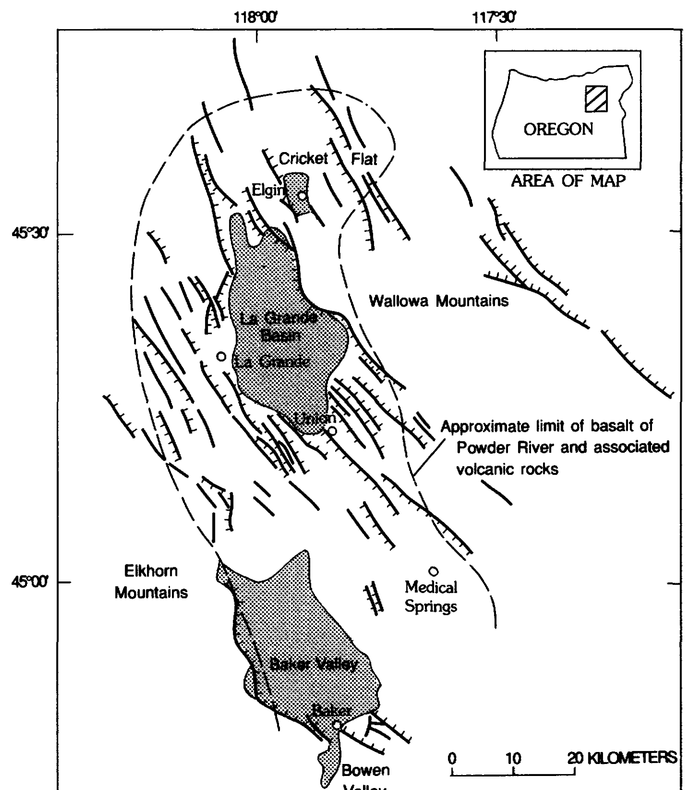


FIGURE 4.19.—Structural map of part of northeastern Oregon, showing Elgin and La Grande Basins and Baker and Bowen Valleys (stippling), and northwest-trending normal faulting with graben-horst development and approximate north limit of the basalt of Powder River and associated volcanic rocks. Modified from Brooks and others (1977) and Walker (1977, 1979). Solid line, normal faults; hachures on downthrown side.

tural expression of that stress regime on either side of the OWL, subsequently accompanied by different styles of volcanism.

The earliest eruptions (Imnaha Basalt) were concentrated in the southeast corner of the Columbia Plateau along the Oregon-Idaho State line (fig. 4.4). The later flood of the Grande Ronde Basalt covered areas increasingly farther to the north and west. To a large extent, this northward to northwestward migration over time was a simple artifact of the progressive uplift of the southeast corner of the Columbia Plateau during the volcanism (Hooper and Camp, 1981; Hooper, 1984a). Magma erupted from the same fissures, but at sites progressively farther north as the points of lowest intersection of the fissures with the ground surface migrated in that direction. We do not know, however, that tilting was the only factor involved.

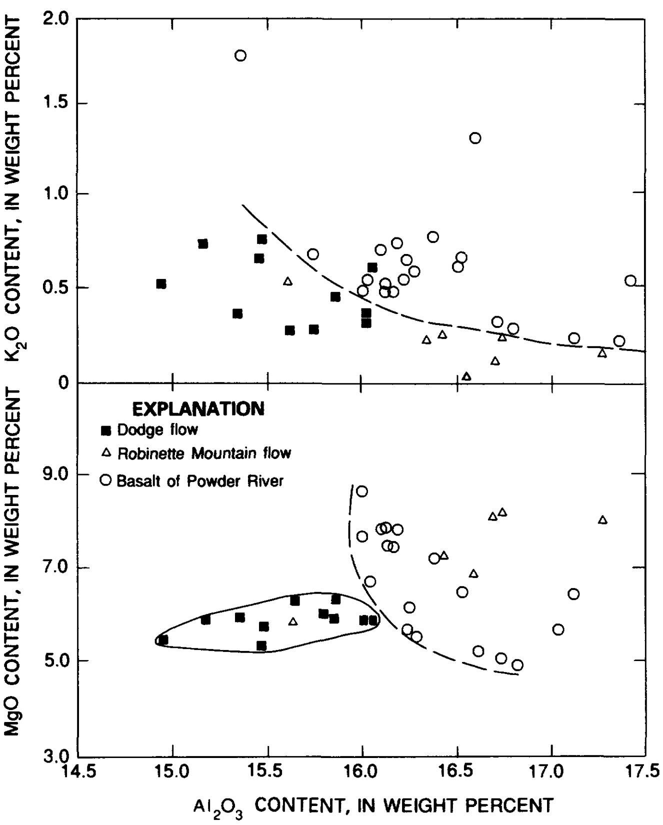


FIGURE 4.20.—Al_2O_3 against K_2O and MgO contents, showing chemical differences between the Dodge and Robinette Mountain flows of the Eckler Mountain Member (of the Wanapum Basalt) and the basalt of Powder River and associated volcanic rocks. Dashed line separates the basalt of Powder River and associated volcanic rocks from all but one of the analyses of the Dodge and Robinette Mountain flows (K_2O versus Al_2O_3 content) and separates the Powder River and associated volcanic rocks from the Dodge flow (MgO versus Al_2O_3 content). Solid line encloses analyses of the Dodge flow (MgO versus Al_2O_3 content).

Feeder dikes for the Grande Ronde Basalt are abundant both in the Cornucopia area and in the lower Grande Ronde River canyon (fig. 4.4). In contrast, the rarity of dikes of the Wanapum and Saddle Mountains Basalts south of the Wallowa-Seven Devils divide leaves open the possible northward migration of the fissures and the magma source over time. Whether or not the source and fissures did migrate, the crustal rocks through which most of the magma had to pass on its way from the reservoir to the surface were similar, consisting of accreted island-arc and oceanic terranes.

The northwestward tilting of the eastern part of the Blue Mountains province continued throughout Columbia River Basalt Group time. The uplift and related tilting are most simply seen as consequences of the continued rise of the Idaho batholith and, to a lesser degree, of the smaller granitic bodies underlying the Seven Devils and Wallowa Mountains and the Nez Perce Plateau (fig. 4.4). The rise of the batholith is documented throughout the Tertiary (Axelrod, 1968) and probably continues to the present day; it seems to have little association with the origin of the basalt.

The regional stress regime (north-northwest horizontal compression, east-northeast horizontal extension) was superimposed on the tilting. A similar stress pattern may have been present as early as the Eocene (Hooper and Camp, 1981) and is still present today (Rohay and Davis, 1983). The regime provides an adequate explanation for most deformation observed on the Columbia Plateau if we take account of significant local control by the pre-Miocene structural grain.

Hooper and Camp (1981) list the main structural elements of the plateau. They are the north-northwest/south-southeast extensional fissures that acted as feeders to the basaltic eruptions, approximately east-west buckles accompanied by reverse faulting, and approximately northwest-southeast right-lateral and northeast-southwest left-lateral strike-slip faults. At a depth corresponding to the base of the crust, potential extensional fissures would result if the vertical stress were large enough to equal approximately the maximum north-northwest horizontal stress, with stress along the east-northeast/west-southwest horizontal axis being significantly less than the other two. Given large magma reservoirs lying near the base of the crust, the magma could use such a potential plane of extension to propagate north-northwest/south-southeast fissures as it forced its way to the surface.

At higher levels in the crust, the vertical stress would be less, and so north-northwest (horizontal) stress would exceed vertical stress, which, in turn, would exceed east-northeast (horizontal) stress. This is the classic condition for development of sets of vertical strike-slip faults (west-northwest right lateral, north-northeast left

lateral). Close to the surface the vertical stress decreases to zero, and so north-northwest (horizontal) stress would exceed east-northeast (horizontal) stress, which, in turn, would exceed vertical stress—the conditions under which approximately east-west buckling and reverse faulting might be expected to develop in brittle basalt flows (Price, 1982). An apparent consequence of this model is that the approximately east-west folds of the Columbia Plateau are near-surface phenomena which probably do not continue to great depth.

At least two complications mar this simple model. First, the suture zone around the cratonic margin forms a major discontinuity that has a different structural pattern; second, clockwise rotation for various parts of the plateau (Basham and Larson, 1978; Magill and others, 1981; Martin, 1984) is now well established, notwithstanding recent comments to the contrary (Fox and Beck, 1985).

The east-west segment of the suture zone passes directly beneath the Lewiston structure (Armstrong and others, 1977; Fleck and Criss, 1985; Mohl and Thiessen, in press). This ancient discontinuity may have provided a weakened zone along which the regional stress found relief in the Miocene. Where the suture extends north-south along the west side of the Idaho batholith (fig. 4.1), it underlies an elongate downwarp. This feature has continued to develop through the past 20 m.y. as a topographic low continually inundated with basalt flows whenever they had access (Camp, 1981; Hooper, 1984b). The cause for the continuing subsidence parallel to the suture zone is unclear but may mildly reflect the same dynamic forces that create trenches above subduction zones.

Progressive clockwise rotation in the western part of the Columbia Plateau is well established (Bentley, 1977, 1979; Magill and others, 1982). It may be explained by movement along northwest-trending right-lateral strike-slip faults or megashears (Bentley, 1977). The most obvious rotation associated with the megashears appears to die out in the main part of the Columbia Plateau east of Pasco Basin (Watkins and Baksi, 1974; Choiniere and Swanson, 1979; Hooper and others, 1979). More detailed recent studies, however, indicate that local clockwise rotation has occurred in the anticlinal crests of the Yakima fold belt (Reidel and others, 1984), within the suture zone on the east side of the Blue Mountains province (Hooper, 1984b), and apparently also on blocks within a zone corresponding to the OWL (Martin, 1984).

More work on the size, age, and amount of rotation of the crustal blocks is needed, but the general tectonic implications are clear. The same forces responsible for the northeastward migration and 60° clockwise rotation of the accreted terranes on the west side of the North American craton continued through the Miocene, albeit in a very subdued form. The greater extension on the south side of the OWL, involving right-lateral motion and probable clockwise rotation of small intervening crustal blocks, created grabens rather than simple fissures to the south of the OWL and may have caused significant crustal thinning. The structural geometry of the OWL appears to be similar to that recorded on the south side of the Lewis and Clark fault zone in northern Idaho and Montana by Sheriff and others (1984). The greater extension and possible crustal thinning is associated in space and time with the change in style of volcanic activity from tholeiitic (Columbia River Basalt Group) to calc-alkaline (Strawberry Volcanics and basalts of Powder River and Weiser).

GENESIS OF THE COLUMBIA RIVER BASALT GROUP

There is as yet no single genetic model for the Columbia River Basalt Group, but the detailed field, petrographic, chemical, and isotopic data now available provide constraints on potential models.

PHYSICAL CONSTRAINTS

(1) The remarkably consistent orientation of feeder dikes throughout the approximately 11-m.y. period (17.5–6.0 Ma) of volcanism implies a nearly constant stress field, not local stress fields related to crustal stretching around high-level magma reservoirs.

(2) The great length of many of the fissure-vent systems, 150 to possibly 300 km, suggests that the fissures extend at least to the base of the crust, 30 to 40 km down. From this, we infer that eruptions occurred when potential fissures resulting from the regional stress regime intersected large magma reservoirs, probably near the crust-mantle boundary (Cox, 1980).

(3) Lava erupted from any one of the long fissure systems is remarkably homogeneous. This homogeneity is the basis of our ability to define and map a detailed flow stratigraphy across the whole plateau (Wright and others, 1973; Swanson, Wright, and others, 1979; Hooper, 1981; Reidel, 1983; Mangan and others, 1986).

(4) The voluminous sheetfloods of basalt resulting from individual eruptions imply rapid discharge from the fissures (Shaw and Swanson, 1970).

The large volumes of homogeneous magma require large reservoirs in which the magma is thoroughly mixed before eruption. It is inconceivable that magma from separate reservoirs, or refills of the same reservoir, would be as homogeneous as that recorded from individual eruptions of the Columbia River Basalt Group. The

rapid discharge, coupled with the homogeneity of the lava, makes it improbable that significant contamination by the country rock occurred between the reservoir and the surface. This inference is supported by the dearth of xenoliths in dikes except in one or two localities, such as the Wallowa Mountains (W.H. Taubeneck, oral commun., 1985).

CHEMICAL CONSTRAINTS

(1) Variations in isotopic ratios of Sr, Nd, O, and Pb (table 4.2; McDougall, 1976; Carlson and others, 1981; Carlson, 1984; Church, 1985) separate the Columbia River Basalt Group into several subdivisions, in which rocks of each subdivision are derived from a different source or combination of sources (fig. 4.3). Carlson (1984) suggested a combination of four or five such sources, including different mantle sources and different types of crustal contaminants. We note that if the origin of the Columbia River Basalt Group is ascribed to multiple sources, then the failure of simple partial-melting or crystal-fractionation models to fully account for the observed chemical variation cannot, of itself, be used as evidence against participation of these processes in the evolution of the magmas. Pb isotopic data indicate that the Picture Gorge Basalt was probably derived from a typical midoceanic-ridge-basalt (MORB)-type depleted mantle (Carlson and others, 1981). The rather high $\delta^{18}O$ values and evolved Pb isotopic ratios of the Grande Ronde Basalt and younger flows imply a sedimentary component, either as a direct crustal contaminant or as a contaminant of the mantle source. The light REE-enriched patterns in these flow units imply a mantle source less depleted than MORB.

(2) No flow composition could have been in equilibrium with a lherzolitic upper mantle. Even the most primitive flows of the Picture Gorge Basalt, as well as the Robinette Mountain flow and the Asotin Member (the most primitive flows in the Wanapum and Saddle Mountains Basalts, respectively), have Mg′ $(Mg/(Mg+Fe_2))$ values, Ni and Cr contents, and olivine phenocryst compositions too evolved to represent primary magma derived from a normal lherzolitic mantle source. Such characteristics require either that the mantle source was an iron-rich pyroxenite (Wilkinson and Binns, 1977; Helz, 1978; Wright and Helz, 1981; Prestvik and Goles, 1985) or that significant crystal fractionation occurred after removal of the magma from its source.

(3) A restricted range in major-element composition and mineralogy characterizes most continental flood-basalt provinces, but it is more marked in the Columbia River Basalt Group than in either the Karoo or the Deccan (Krishnamurthy and Cox, 1977; Cox, 1980; Beane and others, 1986). Neither primitive picritic nor evolved alkalic or silicic varieties are percent. The total range of SiO_2 content for the Columbia River Basalt Group is 48 to 57 weight percent (table 4.1). (Note that, in contrast, the basalts of Powder River and Weiser have a much wider variation of major elements, more akin to that of calc-alkaline rock suites). Again, if a lherzolitic (as opposed to pyroxenitic) mantle source is assumed, then some poorly understood, density-controlled filter system, permitting extrusion only of basalt with a density lowered to some critical level by crystal fractionation (Stolper and Walker, 1980), must also be invoked. Cox (1980) discussed such a density trap involving a large magma reservoir at or near the crust-mantle boundary.

(4) Within the subdivisions defined by similar isotopic ratios, the flows show wide variations in the abundances of incompatible trace elements and smaller variations in major elements. In some samples, the abundances of major and many trace elements have been adequately modeled by crystal fractionation, in which the removal of plagioclase plays a dominant role (Reidel, 1983; Hooper and others, 1984). However, increasing fractionation is commonly accompanied by a small systematic increase in the ratios between different incompatible trace elements (for example, Ba/P_2O_5; Hooper, 1984a). Such changes are unlikely to result either from simple crystal-fractionation or simple partial-melting processes; however, they can occur when these processes are complicated by reservoir recharge or crustal assimilation in open magma systems. The most obvious situation that could cause this interdependence is an open magma system in which the degree of crustal assimilation is controlled by the heat released by crystal fractionation (DePaolo, 1981; O'Hara and Mathews, 1981). If this model is correct, then the crustal contaminant had a $^{87}Sr/^{86}Sr$ ratio nearly identical to that of the magma, and their mixing caused little change in that value. This constraint suggests that the potential contaminant may be recently formed crust, possibly underplated during recent subduction to the bottom of the crust beneath the Blue Mountains province. However, crystal fractionation accompanied by magma recharge (O'Hara and Mathews, 1981) may be adequate to account for the observed chemical changes without recourse to significant crustal assimilation (P.R. Hooper, unpub. data, 1987).

Conversely, sharp changes in chemical composition involving SiO_2, K_2O, and LIL elements between the major subdivisions of the Columbia River Basalt Group appear to coincide with slight changes in the

ratio $^{87}Sr/^{86}Sr$ (fig. 4.3). These changes could reflect recharge of the reservoir by magma of varying chemical and isotopic composition, which might result from various degrees of contamination of the mantle source by hydrous emanations from the subducted slab.

(5) The high $^{87}Sr/^{86}Sr$ ratio of all flows of the Saddle Mountains Basalt could not have been caused by assimilation of the accreted crustal blocks through which most of the Saddle Mountains Basalt flows were extruded. An enriched subcontinental-mantle source appears to be a more probable cause.

(6) Mixing of smaller batches of magma can be demonstrated locally (Hooper, 1985). Mixing between magma batches of similar density, in combination with crystal fractionation, could result in a scatter of elemental abundances that would camouflage the evidence for either process.

Evidently, the origin and evolution of the various magmas contributing to the Columbia River Basalt Group were complex. The isotopic data require different mantle sources, whereas crystal fractionation in the crust played a major role. Simple mixing of different magmas can be clearly demonstrated in some cases and may have been widespread. Assimilation of crustal material has not been established unambiguously, but neither has it been disproved. We still lack convincing evidence to distinguish between basaltic magma contaminated by crustal assimilation, and magma derived either from a mantle source contaminated by subducted crust or from a mantle source enriched by some metasomatic process.

REFERENCES CITED

Anderson, J.L., 1978, The stratigraphy and structure of the Columbia River Basalt in the Clackamas River drainage: Portland, Oreg., Portland State University, M.S. thesis, 136 p.

Anderson, J.L., Beeson, M.H., Bentley, R.D., Fecht, K.R., Hooper, P.R., Niem, A.R., Reidel, S.P., Swanson, D.A., Tolan, T.L., and Wright, T.L., 1987, Distribution maps of stratigraphic units of the Columbia River Basalt Group, *in* Schuster, Eric, ed., Selected papers on the geology of Washington: Olympia, Washington Division of Geology and Earth Resources Bulletin 77, p. 183–196.

Armstrong, R.L., Taubeneck, W.H., and Hales, P.O., 1977, Rb-Sr and K-Ar geochronometry of Mesozoic granitic rocks and their Sr isotopic composition, Oregon, Washington, and Idaho: Geological Society of America Bulletin, v. 88, no. 3, p. 397–411.

Axelrod, D.I., 1968, Tertiary floras and topographic history of the Snake River basin, Idaho: Geological Society of America Bulletin, v. 79, no. 6, p. 713–733.

Bailey, M.M., 1985, Stratigraphy of the Picture Gorge Basalt (PGB), Columbia River Basalt Group, north-central Oregon [abs.]: Geological Society of America Abstracts with Programs, v. 17, no. 6, p. 339.

———1986, The stratigraphy and petrochemistry of the Picture Gorge Basalt, north-central Oregon: Pullman, Washington State University, M.S. thesis, 150 p.

Barrash, Warren, Bond, J.G., Kauffman, J.D., and Venkatakrishnan, Ramesh, 1980, Geology of the La Grande area, Oregon: Oregon Department of Geology and Mineral Industries Special Paper 6, 47 p.

Basham, W.L., and Larson, E.E., 1978, Paleomagnetic evidence for clockwise rotation in western Idaho, eastern Oregon, and northern Nevada [abs.]: Geological Society of America Abstracts with Programs, v. 10, no. 4, p. 210.

Beane, J.E., Turner, C.A., Hooper, P.R., Subbarao, K.V., and Walsh, J.N., 1986, Stratigraphy, composition, and form of the Deccan Basalts, Western Ghats, India: Bulletin of Volcanology, v. 48, no. 1, p. 61–83.

Beeson, M.H., Fecht, K.R., Reidel, S.P., and Tolan, T.L., 1985, Regional correlations within the Frenchman Springs Member of the Columbia River Basalt Group—New insights into the middle Miocene tectonics of northwestern Oregon: Oregon Geology, v. 47, no. 8, p. 87–96.

Beeson, M.H., Perttu, Rauno, and Perttu, Janice, 1979, The origin of the Miocene basalts of coastal Oregon and Washington—An alternative hypothesis: Oregon Geology, v. 41, no. 10, p. 159–166.

Bentley, R.D., 1977, Stratigraphy of the Yakima basalts and structural evolution of the Yakima ridges in the western Columbia Plateau, field trip 12 *of* Geological excursions in the Pacific Northwest: Seattle, Wash., Geological Society of America Annual Meeting, 1977, p. 339–389.

Bentley, R.D., and Anderson, J.L., 1979, Right-lateral strike-slip faults in the western Columbia Plateau, Washington [abs.]: Eos (American Geophysical Union Transactions), v. 60, no. 46, p. 961.

Bentley, R.D., and Cockerham, R.S., 1973, The stratigraphy of the Picture Gorge Basalt (PGB), north-central Oregon [abs.]: Geological Society of America Abstracts with Programs, v. 5, no. 1, p. 9.

Brooks, H.C., Bowen, R.G., Hull, D.A., and Hammitt, R.W., 1977, Preliminary geologic map of the Sawtooth Ridge quadrangle, Oregon: Portland, Oregon Department of Geology and Mineral Industries, scale 1:24,000.

Brooks, H.C., and Vallier, T.L., 1978, Mesozoic rocks and tectonic evolution of eastern Oregon and western Idaho, *in* Howell, D.G., and McDougall, K.A., eds., Mesozoic paleogeography of the Western United States (Pacific Coast Paleogeography Symposium 2): Los Angeles, Society of Economic Paleontologists and Mineralogists, Pacific Section, p. 133–146.

Camp, V.E., 1976, Petrochemical stratigraphy and structure of the Columbia River Basalt, Lewiston Basin area, Idaho-Washington: Pullman, Washington State University, Ph.D. dissertation, 201 p.

———1981, Geologic studies of the Columbia Plateau—Pt. II, Upper Miocene basalt distribution reflecting the source locations, tectonism, and drainage history in the Clearwater embayment, Idaho: Geological Society of America Bulletin, v. 92, pt. 1, no. 9, p. 669–678.

Camp, V.E., and Hooper, P.R., 1981, Geologic studies of the Columbia Plateau—Pt. I, Late Cenozoic evolution of the southeast part of the Columbia River Basalt province: Geological Society of America Bulletin, v. 92, pt. 1, no. 9, p. 659–668.

Camp, V.E., Hooper, P.R., Swanson, D.A., and Wright, T.L., 1984, Columbia River Basalt in Idaho—Physical and chemical characteristics, flow distribution, and tectonic implications, *in* Bonnichsen, Bill, and Breckenridge, R.M., eds., Cenozoic geology of Idaho: Idaho Bureau of Mines and Geology Bulletin 26, p. 55–75.

Carey, S.W., 1958, The tectonic approach to continental drift, *in* Carey, S.W., ed., Continental drift—A symposium: Hobart, Australia, University of Tasmania, p. 117–358.

Carlson, R.W., 1984, Isotopic constraints on Columbia River flood basalt genesis and the nature of the subcontinental mantle: Geochimica et Cosmochimica Acta, v. 48, no. 11, p. 2357–2372.

Carlson, R.W., Lugmair, G.W., and Macdougall, J.D., 1981, "Columbia River volcanism—The question of mantle heterogeneity or crustal contamination" (reply to a comment by D.J. DePaolo, Geochimica

et Cosmochimica Acta, v. 45, no. 12, p. 2483–2499): Geochimica et Cosmochimica Acta, v. 47, no. 4, p. 845–846.

Choiniere, S.R., and Swanson, D.A., 1979, Magnetostratigraphy and correlation of Miocene basalts of the northern Oregon coast and Columbia Plateau, southeast Washington: American Journal of Science, v. 279, no. 7, p. 755–777.

Church, S.E., 1985, Genetic interpretation of lead isotopic data from the Columbia River Basalt Group, Oregon, Washington, and Idaho: Geological Society of America Bulletin, v. 96, no. 5, p. 676–690.

Cox, K.G., 1980, A model for flood basalt volcanism: Journal of Petrology, v. 21, no. 4, p. 629–650.

DePaolo, D.J., 1981, Trace-element and isotopic effects of combined wallrock assimilation and fractional crystallization: Earth and Planetary Science Letters, v. 53, no. 2, p. 189–202.

Dickinson, W.R., and Thayer, T.P., 1978, Paleogeographic and paleotectonic implications of Mesozoic stratigraphy and structure in the John Day inlier of central Oregon, *in* Howell, D.G., and McDougall, K.A., eds., Mesozoic paleogeography of the Western United States (Pacific Coast Paleogeography Symposium 2): Los Angeles, Society of Economic Paleontologists and Mineralogists, Pacific Section, p. 147–161.

Fitzgerald, J.F., 1984, Geology and basalt stratigraphy of the Weiser embayment, west-central Idaho, *in* Bonnichsen, Bill, and Breckenridge, R.M., eds., Cenozoic geology of Idaho: Idaho Bureau of Mines and Geology Bulletin 26, p. 103–128.

Fleck, R.J., and Criss, R.E., 1985, Strontium and oxygen isotopic variations in Mesozoic and Tertiary plutons of central Idaho: Contributions to Mineralogy and Petrology, v. 90, no. 2–3, p. 291–308.

Fox, K.F., and Beck, M.E., Jr., 1985, Paleomagnetic results for Eocene volcanic rocks from northeastern Washington and the Tertiary tectonics of the Pacific Northwest: Tectonics, v. 4, no. 3, p. 323–341.

Fruchter, J.S., and Baldwin, S.F., 1975, Correlations between dikes of the Monument Swarm, central Oregon, and Picture Gorge Basalt flows: Geological Society of America Bulletin, v. 86, no. 4, p. 514–516.

Goles, G.G., 1986, Miocene basalts of the Blue Mountains province in Oregon—Pt. I, Compositional types and their geological settings: Journal of Petrology, v. 27, no. 2, p. 495–520.

Goles, G.G., Brandon, A.D., and Obermiller, Walter, 1985, A plume component in Columbia River basalts? [abs.]: Eos (American Geophysical Union Transactions), v. 66, no. 46, p. 1109.

Hamilton, Warren, 1962, Late Cenozoic structure of west-central Idaho: Geological Society of America Bulletin, v. 73, no. 4, p. 511–515.

Hart, W.K., Aronson, J.L., and Mertzman, S.A., 1984, Areal distribution and age of low-K, high-alumina olivine tholeiite magmatism in the northwestern Great Basin: Geological Society of America Bulletin, v. 95, no. 2, p. 186–195.

Helz, R.T., 1978, The petrogenesis of the Ice Harbor Member, Columbia Plateau, Washington—A chemical and experimental study: University Park, Pennsylvania State University, Ph.D. dissertation, 284 p.

Hooper, P.R., 1981, The role of magnetic polarity and chemical analyses in establishing the stratigraphy, tectonic evolution, and petrogenesis of the Columbia River Basalt, *in* Subbarao, K.V., and Sukheswala, R.N., eds., Deccan volcanism and related basalt provinces in other parts of the world: Geological Society of India Memoir 3, p. 362–376.

———1982, The Columbia River basalts: Science, v. 215, no. 4539, p. 1463–1468.

———1984a, Physical and chemical constraints on the evolution of the Columbia River Basalt: Geology, v. 12, no. 8, p. 495–499.

———1984b, Structural model for the Columbia River Basalt near Riggins, Idaho, *in* Bonnichsen, Bill, and Breckenridge, R.M., eds., Cenozoic geology of Idaho: Idaho Bureau of Mines and Geology Bulletin 26, p. 129–136.

———1985, A case of simple magma mixing in the Columbia River Basalt Group—The Wilbur Creek, Lapwai, and Asotin flows, Saddle Mountains Formation: Contributions to Mineralogy and Petrology, v. 91, no. 1, p. 66–73.

Hooper, P.R., and Camp, V.E., 1981, Deformation of the southeast part of the Columbia Plateau: Geology, v. 9, no. 7, p. 323–328.

Hooper, P.R., Kleck, W.D., Knowles, C.R., Reidel, S.P., and Thiessen, R.L., 1984, Imnaha Basalt, Columbia River Basalt Group: Journal of Petrology, v. 25, no. 2, p. 473–500.

Hooper, P.R., Knowles, C.R., and Watkins, N.D., 1979, Magnetostratigraphy of the Imnaha and Grande Ronde Basalts in the southeast part of the Columbia Plateau: American Journal of Science, v. 279, no. 7, p. 737–754.

Hooper, P.R., and Swanson, D.A., 1987, Evolution of the eastern part of the Columbia Plateau, *in* Schuster, Eric, ed., Selected papers on the geology of Washington: Olympia, Washington Division of Geology and Earth Resources Bulletin 77, p. 197–218.

Hooper, P.R., Webster, G.D., and Camp, V.E., 1985, Geologic map of the Clarkston, Silcott, Asotin, and Rockpile 7.5-minute quadrangles, Washington and Idaho: Washington Division of Geology and Earth Resources, Map GM-31, scale 1:62,500.

Kienle, C.F., Newcomb, R.C., Deacon, R.J., Farooqui, S.M., Bentley, R.D., Anderson, J.L., and Thoms, R.E., 1979, Western Columbia Plateau tectonics structures and their age of deformation [abs.]: Tectonics and Seismicity of the Columbia Plateau Workshop (Rockwell Hanford Operations, Seattle, Wash., Feb., 1977).

Kleck, W.D., 1976, Chemistry, petrography, and stratigraphy of the Columbia River Group in the Imnaha River Valley region, eastern Oregon and western Idaho: Pullman, Washington State University, Ph.D. dissertation, 215 p.

Krishnamurthy, P., and Cox, K.G., 1977, Picrite basalts and related lavas from the Deccan Traps of Western India: Contributions to Mineralogy and Petrology, v. 62, no. 1, p. 53–75.

Kuhns, M.J.P., 1980, Late Cenozoic deposits of the lower Clearwater Valley, Idaho and Washington: Pullman, Washington State University, M.S. thesis, 71 p.

Lindsley, D.H., Smith, Douglas, and Haggerty, S.E., 1971, Petrography and mineral chemistry of a differentiated flow of Picture Gorge Basalt near Spray, Oregon: Carnegie Institution of Washington Year Book 69, p. 264–285.

Long, P.E., and Duncan, R.A., 1982, $^{40}Ar/^{39}Ar$ ages of Columbia River Basalt from deep boreholes in south-central Washington: Rich land, Wash., Rockwell Hanford Operations Report RHO-BW-SA-233P, 11 p.

Magill, J.R., Cox, A.V., and Duncan, Robert, 1981, Tillamook Volcanic Series—Further evidence for tectonic rotation of the Oregon Coast Range: Journal of Geophysical Research, v. 86, no. B4, p. 2953–2970.

Magill, J.R., Wells, R.E., Simpson, R.W., and Cox, A.V., 1982, Post-12-m.y. rotation of southwest Washington: Journal of Geophysical Research, v. 87, no. B5, p. 3761–3776.

Mangan, M.T., Wright, T.L., Swanson, D.A., and Byerly, G.R., 1986, Regional Correlation of Grande Ronde Basalt flows, Columbia River Basalt Group, Washington, Oregon and Idaho: Geological Society of America Bulletin 97, p. 1300–1318.

Martin, B.S., 1984, Paleomagnetism of basalts in northeastern Oregon and west-central Idaho: Pullman, Washington State University, M.S. thesis, 151 p.

McDougall, Ian, 1976, Geochemistry and origin of basalt of the Columbia River Group, Oregon and Washington: Geological Society of America Bulletin, v. 87, no. 5, p. 777–792.

McKee, E.H., Hooper, P.R., and Kleck, W.D., 1981, Age of Imnaha Basalt—Oldest basalt flows of the Columbia River Basalt Group, northwest United States: Isochron/West, no. 31, p. 31–33.

McKee, E.H., Swanson, D.A., and Wright, T.L., 1977, Duration and volume of Columbia River Basalt volcanism, Washington, Oregon, and Idaho [abs.]: Geological Society of America Abstracts with Programs, v. 9, no. 4, p. 463–464.

Mohl, G.B., and Thiessen, R.L., in press, Gravity studies of an island arc-continental suture in west-central Idaho and adjacent Washington, *in* Vallier, T.L., and Brooks, H.C., eds., Geology of the Blue Mountains region of Oregon, Idaho, and Washington—Evolution, accretion, and resources of the Blue Mountains island arc: U.S. Geological Survey Professional Paper 1438.

Nathan, Simon, and Fruchter, J.S., 1974, Geochemical and paleomagnetic stratigraphy of the Picture Gorge and Yakima Basalts (Columbia River Group) in central Oregon: Geological Society of America Bulletin, v. 85, no. 1, p. 63–76.

Nelson, D.O., 1983, Implications of oxygen isotope data and trace-element modeling for a large-scale mixing model for the Columbia River Basalt: Geology, v. 11, no. 4, p. 248–251.

O'Hara, M.J., and Mathews, R.E., 1981, Geochemical evolution in an advancing, periodically replenished, periodically tapped, continuously fractionated magma chamber: Geological Society of London Journal, v. 138, pt. 3, p. 237–277.

Osawa, Masumi, and Goles, G.G., 1970, Trace-element abundances in Columbia River basalts, *in* Gilmour, E.H., and Stradling, Dale, eds., Columbia River Basalt Symposium 2, Wash. [Procs.]: Cheney, Eastern Washington State College Press, p. 55–71.

Patterson, P.V., 1969, Sawtooth Ridge—A northeast Oregon volcanic crater: Ore Bin, v. 31, no. 9, p. 173–183.

Pearce, J.A., and Cann, J.R., 1973, Tectonic setting of basic volcanic rocks determined using trace-element analyses: Earth and Planetary Science Letters, v. 19, no. 2, p. 290–300.

Pearce, T.H., Gorman, B.E., and Birkett, T.C., 1977, The relationship between major-element chemistry and tectonic environment of basic and intermediate volcanic rocks: Earth and Planetary Science Letters, v. 36, no. 1, p. 121–132.

Prestvik, Tore, and Goles, G.G., 1985, Comments on petrogeneses and the tectonic setting of Columbia River basalts: Earth and Planetary Science Letters, v. 72, no. 1, p. 65–73.

Price, E.H., 1982, Structural geometry, strain distribution, and mechanical evolution of eastern Umtanum Ridge and a comparison with other selected localities within Yakima fold structures, south-central Washington: Pullman, Washington State University, Ph.D. dissertation, p. 196.

Reidel, S.P., 1983, Stratigraphy and petrogenesis of the Grande Ronde Basalt from the deep canyon country of Washington, Oregon, and Idaho: Geological Society of America Bulletin, v. 94, no. 4, p. 519–542.

———1984, The Saddle Mountains—The evolution of an anticline in the Yakima fold belt: American Journal of Science, v. 284, no. 8, p. 942–978.

Reidel, S.P., Long, P.E., Myers, C.W., Mase, J., 1982, New evidence for greater than 3.2 km of Columbia River Basalt beneath the central Columbia Plateau [abs.]: Eos (American Geophysical Union Transactions), v. 63, no. 8, p. 173.

Reidel, S.P., Scott, G.R., Bazard, D.R., Cross, R.W., and Dick, Brian, 1984, Post-12-million year clockwise rotation in the central Columbia Plateau, Washington: Tectonics, v. 3, no. 2, p. 251–273.

Riddihough, R.P., 1985, The Roseburg-Pendleton (ROPE) lineament, Oregon [abs.]: Eos (American Geophysical Union Transactions), v. 66, no. 3, p. 24.

Robyn, T.L., 1977, Geology and petrology of the Strawberry Volcanics, northeast Oregon: Eugene, University of Oregon, Ph.D. dissertation, 197 p.

———1979, Miocene volcanism in eastern Oregon—An example of calc-alkaline volcanism unrelated to subduction: Journal of Volcanology and Geothermal Research, v. 5, no. 1–2, p.149–161.

Rohay, A.C., and Davis, J.D., 1983, Contemporary deformation in the Pasco Basin area of the central Columbia Plateau, *in* Caggiano, J.A., and Duncan, D.W., eds., Preliminary interpretation of the tectonic stability of the reference repository location, Cold Creek syncline, Pasco Basin, Richland, Washington: Richland, Wash., Rockwell Hanford Operations Report RHO–BW–ST–19P, p. 6-1 – 6-30.

Ross, M.E., 1978, Stratigraphy, structure, and petrology of Columbia River Basalt in a portion of the Grande Ronde-Blue Mountains area of Oregon and Washington: Moscow, University of Idaho, Ph.D. dissertation, 407 p.

Shaw, H.R., and Swanson, D.A., 1970, Eruption and flow rates of flood basalts, *in* Gilmour, E.H., and Stradling, Dale, eds., Columbia River Basalt Symposium 2, Wash. [Procs]: Cheney, Eastern Washington State College Press, p. 271–300.

Sheriff, S.D., Sears, J.W., and Moore, J.N., 1984, Montana's Lewis and Clark fault zone—An intracratonic transform-fault system [abs.]: Geological Society of America Abstracts with Programs, v. 16, no. 6, p. 653–654.

Shubat, M.A., 1979, Stratigraphy, petrochemistry, petrography, and structural geology of the Columbia River Basalt in the Minam-Wallowa River area, northeast Oregon: Pullman, Washington State University, M.S. thesis, 156 p.

Smith, G.A., and Cushing, P.D., 1985, Preliminary report on revisions to the stratigraphy of the Prineville chemical type, Columbia River Basalt Group [abs.]: Eos (American Geophysical Union Transactions), v. 66, no. 3, p. 25.

Stoffel, K.L., 1984, Geology of the Grande Ronde lignite field, Asotin County, Washington: Washington Division of Geology and Earth Resources Report of Investigations 27, 79 p.

Stolper, E.M., and Walker, David, 1980, Melt density and the average composition of basalt: Contributions to Mineralogy and Petrology, v. 74, p. 7–12.

Swanson, D.A., Anderson, J.L., Camp, V.E., Hooper, P.R., Taubeneck, W.H., and Wright, T.L., 1981, Reconnaissance geologic map of the Columbia River Basalt Group, northern Oregon and western Idaho: U.S. Geological Survey Open-File Report 81–797, 5 sheets, scale 1:250,000.

Swanson, D.A., Bentley, R.D., Byerly, G.R., Gardner, J.N., and Wright, T.L., 1979, Preliminary reconnaissance geologic maps of the Columbia River Basalt Group in parts of eastern Washington and northern Idaho: U.S. Geological Survey Open-File Report 79–534, 9 sheets, scale 1:250,000.

Swanson, D.A., and Wright, T.L., 1981, The regional approach to studying the Columbia River Basalt Group, *in* Subbarao, K.V., and Sukheswala, R.N., eds., Deccan volcanism and related basalt provinces in other parts of the world: Geological Society of India Memoir 3, p. 58–80.

———1983, Geologic map of the Wenaha Tucannon Wilderness, Washington and Oregon: U.S. Geological Survey Miscellaneous Field Studies Map MF–1536, scale 1:48,000.

Swanson, D.A., Wright, T.L., Camp, V.E., Gardner, J.N., Helz, R.T., Price, S.M., Reidel, S.P., and Ross, M.E., 1980, Reconnaissance geologic map of the Columbia River Basalt Group, Pullman and Walla Walla quadrangles, southeast Washington and adjacent Idaho: U.S. Geological Survey Miscellaneous Investigations Series Map I–1139, scale 1:250,000.

Swanson, D.A., Wright, T.L., and Helz, R.T., 1975, Linear vent systems and estimated rates of magma production and eruption for the Yakima Basalt on the Columbia Plateau: American Journal of Science, v. 275, no. 8, p. 877–905.

Swanson, D.A., Wright, T.L., Hooper, P.R., and Bentley, R.D., 1979,

Revisions in stratigraphic nomenclature of the Columbia River Basalt Group: U.S. Geological Survey Bulletin 1457–G, 59 p.

Taubeneck, W.H., 1970, Dikes of Columbia River Basalt in northeastern Oregon, western Idaho, and southeastern Washington, *in* Gilmour, E.H., and Stradling, Dale, eds., Columbia River Basalt Symposium 2, Wash. [Procs]: Cheney, Eastern Washington State College Press, p. 73–96.

Thompson, Michael, and Walsh, J.N., 1983, A handbook of inductively coupled plasma spectrometry: Glasgow, Blackie and Son, 273 p.

Uppuluri, V.R., 1974, Prineville chemical type—A new basalt type in the Columbia River Group: Geological Society of America Bulletin, v. 85, no. 8, p. 1315–1318.

Vallier, T.L., 1985, Petrologic implications for the tectonic evolution of the Blue Mountains island arc, eastern Oregon and western Idaho [abs.]: Geological Society of America Abstracts with Programs, v. 17, no. 4, p. 269.

Walker, G.W., 1973, Contrasting compositions of the youngest Columbia River Basalt flows in Union and Wallowa Counties, northeastern Oregon: Geological Society of America Bulletin, v. 84, no. 2, p. 425–429.

———1977, Geologic map of Oregon east of the 121st meridian: U.S. Geological Survey Miscellaneous Investigations Series Map I–902, scale 1:500,000.

———1979, Reconnaissance geologic map of the Oregon part of the Grangeville quadrangle, Baker, Union, Umatilla, and Wallowa Counties, Oregon: U.S. Geological Survey Miscellaneous Investigations Series Map I–1116, scale 1:250,000.

Waters, A.C., 1961, Stratigraphic and lithologic variations in the Columbia River Basalt: American Journal of Science, v. 259, no. 8, p. 583–611.

Watkins, N.D., and Baksi, A.K., 1974, Magnetostratigraphy and oroclinal folding of the Columbia River, Steens, and Owyhee Basalts in Oregon, Washington, and Idaho: American Journal of Science, v. 274, no. 2, p. 148–189.

Webster, G.D., Kuhns, M.J.P., and Waggoner, G.L., 1984, Late Cenozoic gravels in Hells Canyon and the Lewiston Basin, Washington and Idaho, *in* Bonnichsen, Bill, and Breckenridge, R.M., eds., Cenozoic geology of Idaho: Idaho Bureau of Mines and Geology Bulletin 26, p. 669–684.

Wilkinson, J.F.G., and Binns, R.A., 1977, Relatively iron-rich lherzolite xenoliths of the Cr-diopside suite—A guide to the primary nature of anorogenic tholeiitic andesite magmas: Contributions to Mineralogy and Petrology, v. 65, no. 2, p. 199–212.

Wise, D.U., 1963, An outrageous hypothesis for the tectonic pattern of the North American Cordillera: Geological Society of America Bulletin, v. 74, no. 3, p. 357–362.

Wright, T.L., Black, K.N., Swanson, D.A., and O'Hearn, Tim, 1980, Columbia River Basalt—1978–1979 sample data and chemical analyses: U.S. Geological Survey Open-File Report 80–921, 99 p.

———1982, Columbia River Basalt—1980–1981 sample data and chemical analyses: U.S. Geological Survey Open-File Report 82–532, 51 p.

Wright, T.L., Grolier, M.J., and Swanson, D.A., 1973, Chemical variation related to the stratigraphy of the Columbia River Basalt: Geological Society of America Bulletin, v. 84, no. 2, p. 371–385.

Wright, T.L., and Helz, R.T., 1981, Generation of tholeiitic basalt—A comparative study of three intraplate provinces, *in* Subbarao, K.V., and Sukheswala, R.N., eds., Deccan volcanism and related basalt provinces in other parts of the world: Geological Society of India Memoir 3, p. 419–421.

Wright, T.L., Swanson, D.A., Helz, R.T., and Byerly, G.R., 1979, Major-oxide, trace-element, and glass chemistry of Columbia River Basalt samples collected between 1971 and 1977: U.S. Geological Survey Open-File Report 79–711, 146 p.

5. MIOCENE AND YOUNGER ROCKS OF THE BLUE MOUNTAINS REGION, EXCLUSIVE OF THE COLUMBIA RIVER BASALT GROUP AND ASSOCIATED MAFIC LAVA FLOWS

By GEORGE W. WALKER

CONTENTS

ABSTRACT

Diverse volcanic, volcaniclastic, and sedimentary rocks characterize the Miocene section of the Blue Mountains region; vast volumes of basalt dominate parts of the region, and more restricted andesitic piles are present locally. The basalt was erupted mostly from several widespread dike complexes, and the andesite from central vents. Elsewhere in the region, moderate to large volumes of silicic pyroclastic rocks are present, partly erupted as ash flows and pumiceous ash from recognized vents within the Blue Mountains and from vents in areas to the west, along the Cascade axis, and to the south in the northern Basin and Range. Much of the silicic pyroclastic material, locally intermixed with much palagonitic tuff and breccia, was deposited in several separate and partly ephemeral basins. Some of the widespread Miocene ash-flow tuff layers extend into several adjacent basins, whereas most sedimentary units are geographically restricted.

INTRODUCTION

During the Miocene, while vast volumes of tholeiitic basalt of the Columbia River Basalt Group were being erupted from numerous dike complexes in the northern Blue Mountains, another kind of volcanism was occurring immediately to the south and southeast both within and marginal to the Blue Mountains province. This volcanism, of calc-alkaline affinity, included moderately large volumes of basalt, andesite, and rhyodacite or rhyolite; chemically, it was much like the volcanism that occurred in the Blue Mountains region during Eocene time. It was accompanied by the deposition of large volumes of silicic pyroclastic rocks and silicic to mafic tuffaceous sedimentary rocks in several separate basins; some of these basins were short lived, whereas others persisted into the Pliocene. Some of the persistent basins are graben-like structures resulting from Miocene and younger extensional tectonics that mostly affected areas to the south in the Basin and Range province. Other depositional basins, particularly those along the northern and western margins of the Blue Mountains, are synclinal downwarps in flow sequences of the Columbia River Basalt Group. In most of these basins, the sedimentary fill appears conformable, or nearly so, with underlying Miocene basalts; but in some places along the southern margin of the Blue Mountains, volcaniclastic and sedimentary units are interlayered with flows of the Columbia River Basalt Group or with other middle Miocene flow sequences.

Eruptive centers for this diverse volcanic activity are of several types, including large central-vent complexes, large silicic domal masses, and areally restricted plugs and dikes. Some of the widespread ash-flow tuff sheets of the southern Blue Mountains were erupted from presumed caldera complexes now buried in Harney Basin approximately at the boundary between the Blue Mountains and the Basin and Range province; other tuff deposits may be related to suspected, but as yet largely unsubstantiated, source areas characterized by silicic domal complexes or large silicic intrusions; and still other deposits were erupted from unknown sources, possibly now totally obscured either by the rhyolitic domal complexes or by younger flows and sedimentary rocks.

The large volume of basaltic to rhyolitic eruptive material of Miocene and younger age, exclusive of flows

of the Columbia River Basalt Group, includes both flows and tuffaceous fragmental materials. These sequences, which include large volumes of fine-grained rhyolitic, andesitic, and basaltic pyroclastic ejecta, are partly found in separate depositional basins; volume estimates based on these disconnected sequences are approximate at best and may even be misleading in making comparisons with different types of volcanic activity in adjoining areas. If these isolated sequences of Miocene tuff and sedimentary rocks are erosional remnants of a widespread and continuous sheet, as postulated by Thayer and Brown (1966) for a major part of the section, the volume of rhyolitic and andesitic material erupted would be very large, possibly nearly comparable to that of temporally equivalent flows of the Columbia River Basalt Group. The data indicate, however, that many of these basins were never or only briefly interconnected and that some evolved during deposition; thus, the volume would be greatly diminished. Onlap and offlap relations and the distribution of nearshore and offshore facies in many parts of these basins indicate that at least some of them were separate and resulted from surface deformation related to concurrent tectonism and volcanism.

GEOGRAPHIC DISTRIBUTION

Areas of deposition of the Miocene and younger volcanic and tuffaceous sedimentary rocks are widely distributed, the thicker sections mostly being in basins on the west, south, and southeast margins of the Blue Mountains province and in the Strawberry Mountains (fig. 5.1). Of the many separate basins that have been recognized, Deschutes (or Madras) Basin and the smaller Tygh and The Dalles Basins dominate on the west. On the south, parts of several separate depositional basins extend northward into the Blue Mountains province; these basins include the northern parts of Harney, Juntura, and Harper Basins and the northern part of Bully Creek basin (fig. 5.1). Entirely within the Blue Mountains province near its southern margin are Paulina Basin and Bear and Logan Valleys. All three of these basins contain upper Miocene ash-flow tuff sheets, several of which were erupted from buried calderas in Harney Basin, and interstratified tuffaceous sedimentary rocks. Both John Day Valley and Fox Basin, well within the Blue Mountains province, contain tuff and tuffaceous sedimentary rocks, as well as interlayered fanglomerate, of middle and late Miocene age.

Extensive andesitic to rhyolitic volcanic rocks of middle and early late Miocene age crop out in a large area (more than 4,000 km^2) in and adjacent to the Strawberry Mountains. These volcanic rocks extend tens of kilometers southward from the Strawberry Mountains to the south limits of the Blue Mountains province, eastward into Mormon and Unity Basins, and northward into the headwaters of the Burnt River. On the southeastern and eastern margins of the Blue Mountains province and extending into the north end of the Owyhee Upland province (see fig. 1.2), tuffaceous sedimentary rocks representing northward extensions of units exposed in Boise Basin are present northwest and west of Ontario, Oreg., and Weiser, Idaho (fig. 5.1). Miocene and younger ash-flow tuff, tuffaceous sedimentary rocks, and some interbedded basalt flows are confined to Mormon, Durkee, and Unity Basins, and somewhat similar units are present in several valleys, including Baker Valley and Burnt River and lower Powder River valleys. Structurally depressed areas to the north and northeast of the Blue Mountains axis, including Arlington and Agency Basins, as well as the structurally downdropped block of the La Grande Valley, also contain upper Miocene and younger epiclastic and volcaniclastic deposits, although ash-flow tuff and young basalt flows such as are typical of some of the other basins appear to be absent. Much of the sedimentary fill in these basins appears to be fanglomerate derived from the erosion of volcanic rocks exposed in more elevated parts of the Blue Mountains.

Pasco and Walla Walla Basins lie north of the structurally elevated, northwest-trending block, represented by the Horse Heaven Hills in Washington, that extends southeastward into Oregon (fig. 5.1). Pasco Basin, on the northwest, contains a thick section of terrestrial pyroclastic and sedimentary deposits of Miocene and younger age, and Walla Walla Basin to the southeast contains a thickness of about 180 m of Pliocene(?) and Pleistocene beds of clay, silt, sand, and gravel. Both basins are developed in folded and faulted flows of the Columbia River Basalt Group.

STRATIGRAPHY AND AGE

The stratigraphic relations of the Miocene and younger rocks of the Blue Mountains region, exclusive of the Columbia River Basalt Group, are at present too poorly known to interpret with confidence. Although the volcanic and sedimentary sequences are approximate age equivalents of major parts of the group, they are present either in disconnected depositional basins or in large volcanic piles or domal complexes, and few regionally extensive marker horizons have been recognized. The absence of such marker horizons in most sections makes meaningful regional correlations difficult, although recent radiometric dating has provided greater precision in establishing the temporal equivalence of separate sequences. In the past, correlation of units has relied mostly on the dating of scattered floral and faunal collections, with little consideration to whether the units in which these collections were made were interconnected. Thus, a single

formation name has in places been applied to physically separated Miocene sequences, and in a few places several different formation names have been applied to a single widespread tuff layer that transcends divides between separate depositional basins.

To clarify some of the problems of stratigraphic nomenclature, the Miocene and younger sedimentary sequences are here discussed first by depositional basin or volcanic pile and then by age. Radiometric ages, though few and limited in geographic coverage, permit first approximations in regional correlation of the numerous formation units that have been established and, furthermore, permit age comparisons with major units within the Columbia River Basalt Group (fig. 5.2). The stratigraphic nomenclature of each depositional basin has evolved through independent mapping, and only in a few places do individual units of distinctive ash-flow tuff or related air-fall tuff provide physical ties to deposits of other basins.

BASIN-FILL DEPOSITS

The Miocene and younger basin-fill deposits consist of a great variety of rhyolitic to basaltic volcanic and volcaniclastic rocks, complexly interlayered with tuffaceous sedimentary rocks. These deposits occur in several separate depositional basins, some of which lie wholly within the Blue Mountains region; others, along the margins of the Blue Mountains, extend long distances outside the region.

DESCHUTES BASIN

Deschutes (or Madras) Basin, which is marginal to the Blue Mountains on the west, contains a thick sequence of epiclastic and pyroclastic rocks and a few thin interlayered basalt flows, all incorporated into the Deschutes Formation of Farooqui, Beaulieu, and others (1981) or the Madras Formation (Williams, 1957). The basin fill varies considerably in thickness, locally pinching out on Eocene and Oligocene rocks at both the northern and eastern margins of the basin and apparently resting unconformably on the Grande Ronde Basalt of the Columbia River Basalt Group in the canyon of the Deschutes River about 7 to 8 km northwest of Madras, Oreg. (Waters, 1968). The maximum thickness indicated for this diverse assemblage is about 450 m for a section in the canyon of the Deschutes River south of Warm Springs, Oreg. (Hodge, 1940). Much of the section consists of poorly to moderately well bedded, fine- to coarse-grained clastic sedimentary rocks composed entirely of basaltic, andesitic, and rhyodacite debris (figs. 5.3, 5.4) derived from volcanic rocks to the south and west. Interstratified in the sedimentary sequence are rare basalt and basaltic andesite flows and several partly bedded, pumice-rich layers and elongate lenses of rhyolitic to dacitic ash-flow tuff and laharic tuff. The lithology, bedding, and textural features of some of these pumiceous layers and elongate lenses (fig. 5.5) indicate that they were initially pumiceous ash flows that became mudflows upon encountering and incorporating water and, locally, unconsolidated sediment during flow in stream channels.

Several K-Ar ages on mafic flows in the Deschutes Formation range from 15.9±3.0 Ma to slightly less than 5 Ma (Armstrong and others, 1975). The older age becomes 16.3±3.0 Ma when recalculated using the decay constants acceptable in 1983 (Fiebelkorn and others, 1983). More recently, Smith and Snee (1984) obtained an $^{40}Ar/^{39}Ar$ age of 7.6±0.3 Ma for a basaltic unit identified by them as the Pelton Basalt Member of former usage of the Deschutes Formation. This apparently is the same unit for which Armstrong and others (1975) obtained the K-Ar age of 15.9±3.0 Ma. According to Smith and Snee (1984), their Pelton Basalt Member of the Deschutes Formation represents the basal part of the formation and is separated by an angular unconformity from underlying clastic rocks that contain middle Miocene fossils. Although earlier workers included the middle Miocene rocks in the Deschutes Formation, Smith and Snee assigned these fossil-bearing rocks to their Simtustus unit, which overlies and locally interfingers with flows of the Columbia River Basalt Group. This age range for the Deschutes Formation, from either middle or late Miocene to earliest Pliocene, indicates some temporal overlap with the Columbia River Basalt Group, although the youngest part of the formation apparently postdates any of the flows of the group (fig. 5.2).

TYGH BASIN AND THE DALLES BASIN

Tygh Basin and The Dalles Basin, both northwest of the Blue Mountains, are structural downwarps in flows of the Columbia River Basalt Group filled with indurated epiclastic and volcaniclastic rocks of Miocene and Pliocene(?) age. Rocks in both basins have been described in some detail (Newcomb, 1966; Farooqui, Beaulieu, and others, 1981) as debris fans that spread eastward from the Cascade Range and include, in addition to erosional debris, interbedded lenses and layers of tuff, lahars, and minor basalt flows. Various formation names have been applied to the rock sequences in both Tygh Basin and The Dalles Basin; most recently, they have been named, respectively, Tygh Valley Formation and the Chenoweth Formation by Farooqui, Beaulieu, and others (1981). Both formations are included in their regionally extensive Dalles Group. The two formations vary considerably in thickness; maximum thicknesses reported are 550 m

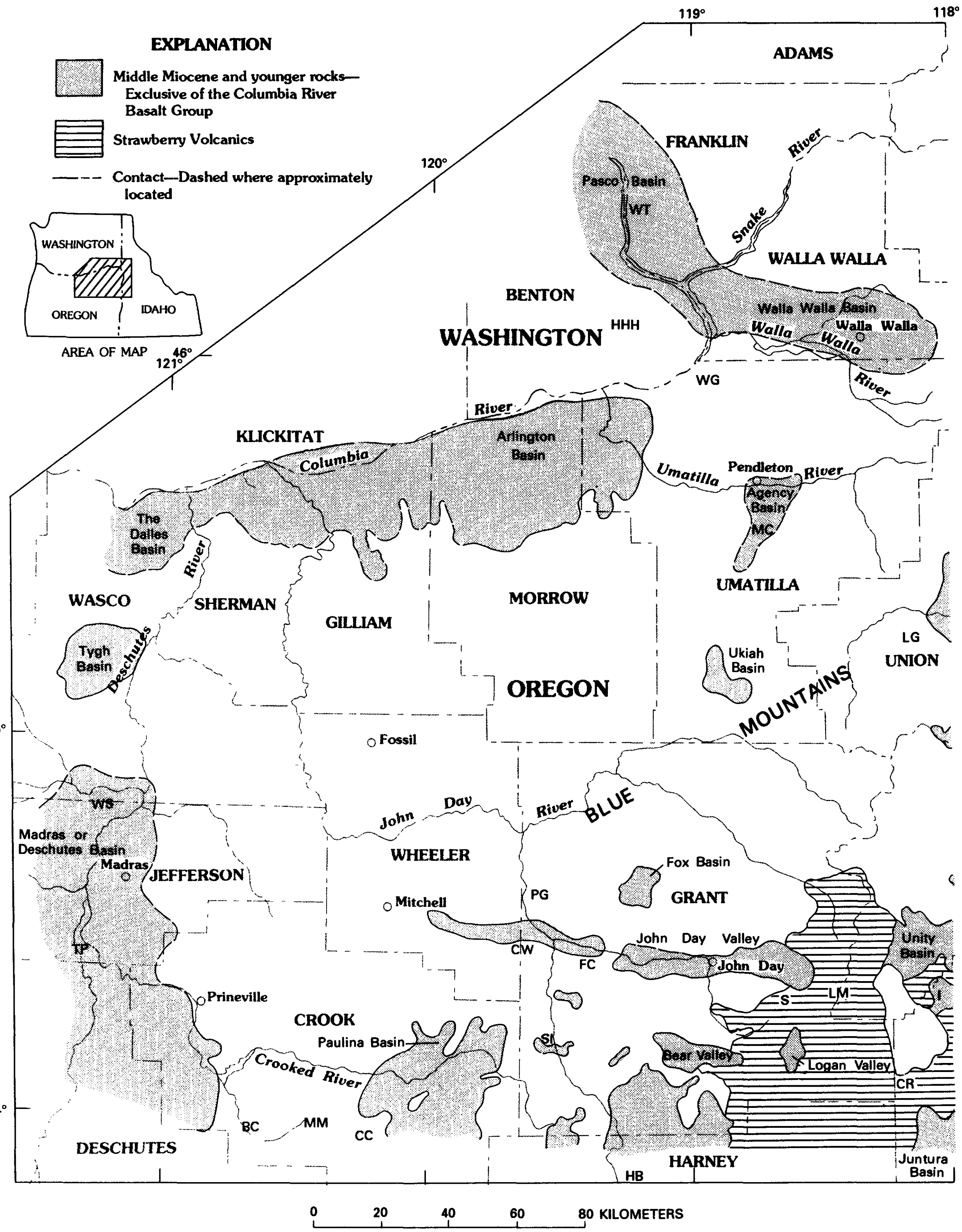
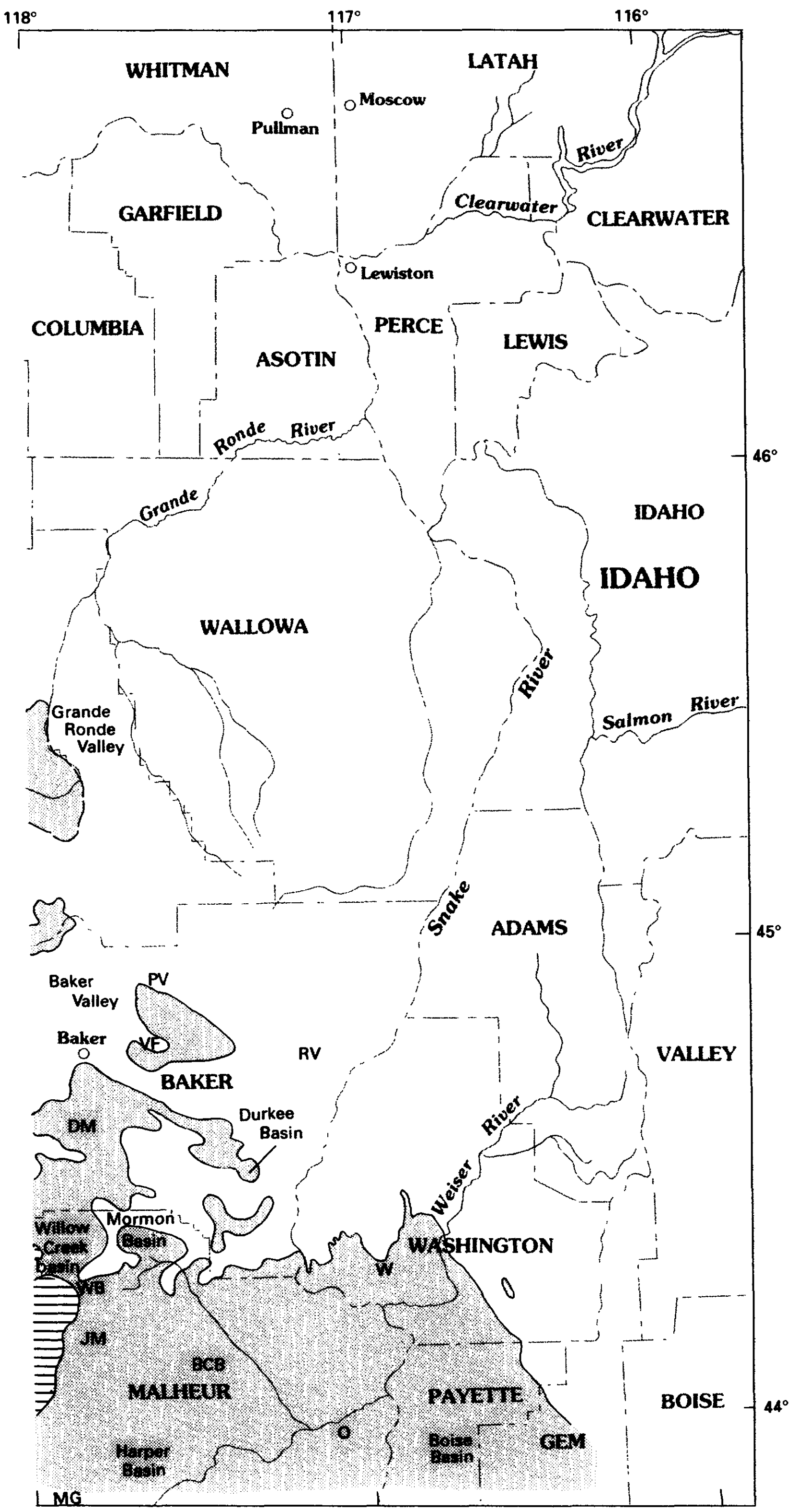


FIGURE 5.1.—Blue Mountains region, showing approximate outcrop area



Index to localities

BC, Bear Creek
BCB, Bully Creek Basin
CC, Camp Creek
CR, Castle Rock
CW, Cottonwood
DM, Dooley Mountain
FC, Flat Creek
HB, Harney Basin
HHH, Horse Heaven Hills
I, Ironside Mountain
JM, Juniper Mountain
LG, La Grande Valley
LM, Lookout Mountain
MC, McKay Creek
MG, Malheur Gorge
MM, Maury Mountains
O, Ontario
PG, Picture Gorge
PV, Powder River valley
RV, Richland Valley
S, Strawberry Mountain
SI, Suplee-Izee area
TP, The Peninsula
VF, Virtue Flat
W, Weiser
WB, Willow Creek basin
WG, Wallula Gap
WS, Warm Springs
WT, White Bluffs

of middle Miocene and younger rocks, exclusive of the Columbia River Basalt Group.

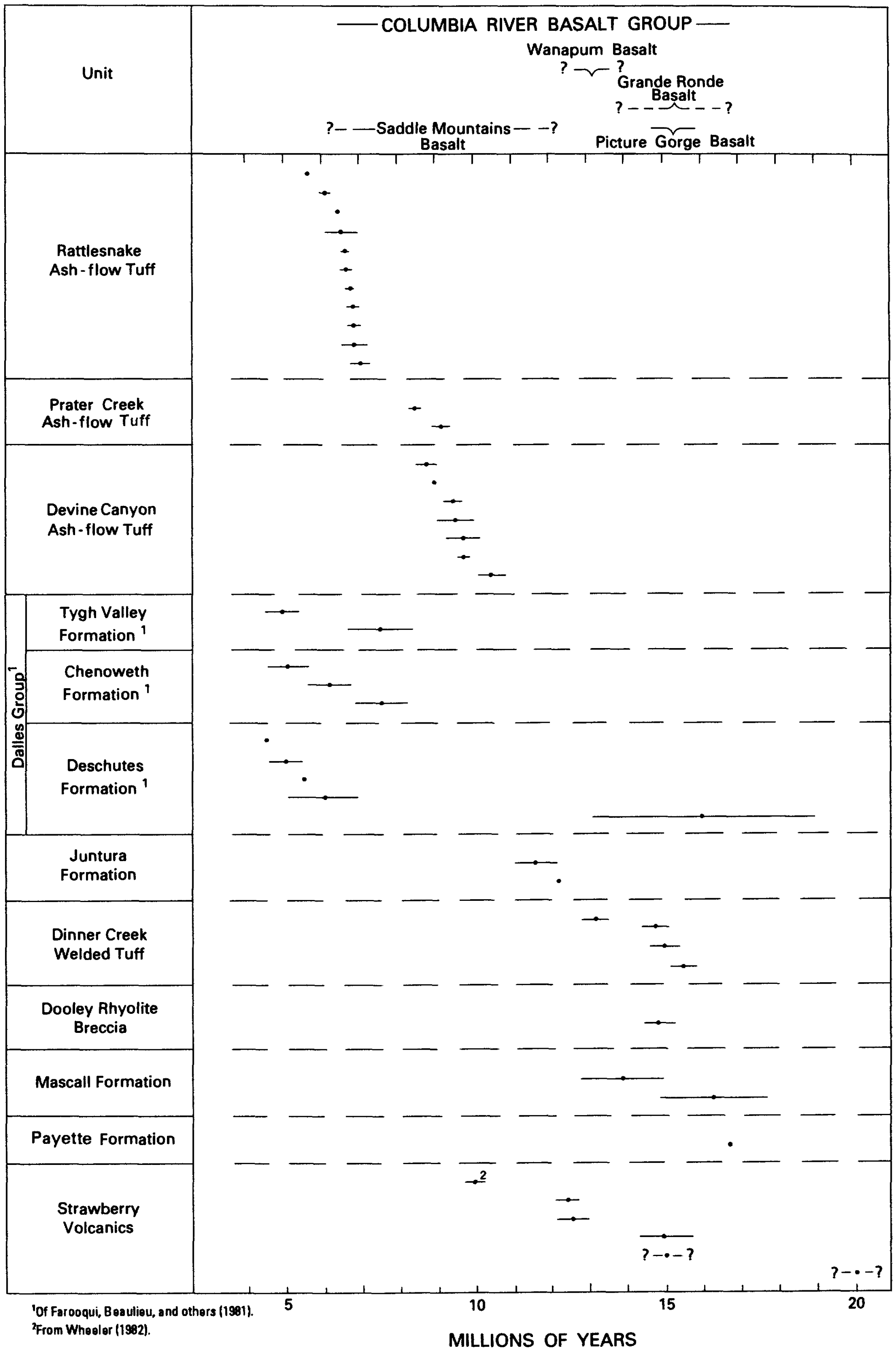


FIGURE 5.2.—K-Ar ages (overall ranges) for predominantly middle and upper Miocene formations in and adjacent to the Blue Mountains. Dots, ages; horizontal bars, analytical error; queried where uncertain, dashed where approximate.

for the volcanic-sedimentary facies in The Dalles Basin (Newcomb, 1969) and nearly 200 m for a reference section in Tygh Basin (Farooqui, Beaulieu, and others, 1981).

FIGURE 5.3.—Miocene strata of the Deschutes Formation of Farooqui, Beaulieu, and others (1981) exposed on northeast side of The Peninsula, Jefferson County, Oreg.; Lake Billy Chinook in foreground. Section is about 185 m thick and composed mostly of pumiceous laharic and ash-flow deposits, bedded volcanic sedimentary rocks, a few interbedded basalt flows, and capping olivine basalt or basaltic andesite.

FIGURE 5.4.—Deschutes Formation of Farooqui, Beaulieu, and others (1981) on west wall of Deschutes River canyon; Lake Billy Chinook in foreground. Section is about 230 m thick and consists mostly of fine- to coarse-grained, poorly to moderately well bedded clastic rocks. Columnar olivine basalt flow, exposed approximately one-third way up canyon wall, was isotopically dated at 5.8±1.0 Ma., and capping olivine basalt flow at 4.9±0.5 Ma (Armstrong and others, 1975).

K-Ar ages on a basalt flow and a tuff in the Tygh Valley Formation are, respectively, 7.5±0.8 and 4.9±0.5 Ma; ages for rocks from the Chenoweth Formation are 7.5±0.7, 5.7±0.6, and 5.1±0.5 Ma (Farooqui, Bunker, and others, 1981). These ages indicate a late Miocene and, possibly, earliest Pliocene age and some apparent temporal overlap with the youngest part of the Columbia River Basalt Group (fig. 5.2).

AGENCY AND ARLINGTON BASINS

Extensions of The Dalles Group of Farooqui, Beaulieu, and others (1981) are found in Agency and Arlington Basins within the Deschutes-Umatilla Plateau at the northern margin of the Blue Mountains (see fig. 1.2). Agency Basin is a structural depression in flows of the Columbia River Basalt Group at the northeast end of the Agency syncline. Arlington Basin lies along the axis of The Dalles-Umatilla syncline and is partly confined on the east by the Service anticline, on the south by the Willow Creek monocline, on the west by the Turner Butte anticline, and on the north by the Columbia Hills anticline. Stratigraphic nomenclature of the sedimentary units in these basins has recently been revised, and new formation names introduced. Within Agency Basin, indurated gravelly sedimentary rocks informally referred to as the McKay beds by Shotwell (1956) and as Pliocene fanglomerate by Hogenson (1964) were later called the McKay Formation by Farooqui, Beaulieu, and others (1981). This formation, which consists of partially cemented basaltic gravel and interbedded tuffaceous sand and silt, is conformable on the Frenchman Springs

FIGURE 5.5.—Laharic and ash-flow deposits at north end of The Peninsula, Jefferson County, Oreg., representing part of the Deschutes Formation of Farooqui, Beaulieu, and others (1981). View westward.

Member of the Wanapum Basalt and on the Grande Ronde Basalt, both part of the Columbia River Basalt Group, along the axis of the syncline; near the margins of the downwarp it shows angular relations with some of the flows. Sections as much as 77 m thick have been penetrated in drill holes in the area between McKay Creek and the Blue Mountains to the south (Gonthier and Harris, 1977, p. 9). Within Arlington Basin, basaltic gravel and tuffaceous sedimentary rocks, as much as 40 m thick, rest on several different units of the Columbia River Basalt Group. These fine- to coarse-grained clastic rocks, which were given various formal and informal names in the past (Alkali lake beds, Shutler Formation, and The Dalles Formation), were later called the Alkali Canyon Formation by Farooqui, Beaulieu, and others (1981).

Although neither of these new formations has been dated radiometrically, vertebrate faunas studied by Shotwell (1956) indicate a Hemphillian or, in current terminology (Berggren and Van Couvering, 1974; Palmer, 1983), late Miocene and early Pliocene age. Within the basins there is evidence of interfingering relations of the clastic rocks with flows of the Columbia River Basalt Group. Whether any part of these sedimentary formations is as old as the Ice Harbor Member (approx. 8.5 Ma) of the Columbia River Basalt Group is not known.

WALLA WALLA BASIN

Walla Walla Basin, about 55 km northeast of Agency Basin, contains a section of unconsolidated to partially consolidated lacustrine and fluvial sedimentary rocks, more than 180 m thick (Newcomb, 1965). As much as 150 m consists of clay that interfingers with sandy to coarse-grained, bouldery alluvial-fan deposits resting unconformably on flows of the Columbia River Basalt Group. Clasts in the fan deposits are exclusively of basaltic material; calcareous cement is locally present in these deposits. Layers of siltstone and conglomerate that overlie the clay and alluvial-fan deposits were tentatively correlated by Newcomb (1965, p. 22) with the Pliocene Ringold Formation of Pasco Basin, a few tens of kilometers to the northwest. Pebbles and cobbles in these beds consist mostly of basalt but include many other rock types. These beds are overlain, in turn, by the eolian Palouse Formation of Pleistocene age, by clay and silt, and by more than 30 m of horizontally bedded silt, fine sand, and gravel making up the Touchet Beds of Flint (1938), also of Pleistocene age. The gravel of the Touchet Beds is distinctive in that the cobbles and boulders are of highly diverse lithology and include granitic, metamorphic, metavolcanic, and metasedimentary rock types. The rocks immediately overlying the Columbia River Basalt Group apparently are not exposed at the surface, but they have been penetrated in many water wells drilled within the basin. On the west, Walla Walla Basin is separated from Pasco Basin by a low structural threshold a few kilometers east of Wallula Gap. Basin fill in Pasco Basin is more than 300 m thick and consists of lacustrine and fluvial sedimentary rocks and air-fall silicic volcanic ash derived from eruptions in the Oregon and Washington Cascades. Approximately half the thickness of this section of clastic rocks is exposed at the surface, principally in the White Bluffs northwest of Pasco; the lower half has been penetrated only in drill holes in deeper parts of the basin. The lower part of the basin fill, which rests on the Saddle Mountains Basalt (Ledgerwood and others, 1978), has been assigned to the Ringold Formation, described in some detail by Strand and Hough (1952) and Newcomb and others (1972), and the overlying rhythmically bedded, pebbly silts of glacial-flood origin (Waitt, 1980) have been assigned to the Touchet Beds.

According to Strand and Hough (1952), the section of the Ringold Formation at White Bluffs consists of a lower unit, about 27 m thick, of iron-stained pebble and cobble conglomerate and an upper, thicker unit of clay, sand, clayey sandstone, laminated siltstone, and some thin beds of ash, diatomite, gravel, and well-cemented flaggy sandstone. A vertebrate fauna collected from the middle part of this section indicated a probable middle to late Pleistocene age to Strand and Hough (1952) and Newcomb (1958). From an analysis of the structural deformation of the Ringold Formation and its relation to underlying flows of the Columbia River Basalt Group, however, Brown and McConiga (1960) inferred a somewhat older age. They indicated that the Ringold sedimentary rocks are locally conformable, or nearly so, with the underlying Miocene flows, implying that the oldest part of the Ringold Formation may possibly be as old as latest Miocene or early Pliocene, rather than Pleistocene. This older age is supported by data on the vertebrate fauna presented by Gustafson (1978), who dated the fossiliferous middle part of the formation as Pliocene and the stratigraphically lowest fossils in the formation as Hemphillian (late Miocene) or older. He considered the younger flows of the Columbia River Basalt Group and their sedimentary interbeds, as well as the lowermost strata of the Ringold, to be early Pliocene in age, which, in current epochal terminology, would be considered latest Miocene. Paleomagnetic studies by Packer (1979) of the Ringold Formation exposed in White Bluff indicate an age older than 0.7 Ma.

The poorly consolidated and disconformably overlying glacial-flood deposits and loess of the Touchet Beds of Flint (1938) are late Pleistocene in age. Ledgerwood and others (1978) indicate that the Touchet Beds are overlain by Mount St. Helens, Glacier Peak, and Mazama ash

beds, all of Holocene age, and Waitt (1980) indicated some interbedding of these tephra units.

PAULINA BASIN

Paulina Basin, on the south flanks of the Ochoco Mountains uplift in the south-central part of the Blue Mountains province, is largely localized by a northeast-trending syncline in Miocene flows of the Columbia River Basalt Group that are generally considered to be part of the Picture Gorge Basalt (Brown and Thayer, 1966; Swanson, 1969). In a study of Paulina Basin with emphasis on ash-flow tuffs, Davenport (1971) indicated that the basin fill has a maximum thickness of slightly more that 300 m and includes part of the middle Miocene Mascall Formation, an overlying sequence of upper Miocene interlayered clastic sedimentary rocks and ash-flow tuff, including the Devine Canyon and Rattlesnake Ash-flow Tuffs and, possibly, the Prater Creek Ash-flow Tuff (Walker, 1979), and upper Miocene or Pliocene rim-forming basalt flows called the Ochoco Basalt by Davenport (1971).

The Mascall Formation, which is at least 138 m thick in Paulina Basin (Forth, 1965), consists of tuffaceous sandstone, conglomerate, a 45-m-thick vitric ash-flow tuff (near the base of the formation), and an interbedded thin flow of olivine basalt. The rhyolitic ash-flow tuff (table 5.1) has yielded a K-Ar age originally reported as 15.8±1.4 Ma (Enlows and Davenport, 1971) but recalculated with new constants at 16.2±1.4 Ma (Fiebelkorn and others, 1983). The distribution and volume of the ash-flow tuff are not known, nor is it clear whether this tuff in Paulina Basin is the same as the Mascall Formation ash-flow tuff, dated at 13.8±0.8 Ma (Fiebelkorn and others, 1983) that crops out southwest of the Maury Mountains, a few kilometers south of Camp Creek. Vents for these ash flows have not been recognized. Interlayering of clastic rocks of the Mascall Formation and flows of the Columbia River Basalt Group was indicated by Evernden and James (1964, fig. 5) for a section exposed in John Day Basin; a K-Ar age on the interlayered basalt is 15.4 Ma. Similar interlayering of Mascall-age clastic rocks, Picture Gorge flows, and flows of the basalt of Prineville compositional type is present southwest and west of Paulina Basin in areas adjacent to the Maury Mountains; parts of these sections also include units of peperite composed of fragmental basalt and silicic tuffaceous sedimentary material.

In addition to the Mascall-age ash-flow tuff, Davenport (1971) recognized three separate and younger ash-flow tuff layers that are separated by layers of tuffaceous sandstone and pebble conglomerate. He considered these ash-flow tuff layers to be of Pliocene age and to form part of the Danforth Formation of Piper and others (1939), comprising sequences of volcanic and sedimentary rocks exposed in Harney Basin, 100 km to the southeast. Regional mapping has shown that these ash-flow tuffs are, indeed, the same as those exposed in Harney Basin, which Walker (1979) redefined as separate formations and named the Devine Canyon, Prater Creek, and Rattlesnake Ash-flow Tuffs. K-Ar ages on these tuff units are about 9.2, 8.4, and 6.5 Ma, respectively, all of late Miocene age.

Collectively, the ash-flow tuff units discontinuously cover a very large area on the south flanks of the Blue Mountains and extend for many tens of kilometers to the south and southeast into and beyond Harney Basin (Walker, 1979). The distinctive, crystal-rich Devine Canyon Ash-flow Tuff can be traced through erosional remnants on butte and ridge tops east of Paulina Basin into the Suplee-Izee area (Dickinson and Vigrass, 1965) and Bear Valley. It has also been recognized in stratigraphic sections many kilometers east and southeast of Bear Valley in Juntura Basin, east of Castle Rock, and along the northern margin of Harper Basin (Walker, 1979). Erosional remnants of the Prater Creek and (or) Rattlesnake Ash-flow Tuff are present in Bear Valley and in Logan Valley, where they locally rest with discordance on the Miocene Strawberry Volcanics.

The units of ash-flow tuff are rhyolitic (Walker, 1979, table 2) and, in the classification of Nockolds (1954), are most like average alkali rhyolite.

TABLE 5.1.—*Chemical analyses of ash-flow tuff from the Mascall Formation*

[Samples: 1, Rhyolite tuff from Belshaws Ranch (Calkins, 1902); 2, Rhyolitic ash-flow tuff from outcrops at confluence of Deer Creek with South Fork of John Day River (Davenport, 1971, table 3). Analysis of sample 1, water free. Oxides in weight percent; n.r., not reported (all Fe for sample 2 reported as FeO); n.d., not determined; ---, undetected]

Sample	1	2
SiO_2	74.22	74.9
Al_2O_3	13.50	13.3
Fe_2O_3	1.09	n.r.
FeO	1.49	2.6
MgO	.19	.5
CaO	1.03	.6
Na_2O	3.69	3.33
K_2O	4.33	4.30
TiO_2	.35	.18
P_2O_5	.11	---
MnO	n.d.	---

The total volume of such upper Miocene ash-flow tuff layers, including only the more densely welded ledge- and rim-forming parts of the flows, is about 450 to 500 km^3 (Walker, 1970); if nonwelded parts of the ash flows are included, the volume could be as much as 1,000 to 1,500 km^3.

Stratigraphically above the clastic basin fill are thin, rim-forming, upper Miocene or Pliocene olivine basalt flows and flow breccia that were erupted from lava cones generally situated around the margins of the basin.

JOHN DAY VALLEY

To the north of Paulina Basin, in John Day Valley, middle Miocene flows of the Picture Gorge Basalt interfinger with and are overlain conformably by clastic rocks of the Mascall Formation, which is, in turn, unconformably overlain locally by gravelly sedimentary rocks and an interbedded layer of ash-flow tuff, collectively referred to by many investigators (Thayer and Brown, 1966; Oles and Enlows, 1971; Enlows, 1976) as the Rattlesnake Formation of late Miocene and Pliocene(?) age (fig. 5.6). Merriam (1901) estimated that 244 to 305 m of Mascall Formation clastic rocks are present, and Enlows (1976) measured a total thickness of about 192 m for the Rattlesnake Formation in the area originally considered the type area.

Because of considerable confusion in stratigraphic nomenclature for upper Miocene rocks of both the John Day region and Harney Basin many kilometers to the south, Walker (1979) redefined several of the stratigraphic units and applied formation names to several widespread ash-flow tuffs that had their source in buried vents, probably calderas, in Harney Basin (Walker, 1970a; Blank and Getting, 1974). The ash-flow tuff interbedded in what had been called the Rattlesnake Formation was elevated to formation status and named the Rattlesnake Ash-flow Tuff. Walker's (1979) justification for redefining this unit includes the following: (1) This ash-flow tuff layer is one of several key marker horizons over thousands of square kilometers of south-central and southeastern Oregon, whereas the locally accompanying fanglomerate deposits are very restricted in areal extent; (2) the exposures in and near John Day Valley are near the distal end of the ash-flow tuff sheet and are not representative of the major part of the unit. The original type area (or locality), on Rattlesnake Creek about 1.6 km west of Cottonwood, Oreg., was redesignated as a reference locality, and a new type locality was established on the west side of U.S. Interstate Highway 395, about 10 km north of Burns, Oreg. (Walker, 1979).

FIGURE 5.6.—West end of John Day valley, Grant and Wheeler Counties, Oreg. Inclined layered flows at right are part of the Picture Gorge Basalt of the Columbia River Basalt Group. Flows are conformably overlain by bedded tuffaceous sedimentary rocks of the middle Miocene Mascall Formation (white outcrops on left) that, in turn, are unconformably overlain by fanglomerate and by the ledge-forming upper Miocene Rattlesnake Ash-flow Tuff on skyline. Buttes on skyline are approximately 300 m high. View westward.

In contrast to Merriam's estimate of 244 to 305 m for the thickness of the Mascall Formation, Thayer and Brown (1966) inferred that the formation at Picture Gorge, at the west end of John Day Valley, was originally 1,830 to 2,130 m thick, although it is now preserved in thinner erosional remnants in isolated structural basins. However, most measured sections of the Mascall Formation are no more than about 300 m thick, and onlap relations and sedimentary facies throughout the region indicate that this is more nearly the correct maximum thickness for the formation. Only slight changes need be postulated in the relation of the Mascall Formation to flows of the Columbia River Basalt Group inferred by Thayer and Brown (1966, fig. 3), in combination with some fault repetition of the basalt section exposed on Flat Creek, as suggested by Swanson and others (1979, p. G13), to drastically reduce the thickness estimates by Thayer and Brown from approximately 2,000 m to only a small fraction of that figure.

Downs (1956, p. 205) briefly described a section of the Mascall Formation at its type locality near Picture Gorge, Grant County, Oreg. This section is 119 m thick and consists of several varieties of altered and unaltered, partly indurated, water-laid and ash-flow tuff composed of glass shards and mineral grains and some laminated and crossbedded tuffaceous sandstone; both the tuff and sandstone locally contain a Barstovian (middle Miocene) vertebrate fauna. Local swampy conditions are indicated in the Mascall Formation by the presence of lenticular lignite beds, as much as 1.5 m thick, a few kilometers to the east (Thayer, 1957). Farther to the east, gravelly sandstone and conglomerate, the pebbles of which are largely 2 to 3 cm in diameter, highly polished, and mostly

of pre-Cenozoic rocks, become more abundant in the formation; a layer of ash-flow tuff, 14 to 15 m thick, locally forms the basal part of the formation. A K-Ar age of 15.4 Ma was determined by Evernden and others (1964, sample KA–1203) for the Mascall Formation of the John Day Valley area. This K-Ar age was confirmed by radiometric dating of a nearby (interbedded?) flow of the Columbia River Basalt Group. On these bases the Mascall Formation is here regarded as middle Miocene in age.

Rocks referred to by previous workers as the Rattlesnake Formation include an upper deposit of coarse-grained fanglomerate of remarkably uniform lithology that disconformably overlies vitric ash-flow tuff, now referred to as the Rattlesnake Ash-flow Tuff (Walker, 1979), and a disconformably underlying lower deposit of fanglomerate with thick lenticular interbeds of sandstone and mudstone (Enlows, 1976).

FOX BASIN

In Fox Basin, about 20 km north of John Day Valley, pebble conglomerate and tuff of the Mascall Formation are preserved in a shallow structural depression partly confined by northwest-trending faults and a gentle synclinal warp (Brown and Thayer, 1966). According to Thayer (1957, p. 236) a series of "Columbia River Basalt flows wedges out southeastward over yellowish rhyolite tuff and Mascall-type pebble gravels in the northwest corner of Fox Valley, and thin basalt flows lie on and have been faulted together with tuffaceous beds in the southern part of the basin."

JUNTURA BASIN

In Juntura Basin, at and just south of the southern margin of the Blue Mountains, a maximum composite thickness of about 670 m of silicic tuff, palagonite tuff and breccia, and fine-grained tuffaceous sedimentary rocks, including ashy diatomite, rests disconformably, possibly with slight angular discordance (Bowen and others, 1963), on Miocene basalt and andesite, in part representing flows of the Strawberry Volcanics. The section of clastic rocks and diatomite thins northward in the Blue Mountains province and, in places, appears to contain larger proportions of palagonitic tuff. The basin deposits, including the Juntura Formation of Barstovian age, the Devine Canyon Ash-flow Tuff, dated at about 9.2 Ma (Walker, 1979), and the Drewsey Formation, which contains a Hemphillian fauna (Shotwell, 1963), are in part equivalent in age to the Strawberry Volcanics and to parts of the Columbia River Basalt Group. Even though the Juntura Basin section impinges on the Strawberry Volcanics and is approximately coeval, clear interfingering or gradational relations have not been recognized.

HARPER BASIN

The northern edge of Harper Basin, which is the next major Miocene and younger(?) basin to the east of Juntura Basin, extends into the southernmost margin of the Blue Mountains south of the Juniper and Ironside Mountains. The basin is a partly faulted downwarp developed in middle Miocene silicic and mafic volcanic rocks that include the Strawberry Volcanics and units named the Littlefield Rhyolite and Hunter Creek Basalt by Kittleman and others (1965). The basin fill consists of fine- and coarse-grained epiclastic and volcaniclastic rocks, diatomite, and several thin basalt flows (fig. 5.7) that together aggregate to nearly 600-m thickness in the deepest part of the basin; the volcanic ash in many beds has been diagenetically altered to zeolites and clay minerals. Kittleman and others (1965) delineated several units, including the Drip Spring and Bully Creek Formations. The section thins northward and pinches out on older Miocene volcanic rocks; along the northern margin of the basin it is dominated by several layers of ash-flow tuff of different lithologic types, and it commonly occurs in disconnected patches (Brooks and others, 1976). The distinctive, crystal-rich Devine Canyon Ash-flow Tuff, dated at about 9.2 Ma, is present in the section, as well as isolated patches of an older welded tuff, possibly equivalent to the Dinner Creek Welded Ash-flow Tuff of Haddock (1967, 1968) (equivalent to the Dinner Creek Welded Tuff as used by Greene and others, 1972), and one or more younger pumiceous, vitric ash-flow tuff layers, possibly representing distal ends of the Prater Creek and (or) Rattlesnake Ash-flow Tuffs of Harney Basin, 100 km to the southwest. The Dinner Creek Welded Ash-flow Tuff, which Haddock (1966) considered to have a volume of more than 25 km^3 and to have been

FIGURE 5.7.—Bedded tuffaceous sedimentary rocks, partly diagenetically altered to clay minerals and zeolites, exposed in northern part of Harper Basin, northern Malheur County, Oreg. Part of the upper Miocene Bully Creek Formation. Bluff in background is about 100 m high. View northward.

erupted from a fissure now occupied by a rhyolite dike at Castle Rock, is well exposed in Malheur Gorge (fig. 5.8) and in discontinuous patches to the north and northeast through Willow Creek basin and into Unity Basin. It has been dated by K-Ar methods at 15.3±0.4, 14.9±0.4, 14.7±0.4, and 13.1±0.4 Ma from widely separated areas on the south flanks of the Blue Mountains. Considering these discordant ages, more than one ash flow may have been identified by various workers in the region as the Dinner Creek Welded Ash-flow Tuff. If so, one or more of these tuff layers may be the same as some of those recognized in the Mascall Formation in Paulina Basin and the John Day Valley area (fig. 5.2).

BOISE BASIN

Along the southeastern margin of the Blue Mountains and in the adjacent Owyhee Upland and Columbia Intermontane provinces are extensive outcrop areas of Miocene and younger sedimentary rocks and minor volcanic rocks that represent the northward and northwestward extensions of sedimentary and volcanic sequences of Boise Basin (Bond, 1978). Many formation names have been applied to these rocks, depending on their geographic position within the basin and their lithology, and several regional correlations have been attempted (Malde and Powers, 1962; Kittleman and others, 1965; Armstrong, 1975). In Oregon, they generally have been referred to as the northward extensions of the Miocene Deer Butte and Pliocene Chalk Butte Formations of Corcoran and others (1962), or of the Bully Creek and Grassy Mountain Formations of Kittleman and others (1965); in Idaho, they have been referred to as the Payette (Lindgren, 1898; Russell, 1902), Poison Creek, Chalk Hills, Glenns Ferry, Bruneau, and Idaho Formations, the "Caldwell-Nampa sediments" of Mitchell and

FIGURE 5.8.—Faulted middle Miocene Dinner Creek Welded Tuff (arrows) in Malheur Gorge, northern Malheur County, Oreg. Highest bluff in background is about 500 m high. View westward.

Bennett (1979), and the Cenozoic "undifferentiated rocks" of Malde and Powers (1962). Collectively, most of the middle Miocene and Pliocene sedimentary rocks and interbedded flows have been considered to be a part of what is referred to as the Idaho Group, and an overlying Pleistocene and Holocene sequence of interstratified basalt flows and gravels has been called the Snake River Group.

The basin fill ranges in thickness from about 900 m in areas to the south, in the deepest parts of Boise Basin, to near 0 along the northern margin of the basin. There it pinches out, principally by onlap on middle Miocene volcanic rocks, including basalt flows mapped as part of the Columbia River Basalt Group in Idaho, and on basaltic to rhyolitic volcanic rocks identified as the Strawberry Volcanics, Littlefield Rhyolite, Dinner Creek Welded Tuff, or associated unnamed rocks in Oregon.

Most of the formations of the Idaho Group consist of siliceous volcanic ash and fine-grained tuffaceous detrital material in both bedded and massive sequences, particularly toward the central part of the basin, and of somewhat coarser grained, commonly conglomeratic and partly locally derived detritus, intermixed with tuffaceous sandstone and siltstone, near the basin margins. Both lacustrine and fluviatile deposits are represented. Locally, thin basalt flows are interbedded in the Idaho Group; along the northwestern margin of the basin, near the boundary of the Blue Mountains and the Owyhee Upland province, several thin flows of diktytaxitic, subophitic to intergranular olivine basalt and basaltic andesite cap the Miocene and Pliocene(?) sedimentary rocks (Brooks and others, 1976). The olivine basalt flows were erupted from several small cinder and lava cones along the margin of the basin. The capping basalts are undated but are probably either late Miocene or early Pliocene in age. Other parts of the section have been dated as Miocene, Pliocene, and Pleistocene from contained vertebrate fossils and from K-Ar ages (Armstrong, 1975) that range from about 13 to less than 1 Ma.

Much of the fine-grained debris in the basin fill is siliceous volcanic ash or its alteration products, including clays, zeolites, and secondary silica minerals. Most of this detrital material is related to large-volume eruptions of tuff from calderas in Harney Basin during the late Miocene. Some of the ash may be related to middle Miocene eruptions of ash flows now recognized as part of either the Mascall Formation or the Dinner Creek Welded Tuff; some may be related to Miocene eruptions in the McDermitt-White Horse caldera complex, nearly 250 km south-southwest of the central part of Boise Basin and to Miocene calderas south of Ontario, Oreg. The total volume of siliceous debris in these sedimentary sequences in the basin is not known; detailed information on the

configuration of the basin floor and the thickness of the basin fill apparently is almost nonexistent, and so volume estimates are approximate at best. Furthermore, the relative proportions of silicic ash and other detrital constituents are unknown. Certainly, thousands of cubic kilometers of sediment was deposited in Boise Basin, and much of it still remains after extensive erosion. As much as 75 percent of the fill is composed of fresh or altered vitric ash and pumice. The very large total volume of siliceous volcanic ash indicates voluminous silicic volcanic activity in adjoining areas to the west and north and possibly to the south.

BASINS NORTH OF HARPER AND BOISE BASINS

Several isolated and comparatively large depositional basins are present to the north of Harper and Boise Basins in and adjacent to the Blue Mountains, including Mormon, Unity, and Durkee Basins and Willow Creek basin (fig. 5.1), all containing middle Miocene and younger volcanic and volcaniclastic rocks, as well as some coarse-grained conglomeratic fluviatile rocks. Additional, structural and erosional depressions filled with epiclastic and volcaniclastic rocks and a few interstratified basalt and andesite flows have been recognized in the Burnt River valley and along several other drainages in the region.

In and adjacent to Mormon Basin, Wolff (1965) reported 250 to 300 m of Miocene volcanic rocks, including the Dinner Creek Welded Ash-flow Tuff, coarse-grained mudflow deposits, and tuffaceous sedimentary rocks, as well as several interbedded basalt flows and rhyolite flow breccia, all probably temporal equivalents of parts of the Strawberry Volcanics. Disconformably overlying this volcanic assemblage are coarse-grained fluvial conglomerate, sandstone, and mudstone deposits that Wolff (1965, p. 148) considered to be of Miocene age. Unconformably overlying these coarse-grained clastic rocks is a unit of partly lacustrine, light-colored tuffaceous siltstone and sandstone with interlayered lenses of diatomite; coarse-grained conglomeratic facies are present near the basin margins. The tuffaceous siltstone and sandstone unit has an aggregate maximum thickness of more than 180 m, but it thins and pinches out on older rocks along the margins of the basin. Wolff (1965, p. 173) considered these rocks to be of Pliocene age, partly on the basis of correlation with tuffaceous sedimentary rocks in adjacent areas that contain a vertebrate fauna originally identified as early Pliocene in age. In current geochronologic terminology, that vertebrate fauna would most likely be considered of late Miocene age. Thin, gently dipping olivine-bearing basalt and basaltic andesite flows that were erupted from small central vents cap the sedimentary basin fill in places.

Unity Basin, 15 to 20 km west-northwest of Mormon Basin, and the Willow Creek basin (fig. 5.1), west of Mormon Basin, are largely fault-controlled depressions that contain fill a few tens to several hundreds of meters thick consisting of beds of conglomerate, claystone, and fine-grained water-laid tuff (fig. 5.9) rich in glass and pumice (Thayer and Brown, 1966). Miocene and Pliocene fossils considered of possible late Miocene age have been recovered from beds in Unity Basin (Thayer, 1957, p. 238). East of Unity Basin, along the Burnt River valley, fluvial gravel, indurated ash beds, tuffaceous sedimentary rocks, and one or more thin layers of vitric ash-flow tuff have been incorporated in a unit designated as Miocene and Pliocene "tuffaceous sedimentary rocks" by Brooks and others (1976).

Durkee Basin, about 30 km northeast of Mormon Basin (fig. 5.1), is a grabenlike structure partly confined by northwest-trending normal faults. The basin contains sedimentary and volcanic deposits that total more than 350 m in thickness; epiclastic deposits, which coarsen toward the basin margins and locally are conglomeratic, in places contain vertebrate fossils of Clarendonian (late Miocene) age. Prostka (1967) described the basin fill as mostly fine- to medium-grained, well-bedded, poorly consolidated sand, silt, and varying amounts of intermixed ash and pumice. Both fluvial and lacustrine deposits are present. Also recognized are ash beds, in part diagenetically altered to zeolites, potassium feldspar, clay minerals, and silica minerals (Gude and Sheppard, 1978), as well as diatomite and one or more layers of ash-flow tuff. One welded tuff locally underlies and elsewhere is interbedded in the sedimentary sequence. According to Gude and Sheppard (1978), the diagenetic

FIGURE 5.9.—Poorly to moderately well bedded upper Miocene tuffaceous sedimentary rocks in Unity Basin, Baker County, Oreg. Bluff in background is about 50 m high. View northeastward.

minerals indicate that a closed, saline, alkaline lake occupied part of Durkee Basin during the Pliocene (late Miocene time in current terminology). Local thin flows of platy olivine basalt, erupted from small central vents, are present in areas marginal to the basin and appear to predate the sedimentary basin fill.

BASINS NORTH AND NORTHWEST OF DURKEE BASIN

North and northwest of Durkee basin, silicic pyroclastic and tuffaceous sedimentary rocks fill several separate structural basins developed in flows of the Columbia River Basalt Group and pre-Cenozoic rocks. These structural basins, including Baker Valley, lower Powder River valley, and Richland Valley, contain sections about 150 m or more thick. As pointed out by Brooks and others (1976), the clastic basin fill apparently does not interfinger with the flows of the Columbia River Basalt Group; but if the Clarendonian (late Miocene) age postulated by Prostka (1967) is correct, some of these clastic rocks may be temporal equivalents of younger late Miocene flows of the central part of the Columbia Plateau.

In general, the basin fill consists of a lower lacustrine facies of thin-bedded, light-colored tuffaceous claystone, siltstone, and sandstone, tuff, and diatomite, and an upper fluvial facies of pebble conglomerate interbedded with fine-grained tuffaceous sandstone and siltstone. In places, much of the volcanic ash and pumice is diagenetically altered to zeolites, clay minerals, and secondary silica minerals. Several thin (maximum 8-m-thick) layers of vitric ash-flow tuff, mostly composed of abundant well-sorted, pinkish-brown to buff and nearly colorless glass shards, and some pumice lapilli and rare feldspar, biotite, and rock fragments, are interstratified with the sedimentary rocks (fig. 5.10). The ash-flow tuff layers show varying degrees of compaction and welding of shards and pumice, in a few places exhibiting well-defined eutaxitic texture and elsewhere showing essentially no flattening or deformation of shards and lapilli. The source of the ash-flow tuff deposit is not known; lithologically, they appear to be most like the upper Miocene Prater Creek and Rattlesnake Ash-flow Tuffs of Harney Basin (Walker, 1979), about 8.4 and 6.5 Ma, respectively, and they appear to be in a similar or near similar stratigraphic position. However, much additional fieldwork and geochemical and radiometric-age data are necessary to either firmly establish or reject these possible correlations.

Grande Ronde Valley, to the north of Baker Valley, is a complex graben developed in flows of the Columbia River Basalt Group; the graben contains as much as 600 m of fine-grained, poorly consolidated lacustrine deposits, mostly claystone, siltstone, and sandstone (Hampton and Brown, 1964). Poorly sorted, bouldery alluvial-fan deposits are present in many places along the precipitous margins of the graben. A Pleistocene(?) age was assigned to the lacustrine deposits by Hampton and Brown (1964), but in current age terminology some of the earliest of these deposits probably would be considered Pliocene and, possibly, even latest Miocene.

MIOCENE VOLCANIC ACCUMULATIONS

Interspersed among the sedimentary basins along the south and southeast flanks of the Blue Mountains are accumulations of Miocene eruptive rocks in the form of shield volcanos, cones transitional between shields and stratovolcanos, large complex domal masses, local flow sequences, and several kinds of related intrusions. In places, these volcanic rocks impinge on and intermingle with the Miocene sedimentary basin-fill deposits, but elsewhere they form separate and distinct eruptive piles. The rocks are of several types of eruptive materials, including basaltic to rhyolitic calc-alkaline flows and flow breccia of the Strawberry Volcanics, rhyolitic flows, breccia, tuff, and intrusions of the igneous complex at Dooley Mountain, basaltic to andesitic peperite, palagonitic tuff, minor flows of high-alumina (plagioclase-rich) ophitic basalt, and several unnamed domal and intrusive complexes of dacite and rhyolite.

STRAWBERRY VOLCANICS

The largest and most diverse group of these eruptive rocks, named the Strawberry Volcanics by Thayer

FIGURE 5.10.—Poorly exposed, fine-grained, tuffaceous sedimentary rocks containing interlayers of pumiceous ash-flow tuff at Virtue Flat, Baker County, Oreg. Vertebrate fossils indicate middle to late Miocene age for the sedimentary rocks. Source of ash flow is unknown; lithologically, it resembles distal end of either the Prater Creek or Rattlesnake Ash-flow Tuff of Harney Basin. Outcrops range in height from about 3 to 12 m. View westward.

(1957), is mostly coeval with the Columbia River Basalt Group; radiometric ages mostly indicate an age range of about 20 to 10 Ma (fig. 5.2) for the formation. There is some question as to whether the K-Ar ages older than about 16 Ma were obtained on rocks representing lower parts of the Strawberry Volcanics or on some older unit. The Strawberry Volcanics is here considered to be middle and late Miocene in age. The Strawberry Volcanics is exposed widely in southeastern Grant County and parts of adjoining counties on the south flanks of the Blue Mountains (fig. 5.1; Brown and Thayer, 1966; Robyn, 1977, 1979). Most of the volcanic rocks are flows and flow breccia of olivine basalt, basaltic andesite, and pyroxene andesite, although they also include local large masses of dacite or rhyodacite. Chemically, they show great diversity (fig. 5.11); SiO_2 contents range from about 51 to 72 weight percent. Comparisons with average compositions of Rock Creek, Frenchman Springs, and Pomona chemical types (Swanson and others, 1979), all parts of the Columbia River Basalt Group and all about the same age as the Strawberry Volcanics, show (fig. 5.11) that the Strawberry Volcanics is almost invariably lower in CaO, TiO_2, and MgO contents and that its rocks tend to be higher in Na_2O and K_2O contents for comparable SiO_2 contents. Comparisons with the average composition of the Prineville chemical type, also apparently coeval with part of the Strawberry Volcanics, show similar alkali, higher Al_2O_3, and lower TiO_2 and Fe_2O_3 contents.

Numerous vents for these rocks are present in the vicinity of Strawberry and Lookout Mountains, 25 km or more southeast of John Day, and several large masses of silicic rocks were erupted from local large domal complexes east and southeast of Bear Valley.

The total volume of the calc-alkaline Strawberry Volcanics is difficult to estimate, largely because the amount of relief on the surface on which these eruptive rocks were deposited is poorly known. As shown by Brown and Thayer (1966), they cover an area north of lat 44° N. of more than 3,400 km^2 and locally are much more than 2,000 m thick. Thayer (1957) recognized a section more than 2,200 m thick dominated by basaltic to andesitic hypersthene-bearing lava on the north slopes of Strawberry Mountain; Thayer and Brown (1973) indicated a total maximum thickness of 2,700 to 3,000 m for a section composed mainly of rhyolite and andesite in Ironside Mountain, a few kilometers southwest of Willow Creek basin. Sections hundreds of meters thick also are exposed on canyon walls in areas south, southwest, and southeast of Logan Valley.

If the average thickness of these volcanic rocks over 3,400 km^2 of the outcrop area north of lat 44° N. is assumed to be 2 km, then their total volume is 6,800 km^3. An additional volume of these rocks is present south of lat 44° N. on the south flank of the Blue Mountains.

OTHER VOLCANIC ACCUMULATIONS

In Paulina Basin and westward around the margins of the Maury Mountains, flows, breccia, and local peperite of basalt and basaltic andesite are interstratified with tuffaceous sedimentary rocks that in places contain a Barstovian (Mascallian) fauna. Apparently, none of these rocks has been dated by radiometric methods, but their Barstovian faunal age and stratigraphic relations indicate that they are coeval with parts of the Columbia River Basalt Group. Although some of these flows show chemical similarities with flows of the group, others are chemically quite different and, as shown by Osawa and Goles (1970), contain trace-element abundances that differ significantly from any known abundances in the Columbia River Basalt Group.

Southwest of Paulina, on Camp Creek, basalt flows with good columnar jointing, presumably correlative with the flows at Picture Gorge, are sandwiched between silicic tuff and tuffaceous sedimentary rocks; the basalts pinch out southward on underlying sedimentary rocks of the John Day Formation and are, in turn, overlain by bedded tuff and tuffaceous sedimentary rocks that contain a Barstovian vertebrate fauna. An analysis of one of the flows shows 48.8 weight percent SiO_2, 15.3 weight percent Al_2O_3, 10.6 weight percent CaO, and low K_2O and TiO_2 contents. This composition is most like the average composition of flows of the Picture Gorge Basalt given by Swanson and others (1979, table 2, column 1), although it is somewhat less silicic.

Flows of ophitic, high-alumina (plagioclase-rich) basalt are interstratified in the middle Miocene part of the sedimentary section both in the west part of Paulina Basin and on Bear Creek at the west end of the Maury Mountains. In major-oxide chemistry (fig. 5.11) these rocks are unlike any of the chemical averages of flows of the Columbia River Basalt Group (Swanson and others, 1979, table 2). A few flows at the west end of the Maury Mountains are like Picture Gorge flows in their trace-element abundances, but interstratified ophitic flows are chemically unlike any flows of the Columbia River Basalt Group (Osawa and Goles, 1979, table 4, samples 27A–27E). The major-oxide chemistry of the ophitic flows is more nearly comparable to that of flows in the Steens Basalt (Gunn and Watkins, 1970; Walker, 1970b) of southeastern Oregon, particularly in their high Al_2O_3 and CaO and relatively low K_2O contents.

Middle Miocene silicic volcanic rocks are widely exposed in the mountain ranges south of Baker, particularly in the Dooley Mountain area and in ranges both north and south of Mormon Basin. The silicic rocks consist of a wide variety of flows, breccia, intrusive rocks, and related air-fall and ash-flow tuff, presumably all of rhyolitic composition. Gilluly (1937) named the silicic rocks on

Dooley Mountain and in adjacent areas the Dooley Rhyolite Breccia and considered those of Dooley Mountain to mark the site of a volcanic eruptive center.

These rocks have been radiometrically dated in only one place on Dooley Mountain, so that their age range and precise correlation with other Miocene units are not

EXPLANATION

- + Analyses of Strawberry Volcanics from Thayer (1957)
- • Analyses of Strawberry Volcanics from Robyn (1977)
- × Analyses of Strawberry Volcanics at Ironside Mountain (from Thayer and Brown, 1973)
- ● Unpublished (USGS) analyses of Miocene flows adjacent to the Maury Mountains, along south margin of Blue Mountains province
- ◉ Average Prineville chemical type of basalt from Uppuluri (1974)
- ▲ Average (68 analyses) of Rock Creek chemical type, Columbia River Basalt Group (Swanson and others, 1979)
- + Average (8 analyses) of Frenchman Springs chemical type, Columbia River Basalt Group (Swanson and others, 1979)
- ▽ Average (30 analyses) of Pomona chemical type, Columbia River Basalt Group (Swanson and others, 1979)
- × Average (18 analyses, including several of Steens Basalt) of Miocene basalt flows of southeastern Oregon (Walker, 1979)

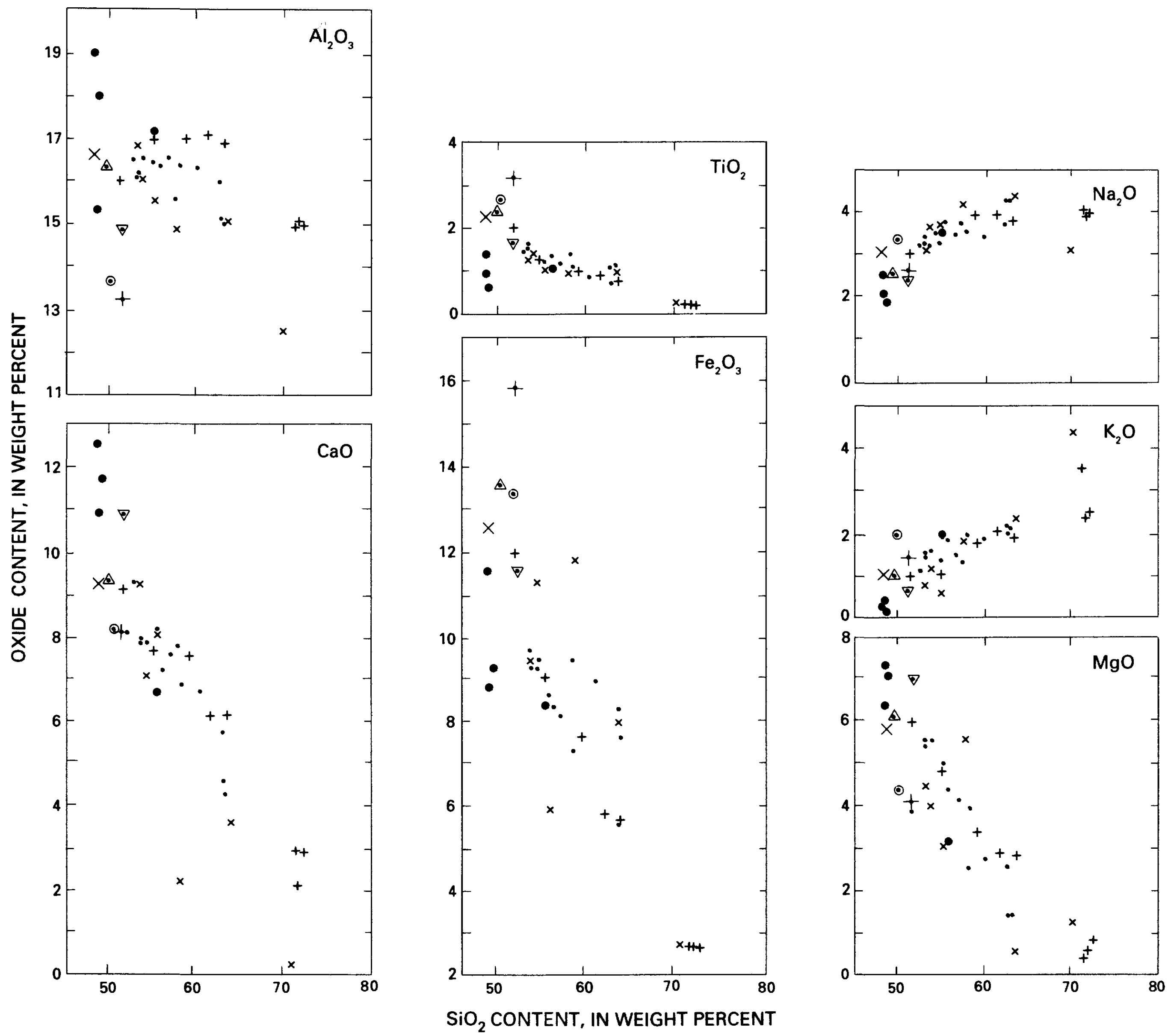


FIGURE 5.11.—Harker diagrams of the Strawberry Volcanics and temporally related Miocene flows along south margins of Blue Mountains, showing comparisons with selected average compositions of chemical types within the Columbia River Basalt Group.

known. A K-Ar age of 14.7±4 Ma (Fiebelkorn and others, 1983) on intrusive rhyolite exposed in a quarry near the summit of Dooley Mountain indicates approximate temporal equivalence with the Dinner Creek Welded Tuff, the Mascall Formation, and both the Picture Gorge and Grande Ronde Basalts of the Columbia River Basalt Group.

REFERENCES CITED

Armstrong, R.L., 1975, The geochronometry of Idaho: Isochron/West, no. 14, 50 p.

Armstrong, R.L., Taylor, E.M., Hales, P.O., and Parker, D.J., 1975, K-Ar dates for volcanic rocks, central Cascade Range of Oregon: Isochron/West, no. 13, p. 5–10.

Berggren, W.A., and Van Couvering, J.A., 1974, The late Neogene—Biostratigraphy, geochronology, and paleoclimatology of the last 15 m.y. in marine and continental sequences: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 16, no. 1–2, p. 1–216.

Blank, H.R., and Getting, M.E., 1974, Geophysical evidence of caldera structures in the Harney Basin of central eastern Oregon [abs.]: Eos (American Geophysical Union Transactions), v. 55, no. 5, p. 557.

Bond, J.C., 1978, Geologic map of Idaho: Moscow, Idaho Bureau of Mines and Geology, scale 1:500,000.

Bowen, R.G., Gray, W.L., and Gregory, D.C., 1963, General geology of the northern Juntura Basin, chap. 2 *of* Shotwell, J.A., ed., The Juntura Basin—Studies in earth history and paleoecology: American Philosophical Society Transactions, v. 53, pt. 1, p. 22–34.

Brooks, H.C., McIntyre, J.R., and Walker, G.W., 1976, Geology of the Oregon part of the Baker 1° by 2° quadrangle: Oregon Department of Geology and Mineral Industries Geologic Map Series Map GMS–7, scale 1:250,000.

Brown, C.E., and Thayer, T.P., 1966, Geologic map of the Canyon City quadrangle, northeastern Oregon: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–447, scale 1:250,000.

Brown, R.E., and McConiga, M.W., 1960, Some contributions to the stratigraphy and indicated deformation of the Ringold Formation: Northwest Science, v. 34, no. 1, p. 43–54.

Calkins, F.C., 1902, A contribution to the petrography of the John Day Basin: University of California Publications, Department of Geology Bulletin, v. 3, p. 109–172.

Corcoran, R.E., Doak, R.A., Porter, P.W., Prichett, F.I., and Privrasky, N.C., 1962, Geology of the Mitchell Butte quadrangle, Oregon: Oregon Department of Geology and Mineral Industries Geologic Map Series Map GMS–2, scale 1:125,000.

Davenport, R.E., 1971, Geology of the Rattlesnake and older ignimbrites in the Paulina Basin and adjacent area, central Oregon: Corvallis, Oregon State University, Ph.D. dissertation, 132 p.

Dickinson, W.R., and Vigrass, L.W., 1965, Geology of the Suplee-Izee area, Crook, Grant, and Harney Counties, Oregon: Oregon Department of Geology and Mineral Industries Bulletin, v. 58, 109 p.

Downs, Theodore, 1956, The Mascall fauna from the Miocene in Oregon: University of California Publications in Geological Sciences, v. 31, no. 5, p. 199–354.

Enlows, H.E., 1976, Petrography of the Rattlesnake Formation at the type area, central Oregon: Oregon Department of Geology and Mineral Industries Short Paper 25, 34 p.

Enlows, H.E., and Davenport, R.E., 1971, Tertiary ignimbrites in central Oregon [abs.]: Oregon Academy of Science Proceedings, v. 7, no. 6, p. 75.

Evernden, J.F., and James, G.T., 1964, Potassium-argon dates and the Tertiary floras of North America: American Journal of Science, v. 262, no. 8, p. 945–974.

Evernden, J.F., Savage, D.W., Curtis, G.H., and James, G.T., 1964, Potassium-argon dates and the Cenozoic mammalian chronology of North America: American Journal of Science, v. 262, no. 2, p. 145–198.

Farooqui, S.M., Beaulieu, J.D., Bunker, R.C., Stensland, D.E., and Thoms, R.E., 1981, Dalles Group—Neogene formations overlying the Columbia River Basalt Group in north-central Oregon: Oregon Geology, v. 43, no. 10, p. 131–140.

Farooqui, S.M., Bunker, R.C., Thoms, R.E., and Clayton, D.C., 1981, Post-Columbia River Basalt Group stratigraphy and map compilation of the Columbia Plateau, Oregon: Oregon Department of Geology and Mineral Industries Open-File Report O–81–10, 79 p.

Fiebelkorn, R.B., Walker, G.W., MacLeod, N.S., McKee, E.H., and Smith, J.G., 1983, Index to K-Ar determinations for the State of Oregon: Isochron/West, no. 37, 60 p.

Flint, R.F., 1938, Origin of the Cheney-Palouse scabland tract, Washington: Geological Society of America Bulletin, v. 49, no. 3, p. 461–523.

Forth, Michael, 1965, Geology of the southwest quarter of the Dayville quadrangle, Oregon: Corvallis, Oregon State University, M.S. thesis, 75 p.

Gilluly, James, 1937, Geology and mineral resources of the Baker quadrangle, Oregon: U.S. Geological Survey Bulletin 879, 119 p.

Gonthier, J.B., and Harris, D.D., 1977, Water resources of the Umatilla Indian Reservation, Oregon: U.S. Geological Survey Water Resources Investigations Report 77–3, 112 p.

Greene, R.C., Walker, G.W., and Corcoran, R.E., 1971, Geologic map of the Burns quadrangle, Oregon: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–680, scale 1:250,000.

Gude, A.J., III, and Sheppard, R.A., 1978, Chabazite in siliceous tuffs of a Pliocene lacustrine deposit near Durkee, Baker County, Oregon: U.S. Geological Survey Journal of Research, v. 6, no. 4, p. 467–472.

Gunn, B.M., and Watkins, N.D., 1970, Geochemistry of the Steens Mountain basalts, Oregon: Geological Society of America Bulletin, v. 81, no. 5, p. 1497–1515.

Gustafson, E.P., 1978, the vertebrate faunas of the Pliocene Ringold Formation, south-central Washington: Eugene, University of Oregon, Museum of Natural History Bulletin 23, 62 p.

Haddock, G.H., 1966, Vent location by investigation of lateral and vertical characteristics of a welded ash-flow tuff [abs.], *in* Abstracts for 1965: Geological Society of America Special Paper 87, p. 206–207.

———1967, The Dinner Creek Welded Ash-flow Tuff of the Malheur Gorge area, Malheur County, Oregon: Eugene, University of Oregon, Ph.D. dissertation, 111 p.

Hampton, E.R., and Brown, S.G., 1964, Geology and ground-water resources of the upper Grande Ronde River basin, Union County, Oregon: U.S. Geological Survey Water-Supply Paper 1597, 99 p.

Hodge, E.T., 1940, Geology of the Madras quadrangle: Corvallis, Oregon State College Studies in Geology, no. 1, scale 1:25,000.

Hogenson, G.M., 1964, Geology and ground water of the Umatilla River basin, Oregon: U.S. Geological Survey Water-Supply Paper 1620, 162 p.

Kittleman, L.R., Green, A.R., Hagood, A.R., Johnson, A.M., McMurray, J.M., Russell, R.G., and Weeden D.A., 1965, Cenozoic stratigraphy of the Owyhee region, southeastern Oregon: Eugene, University of Oregon, Museum of Natural History Bulletin 1, 45 p.

Ledgerwood, R.K., Myers, C.W., and Cross, R.W., 1978, Pasco Basin stratigraphic nomenclature: Richland, Wash., Rockwell Hanford Operations Report RHO–BW1–LD–L, 23 p.

Lindgren, Waldemar, 1898, The mining districts of the Idaho Basin and the Boise Ridge, Idaho: U.S. Geological Survey Annual Report, 18th, pt. 3, p. 617–719.

Malde, H.E., and Powers, H.A., 1962, Upper Cenozoic stratigraphy of western Snake River Plain, Idaho: Geological Society of America Bulletin, v. 73, no. 10, p. 1197–1219.

Merriam, J.C., 1901, A contribution to the geology of the John Day Basin: University of California Publications, Department of Geology Bulletin, v. 2, p. 269–314.

Mitchell, V.E., and Bennett, E.H., 1979, Geologic map of the Baker quadrangle, Idaho: Idaho Bureau of Mines and Geology Geologic Map Series, scale 1:250,000.

Newcomb, R.C., 1958, Ringold Formation of Pleistocene age in type locality, the White Bluffs, Washington: American Journal of Science, v. 256, no. 5, p. 328–340.

———1965, Geology and ground-water resources of the Walla Walla River basin, Washington-Oregon: Washington Division of Water Resources Water Supply Bulletin, no. 21, 151 p.

———1966, Lithology and eastward extension of The Dalles Formation, Oregon and Washington, *in* Geological Survey research, 1966: U.S. Geological Survey Professional Paper 550-D, p. 59–63.

———1969, Effect of tectonic structure on the occurrence of ground water in the basalt of the Columbia River Group of The Dalles area, Oregon and Washington, 1966: U.S. Geological Survey Professional Paper 383–C, 33 p.

Newcomb, R.C., Strand, J.R., and Frank, F.J., 1972, Geology and ground-water characteristics of the Hanford Reservation of the U.S. Atomic Energy Commission, Washington: U.S. Geological Survey Professional Paper 717, 78 p.

Nockolds, S.R., 1954, Average chemical compositions of some igneous rocks: Geological Society of America Bulletin, v. 65, no. 10, p. 1007–1032.

Oles, K.F., and Enlows, H.E., 1971, Bedrock geology of the Mitchell quadrangle, Wheeler County, Oregon: Oregon Department of Geology and Mineral Industries Bulletin 72, 62 p.

Osawa, Masumi, and Goles, G.G., 1970, Trace-element abundances in Columbia River basalts, *in* Gilmour, E.H., and Stradling, Dale, eds., Columbia River Basalt Symposium 2, Wash. [Procs.]: Cheney, Eastern Washington State College Press, p. 55–71.

Packer, D.R., 1979, Paleomagnetism and age dating of the Ringold Formation and loess deposits in the State of Washington: Oregon Geology, v. 41, no. 8, p. 119–132.

Palmer, A.R., 1983, The Decade of North American Geology 1983 geologic time scale: Geology, v. 11, no. 9, p. 503–504.

Piper, A.M., Robinson, T.W., Jr., and Park, C.F., Jr., 1939, Geology and ground-water resources of the Harney Basin, Oregon: U.S. Geological Survey Water-Supply Paper 841, 189 p.

Prostka, H.J., 1967, Preliminary geologic map of the Durkee quadrangle, Oregon: Oregon Department of Geology and Mineral Industries Geologic Map Series Map GMS–3, scale 1:62,500.

Robyn, T.L., 1977, Geology and petrology of the Strawberry Volcanics, NE Oregon: Eugene, University of Oregon, Ph.D. dissertation, 197 p.

———1979, Miocene volcanism in eastern Oregon—An example of calc-alkaline volcanism unrelated to subduction: Journal of Volcanology and Geothermal Research, v. 5, p. 149–161.

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

Shotwell, J.A., 1956, Hemphillian mammalian assemblage from northeastern Oregon: Geological Society of America Bulletin, v. 67, no. 6, p. 717–738.

———1963, The Juntura Basin—Studies in earth history and paleoecology: American Philosophical Society Transactions, v. 53, pt. 1, 77 p.

Smith, G.A., and Snee, L.W., 1984, Revised stratigraphy of the Deschutes Basin, Oregon—Implications for the Neogene development of the central Oregon Cascades [abs.]: Eos (American Geophysical Union Transactions), [newspaper edition], v. 65, no. 17, p. 330.

Strand, J.R., and Hough, M.J., 1952, Age of the Ringold Formation: Northwest Science, v. 26, no. 4, p. 152–154.

Swanson, D.A., 1969, Reconnaissance geologic map of the east half of the Bend quadrangle, Crook, Wheeler, Jefferson, Wasco, and Deschutes Counties, Oregon: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–568, scale 1:250,000.

Swanson, D.A., Wright, T.L., Hooper, P.R., and Bentley, R.D., 1979, Revisions in stratigraphic nomenclature of the Columbia River Basalt Group: U.S. Geological Survey Bulletin 1457-G, 59 p.

Thayer, T.P., 1957, Some relations of later Tertiary volcanology and structure in eastern Oregon, *in* v. 1 of Vulcanología del Cenozóico: International Geological Congross, 20th, Mexico City, 1956, [Trabajos], sec. 1, p. 231–245.

Thayer, T.P., and Brown, C.E., 1966, Local thickening of basalts and late Tertiary silicic volcanism in the Canyon City quadrangle, northeastern Oregon, *in* Geological Survey research, 1966: U.S. Geological Survey Professional Paper 550–C, p. 73–78.

———1973, Ironside Mountain, Oregon—A late Tertiary volcanic and structural enigma: Geological Society of America Bulletin, v. 84, no. 2, p. 489–498.

Uppuluri, V.R., 1974, Prineville chemical type—A new basalt type in the Columbia River Group: Geological Society of America Bulletin, v. 85, no. 2, p. 1315–1318.

Waitt, R.B., Jr., 1980, About forty last-glacial Lake Missoula jökulhlaups through southern Washington: Journal of Geology, v. 88, no. 6, p. 653–679.

Walker, G.W., 1970a, Cenozoic ash-flow tuffs of Oregon: Ore Bin, v. 32, no. 6, p. 97–115.

———1970b, Some comparisons of basalts of southeast Oregon with those of the Columbia River Group, *in* Gilmour, E.H., and Stradling, Dale, eds., Columbia River Basalt Symposium 2, Wash. [Procs.]: Cheney, Eastern Washington State College Press, p. 223–237.

———1979, Revisions to the Cenozoic stratigraphy of Harney Basin, Southeastern Oregon: U.S. Geological Survey Bulletin 1475, 35 p.

Waters, A.C., 1968, Reconnaissance geologic map of the Madras quadrangle, Jefferson and Wasco Counties, Oregon: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–555, scale 1:125,000.

Wheeler, Greg, 1982, Problems in the regional stratigraphy of the Strawberry Volcanics: Oregon Geology, v. 44, no. 1, p. 3–7.

Williams, Howel, 1957, A geologic map of the Bend quadrangle and a reconnaissance geologic map of the central portion of the High Cascade Mountains: Oregon Department of Geology and Mineral Industries, in cooperation with the U.S. Geological Survey, scales 1:125,000 and 1:250,000.

Wolff, E.N., 1965, The geology of the northern half of the Caviness quadrangle, Oregon: Eugene, University of Oregon, Ph.D. dissertation, 200 p.

6. CENOZOIC TECTONISM AND VOLCANISM OF THE BLUE MOUNTAINS REGION

By GEORGE W. WALKER and PAUL T. ROBINSON[1]

CONTENTS

ABSTRACT

The extensive record of Cenozoic volcanism preserved in the Blue Mountains region of northeastern Oregon and adjacent Washington and Idaho indicates a more or less continuous eruptive history interspersed with a few apparent brief hiatuses and several periods of deformation. The types and volumes of volcanic products varied both geographically and over time. The earliest volcanism, represented mostly by the dominantly Eocene Clarno Formation in Oregon and the partly coeval Challis Volcanics in Idaho, has been attributed by some workers to a west-northwest-trending volcanic arc, commonly identified as the Challis arc. The distribution of the Clarno and coeval rocks in Oregon seem to indicate, however, that they are more likely related to a north-south-trending volcanic arc, the axis of which lay to the east of the present Cascade Range.

Subsequent volcanic activity both in and marginal to the Blue Mountains appears to be related to (1) younger arc volcanism, with the axis of eruptive activity in or near the Cascade axis; (2) extensional tectonics that generated bimodal volcanism along the southern margin of the Blue Mountains province and large volumes of Miocene flood basalt erupted principally from two major dike swarms located within the province; and (3) numerous isolated eruptive centers within the province that probably also are related to extensional tectonism.

INTRODUCTION

Since 1980, several summaries have synthesized the vast amount of data on the distribution, composition, and age of igneous activity in the western Cordillera, as well as on its structural evolution (Hamilton and Myers, 1966; Christiansen and Lipman, 1972; Lipman and others, 1972; Snyder and others, 1976; Armstrong, 1978; Christiansen and McKee, 1978; Smith, 1978). The purpose of this chapter is to consider in more detail the tectonic and volcanic history of the Blue Mountains region of Oregon and contiguous areas of Washington and Idaho. We examine the relations of Cenozoic igneous activity and coeval structural deformation of the region, review evidence for the persistence and continuity of volcanism, discuss the types, ages, and distribution of volcanic products, including volume relations where data permit, and consider how the underlying pre-Cenozoic basement affected the Tertiary volcanism. The preceding chapters in this volume clearly demonstrate that some parts of the Cenozoic column are far better known than others and that our knowledge of the age and distribution of different rock types is uneven.

The Blue Mountains region is a complex uplifted crustal block composed of Cenozoic and older rocks fringed on the north, west, and south by Miocene and younger volcanic rocks and terrestrial sedimentary rocks. The Blue Mountains crustal block has no well-defined eastern topographic or structural boundary and appears to merge into the uplifted Columbia Intermontane province of Idaho.

On the south, the Blue Mountains block is bounded partly by the northwest-trending Brothers fault zone, which separates the Blue Mountains from the High Lava Plains part of the Basin and Range province. The fault zone is expressed at the surface by closely spaced (commonly 0.5–3 km apart) en echelon normal faults of small to moderate displacement, mostly 10 to 100 m. Basaltic and rhyolitic vents of late Miocene through Pleistocene age are abundant within and adjacent to the fault zone. The ages of these vents indicate that recurrent crustal breaking has taken place along the zone since at least Miocene time. The normal faults of the zone and the many volcanic vents are interpreted as the surface manifestations of a large, deeply buried structure. Some vertical offset apparently has occurred along this zone, inasmuch as pre-Cenozoic and early Cenozoic rocks occur at the surface only in the block to the north; the pattern

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of normal faults within and near the fault zone and the relation of many small monoclinal folds to the faults also indicate some lateral displacement. The normal faults denote only adjustment of near surface brittle volcanic and tuffaceous sedimentary rocks to movement on the deeply buried structure. Analysis of surface features permits either right- or left-lateral displacement of the deep-crustal fault, although right-lateral (dextral) movement seems more likely (Walker and others, 1967; Walker, 1969; Lawrence, 1976). The zone is interpreted as a transcurrent-fault zone separating a major block (or blocks) on the south, in which significant east-west extension has occurred, from the Blue Mountains block to the north, which has comparatively less extension and more restricted late Cenozoic volcanism.

The less well defined northern margin of the Blue Mountains block consists of a series of northeast-trending, faulted monoclines and associated anticlines and synclines that extend from near the Deschutes River on the west to the vicinity of Pendleton and Walla Walla on the east (fig. 6.1). In many places, the north boundary is represented by abrupt changes in dip on flows of the Columbia River Basalt Group; west of Lewiston, Idaho, the inclination of flows is generally to the west, although some local flexures disrupt the regional dip. The northeast-trending monoclines and faults are locally intersected by northwest-trending folds and faults, including such enigmatic structures as the Olympic-Wallowa lineament. Most of the northwest-trending faults within the Blue Mountains block exhibit dip-slip movement, but some show strike-slip, mostly dextral, movement.

Bounding structures on the southern and northern margins of the Blue Mountains block disrupt Miocene and older rocks; toward the west end of the Brothers fault zone, on the southern margin of the block, upper Miocene, Pliocene, and, locally, lower Pleistocene rocks also are disrupted.

Along its western margin the block is onlapped by upper Miocene and Pliocene(?) sedimentary rocks, volcaniclastic rocks, and basalt flows of the Madras Formation (the Deschutes Formation of Farooqui and others, 1981) and by overlying Quaternary flows, poorly consolidated sedimentary rocks, and ash. The time of inception of the bounding structures is poorly delimited, but these structures are thought to reflect interactions at the boundary between the North American plate and several crustal plates lying to the west during Miocene and more recent time.

BASEMENT BENEATH THE CENOZOIC ROCKS

The pre-Tertiary basement rocks in the Blue Mountains region vary lithologically and represent both allochthonous and autochthonous terranes (Jones and others, 1977; Brooks and Vallier, 1978; Dickinson, 1979b). The entire region appears to represent a zone, at least 500 km wide, that is transitional between regions underlain by old continental crust on the east and by young oceanic crust on the west. Within this transition zone, several petrologic-tectonic terranes have been recognized, all of which have been partly metamorphosed and locally intruded by Mesozoic granitic plutons. The different terranes are arranged in belts subparallel to a northeast-trending regional gravity gradient (Berg and Thiruvathukal, 1967; Thiruvathukal and others, 1970, fig. 1) and to the trend of the Blue Mountains uplift itself. Data from a single refraction profile extending approximately north-south across this gravity gradient were interpreted by Hill (1972; 1978, p.160) to indicate a change in crustal thickness from approximately 25 km beneath Pasco Basin on the north to about 45 km beneath the Blue Mountains uplift to the south (Thiruvathukal and others, 1970). However, recent detailed seismic-reflection and deep-refraction studies in Pasco Basin indicate a crustal thickness of about 42 km, rather than 25 km (W.D. Mooney, oral commun., 1985).

Initial $^{87}Sr/^{86}Sr$ ratios of Cenozoic rocks increase from about 0.703–0.704 in the Cenozoic volcanic region of east-central Oregon and in the Cascade Range on the west to about 0.706–0.709 in the Challis Volcanics and Absaroka Volcanic Supergroup on the east (Leeman, 1982). Both the Absaroka Volcanic Supergroup and Challis Volcanics are east of the line representing an initial $^{87}Sr/^{86}Sr$ ratio of 0.706, which has been inferred to be related to the edge of Precambrian continental crust in the Western United States (fig. 6.2). The Cenozoic volcanic rocks of the Blue Mountains and coeval volcanic rocks of the Cascade Range in Washington are west of this line and are underlain in part or entirely by accreted terranes and dismembered oceanic crust. Armstrong and others (1977) determined an initial $^{87}Sr/^{86}Sr$ ratio in Mesozoic plutons of the region, and Fleck (1983) recognized an abrupt increase in the $^{87}Sr/^{86}Sr$ ratio from less than 0.704 to greater than 0.710 within a distance of 7 km across the boundary between the Seven Devils terrane (Seven Devils Group) on the west and the Belt/Yellowjacket terrane (Belt Supergroup and Yellowjacket Formation) on the east.

Within this 500-km-wide zone of transition, four major pre-Tertiary petrologic-tectonic terranes have been delineated, all of which may represent allochthonous sequences that have collided with and been accreted to the North American cratonic margin during Mesozoic time (Brooks and Vallier, 1978; Dickinson and Thayer, 1978; Dickinson, 1979b; Silberling and others, 1984). The Cenozoic volcanic and sedimentary pile rests discordantly on these different terranes, which include sedimentary and volcanic rocks of Devonian to Cretaceous (mostly

Permian to Jurassic) age, and on the late Mesozoic granitic plutons that invade them. The terranes were named by Brooks and Vallier (1978) the Wallowa Mountains-Seven Devils Mountains volcanic-arc terrane, dismembered oceanic-crust terrane, Jurassic flysch, and Juniper Mountain-Cuddy Mountain volcanic-arc terrane, and by Dickinson (1979b) the Seven Devils, central melange, Mesozoic clastic, and Huntington terranes. More recently, these sequences were named by Silberling and others (1984) the Wallowa, Baker, Izee, and Olds Ferry terranes.

The relations of these terranes to each other are poorly understood. As summarized by Brooks and Vallier (1978, p. 133), the Wallowa Mountains-Seven Devils Mountains and Juniper Mountain-Cuddy Mountain volcanic-arc terranes "may merge beneath Cenozoic lavas south of the Seven Devils Mountains, or they may be joined along a major suture. The dismembered ocean-crust terrane and the Juniper Mountain-Cuddy Mountain volcanic-arc terrane were joined tectonically in Late Triassic and Early Jurassic time, and flysch derived mainly from erosion of the Juniper Mountain-Cuddy Mountain volcanic-arc terrane was deposited across the junction."

According to Brooks and Vallier (1978), compressional deformation of the flysch and closure of the Jurassic basin probably occurred during the late phases of the accretionary process. Final emergence of the Mesozoic rocks occurred in Late Jurassic or Early Cretaceous time, some time after flysch deposition ceased. Emergence of these terranes provided a source for the upper Albian to Cenomanian epiclastic, commonly conglomeratic sedimentary rocks in the Mitchell and Bernard Ranch areas (fig. 6.1). Recent studies by Thompson and others (1984) of pollen and foraminifers from samples obtained from several wells south of the Mitchell area indicate the presence of Upper Cretaceous (Cenomanian through Maestrichtian) strata in the subsurface. Deposition of these Cretaceous near-shore fluvial and deltaic marine assemblages was followed by further uplift and erosion.

CENOZOIC ERA

Discussion of the Cenozoic tectonic and volcanic history of the Blue Mountains and adjacent areas has generally considered the rocks in major groups defined largely by age. These broad categories generally correspond to the coverage in the preceding chapters: (1) Paleocene(?) and Eocene rocks, (2) Oligocene and lower Miocene rocks, and (3) middle Miocene and younger rocks. Both volcanic and terrestrial sedimentary processes were active throughout the region for most of the Cenozoic, and both were affected by recurrent tectonism. Sites of volcanism and sedimentation have shifted geographically throughout the Cenozoic, in places merging in common depositional basins, elsewhere or at other times being separated by considerable distances.

Volcanism appears to have been more or less continuous within the province, although the types of volcanic products varied both geographically and over time. However, when volumes of erupted material are considered, the volcanism is seen to have been significantly episodic. Major outpourings of calc-alkaline andesitic and basaltic rocks occurred during the period from about 54 to 38 Ma. Chiefly rhyolitic and dacitic rocks were erupted during the period from about 37 to 23 Ma, with principal vents near or to the west of the western margin of the Blue Mountains province. Large volumes of tholeiitic basalt and moderate volumes of calc-alkaline andesitic and rhyolitic rocks were erupted during the period from about 20 to 7 Ma. Within this youngest group, different coeval chemical groups of rocks reflect different tectonic settings. Many of the lower(?) and middle Tertiary volcanic rocks, particularly those of the John Day Formation and upper Miocene ash-flow tuff units, originated from vents outside the Blue Mountains province.

Pliocene and younger volcanism, though widespread in south-central Oregon and in the Cascade Range, has been confined to comparatively small eruptions of basalt along the western and southern margins of the Blue Mountains province (see Swanson, 1969; Brooks and others, 1976). Small to moderate volumes of silicic volcanic rocks were erupted in the Blue Mountains about 12 Ma, and a small dacite intrusion northwest of John Day has been radiometrically dated at 3.1 Ma by Robyn (1977). Young silicic volcanism occurred extensively in the Brothers fault zone about 50 km to the south, as well as in areas still farther south. Among its products are upper Miocene ash-flow tuff deposits of moderately large volume that were erupted from calderas in Harney Basin and extend long distances northward into the Blue Mountains province.

PALEOCENE(?) AND EOCENE

Our knowledge of the earliest Cenozoic history of the Blue Mountains region relies heavily on fragmentary information derived from poorly exposed arkosic, carbonaceous, and conglomeratic sedimentary rocks along the axis of the Blue Mountains uplift. Deeply weathered, poorly exposed, and poorly dated coarse-grained conglomerate in areas adjacent to the Wallowa and Elkhorn Mountains may also be part of the lowermost Cenozoic sequence.

The arkosic sedimentary rocks, which rest discordantly on pre-Cenozoic basement rocks and have been paleobotanically dated as Paleocene(?) age (see chap. 2), consist of well-bedded, moderately well sorted, locally carbonaceous terrestrial sandstone, siltstone, and shale that represent nearshore, partly deltaic deposits. They

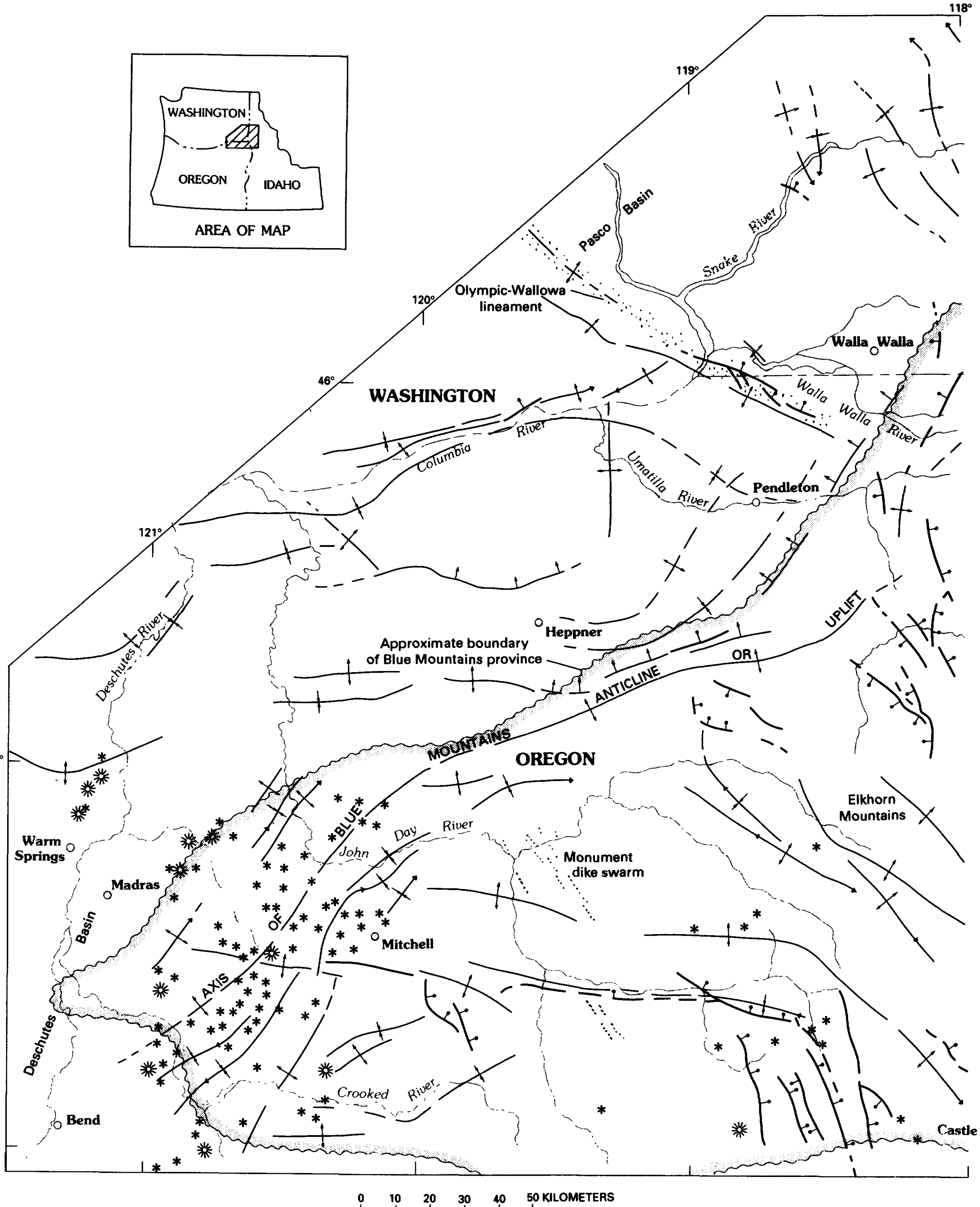
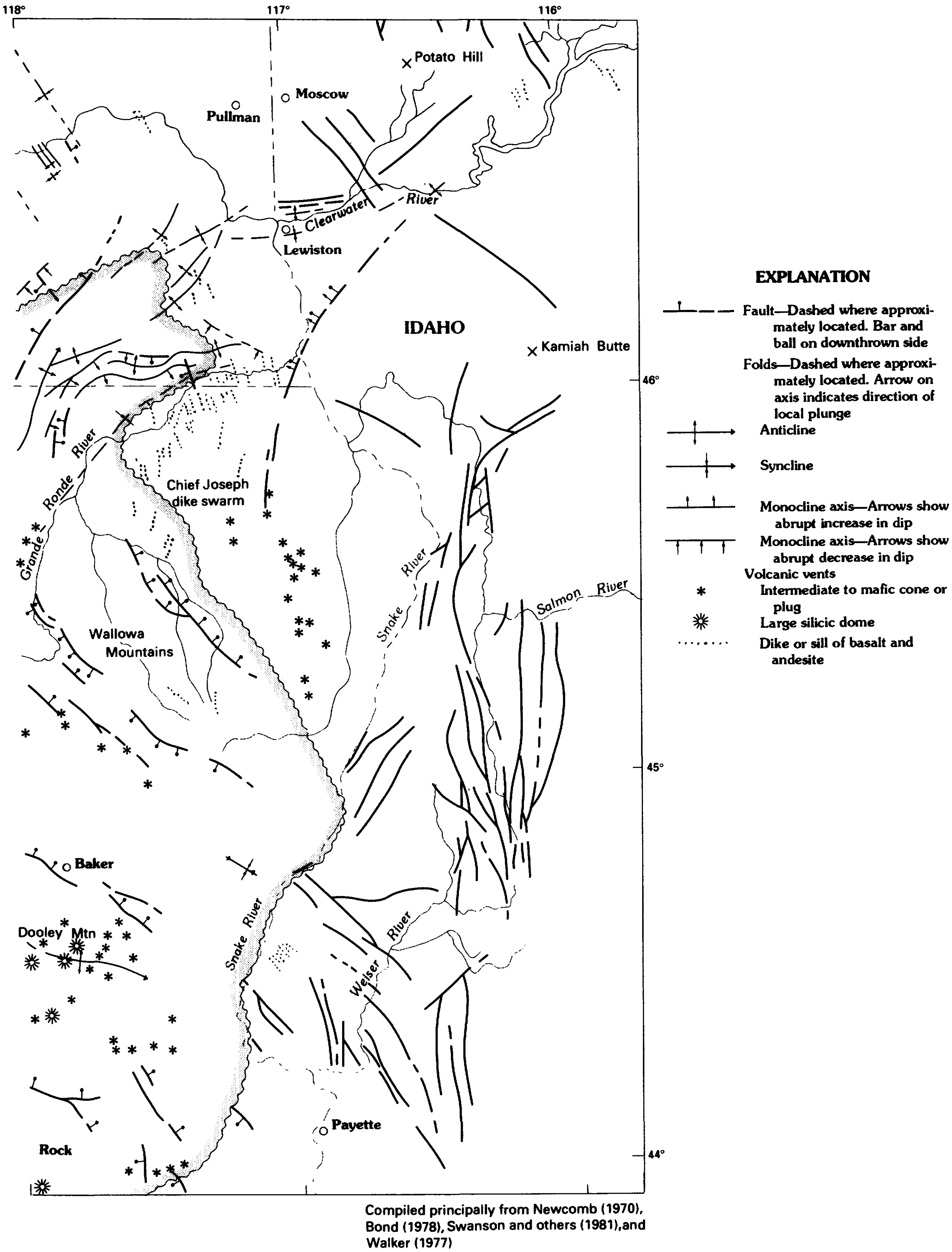


FIGURE 6.1.—Cenozoic tectonic map of the



Blue Mountains province and adjacent areas.

contain detritus apparently derived from a granitic or metamorphic terrane, presumably lying to the east, and little, if any, Cenozoic volcanic debris. Coarse-grained conglomerate is discontinuously exposed in areas east of the arkosic sedimentary rocks and consists largely of rounded cobbles and boulders of quartzite, some metavolcanic rocks (greenstone), and rare granitoid clasts. It does not contain clasts of lower Cenozoic volcanic rocks and is interpreted to be essentially coeval with the arkosic sedimentary rocks and possibly to represent an onshore, coarser grained fluvial facies. All of these terrestrial sedimentary rocks were deposited on an eroded and locally weathered surface that had tens or hundreds of meters of relief. The mineralogy and lithology of the detritus, and its apparent eastward coarsening, indicate probable derivation from outcrops in the vicinity of the Idaho batholith and thus a possible relation to its uplift and subsequent deroofing. Granitic clasts in the Cretaceous (Cenomanian) Bernard Formation (Dickinson and Vigrass, 1965, p. 68) imply that the batholithic masses of northeastern Oregon were initially deroofed during Late Cretaceous rather than Paleocene or early Eocene time. Along the axis of the Blue Mountains uplift, south of Heppner, the arkosic sedimentary rocks are overlain, apparently conformably (Pigg, 1961, p. 16), by porphyritic andesite flows of the Clarno Formation.

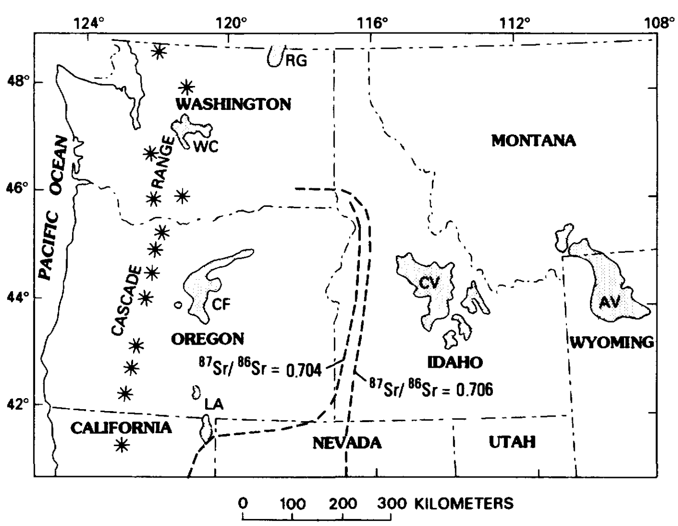


FIGURE 6.2.—Regional distribution of Eocene and lower Oligocene calc-alkaline volcanic rocks in and east of the Cascade Range. Separate volcanic fields include: AV, Absaroka Volcanic Supergroup (Eocene); CF, Clarno Formation (Eocene and lowermost Oligocene?); CV, Challis Volcanics (Eocene); LA, volcanic rocks of Lakeview area, which locally are intruded by 34- to 33-Ma adamellite stocks (Muntzert and Field, 1969; Fiebelkorn and others, 1983) and are unconformably overlain by tuffs and tuffaceous sedimentary rocks correlative with part of the uppermost Eocene(?) to lower Miocene John Day Formation; RG, volcanic rocks of Republic graben; WC, volcanic rocks of Washington Cascades. Lines of $^{87}Sr/^{86}Sr=0.704$ and 0.706 from Leeman (1982).

Cenozoic volcanism in the Blue Mountains commenced earlier than 50 Ma (fig. 6.3) near what is now the axis of Blue Mountains uplift. This earliest known volcanism, represented by altered plagioclase-phyric andesite of the Clarno Formation, is of early Eocene age: 53.7±1.0 Ma (P.T. Robinson and E.H. McKee, quoted in Fiebelkorn and others, 1983). Calc-alkaline andesitic and dacitic volcanism continued mostly near the axis of the uplift and in several areas to the southeast (see fig. 2.1) until latest Eocene time. Radiometric ages on these rocks range from about 54 to 38 Ma, but the age range depends on what rocks are included in the Clarno Formation (see chap. 2; fig. 2.3).

The Challis Volcanics of central Idaho, which are at least partly coeval with the Clarno Formation, were erupted between about 50 and 39 Ma (Armstrong, 1975). They consist mostly of rhyodacitic and rhyolitic ash-flow tuff related to cauldron complexes, but they include flows of dacite, latite, andesite, and basalt. The isolated outcrops of trachytic andesite flows and related breccia at Kamiah Butte and the well-indurated breccia at Potato Hill, both in western Idaho (fig. 6.1), may represent outlying parts of either the Challis Volcanics or the Clarno Formation, or they may be unrelated. Partly coeval volcanism, also mostly andesitic but including basaltic, dacitic, and rhyodacitic material, was occurring in other adjacent regions (fig. 6.2), for example, to the north in and adjacent to the Republic graben near the United States-Canada border (Muessig, 1962; Armstrong, 1978; Ewing, 1980), to the south near Lakeview, Oreg. (Walker, 1980), and in the Cedarville area, Calif., as well as in the Cascade Range in Washington (fig. 6.2; Vance, 1982; Ort and others, 1983; Tabor and others, 1984). Basaltic and andesitic volcanism that is partly coeval also was occurring in areas west of the Cascade Range, as represented by mafic flows and breccia of the upper Eocene and lower Oligocene Goble Volcanics (Livingston, 1966; Beck and Burr, 1979), upper Eocene basalt and basaltic andesite of the Tillamook Volcanics (Wells and others, 1983), and, possibly, parts of the Eocene to Miocene(?) Skamania Volcanic Series of Trimble (1963), which strata Wise (1970) included in his Ohanapecosh Formation, and the volcanic parts of the upper Eocene and Oligocene Fisher Formation.

Several workers (Snyder and others, 1976; Armstrong, 1978; Vance, 1982) have suggested that the Clarno, Challis, and Absaroka volcanic rocks are related to an extensive magmatic arc, generally referred to as the Challis arc, that extended westward from Wyoming and Montana, through east-central Oregon, and northwestward to west-central Washington. Magma for the Challis Volcanics and Absaroka Volcanic Supergroup appears to

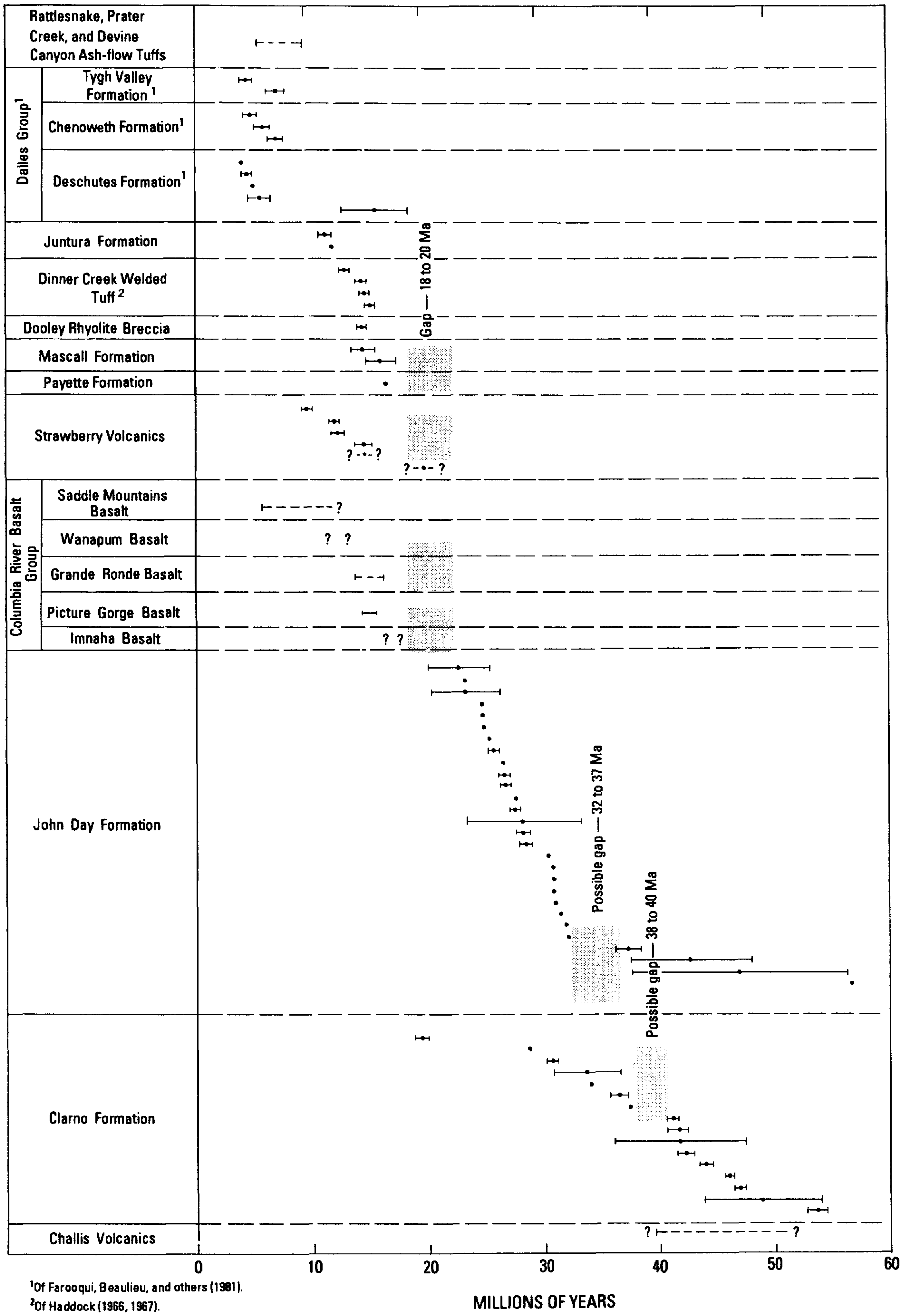


FIGURE 6.3.—K-Ar ages of volcanic and volcaniclastic rocks, showing apparent gaps in volcanic activity (stippling). Within formations, dated samples are not necessarily arranged according to stratigraphic position. Dots, calculated age; bars, analytical error; dashed lines, approximate K-Ar age range; ?, questionable age limits.

have originated in old continental lithosphere, whereas magma for the Clarno Formation and, probably, for the Cascade Range originated in oceanic lithosphere that had previously been somewhat dismembered during amalgamation and accretion at the edge of the continent. Noblett (1981) considered rocks of the Clarno Formation to be a product of subduction-related volcanism along a continental margin and to be a late phase of Challis arc volcanism. His model postulates that the Challis arc migrated westward during Eocene time from Idaho and Montana, across central Oregon, to and through the locus of Clarno volcanism. However, the similar ages of volcanism in these widely separated areas, as established by radiometric-age determinations, seems to refute the concept of time-transgressive Eocene arc volcanism. Also, the partly coeval calc-alkaline andesitic to rhyolitic volcanic rocks in the Paisley Hills and the Lakeview-Cedarville area, 150 km south of the Blue Mountains near the Oregon-California State line, and those in western Oregon and Washington do not fit the geometry of the proposed Challis arc, although they may constitute completely separate volcanic piles of different origin. Rogers and Novitsky-Evans (1977) considered the calc-alkaline basaltic andesite and andesite of the Clarno Formation to be a continental-margin assemblage formed above a subduction zone.

We might speculate that some of the outcrops in the Eocene volcanic belt have been brought fortuitously into position by major clockwise rotation during post-Eocene time. As much as 70° of rotation has been documented for rocks of the Coast Range of Oregon and Washington (Simpson and Cox, 1977; Beck and Plumley, 1980; Magill and Cox, 1980). These rocks could also have migrated a significant distance to the northeast from their original site of deposition. The rotation of the coeval rocks of the Clarno Formation apparently is considerably less, possibly no more than $17° \pm 10°$ (Beck and others, 1978; C.S. Grommé, oral commun., 1984). This relatively small amount of rotation should have had little affect on the geographic position of the Clarno Formation. Even where rotation of crustal blocks has been significant, there may be little or no change in geographic position of the rocks if the rotation is confined largely to microblocks within a zone of transition between major crustal plates.

Any model for the origin of the various Eocene calc-alkaline rocks of the northwestern United States must take into account the generally synchronous occurrence of the volcanism, its chemical similarities and differences, and, most importantly, the geometry of possible volcanic belts and their relation to the position and orientation of the continental margin. The affects of rotation and changes in the geographic position of crustal blocks along the western margin of the region must also be considered. We believe that the available data argue against the proposed west- to northwest-trending Challis arc. If such an arc existed, the distribution of the Eocene rocks virtually mandates a fore-arc trench and related subduction zone parallel to the arc on the south or southwest side. Such a model would require essentially north-south convergence of the North American plate and an oceanic plate in the Pacific over a front more than 1,000 km long in Eocene time. The orientation of this proposed arc is at a large angle to the continental margin, and so the trench and subduction zone would also have been nearly at right angles to the present margin. This requirement alone seems to refute the concept of a trench, subduction zone, and related island or magmatic arc all trending west-northwestward to westward across the region. Dickinson (1979a, fig. 1), Engebretson (1982), and Wells and others (1984) inferred that during Eocene time the direction of convergence was more northerly than in either Cretaceous or Oligocene time, but their convergence directions are still at a small angle to the postulated west-northwestward to westward trend of arc and trench.

Taylor (1977) considered that the Clarno volcanic rocks of east-central Oregon reflect a northeast-southwest andesite belt that at the time of formation was undergoing northwest-southeast compression. Such a belt trending northeast-southwest is necessary if the Eocene andesites of the Lakeview-Cedarville area are related to those of the Clarno Formation. This model requires a fore-arc trench approximately 100 km to the northwest of the arc, with both arc and trench trending generally parallel to the present continental margin. Plate convergence could be essentially east-west, in accord with convergence directions as postulated by Dickinson (1979a). This model would almost certainly exclude a genetic connection along a single subduction zone between the Clarno Formation and coeval volcanic rocks to the east in Idaho, Montana, and Wyoming; it instead implies different origins for these similar and coeval volcanic assemblages.

From our evaluation of the data, we conclude that the Eocene volcanism of the Blue Mountains region and adjacent areas can best be explained by eruptions along a north- to northeast-trending arc that was situated above a subduction zone oriented nearly parallel to the present continental margin. The Eocene andesites of the Lakeview-Cedarville area near the California-Oregon State line are inferred to represent the southward extension of this belt, although some left-lateral offset along such features as the Brothers fault zone appears necessary to bring them into line with coeval rocks in the Blue Mountains. Alternatively, the geometry of a trench-arc assemblage may have been modified by later tectonic dismembering, particularly in the region south of the volcanic belt in the zone of significant crustal extension.

OLIGOCENE AND EARLY MIOCENE

Oligocene and lower Miocene sequences in the Blue Mountains region are chiefly intermediate to silicic air-fall tuff, ash-flow tuff, and tuffaceous sedimentary rocks mostly representing part of the John Day Formation. Locally, flows of alkali olivine basalt and trachyandesite are interlayered with the pyroclastic deposits, and rhyolite flows and domes are common along the western margin of the Blue Mountains province. Included in this assemblage are several local piles of plagioclase-, pyroxene-, and olivine-phyric andesitic rocks, some of which have been previously assigned to the Clarno Formation (see chaps. 2, 3). However, many of these rocks are less altered than typical Clarno Formation rocks and have radiometric ages equivalent to those of the John Day Formation.

Isotopic ages of the Oligocene and lower Miocene rocks range from about 37 to 20 Ma (fig. 6.3). Vertebrate fossils from the uppermost part of the John Day Formation in the Mutton Mountains area, about 10 km north of Warm Springs, are Hemingfordian in age (Woodburne and Robinson, 1977), probably about 20 to 19 Ma. Andesitic rocks that predate the Columbia River Basalt Group and are radiometrically dated at 20 to 19 Ma are included in the section of Miocene and younger rocks.

Volcanism in the Blue Mountains region during Oligocene and early Miocene time appears to have been much more restricted geographically than Eocene volcanism. The great bulk of the pyroclastic material making up the John Day Formation apparently was derived from vents west of the Blue Mountains province, in or near the present Cascade Range (Robinson and others, 1984). Within the Blue Mountains province itself, small flows of alkali olivine basalt and trachyandesite (minimum 30 Ma) were erupted from several vents, and more voluminous flows and domes of rhyolite were emplaced, particularly along the western margin of the region. The local andesite flows and breccia of John Day age, which are lithologically similar to Clarno sequences, may simply represent a waning phase of the Eocene volcanic activity. However, some of these rocks are as young as 28 to 27 Ma and thus are 9 to 10 m.y. younger than the 37-Ma base of the John Day Formation, as established by Swanson and Robinson (1968). Very few vents have been identified for this andesitic volcanism, but they apparently lie chiefly in areas to the east of the silicic vents.

Distribution of the Oligocene and lower Miocene rocks demonstrates that early Cenozoic structural trends, particularly the Blue Mountains uplift, developed before their eruption and deposition; subsequent deformation followed these earlier Cenozoic structural trends. An episode of regional uplift and consequent erosion in latest Eocene time produced an erosion surface on the Clarno Formation with several hundred meters of relief. In many areas, a red saprolite formed on this surface, and this paleosol has been used where present to mark the boundary between the Clarno and John Day Formation (Waters, 1954). The uplift was concentrated along pre-existing structural trends and apparently did not affect the bounding structures on the northern and southern margins of the Blue Mountains uplift.

Fisher (1967, p. 120) recognized both northeast- and northwest-trending folds within the Blue Mountains uplift that are either pre-Oligocene or earliest Oligocene in age. The northeast-trending folds are dominant and are attributed to northwest-southeast compression (Taylor, 1981, p. 112). Much of this deformation apparently occurred during Eocene time; in the area between Madras and Mitchell, Oreg., however, it partly postdates eruption of the basal ash-flow sheet of the John Day Formation, which has been dated at about 37 Ma (Swanson and Robinson, 1968). The distribution and thickness variation of the John Day Formation and its stratigraphic relations to the underlying Clarno Formation indicate that a major topographic barrier existed along the Blue Mountains axis throughout most of John Day time.

Folds that developed during the Oligocene and early Miocene mostly trend northeast, parallel to the older structures. These northeast-trending folds also are inferred to indicate northwest-southeast compression, although the stress regime appears to have weakened after Eocene deformation.

The initiation of John Day volcanism about 37 Ma signified a major change in the composition and locus of volcanic activity in the region. Shortly after cessation of Clarno volcanism in the Blue Mountains, volcanos in the vicinity of the present-day Cascade Range began erupting andesitic, dacitic, and rhyolitic pyroclastic material, which was deposited in the John Day area as ash falls. Simultaneously, vents between the Cascade Range and the Blue Mountains uplift were erupting rhyolitic ash flows, lava flows, and minor air-fall pyroclastic material. Most of the ash-flow tuff was derived from vents west of the present-day outcrops in the Blue Mountains, whereas silicic lava flows and domes represent local eruptions within the western part of the Blue Mountains region. Other vents, mostly east of the major rhyolitic volcanos, erupted alkaline basalt and trachyandesite. Partly coeval dacitic and rhyolitic rocks indicate that volcanism was occurring in several parts of the Basin and Range many tens of kilometers to the south near Lakeview, in the Coyote and Rabbit Hills and at Hart Mountain, about 50 km northeast of Lakeview, and in the Steens Mountains, 80 km northeast of Lakeview.

This pattern of volcanism continued until approximately 25 to 20 Ma, when rhyolitic, basaltic, and trachyandesitic eruptions within the Blue Mountains ceased.

Andesitic, dacitic, and rhyolitic eruptions continued in the Cascade Range; air-fall tuff from this volcanism forms the upper part of the John Day Formation. Deposition of the John Day Formation ceased at about 20 to 18 Ma, coincident with the start of a probable hiatus in Cascade Range volcanism (McBirney and others, 1974); after a short period of folding and erosion, the Columbia River Basalt Group was erupted, beginning about 17 Ma (see chap. 4; Watkins and Baksi, 1974; McKee and others, 1977).

The westward shift in the locus of volcanic activity from Clarno to John Day time is interpreted as the result of a westward migration and possible reorientation of the early Tertiary subduction zone along the continental margin. The postulated north- to northeast-trending volcanic arc of Eocene time apparently passed through the western part of the Blue Mountains. In latest Eocene time activity along this trend diminished significantly, and a new arc was developed in the vicinity of the present-day Cascade Range, possibly as early as 41 Ma (Lux, 1982).

MIOCENE TO HOLOCENE

The compressional tectonism of Mesozoic and early Cenozoic time was replaced in middle Miocene time by mostly extensional tectonism, with only local compression. This change in tectonic style, presumably related to the development of transform-fault systems along parts of western North America, significantly altered the character of volcanism both within the Blue Mountains region and in areas to the south. The Brothers fault zone (fig. 6.4), lying between the Blue Mountains province and the Basin and Range province to the south, has been interpreted as a transcurrent- (transform?-) fault zone that separates crustal blocks characterized by appreciable differences in the amount of extension, in structural pattern, and in type of volcanism. Some investigators also have speculated that the Olympic-Wallowa lineament north of the Blue Mountains (fig. 6.4) is the surface manifestation of another, older transcurrent-fault zone, separating continental crust on the north from dismembered oceanic crust on the south.

Stewart (1978) summarized current hypotheses on the origin of this Neogene extensional stress regime and its relation to regional tectonics, particularly as applied to the Basin and Range. One hypothesis relates to "oblique tensional fragmentation within a broad belt of right-lateral movement and distributed extension along the west side of the North American plate," another to "spreading caused by upwelling from the mantle behind an active subduction zone (back-arc spreading)," and a third to presumed subduction of the East Pacific Rise beneath part of North America. Eaton and others (1978, p. 86) explained the extensional tectonics in the western Cordillera by "the rise, and probably the lateral divergent flow, of hot asthenospheric material accompanying major mantle upwelling." This model, although it accounts for many of the phenomena that characterize the Basin and Range province, such as normal faults, high heat flow, relatively thin crust, and low P_n seismic velocity, is inconsistent with apparent north-south compressional features north of the Brothers fault zone and with the apparent thickening of crust from the Columbia Plateau and Pasco Basin southward beneath the Blue Mountains (Hill, 1978).

By middle Miocene time, calc-alkaline volcanism in the Cascade Range was well established. The presence of this persistent, mostly andesitic and basaltic volcanic activity, along a belt several tens of kilometers wide and more than 1,000 km in length from northern California to the United States-Canada border, is persuasive evidence for the existence of a subduction zone along western North America. The Blue Mountains region is geographically behind (east of) this magmatic arc, and the pattern of Miocene and younger deformation in and adjacent to

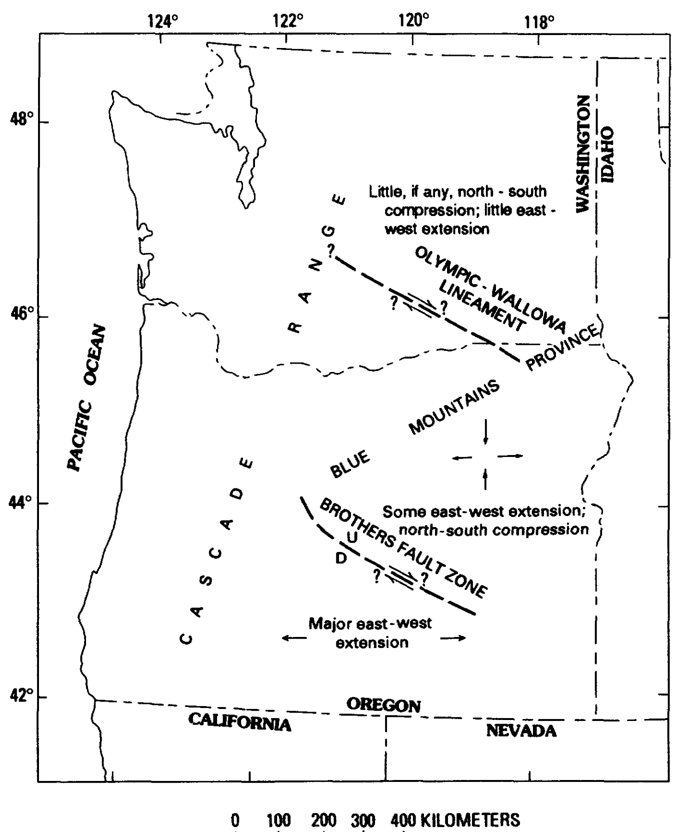


FIGURE 6.4.—Geographic relations of the Blue Mountains province to the Brothers fault zone, Olympic-Wallowa lineament, and Cascade Range. Queried arrows on lineament and fault zone, uncertain direction of movement. D, downthrown side; U, upthrown side.

this region appears to be most consistent with a combination of back-arc spreading and right-lateral movement of crustal blocks in the transition zone between the Pacific and North American plates. Transcurrent movement along transform zones was probably responsible for creating local compressional stresses.

Major crustal extension commenced at about 17 to 16 Ma and has continued throughout the late Cenozoic. Some deformational evidence, as well as fault-plane solutions of recent seismic events, indicates that the regional stress regime responsible for this warping and block faulting still persists (Couch and Lowell, 1971). The Blue Mountains province was less affected by this extension than were areas in the Basin and Range province to the south; the consequent differential extension was apparently accommodated along the Brothers fault zone. Faults and folds within the Blue Mountains generally parallel preexisting northeast-trending structures, but northwest-, northeast-, and east-trending structures are also present. Thayer (1957) concluded that structures in the John Day region resulted principally from north-south compression and east-west extension.

Robyn and Hoover (1982) identified three separate types and ages of Cenozoic deformational structures: (1) northeast- and east-trending fold axes and reverse faults generated during the period 36 to 17 Ma; (2) north-northwest-trending fissures and north-trending strike-slip faults generated from about 17 to 10 Ma; and (3) northwest- and northeast-trending conjugate faults that formed from 10 Ma to the present. Robyn and Hoover (1982) related all of these structures to interaction between the northwestward-moving lithosphere of the Basin and Range and a relatively immobile block of lithosphere beneath the Blue Mountains. According to this interpretation, many of the Miocene and younger faults and folds can be best explained by adjustment of brittle surface and near-surface layers of volcanic and tuffaceous sedimentary rocks to transcurrent movement along bounding faults.

In the northernmost Basin and Range, south of the Brothers fault zone, major east-west extension was accompanied by voluminous basaltic and rhyolitic (bimodal) volcanism from the middle to late Pliocene and, locally, into the Pleistocene. The basalts are dominantly olivine tholeiite or plagioclase-rich high-alumina types. Vents for the mafic rocks include numerous small lava cones or shields, cinder cones, and localized dike swarms; numerous domes and intrusions and several calderas mark the sites of eruption for the silicic rocks.

In the Blue Mountains province north of the Brothers fault zone, where east-west extension was much less, Miocene volcanism produced voluminous tholeiitic flood basalts, moderate volumes of calc-alkaline andesitic to rhyolitic lava, and indeterminate volumes of rhyolitic lava, tuff, and tuffaceous sedimentary rocks. Younger, upper Cenozoic volcanic deposits similar to those in the northern Basin and Range are absent. According to Robyn (1979) the high-alumina basalt, tholeiite, and olivine tholeiite of eastern Oregon show a southward decrease of normative quartz similar to compositional variations observed across linear volcanic belts associated with subduction. Much of the Miocene silicic ejecta was incorporated into bedded pumiceous and ashy sedimentary rocks and interlayered ash-flow tuff; lesser amounts occur in domal masses, such as those at Dooley Mountain and the rhyolitic complexes within the Strawberry Volcanics. The total original volume of this rhyolitic to dacitic material was very large, possibly nearly comparable to that of the coeval flood basalts, if the thickness estimates by Thayer and Brown (1966) are accurate. Vents for some of the silicic rocks within the Blue Mountains province, including those for middle Miocene ash-flow tuff, apparently are buried beneath rhyolitic complexes like that at Dooley Mountain, 20 km south of Baker; others are represented by such intrusions as those at Castle Rock and at Black Butte, a few kilometers northwest of Castle Rock. Most of the Miocene andesitic rocks are platy and contain phenocrysts of augite and hypersthene; some of the more mafic flows associated with the andesites are olivine bearing. Hornblende-porphyritic andesite flows and breccia of early to middle Miocene age (approx. 20–19 Ma) occur in several ranges north and east of Unity Basin; commonly, these rocks have been misidentified as representing part of the Eocene and Oligocene(?) Clarno Formation.

Along the northern margins of the Blue Mountains and in areas to the north, the amount of east-west extension appears to be even less than in areas to the south, and Miocene volcanism was largely basaltic. Tholeiitic flood basalt of the Columbia River Basalt Group is localized mostly north of the axis of the Blue Mountains uplift, although flow sequences extend in places far south of the axis. All major vent systems for the voluminous flows of the Columbia River Basalt Group apparently lie within the area underlain by dismembered oceanic crust and accreted terranes rather than continental crust, as indicated by Thompson (1977). The feeder dikes for these flows are principally in two major dike swarms (fig. 6.1). The northwest-trending Monument swarm in John Day Basin was the main source for flows of the Picture Gorge Basalt, and the north- to northwest-trending Chief Joseph swarm, exposed mainly in drainages tributary to the Grande Ronde River, was the main source of flows of the Grande Ronde, Wanapum, and Saddle Mountains Basalts. Isolated small groups of basalt feeder dikes also have been identified in many other parts of the Blue Mountains, in several widely separated areas in western Idaho (fig. 6.1), and along the Snake River in Washing-

ton. Vents for the Prineville type of flood basalt and for plagioclase-rich, ophitic basalt flows along the southern margin of the Blue Mountains uplift have not yet been recognized and presumably are buried beneath younger rocks in Deschutes Basin or, possibly, in or near the Brothers fault zone.

Robyn (1979) recognized three centers of Miocene calc-alkaline hypersthene andesite volcanism in the Blue Mountains (fig. 6.5; Sawtooth Crater, Strawberry Volcano, and Dry Mountain), all in a straight line or belt with a northeastward trend and with 100-m spacing between centers. A fourth center, also composed largely of middle to upper Miocene hypersthene andesite, has been recognized at Gearhart Mountain, lying precisely on the southwestward extension of this linear belt at approximately the same spacing as between the other centers (fig. 6.5). This uncommonly straight, narrow linear belt or line of volcanos, all of middle or late Miocene age, does not appear to coincide with any single throughgoing structure or group of structures recognizable in surface exposures. Robyn (1979) suggested that this line of volcanos is situated at the intersection of two geologic provinces and generally parallels and lies to the south of the northeast-trending gravity gradient (Thiruvathukal and others, 1970; Hildenbrand and others, 1982). Therefore, the belt appears to be underlain by crust that is somewhat thicker than that in areas to the west and northwest. Whether this Miocene hypersthene andesite belt extends farther southwestward into northern California remains to be ascertained. The possible effect of the Brothers fault zone, which appears to transect the belt, must also be considered. Robyn (1979) noted that the oblique relation of this linear trend to Miocene andesitic centers in the western Cascades is similar to that of the volcanos adjacent to the Ryukyu and Izu Trenches in the northwestern Pacific. Differences in the structures and volcanic features of these widely separated areas indicated to Robyn (1979, p. 159), however, that the Miocene andesite volcanos of eastern Oregon, unlike those of the modern northwestern Pacific, are not subduction related. He proposed, instead, that the Oregon lava was generated by small degrees of melting to form high-alumina basalt magma which subsequently differentiated to form the calc-alkaline andesite and associated dacite. Their relation to tectonic events is unclear, although Robyn (1979, p. 159) inferred that the andesite and related rocks were generated at a major crustal boundary.

Basaltic flows and flow breccia of Pliocene age are present in a few restricted areas in northern Malheur County and adjacent parts of Baker County, Oreg., on the northwestward projection of the Snake River graben, and in and adjacent to Deschutes Basin. The precise age of these rocks is not well documented, but they generally rest on or are interbedded with sedimentary beds containing Pliocene vertebrate faunas. Most of the flows show little evidence of deformation. They consist mostly of thin, open-textured (diktytaxitic), commonly plagioclase-rich olivine basalt and some basaltic andesite, and many of them can be related to small, well-defined shield volcanos or cones of mafic cinders and agglutinate. In the western part of the region, near the boundary between the Blue Mountains province and the Deschutes-Umatilla Plateau province (see fig. 1.1), upper Pliocene or Pleistocene flows of diktytaxitic olivine basalt overlie the Deschutes Formation (Swanson, 1969; Robinson, 1975; Peterson and others, 1976; Robinson and Stensland, 1979). Many of these flows form intracanyon fill in the deeply incised canyons of the Deschutes and Crooked Rivers. Some of these flows and flow breccias of small to moderate volume appear to be the terminal phases of volcanism in the Blue Mountains province, although calc-alkaline basaltic, andesitic, and, locally, dacitic and rhyolitic volcanism persisted into the Holocene in areas of the northwestern Basin and Range to the south of the Blue Mountains and in the Cascade Range to the west.

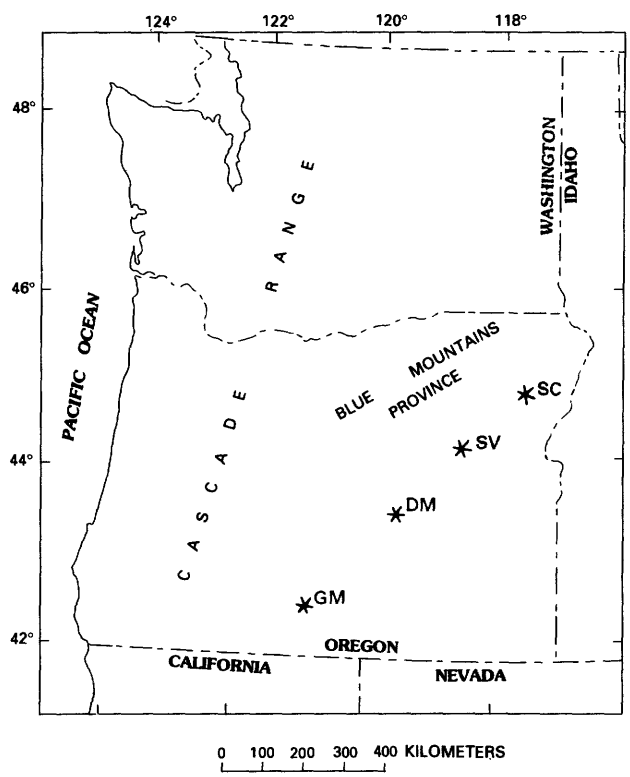


FIGURE 6.5.—Middle and upper Miocene volcanic centers of eastern Oregon that are partly characterized by hypersthene andesite. DM, Dry Mountain; GM, Gearhart Mountain; SC, Sawtooth Crater; SV, Strawberry Volcanics. Modified from Robyn (1979, fig. 1).

SUMMARY AND CONCLUSIONS

Proposed models for the Cenozoic volcanism and tectonism in the northwestern United States are made uncertain by the paucity of information concerning the gross structure, thickness, and lithology of the crust beneath the Cenozoic cover, as well as data on the age, geographic distribution, volume, and geochemistry of the Cenozoic units that make up that cover. The Blue Mountains province and adjacent parts of Washington and Idaho appear to lie largely, if not entirely, to the west of the line marking an initial $^{87}Sr/^{86}Sr$ ratio of 0.706, implying that old (Precambrian) crust is absent in the subsurface. Isolated surface exposures indicate that basement to the Cenozoic rocks consists mostly of accreted terranes of Devonian through Jurassic age comprising marine sedimentary and volcanic island-arc sequences, ophiolite assemblages, granitic plutons, and upper Mesozoic marine sedimentary rocks, all interpreted to rest on dismembered oceanic crust. Within the Blue Mountains region, deformation and uplift of these older rocks was followed in earliest Tertiary time by volcanism and terrestrial sedimentation that have persisted throughout the Cenozoic.

Few reliable data exist on the ages of initial structural uplift of the Blue Mountains province or development of the Cenozoic fold and fault structures. The stress regimes that initiated uplift of the Blue Mountains and created fold and fault structures within the region presumably result from interactions between crustal blocks in a zone of deformation 1,000 to 1,500 km wide that is related to oblique convergence of the Pacific and North American plates; near-surface, local stresses also developed within and between these crustal blocks. Paleomagnetic data indicate that crustal blocks in this wide zone were rotated clockwise during the Cenozoic and that the greatest rotation occurred on blocks west of the Blue Mountains in the Cascade and Coast Ranges of Oregon and parts of Washington. The Blue Mountains province is characterized by regional uplift and is in part bounded by linear structures that at least locally exhibit transcurrent movement and may represent transform zones. The region has been subjected to several compressional regimes of different orientations and has undergone much less extension than the northern Basin and Range province. Apparent gaps in the Tertiary volcanic record appear to coincide with changes in tectonic style, such as that from mostly compressional to mostly extensional stress regimes during early to middle Miocene time; however, dating of many rock units of the region is not adequate to support definite models for episodic volcanism. Furthermore, the absence of reliable volume estimates for most of these intermediate to silicic volcanic rocks prevents valid comparisons of the magnitudes of different kinds of volcanism by age and their significance in petrogenetic modeling.

The chemical composition of several Cenozoic sequences of the Blue Mountains province, including rocks of the Clarno Formation, flows of the Columbia River Basalt Group, and a geographically restricted part of the Strawberry Volcanics, is reasonably well established, but little is known of the Oligocene andesites that are coeval with part of the John Day Formation or of the silicic volcanic rocks of middle and late Miocene age. Estimates of the relative volumes of the different chemical types vary in their precision among the different age groups. Virtually nothing is known of the volume of the Paleocene(?) terrestrial sedimentary rocks. In addition, although the volume of Eocene calc-alkaline volcanic rocks can be approximated from their areal distribution and approximate thickness, the extent of these rocks beneath younger cover and the proportions of intermediate to silicic rock types on a regional basis are largely unknown. From the distribution of surface exposures and the thicknesses of measured sections, we estimate the total volume of these early Cenozoic calc-alkaline volcanic and volcaniclastic rocks to be in the range 10,000 to 20,000 km^3. Oligocene and early Miocene volcanism was largely rhyolitic and was confined to the western part of the province, and most of the vents for these rocks are represented by large domal complexes. Most of the andesitic and dacitic pyroclastic material of the John Day Formation was erupted from vents west of the Blue Mountains province, probably from the site of the present Cascade Range. Some coeval andesitic volcanism occurred in the central and eastern parts of the province, but there are few data to establish its precise regional distribution or the volume of products. The distribution of Oligocene and lower Miocene rocks and random measurements of their sections indicate a total volume of intermediate and silicic volcanic and volcaniclastic rocks well in excess of 5,000 km^3; much of this volume originated in vents along the western margin of the Blue Mountains or in the Cascade Range.

Miocene volcanism was apparently more diverse. Tens of thousands of cubic kilometers of tholeiitic basalt was erupted mostly from fissure-dike swarms near the axis of the Blue Mountains uplift, possibly equal volumes of bimodal basalt and rhyolite were erupted along the southern margin of the Blue Mountains and to the south in the northern Basin and Range, and somewhat lesser volumes of calc-alkaline andesite, dacite, and rhyolite were erupted in the southeastern and eastern parts of the province. The original volume of Miocene rhyolitic flows, breccia, domes, and ash-flow and air-fall tuff erupted in the region is virtually impossible to calculate: (1) Great volumes of volcaniclastic material have been removed by erosion, and some areas were completely stripped of

these rocks; (2) some, possibly much, of the silicic clastic material may have been derived from erosion of rocks of the Clarno and John Day Formations or introduced as airborne material derived from volcanos in the Cascades to the west; and (3) thickness estimates for this part of the section differ significantly among different investigators. The volume of silicic volcanic material was large, certainly totaling many thousands of cubic kilometers.

In spite of these major gaps in data, many geologists have speculated on the relation between tectonism and volcanism in this region, on the sequential structural development of the region within the framework of plate tectonics, and on the genesis of the different volcanic products. We conclude that conceptual models relating the Eocene Blue Mountains volcanism to widespread coeval activity to the east along a northwest-trending Challis arc need to be modified because they appear to conflict with currently available data. The overall distribution of these Eocene calc-alkaline rocks and the geometric relations of such a postulated arc to the continental margin and to presently accepted plate-convergence directions are all inconsistent with such a model. We propose, instead, that the Clarno volcanism is related to a north- or northeast-trending arc passing through the Blue Mountains and that the Absaroka Volcanic Supergroup and Challis Volcanics, lying tens to hundreds of kilometers to the east, resulted from entirely different processes. We also suggest that the Clarno arc persisted until the earliest Oligocene(?), when the locus of volcanic activity shifted westward to the site of the present Cascade Range. Many problems concerning the relations of Cenozoic volcanism to tectonism in the Blue Mountains region are beyond analysis using available data. Some of the more critical questions that require additional information for comprehensive study and review include the following:

(1) What are the thickness and composition of the crust beneath the Blue Mountains and adjacent areas? Seismic and gravity data indicate that it thickens southward from Pasco Basin toward the axial region of the Blue Mountains uplift. Does the thickened crust beneath the uplifted block result from stacking of accreted terranes, from incorporation of material other than dismembered oceanic crust, or from other causes? Does it thin appreciably south of the Brothers fault zone in the northern Basin and Range as a result of extension and stretching?

(2) What role has been played by tectonic rotation of blocks in the region? Paleomagnetic data imply various amounts of rotation of crustal blocks, especially before middle Miocene time, but the amounts of rotation and their geographic extent are still poorly documented. The impact of this apparent rotation on regional tectonic analysis is uncertain, and we do not even know if any of the earliest Cenozoic terranes originated in other than their present position. How far have these terranes moved, if at all, or have they rotated approximately in place? What were their original geometric relations to early Cascade volcanism?

(3) What are the age and origin of the enigmatic linear structures, such as the Brothers fault zone and the Olympic-Wallowa lineament, and how do they relate to the development of the uplifted Blue Mountains block? The Olympic-Wallowa lineament is traceable in surface physiographic features for many tens of kilometers, and it locally disrupts middle Miocene rocks. These surface features are generally considered to represent only minor adjustments of surficial rocks that overlie a deeply buried structure. Some post-Miocene movement is implied by deformation of Miocene basalt flows (Tabor and others, 1984, p. 43). Surface structures along the Brothers fault zone disrupt rocks of middle Miocene (approx. 15-Ma) to Pleistocene (approx. 1-Ma) age. There are virtually no data, however, to indicate when this zone originated or whether it overlies an older, deeply buried crustal structure. Are these linear structures transform zones separating the Blue Mountains uplift from adjacent terranes, or are they tangential shears related to oblique convergence of the Pacific and North American plates? Are the Brothers fault zone and the Olympic-Wallowa lineament the same type of structure with somewhat different surface expression? Surface structures related to the Brothers fault zone can be explained by stresses related to either right- or left-lateral movement on a much deeper buried linear structure, but no convincing field evidence has as yet been recognized to support either dextral or sinistral displacement.

(4) When was the uplift of the Blue Mountains initiated, and when did uplift terminate? Is deformation still in progress?

(5) Could the Blue Mountains uplift be the result solely of isostatic adjustments where accretionary wedges of comparatively light-weight rocks have been crammed against the North American plate?

(6) Do apparent gaps in the volcanic record indicate episodic volcanism? If so, do they correspond with changes in tectonic regime? The significance of episodic volcanism can only be evaluated when more comprehensive data are available on the volumes of different chemical kinds of volcanic products.

(7) Where are the vents (calderas?) for the moderate volumes of ash-flow tuff and air-fall pumice lapilli tuff in the Clarno and John Day Formations? Presumably, most of these vents are west of the axis of the Blue Mountains uplift, in and adjacent to Deschutes

Basin; all the ones known are now represented by plugs, domes, and related flows, and none is known to be related to ring fracture zones or to show evidence of cauldron subsidence. How do these silicic volcaniclastic rocks relate genetically to contemporaneous silicic ash flows and tuff in the western Cascades?

(8) Where are the vents for ash-flow tuff in the middle Miocene part of the section? Haddock (1966, 1967) postulated that the Dinner Creek Welded Ash-flow Tuff was erupted from fissure vents at and near Castle Rock, but where are the vents for other ash-flow tuff in the Mascall Formation? Are there ash-flow tuff vents related to the Dooley Rhyolite Breccia or to other rhyolitic bodies associated with the Strawberry Volcanics?

(9) How do the Miocene hypersthene andesite volcanos aligned along a northeastward trend relate to the apparent thickened crust underlying the Blue Mountains? Is their position controlled by some throughgoing structure, or do they simply signify a common type of crust in the Blue Mountains and areas to the south near Gearhart Mountain, which is within but near the margin of the Basin and Range province?

REFERENCES CITED

Armstrong, R.L., 1975, The geochronometry of Idaho: Isochron/West, no. 14, 50 p.

———1978, Cenozoic igneous history of the U.S. Cordillera from lat 42° to 49° N., chap. 12 *of* Smith, R.B., and Eaton, G.P., eds., Cenozoic tectonics and regional geophysics of the western Cordillera: Geological Society of America Memoir 152, p. 265–282.

Armstrong, R.L., Taubeneck, W.H., and Hales, P.O., 1977, Rb-Sr and K-Ar geochronometry of Mesozoic granitic rocks and their Sr isotopic composition, Oregon, Washington, and Idaho: Geological Society of America Bulletin, v. 88, no. 3, p. 397–411.

Beck, M.E., Jr., and Burr, C.D., 1979, Paleomagnetism and tectonic significance of the Goble Volcanic Series, southwestern Washington: Geology, v. 7, no. 4, p. 175–179.

Beck, M.E., Jr., Engebretson, D.C., Grommé, C.S., Taylor, E.M., and Whitney, J.W., 1978, Paleomagnetism of the middle Tertiary Clarno Formation, north-central Oregon—Constraints on models for tectonic rotation [abs.]: Eos (American Geophysical Union Transactions), v. 59, no. 12, p. 1058.

Beck, M.E., Jr., and Plumley, P.W., 1980, Paleomagnetism of intrusive rocks in the Coast Range of Oregon—Microplate rotations in middle Tertiary time: Geology, v. 8, no. 12, p. 573–577.

Berg, J.W., and Thiruvathukal, J.V., 1967, Complete Bouguer gravity anomaly map of Oregon: Oregon Department of Geology and Mineral Industries Geologic Map Series GMS–4b, scale 1:500,000.

Bond, J.G., 1978, Geologic map of Idaho: Moscow, Idaho Bureau of Mines and Geology, scale 1:500,000.

Brooks, H.C., McIntyre, J.R., and Walker, G.W., 1976, Geology of the Oregon part of the Baker 1° by 2° quadrangle: Oregon Department of Geology and Mineral Industries Geologic Map Series Map GMS–7, scale 1:250,000.

Brooks, H.C., and Vallier, T.L., 1978, Mesozoic rocks and tectonic evolution of eastern Oregon and western Idaho, *in* Howell, D.G., and McDougall, K.A., eds., Mesozoic paleogeography of the Western United States (Pacific Coast Paleogeography Symposium 2): Los Angeles, Society of Economic Paleontologists and Mineralogists, Pacific Section, p. 133–145.

Christiansen, R.L., and Lipman, P.W., 1972, Cenozoic volcanism and plate tectonic evolution of the Western United States—Pt. II, Late Cenozoic: Royal Society of London Philosophical Transactions, ser. A, v. 271, no. 1213, p. 249–284.

Christiansen, R.L., and McKee, E.H., 1978, Late Cenozoic volcanic and tectonic evolution of the Great Basin and Columbia Intermontane regions, chap. 13 of Smith, R.B., and Eaton, G.P., eds., Cenozoic tectonics and regional geophysics of the western Cordillera: Geological Society of America Memoir 152, p. 283–311.

Couch, R.W., and Lowell, R.P., 1971, Earthquakes and seismic energy release in Oregon: Ore Bin, v. 33, no. 4, p. 61–84.

Dickinson, W.R., 1979a, Cenozoic plate tectonic setting of the Cordilleran region in the United States, *in* Armentrout, J.M., Cole, M.R., and TerBest, Harry, Jr., eds., Cenozoic paleogeography of the Western United States (Pacific Coast Paleogeography Symposium 3): Los Angeles, Society of Economic Paleontologists and Mineralogists, Pacific Section, p. 1–13.

———1979b, Mesozoic fore-arc basin in central Oregon: Geology, v. 7, no. 4, p. 166–170.

Dickinson, W.R., and Thayer, T.P., 1978, Paleogeographic and paleotectonic implications of Mesozoic stratigraphy and structure in the John Day inlier of central Oregon, *in* Howell, D.G., and McDougall, K.A., eds., Mesozoic paleogeography of the Western United States (Pacific Coast Paleogeography Symposium 2): Los Angeles, Society of Economic Paleontologists and Mineralogists, Pacific Section, p. 147–161.

Dickinson, W.R., and Vigrass, L.W., 1965, Geology of the Suplee-Izee area, Crook, Grant, and Harney Counties, Oregon: Oregon Department of Geology and Mineral Industries Bulletin 58, 109 p.

Eaton, G.P., Wahl, R.R., Prostka, H.J., Mabey, D.R., and Kleinkopf, M.D., 1978, Regional gravity and tectonic patterns—Their relation to late Cenozoic epeirogeny and lateral spreading in the western Cordillera, chap. 3 *of* Smith, R.B., and Eaton, G.P., eds., Cenozoic tectonics and regional geophysics of the western Cordillera: Geological Society of America Memoir 152, p. 51–91.

Engebretson, D.C., 1982, Relative motions between oceanic and continental plates in the Pacific Basin: Stanford, Calif., Stanford University, Ph.D. dissertation, 211 p.

Ewing, T.E., 1980, Paleogene tectonic evolution of the Pacific Northwest: Journal of Geology, v. 88, no. 6, p. 619–638.

Farooqui, S.M., Beaulieu, J.D., Bunker, R.C., Stensland, D.E., and Thoms, R.E., 1981, Dalles Group—Neogene formations overlying the Columbia River Basalt Group in north-central Oregon: Oregon Geology, v. 43, no. 10, p. 131–140.

Fiebelkorn, R.B., Walker, G.W., MacLeod, N.S., McKee, E.H., and Smith, J.G., 1983, Index to K-Ar determinations for the State of Oregon: Isochron/West, no. 37, 60 p.

Fisher, R.V., 1967, Early Tertiary deformation in north-central Oregon: American Association of Petroleum Geologists Bulletin, v. 51, no. 1, p. 111–123.

Fleck, R.J., 1983, Strontium isotope variations in western Idaho and eastern Oregon and Washington [abs.]: Geological Society of America Abstracts with Programs, v. 15, no. 5, p. 372.

Haddock, G.H., 1966, Vent location by investigation of lateral and vertical characteristics of a welded ash-flow tuff [abs.], *in* Abstracts for 1965: Geological Society of America Special Paper 87, p. 206–207.

———1967, The Dinner Creek Welded Ash-flow Tuff of the Malheur Gorge area, Malheur County, Oregon: Eugene, University of Oregon, Ph.D. dissertation, 111 p.

Hamilton, Warren, and Myers, W.B., 1966, Cenozoic tectonics of the Western United States: Reviews of Geophysics, v. 4, no. 4, p. 509–549.

Hildenbrand, T.G., Simpson, R.W., Godson, R.H., and Kane, M.F., 1982, Digital colored residual and regional Bouguer gravity maps of the conterminous United States with cutoff wavelengths of 250 km and 1000 km: U.S. Geological Survey Geophysical Investigations Map GP-953-A, 2 sheets, scale 1:7,500,000.

Hill, D.P., 1972, Crustal and upper-mantle structure of the Columbia Plateau from long-range seismic-refraction measurements: Geological Society of America Bulletin, v. 83, no. 6, p. 1639-1648.

———1978, Seismic evidence for the structure and Cenozoic tectonics of the Pacific Coast States, chap. 7 *of* Smith, R.B., and Eaton, G.P., eds., Cenozoic tectonics and regional geophysics of the western Cordillera: Geological Society of America Memoir 152, p. 145-174.

Jones, D.L., Silberling, N.J., and Hillhouse, John, 1977, Wrangellia—A displaced terrane in northwestern North America: Canadian Journal of Earth Sciences, v. 14, no. 11, p. 2565-2577.

Lawrence, R.D., 1976, Strike-slip faulting terminates the Basin and Range province in Oregon: Geological Society of America Bulletin, v. 87, no. 6, p. 846-850.

Leeman, W.P., 1982, Tectonic and magmatic significance of strontium isotope variations in Cenozoic volcanic rocks from the Western United States: Geological Society of America Bulletin, v. 93, no. 6, p. 487-503.

Lipman, P.W., Prostka, H.J., and Christiansen, R.L., 1972, Cenozoic volcanism and plate tectonic evolution of the Western United States—Pt. I, Early and middle Cenozoic: Royal Society of London Philosophical Transactions, ser. A, v. 271, no. 1213, p. 217-248.

Livingston, V.E., Jr., 1966, Geology and mineral resources of the Kelso-Cathlamet area, Cowlitz and Wahkiakum Counties, Washington: Washington Division of Mines and Geology Bulletin 54, 110 p.

Lux, D.R., 1982, K-Ar and ^{40}Ar-^{39}Ar ages of mid-Tertiary volcanic rocks from the western Cascade Range, Oregon: Isochron/West, no. 33, p. 27-32.

Magill, J.R., and Cox, A.V., 1980, Tectonic rotation of the Oregon western Cascades: Oregon Department of Geology and Mineral Industries Special Paper 10, 67 p.

McBirney, A.R., Sutter, J.F., Naslund, H.R., Sutton, K.G., and White, C.M., 1974, Episodic volcanism in the central Oregon Cascade Range: Geology, v. 2, no. 12, p. 585-589.

McKee, E.H., Swanson, D.A., and Wright, T.L., 1977, Duration and volume of Columbia River Basalt volcanism, Washington, Oregon, and Idaho [abs.]: Geological Society of America Abstracts with Programs, v. 9, no. 4, p. 463-464.

Muessig, Siegfried, 1962, Tertiary volcanic and related rocks of the Republic area, Ferry County, Washington, art. 135 of Short papers in geology, hydrology, and topography: U.S. Geological Survey Professional Paper 450-D, p. 56-58.

Muntzert, J.K., and Field, C.W., 1969, Geology and mineral deposits of the Brattain district, Lake County, Oregon [abs.], *in* Abstracts for 1968: Geological Society of America Special Paper 121, p. 616-617.

Newcomb, R.C., 1970, Tectonic structure of the main part of the basalt of the Columbia River Group, Washington, Oregon, and Idaho: U.S. Geological Survey Miscellaneous Geologic Investigations Map I-587, scale 1:500,000.

Noblett, J.B., 1981, Subduction-related origin of the volcanic rocks of the Eocene Clarno Formation near Cherry Creek, Oregon: Oregon Geology, v. 43, no. 7, p. 91-99.

Ort, K.M., Tabor, R.W., and Frizzell, V.A., Jr., 1983, Chemical analyses of selected Tertiary and Quaternary volcanic rocks, Cascade Range, Washington: U.S. Geological Survey Open-File Report 83-1, 14 p.

Peterson, N.V., Groh, E.A., Taylor, E.M., and Stensland, D.E., 1976, Geology and mineral resources of Deschutes County, Oregon: Oregon Department of Geology and Mineral Industries Bulletin 89, 66 p.

Pigg, J.H., Jr., 1961, The lower Tertiary sedimentary rocks in the Pilot Rock and Heppner areas, Oregon: Eugene, University of Oregon, M.S. thesis, 67 p.

Robinson, P.T., 1975, Reconnaissance geologic map of the John Day Formation in the southwestern part of the Blue Mountains and adjacent areas, north-central Oregon: U.S. Geological Survey Miscellaneous Investigations Series Map I-872, scale 1:125,000.

Robinson, P.T., Brem, G.F., and McKee, E.H., 1984, The John Day Formation of Oregon—A distal record of early Cascade volcanism: Geology, v. 12, no. 4, p. 229-232.

Robinson, P.T., and Stensland, D.H., 1979, Geologic map of the Smith Rock area, Jefferson, Deschutes, and Crook Counties, Oregon: U.S. Geological Survey Miscellaneous Investigations Series Map I-1142, scale 1:48,000.

Robyn, T.L., 1977, Geology and petrology of the Strawberry Volcanics, NE Oregon: Eugene, University of Oregon, Ph.D. dissertation, 197 p.

———1979, Miocene volcanism in eastern Oregon—An example of calc-alkaline volcanism unrelated to subduction: Journal of Volcanology and Geothermal Research, v. 5, no. 1-2, p. 149-161.

Robyn, T.L., and Hoover, J.D., 1982, Late Cenozoic deformation and volcanism in the Blue Mountains of central Oregon—Microplate interactions?: Geology, v. 10, no. 11, p. 572-576.

Rogers, J.J.W., and Novitsky-Evans, J.M., 1977, The Clarno Formation of central Oregon, U.S.A.—Volcanism on a thin continental margin: Earth and Planetary Science Letters, v. 34, no. 1, p. 56-66.

Silberling, N.J., Jones, D.L., Blake, M.C., Jr., and Howell, D.G., 1984, Lithotectonic terrane map of the western conterminous United States, *in* Silberling, N.J., and Jones, D.L., eds., Lithotectonic terrane maps of the North American Cordillera: U.S. Geological Survey Open-File Report 84-523, p. C-1 to C-43.

Simpson, R.W., and Cox, A.V., 1977, Paleomagnetic evidence for tectonic rotation of the Oregon Coast Range: Geology, v. 5, no. 10, p. 585-589.

Smith, R.B., 1978, Seismicity, crustal structure, and intraplate tectonics of the interior of the western Cordillera, chap. 6 *of* Smith, R.B., and Eaton, G.P., eds., Cenozoic tectonics and regional geophysics of the western Cordillera: Geological Society of America Memoir 152, p. 111-144.

Snyder, W.S., Dickinson, W.R., and Silberman, M.L., 1976, Tectonic implications of space-time patterns of Cenozoic magmatism in the Western United States: Earth and Planetary Science Letters, v. 32, no. 1, p. 91-106.

Stewart, J.H., 1978, Basin-range structure in western North America—A review, chap. 1 *of* Smith, R.B., and Eaton, G.P., eds., Cenozoic tectonics and regional geophysics of the western Cordillera: Geological Society of America Memoir 152, p. 1-31.

Swanson, D.A., 1969, Reconnaissance geologic map of the east half of the Bend quadrangle, Crook, Wheeler, Jefferson, Wasco, and Deschutes Counties, Oregon: U.S. Geological Survey Miscellaneous Geologic Investigations Map I-568, scale 1:250,000.

Swanson, D.A., Anderson, J.L., Camp, V.E., Hooper, P.R., Taubeneck, W.H., and Wright, T.L., 1981, Reconnaissance geologic map of the Columbia River Basalt Group, northern Oregon and western Idaho: U.S. Geological Survey Open-File Report 81-797, scale 1:250,000.

Swanson, D.A., and Robinson, P.T., 1968, Base of the John Day Formation in and near the Horse Heaven mining district, north-central Oregon, *in* Geological Survey research, 1968: U.S. Geological Survey Professional Paper 600-D, p. 154-161.

Tabor, R.W., Frizzell, V.A., Jr., Vance, J.A., and Naeser, C.W., 1984, Ages and stratigraphy of lower and middle Tertiary sedimentary

and volcanic rocks of the central Cascades, Washington—Application to the tectonic history of the Straight Creek fault: Geological Society of America Bulletin, v. 95, no. 1, p. 26–44.

Taylor, E.M., 1977, The Clarno Formation—A record of early Tertiary volcanism in central Oregon [abs.]: Geological Society of America Abstracts with Programs, v. 9, no. 6, p. 768.

———1981, A mafic-dike system in the vicinity of Mitchell, Oregon, and its bearing on the timing of Clarno-John Day volcanism and early Oligocene deformation in central Oregon: Oregon Geology, v. 43, no. 8, p. 107–112.

Thayer, T.P., 1957, Some relations of later Tertiary volcanology and structure in eastern Oregon, *in* v. 1 of Vulcanología del Cenozóico: International Geological Conference, 20th, Mexico City, 1956, [Trabajos], sec. 1, p. 231–245.

Thayer, T.P., and Brown, C.E., 1966, Local thickening of basalts and late Tertiary silicic volcanism in the Canyon City quadrangle, northeastern Oregon, *in* Geological Survey research, 1966: U.S. Geological Survey Professional Paper 550–C, p. 73–78.

Thiruvathukal, J.V., Berg, J.W., Jr., and Heinrichs, D.F., 1970, Regional gravity of Oregon: Geological Society of America Bulletin, v. 81, no. 3, p.725–738.

Thompson, G.G., Yett, J.R., and Green, K.E., 1984, Subsurface stratigraphy of the Ochoco Basin, Oregon: Oregon Department of Geology and Mineral Industries Oil and Gas Investigation 8, 22 p.

Thompson, R.N., 1977, Columbia/Snake River-Yellowstone magmatism in the context of western U.S.A. Cenozoic geodynamics: Tectonophysics, v. 39, no. 4, p. 621–636.

Trimble, D.E., 1963, Geology of Portland, Oregon, and adjacent areas: U.S. Geological Survey Bulletin 1119, 119 p.

Vance, J.A., 1982, Cenozoic stratigraphy and tectonics of the Washington Cascades [abs.]: Geological Society of America Abstracts with Programs, v. 14, no. 4, p. 241.

Walker, G.W., 1969, Geology of the High Lava Plains province, *in* Mineral and water resources of Oregon: Oregon Department of Geology and Mineral Industries Bulletin 64, p. 77–79.

———1977, Geologic map of Oregon east of the 121st meridian: U.S. Geological Survey Miscellaneous Investigations Series Map I–902, scale 1:500,000.

———1980, Preliminary report on the geology of the Lakeview uranium area, Lake County, Oregon: U.S. Geological Survey Open-File Report 80–532, 33 p.

Walker, G.W., Peterson, N.V., and Greene, R.C., 1967, Reconnaissance geologic map of the east half of the Crescent quadrangle, Lake, Deschutes, and Crook Counties, Oregon: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–493, scale 1:250,000.

Waters, A.C., 1954, John Day Formation west of its type locality [abs.]: Geological Society of America Bulletin, v. 65, no. 12, pt. 2, p. 1320.

Watkins, N.D., and Baksi, A.K., 1974, Magnetostratigraphy and oroclinal folding of the Columbia River, Steens and Owyhee basalts of Oregon, Washington, and Idaho: American Journal of Science, v. 274, no. 2, p. 148–189.

Wells, R.E., Engebretson, D.C., Snavely, P.D., Jr., and Coe, R.S., 1984, Cenozoic plate motions and the volcanotectonic evolution of western Oregon and Washington: Tectonics, v. 3, no. 2, p. 275–294.

Wells, R.E., Niem, A.R., MacLeod, N.S., Snavely, P.D., Jr., and Niem, W.A., 1983, Preliminary geologic map of the west half of the Vancouver (Washington-Oregon) 1° by 2° quadrangle, Oregon: U.S. Geological Survey Open-File Report 83–591, scale 1:250,000.

Wise, W.S., 1970, Cenozoic volcanism in the Cascade Mountains of southern Washington: Washington Division of Mines and Geology Bulletin 60, 45 p.

Woodburne, M.O., and Robinson, P.T., 1977, A new late Hemingfordian mammal fauna from the John Day Formation, Oregon, and its stratigraphic implications: Journal of Paleontology, v. 51, no. 4, p. 750–757.

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