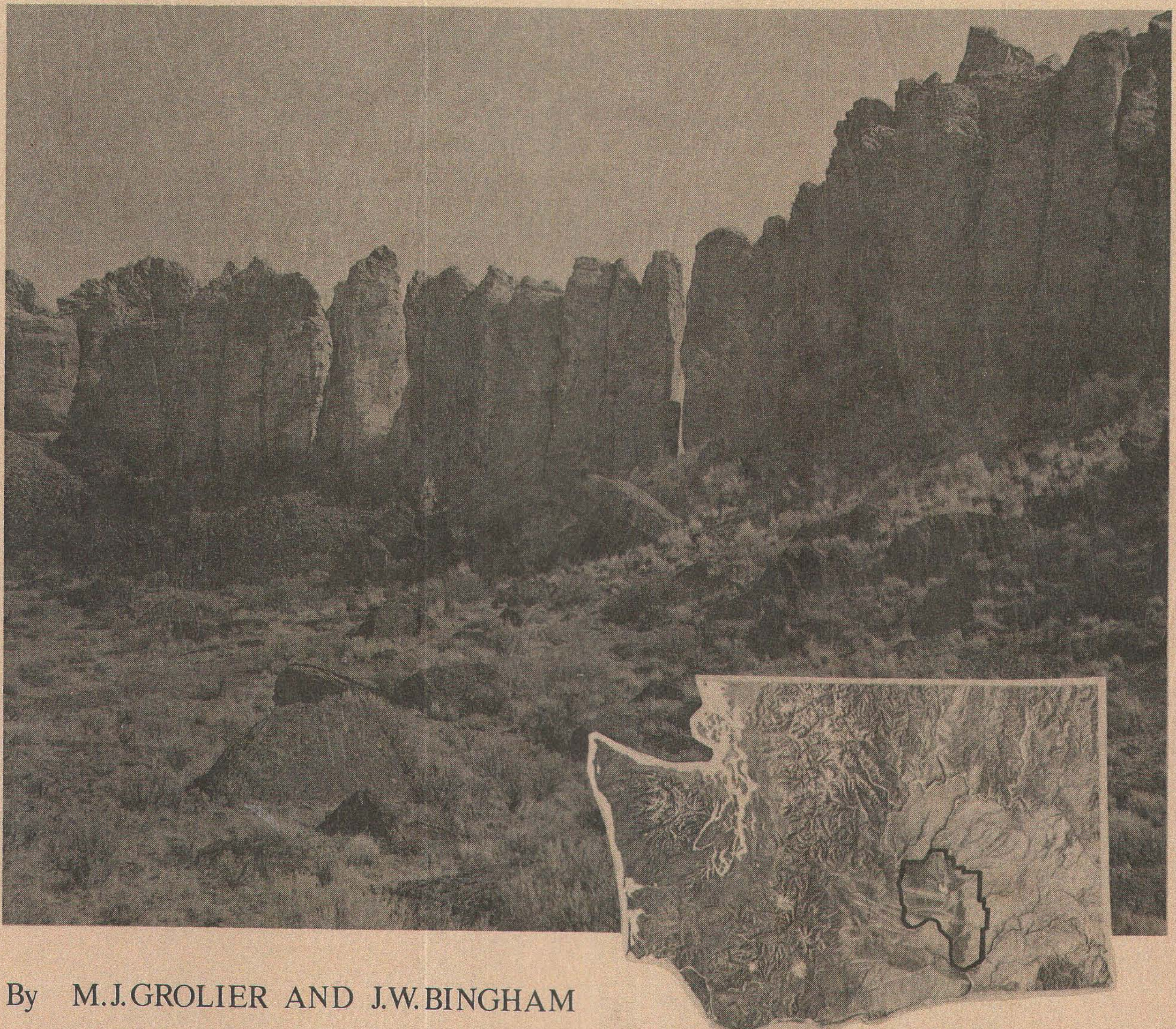
Bulletin 71
Editor

BULLETIN NO. 71

GEOLOGY OF PARTS OF GRANT, ADAMS, AND FRANKLIN COUNTIES, EAST-CENTRAL WASHINGTON



By M. J. GROLIER AND J. W. BINGHAM

DEPARTMENT OF NATURAL RESOURCES
DIVISION OF GEOLOGY AND EARTH RESOURCES
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COVER PHOTO

Colonnade in the Roza basalt, Frenchman Springs Coulee, Washington

GEOLOGY OF PARTS OF GRANT, ADAMS, AND FRANKLIN COUNTIES, EAST-CENTRAL WASHINGTON

FRONTISPIECE

"Dry Falls When Wet"

This late Pleistocene scene of Grand Coulee, a few miles north of the study area, shows the course of the Columbia River down the coulee and over the present Dry Falls. The previously established (and present) course of the river around the northern margin of the Columbia Plateau lava flows was blocked temporarily—and several times—by the Okanogan Lobe of the Cordilleran Ice Sheet that entered Washington from Canada during the ice age. The illustration shows the ice sheet in the background and the large melt-water discharge of the Columbia River, which cut Grand Coulee and formed the Dry Falls escarpment by the headward (northward) plucking away of the layers of columnar-jointed basalt. From here the water discharged across the study area, and returned to the established course of the Columbia River by way of the Quincy Basin, Lower Crab Creek valley, and Pasco Basin.

From an oil painting by Dee Molenaar,
courtesy of the Washington State
Parks and Recreation Commission.

STATE OF WASHINGTON

DEPARTMENT OF NATURAL RESOURCES

BERT L. COLE, Commissioner of Public Lands
RALPH A. BESWICK, Supervisor

DIVISION OF GEOLOGY AND EARTH RESOURCES

VAUGHN E. LIVINGSTON, JR., State Geologist

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GEOLOGY OF PARTS OF GRANT, ADAMS, AND FRANKLIN COUNTIES, EAST-CENTRAL WASHINGTON

BY

M. J. GROLIER and J. W. BINGHAM

Prepared in cooperation with the
U.S. Geological Survey, U.S. Department of Energy,
and the
Washington State Department of Ecology

1978

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CONTENTS

CONTENTS

CONTENTS

ILLUSTRATIONS

Page

PREFACE

This report was approved by the Director of the U.S. Geological Survey in 1966. The original report included a geologic map and cross sections of the area which subsequently were published, separately from the text, in 1971 as U.S. Geological Survey Investigations Map I-589. The text and illustrations, which were intended to supplement Map I-589, remained unpublished for more than a decade primarily because of shortage of publication funds. However, in 1977 the U.S. Department of Energy provided funds for publication of the report by the State of Washington Department of Natural Resources, Division of Geology and Earth Resources.

The report is printed basically as approved in 1966, except for the absence of the geologic map and cross sections (Map I-589). There are minor footnoted revisions in the text and revisions of the stratigraphic diagram (fig. 9) in response to suggestions offered by the Geological Survey's Geologic Names Committee (Rudy Kopf, oral and written communications, 1977). Also, the bibliography has been expanded to include—besides references cited in the original report—a list of selected reports, (mostly since 1966) that have added to the geologic knowledge of the Columbia Plateau and which include updated interpretations of the rock units and their stratigraphic relationships.

GEOLOGY OF PARTS OF GRANT, ADAMS, AND FRANKLIN COUNTIES, EAST-CENTRAL WASHINGTON

by

M. J. Grolier and J. W. Bingham[1/]

ABSTRACT

The study area, which is coextensive with the Columbia Basin Irrigation Project area, covers 3,635 square miles within the northern part of the Columbia Plateau, in the arid to semiarid part of east-central Washington. The area generally has low relief except where the regional southwestward slope is interrupted by canyons and anticlinal ridges. Two main ridges, Frenchman Hills and Saddle Mountains, separate the project area into three principal basins: the Quincy, Othello, and Pasco Basins.

The geologic events that produced the present landforms of the project area began during late Cenozoic time. Large volumes of highly fluid basaltic magma were extruded and spread in successive layers over large areas. The lava accumulated to an overall thickness that exceeded 5,000 feet, and possibly 10,000 feet, in the central part of the Columbia Plateau. During intervals between lava extrusions, fluvial and lacustrine sediments were deposited in depressions on the basalt surface and across areas along margins of the flows.

The entire sequence of lava flows and associated sedimentary interbeds has been termed the Columbia River Basalt Group. The top several thousand feet of the group in this area has been called the Yakima Basalt of late Miocene and early Pliocene age (see footnote to figure 9). Within that formation, the top 1,200 feet includes five separate and largely conformable members, defined (from oldest to youngest) as follows: the Vantage Sandstone Member; the Frenchman Springs Member, which locally includes the Squaw Creek Diatomite Bed; the Roza Member; the Priest Rapids Member, which locally includes the Quincy Diatomite Bed; and the Saddle Mountains Member, which interfingers with the Beverly Member of the Ellensburg Formation. The basalt members, each of which consists of one to four lithologically similar flows, are correlated in the field by their frequency and size of phenocrysts, color, jointing habits, texture, vesicularity, and stratigraphic position. The sedimentary materials that are interbedded with the basalt flows were deposited in lakes and streams. Each deposit was spread widely over the surface of the preceding basalt flow.

Most folding in the project area occurred after extrusion of the Yakima Basalt and created several shallow basins in which sedimentary materials were deposited, in some places to maximum thicknesses of about 1,000 feet. These deposits, in part named the Ringold Formation, consist of interbedded lacustrine clay, silt, and sand; fluviatile sand and gravel; eolian sand and silt, and alluvial-fan deposits, with varying amounts of pyroclastic debris.

The topography of the project area was greatly affected by melt-water runoff and catastrophic flooding associated with the continental glaciations that extended into northern Washington several times during late Pleistocene time. Many gorges were cut where floodwaters were diverted across the area, subfluvial gravels were deposited as bars in channel-ways, and deltaic and lacustrine deposits accumulated in temporary lakes that occupied the Quincy, Othello, and Pasco structural basins.

After recession of the continental glaciers from the area, a dry, windy climate prevailed and promoted widespread deposition of loess and sand. The material was derived primarily from fine-grained parts of the glaciofluvial deposits.

1/ Authors are with U.S. Geological Survey.

INTRODUCTION

LOCATION

The area described in this report covers 3,635 square miles in parts of Grant, Adams, and Franklin Counties, in east-central Washington (fig. 1). It lies within that part of the Columbia Plateau known as the "Big Bend Country," and has boundaries coextensive with that of the U.S. Bureau of Reclamation's Columbia Basin Irrigation Project as of 1960. Cultural features within the study area are shown in figure 2.

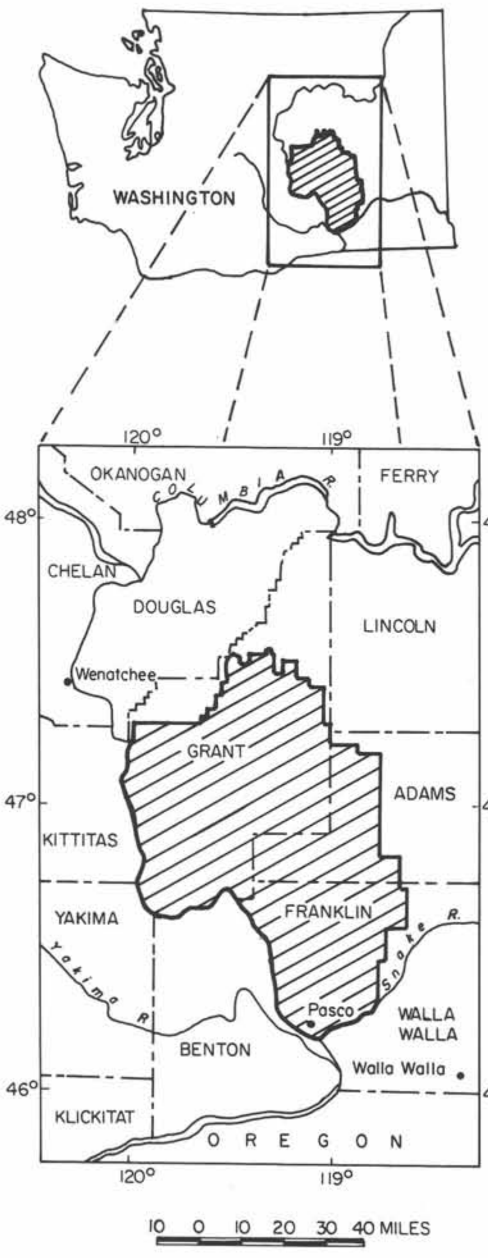

FIGURE 1.—Location of the project area.

BACKGROUND OF THE STUDY

Geologic and hydrologic studies associated with the Columbia Basin Irrigation Project have spanned more than three decades. Studies by the U.S. Geological Survey have been conducted in cooperation with the Bureau of Reclamation and with the water-management agencies of the State of Washington. The various cooperative geohydrologic studies have been largely interrelated and have evolved in two stages: (1) a program of collection of hydrologic and geologic data for evaluating potential ground-water problems resulting from irrigation development using surface water imported from the Columbia River behind Grand Coulee Dam (A. M. Piper, written communication, 1939; Warne and Barrows, 1941, p. 23); and (2) an investigation of the geology of the basin relative to the mode of occurrence and movement of ground water within the various stratigraphic units and as controlled by geologic structures, and as later affected by project irrigation.

The short-term objective of the cooperative activities in the first stage was the gathering of basic ground-water data for an evaluation of ground-water supplies for domestic use on farmsteads and of waterlogging and drainage problems that might ensue from irrigation. The first stage of the study began in 1939 and included collection of basic hydrologic data both before and during the onset of irrigation. Data collection was accomplished cooperatively by the Geological Survey, the Bureau of Reclamation, and the

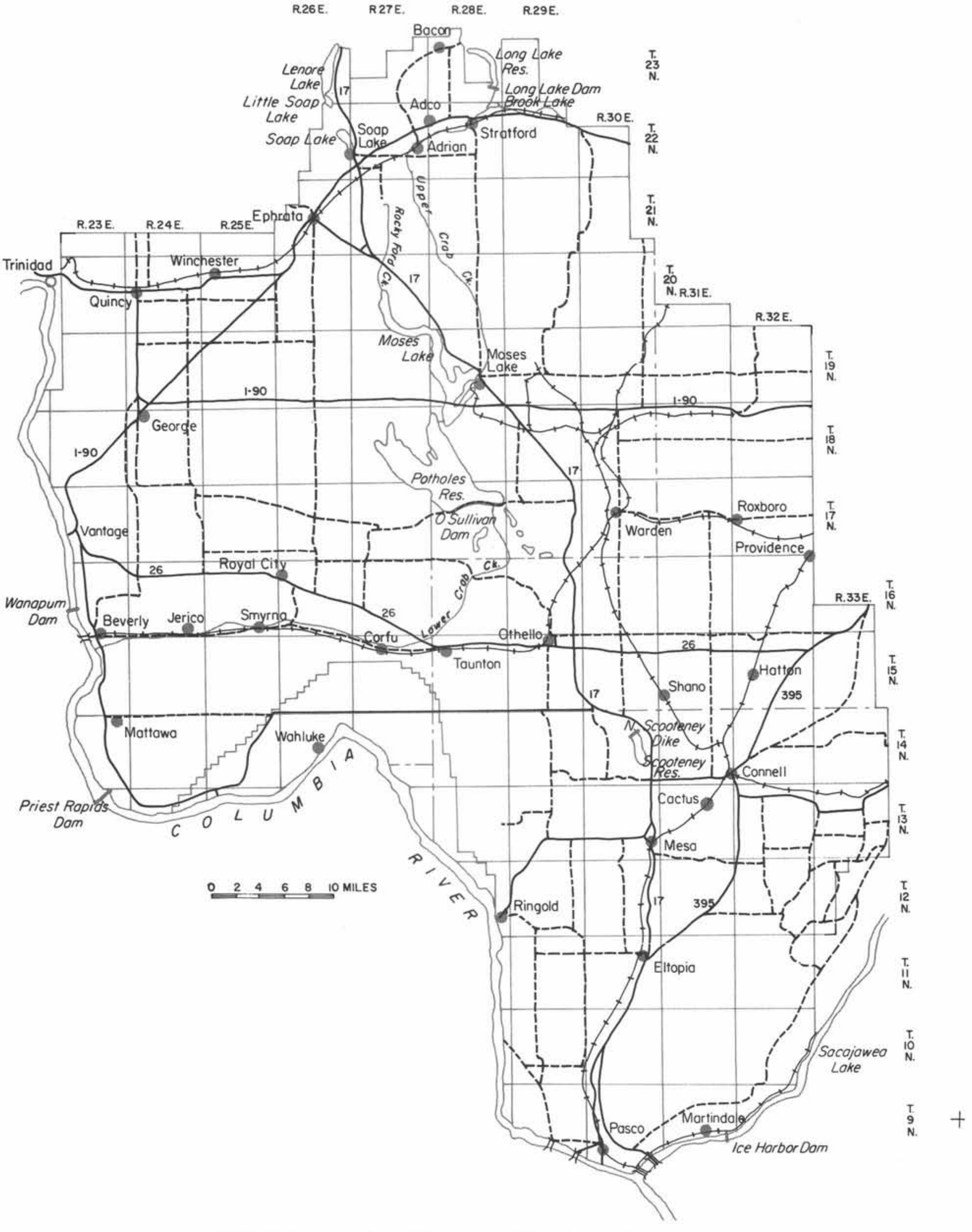

FIGURE 2.—Cultural features within the project area.

Washington Division of Water Resources (functions now included in the Department of Ecology). Four data reports resulted and were open filed by the U.S. Geological Survey (Taylor, 1941, 1944, 1948; Mundorff and others, 1952). A compilation of the data through December 1958 was published by the Washington State Division of Water Resources (Walters and Grolier, 1960). An open-file report was subsequently released by the U.S. Geological Survey (Bingham, 1965).

The second stage of the study was begun in August 1959 by the Geological Survey in cooperation with the State Division of Water Resources and continued in cooperation with the Department of Ecology. Specific objectives were twofold: (1) to delineate the geologic structures and pertinent stratigraphy of the project area in relation to the source, mode of occurrence, and movement of ground water; and (2) to evaluate effects on the ground-water regimen caused by the importation of large volumes of surface water for irrigation in the project area. Results of the first objective were published by the Geological Survey as a set of geologic maps and sections (Grolier and Bingham, 1971), and results of the second objective were published by the State of Washington Department of Ecology (Tanaka, Hansen, and Skrivan, 1974).

This present report is published by the State of Washington Division of Geology and Earth Resources in response to requests for a detailed discussion of the geology that is depicted graphically by the geologic maps and sections of Grolier and Bingham (1971). In the geologic investigation, field mapping was concentrated on the stratigraphy of the basalt sequence and the glaciofluvial gravel. The basalt is the one areawide source of ground water, and the glaciofluvial gravel, where saturated, is the aquifer of the greater yield. With the segregation of the basalt stratigraphically, it becomes possible not only to correlate water-bearing zones over wider areas and to better understand the regional hydrologic conditions, but also to help resolve many long-held questions on the geologic history and structure of the project area and other parts of the vast Columbia Plateau.

ACKNOWLEDGMENTS

The U.S. Geological Survey acknowledges the cooperation of the U.S. Bureau of Reclamation, through P. R. Nalder and W. E. Rawlings, former Project Managers for the Columbia Basin Irrigation Project. Geologic information was obtained from unpublished Bureau of Reclamation reports by F. O. Jones, F. A. Nickell, George Neff, and W. E. Walcott.

Unpublished soils maps and aerial photographs were loaned by the Ephrata and Pasco offices of the U.S. Soil Conservation Service. The writers were aided in their study of basalt stratigraphy by discussions with the late J. Hoover Mackin and with Aaron C. Waters.

Personnel of the Irrigation and Drainage Section and of the Hydrography Section in the Bureau of Reclamation Project headquarters furnished many well logs and quantitative hydrologic data, as well as the base maps used for this report. Of special assistance in this regard were E. H. Neal, W. O. Watson, James Ellingboe, and Edwin Nasburg.

The U.S. Atomic Energy Commission, through J. E. Travis, permitted access to the part of the project area under control of the Hanford Operations Office, and the General Electric Co., Hanford Laboratories Operation, made available pertinent hydrologic and climatic data through R. E. Brown and J. J. Fuquay.

The Walla Walla District of the U.S. Army Corps of Engineers and the Grant County Public Utility District made available unpublished records of the areas of their investigations along the Columbia and

Snake Rivers at dams they operate. Several municipalities, irrigation districts, and industrial firms provided logs and pumping records of wells.

Of benefit to the final report were technical reviews of the manuscript by A. M. Piper and R. C. Newcomb, both now retired from the U.S. Geological Survey.

GEOGRAPHIC SETTING

Landforms

The Columbia Basin Irrigation Project area lies in the northern part of the Columbia Plateau province (Fenneman, 1931, p. 225). In the project area, the province is generally diversified by several broad basins and flats separated by anticlinal ridges and transected by stream channels and flat-bottomed coulees. Altitudes range from about 350 feet to about 2,700 feet. Figure 3 shows the general land-surface altitudes in the project area, and figure 4 outlines and names the principal physiographic features.

The eastern part of the area is a loess-mantled upland that, in places, has been dissected into many flat-bottomed coulees which have floors that merge with the basins and plains to the west; the deposits in the basins block and fill the lower parts of the coulees. In the western and southern parts of the area, several types of plains are apparent from the topography and geology shown in the geologic map of Grolier and Bingham (1971); these include the gravel plain in the northeastern Quincy Basin, the remainder of the Quincy Basin with its low relief and predominantly sandy soils, and the Othello and Pasco Basins. The three basins all have dissected plains, remnants of early Pleistocene lake deposits.

The project area and adjacent areas contain many canyons, coulees, and scabland tracts. These are features that have been scoured and modified by glacial melt-water rivers and floods during the Pleistocene Epoch (Bretz, 1930). Many of the major coulees entering the basins from the east trend in a westerly direction whereas their tributaries trend southwesterly and are concentrated along the north side of the east-west coulees.

All perennial and intermittent streams in the project area, including the Columbia and Snake Rivers, flow in flat-bottomed coulees or scabland channels whose great sizes indicate occupancy by considerably larger ancestral streams. The most spectacular of these coulees is the Grand Coulee, a 50-mile-long channel divided into two (upper and lower) cliff-walled segments that were formed by cataract recession. A part of the Lower Grand Coulee is included in the project area (see frontispiece). Two of the other many remarkable coulees and cataracts within the project area are Potholes Coulee (fig. 5) and Frenchman Springs Coulee (fig. 6). These served as spillways from the Quincy Basin during glacial flooding.

In addition to the anticlinal ridges, basins, and coulees, several other features characterize the area. Landslides are prominent along the White Bluffs and the north side of the Saddle Mountains. Both active and stabilized sand dunes also are prominent features of the basin. Large areas contain stabilized sand dunes that were formed by reworking of the fine-grained components of the glacial melt-water deposits.

Drainage

The present drainage of the project area is controlled by Pleistocene or older drainageways. These relict valleys have been only slightly modified during Holocene time by the underfit and intermittent streams that occupy them today. The project area has very little surface runoff except under certain conditions, such as when heavy rainfall occurs on frozen or saturated soil, and when snow melts rapidly above frozen soil.

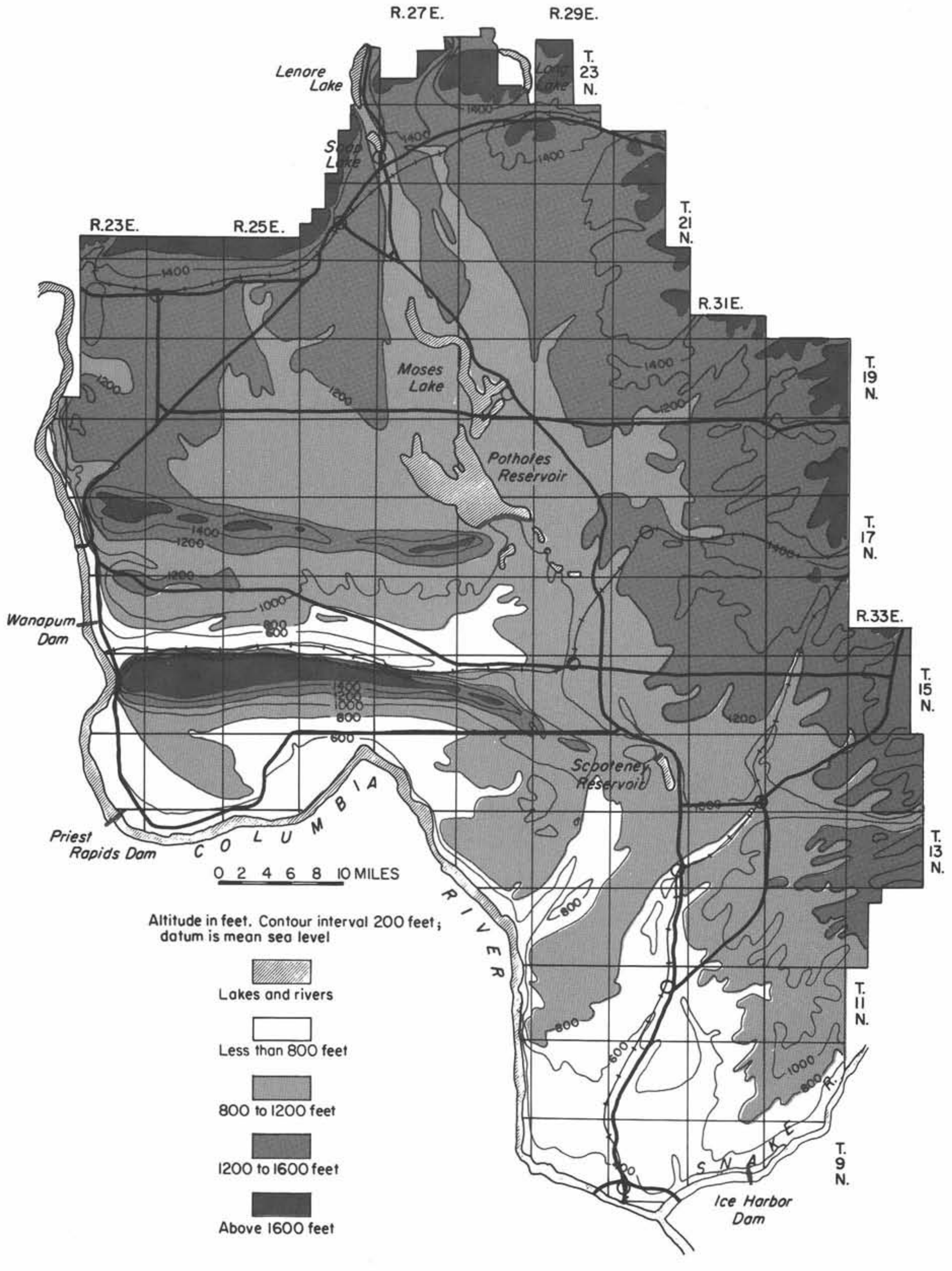

FIGURE 3.—Generalized land-surface altitudes within the project area.

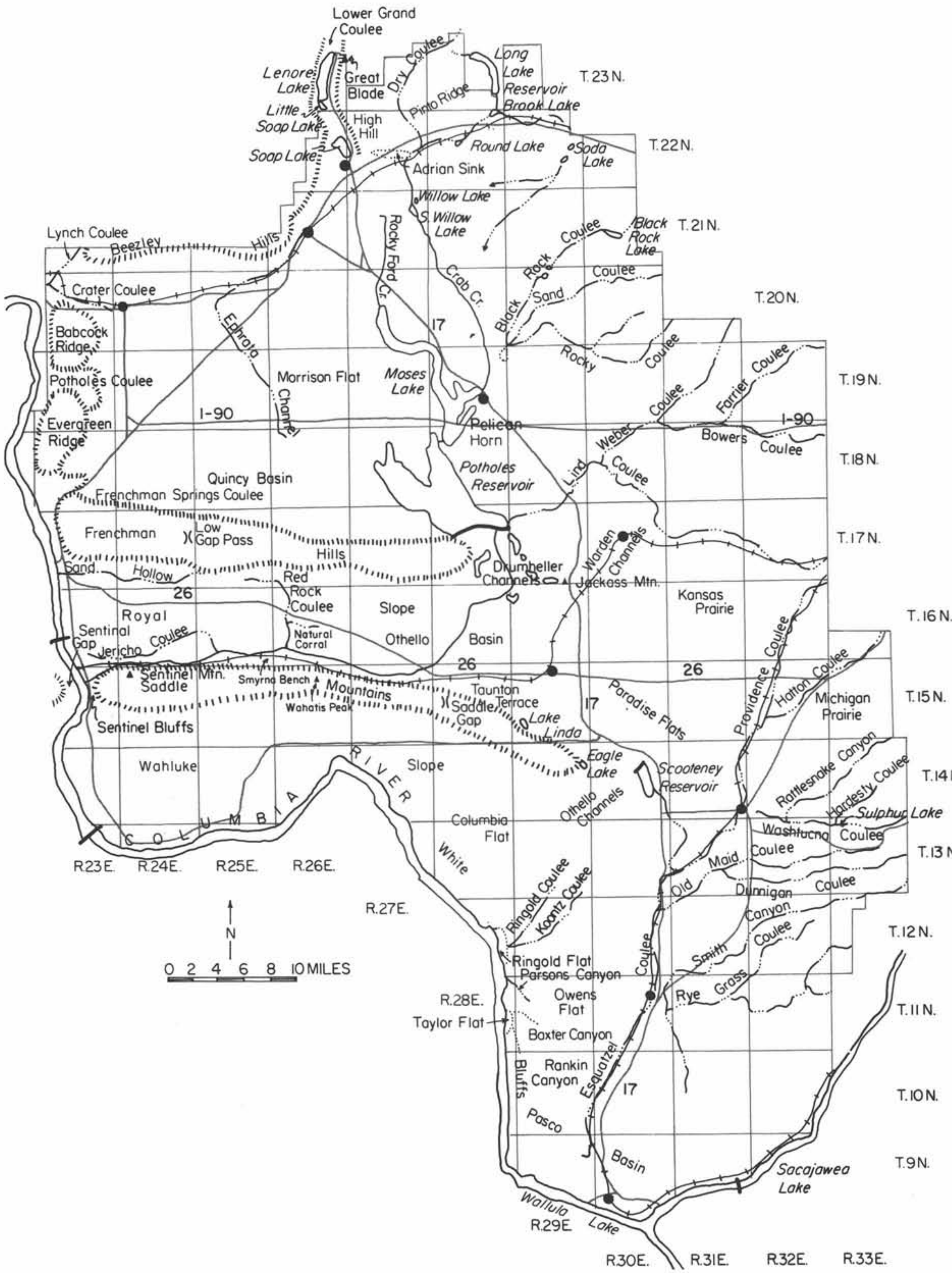

FIGURE 4.—Physiographic features of the project area.

FIGURE 5.—Aerial view of Potholes Coulee, looking east into the Quincy Basin. Columbia River in foreground. Note gravel terraces and bars in alcoves and along river. (Grant County PUD photo, 1961.)

Crab Creek is the major stream in the area. Its drainage area extends north and east of the project area and is very large (4,464 square miles above O'Sullivan Dam and 4,864 square miles above the mouth at Beverly). In this report, Crab Creek proper is defined as that part of the stream above Potholes Reservoir, while Lower Crab Creek is defined as that part below the reservoir. The water in Crab Creek entering the project area overflows into Adrian Sink during flood stage. The water that spills over the control dams on Moses Lake then flows between the sand dunes, which form the lake's natural dam, into Potholes Reservoir. (Prior to 1952, when O'Sullivan dam was constructed, the flow of Crab Creek continued through Drumheller Channels down Lower Crab Creek to the Columbia River.) Since 1952, the flow of Lower Crab Creek is nearly all from irrigation runoff in the area south of the dam.

The southern part of the project area is drained to Washtucna and Esquatzel Coulees; runoff from these coulees does not contribute directly to the Columbia River as does that from Lower Crab Creek. The Washtucna drainage ends in Sulphur Lake while the Esquatzel drainage ends in a closed depression just north of Pasco. The remaining peripheral parts of the project area are drained directly to the Columbia and Snake Rivers by short intermittent streams.

CLIMATE

The project area is partly arid and partly semiarid, based on the definition by the American

FIGURE 6.—Aerial view of Frenchman Springs Coulee, looking northeast into Quincy Basin. White patches beyond alcove scarp are quarries in the Quincy Diatomite Bed. Coulee is type area of Frenchman Springs Member. Roza Member is crinkly columnar basalt capping narrow rib between alcoves and scarp in upper right. (Grant County PUD photo, 1961.)

Geological Institute (1960, p. 15). Average annual precipitation ranges from less than 6 inches near Priest Rapids Dam to more than 10 inches in the far eastern part of the area. Like that in most of eastern Washington, the precipitation occurs largely during the winter.

The monthly distribution of precipitation reflects the maritime influence in the winter, when global atmospheric circulation creates a large low-pressure center in the Gulf of Alaska from where moisture-laden storms move into the Pacific Northwest. The little rainfall that occurs in spring and summer is chiefly from local thunderstorms. Precipitation during the winter generally falls as light rain, or as snow which may accumulate up to about 12 inches on the ground. Chinook winds are common and cause rapid snowmelt; this in turn results in appreciable surface runoff when the ground is frozen.

The continental environment of the project area is reflected by large daily and seasonal temperature fluctuations. In contrast, the marine effect is reflected by generally mild winters, except when large polar airmasses occasionally spill cold air over the Canadian Rockies into central Washington and cause extreme low temperatures. Mean annual temperatures recorded in the study area range from 9°C (48.4°F) at Moses Lake to 12°C (54.5°F) at Wahluke.

The average annual evaporation in the project area, as measured from a U.S. Weather Bureau Class-A pan, ranges from about 50 to 70 inches (Kohler and others, 1959, pls. 1 and 5).

GEOLOGIC HISTORY

The major events in the geologic history of the project area were (1) the extrusion of basaltic lavas in the Miocene and early part of the Pliocene Epochs, (2) the tectonic deformation during the Pliocene and early Pleistocene Epochs, (3) the lacustrine, fluvial, and eolian deposition, and intermittent deformation during the Pleistocene Epoch, and (4) the glaciofluvial erosion and deposition during the late Pleistocene Epoch.

MIOCENE AND PLIOCENE EPOCHS

Starting in early(?) Miocene time and continuing into the early Pliocene, tremendous volumes of basaltic lava, named the Columbia River Basalt Group (Griggs, 1976), were extruded across the Columbia Plateau region. The molten lava flowed over thousands of square miles, and lavas of some flows are now believed to have traveled over 200 miles from their source vents. The continuity of individual flows over many hundreds of square miles indicates that there were virtually no obstructions to the flowage in the area.

Along the marginal areas of the expansive lava field, the flows dammed the peripheral river valleys. The sediments carried by rivers in these valleys were then spread over the edges of the flow, only to be buried by the succeeding flow. The zones of pillows and palagonite at the base of many flows and the silicified logs or stumps between pillows or within the palagonite all show that the lava flows advanced across lakes, streams, and forests. The lakes in which sedimentation took place owed their existence to disruption of drainage by advancing lava flows, and possibly to subsidence of the lava field as well.

The many layers of basalt that underlie the project area are identified as part of the Yakima Basalt of late Miocene and early Pliocene age (Fiske and others, 1963, p. 63). Figure 9 shows the age, name, and thickness of the various Yakima Basalt units. The thicknesses cited in figure 9 apply only along the Columbia River between Frenchman Springs Coulee and Priest Rapids Dam, where they have been measured by Mackin (1961) and others. The geologic map by Grolier and Bingham (1971) shows the areal distribution of the several members within the project area. The geologic map also shows that the oldest exposed flows of Yakima Basalt in the area crop out along the northwestern and northern margins of the area, where they are exposed in the walls of the Columbia River canyon and in the west wall of Lower Grand Coulee. In contrast, the younger basalt crops out in the southern part of the project area. Great dike swarms thought to represent the feeders of at least some of the lava flows occur outside the area (Waters, 1961, p. 586).

Sedimentary deposits were laid down atop basalt surfaces, and then covered by subsequent flows. These interbeds are found principally in the upper 1,200 feet of the Yakima Basalt. The principal deposits are the Vantage Sandstone Member, the Squaw Creek and Quincy Diatomite Beds, and the Beverly Member of the Ellensburg Formation.

The earliest sedimentary unit mapped is the Vantage Sandstone Member of late Miocene age. The Vantage consists chiefly of quartz-feldspar-mica sand, tuffaceous sand, and clay (Mackin, 1961, p. 12) that belong to a mineralogic province foreign to that of the basalt. Thus, the deposits must have had a source outside the lava field. Mackin (1961, p. 12) suggests that they might have been derived from older granitic rocks or from the arkose of the Swauk Formation of early Tertiary age in the northern part of the Cascade Range. The occurrence of Vantage Sandstone Member west of the Columbia River, and its thinning eastward from the Columbia River, further suggest that the sediments were deposited across the edge of the basalt flows and over a broad alluvial plain that extended from the northwest toward the center and around the west edge of the lava field. The ancestral Columbia River was probably much farther east than its present position, beyond the edges of the southeastward extent of the Vantage sediments.

Both the Squaw Creek and Quincy Diatomite Beds are younger than the Vantage Sandstone Member. The Squaw Creek lies below and the Quincy lies above the Roza Member. Each contains gravel, sand, or clay in addition to the predominant diatomite. The diatoms apparently flourished in shallow lakes dammed by the lava flows. The deposition apparently occurred at two different places. The Columbia River probably did not cross the project area during the deposition of these diatomites, because no quartzite-bearing gravels have been found within either deposit.

The Beverly Member of the Ellensburg Formation was deposited during the interval between the solidification of the Priest Rapids flow and the arrival of the lava of the last Saddle Mountains flow. The conglomerate in the Beverly Member, a principal facies at Sentinel Gap, consists largely of quartzite-bearing gravel and sand deposited in the channel of a large stream. The quartzite and most other pebbles and cobbles in the conglomerate near Sentinel Gap are foreign to the basalt province, but are similar to those in the bed of the present Columbia River. This suggests that the ancestral Columbia River occupied a position close to its present course at Sentinel Gap before the extrusion of the last Saddle Mountains flow. The quartzite-bearing conglomerate in the Beverly Member is widespread. It extends southward and westward, outside the project area (Mackin, 1961, p. 3, fig. 1), and eastward from Sentinel Gap (Grolier and Bingham, 1971).

Analyses of fossil leaves and pollen in the sedimentary deposits interfingered with the basalt flows, together with silicified wood within the flows, provide an indication of the climate that prevailed during the late Miocene and early Pliocene time (Estella Leopold and Anne Davis, U.S. Geological Survey, written communication; Prakash and Barghoorn, 1961a and 1961b; Smiley, 1963). The climate appears to have been humid, warm, and temperate, with an annual precipitation of 40 to 45 inches uniformly distributed throughout the year.

Subsidence began prior to the extrusion of the Saddle Mountains basalt, and the surface of the Priest Rapids Member became subject to erosion and deposition. This activity is indicated by the thickness and extent of the Saddle Mountains Member, the peripheral location of the Saddle Mountains flow tongues, and the presence of fanglomerate below the Saddle Mountains flows in some outcrops. In the project area the flows of the Saddle Mountains Member are restricted to the Pasco Basin and the area herein referred to as the Othello Basin; the flow tongues fill erosional valleys cut into the underlying Priest Rapids and Roza flows.

The eastward-trending folds that today are an important part of the physiographic setting apparently began to develop during the period before extrusion of the Saddle Mountains flows. A discordance on the south flank of the Saddle Mountains anticline indicates that the rise of the Saddle Mountains, and possibly of other eastward-trending anticlines in the area as well, began soon after extrusion of the Priest Rapids basalt (Mackin, 1961, p. 31).

As the extrusion of Yakima Basalt waned, the ancestral Columbia River had become fixed around the north side of the lava field. Presumably a permanent drainage system began to develop on the lava field. However, the folding of the basalt in Pliocene and Pleistocene time prevented most of the small local streams from maintaining their courses. Even the Columbia River, which was able to maintain its course across the rising Frenchman Hills and the Saddle Mountains, was deflected by the rise of Umtanum Ridge, outside the project area (Calkins, 1905, p. 41; Mackin, 1961, p. 38).

Events that occurred in the project area after extrusion of the Saddle Mountains basalt, but before early Pleistocene time, are poorly known. However, geomorphic evidence suggests that much of the landscape in the project area is related to erosion in late Pliocene time—the westward-trending valleys of Crab Creek, Rocky Coulee, and Lind Coulee are probably the sites of former master streams deeply entrenched in basalt. These streams presumably flowed across the Quincy Basin, though their former courses cannot be traced because of burial by fill of Pleistocene age. Lind Coulee contains the remnants of a once-extensive fill assigned to the Ringold Formation of Pliocene and Pleistocene age.[1/] The principal exposure consists of flat-lying lacustrine clay which lies against the south wall of the coulee just above the coulee floor in secs. 28 and 34, T. 17 N., R. 32 E. The presence of the Ringold deposits here proves that Lind Coulee had already been excavated to its present depth in pre-Pleistocene time.

Additional examples of the effect that the shape of the basalt surface had on present-day topography are common. Two unnamed southwestward-trending coulees heading in Kansas Prairie change direction abruptly from a southwesterly course, developed upon sloping basalt, to a westerly course at the eastern limit of the Ringold clay. The changes in direction occur on Paradise Flats, about three-quarters of a mile and $4\frac{1}{4}$ miles, respectively, north of Shano (figs. 2 and 4). A third southwestward-trending coulee bends eastward abruptly in the SW$\frac{1}{4}$ sec. 16, T. 14 N., R. 31 E., and follows a course almost parallel to the strike of the underlying basalt and along the lateral boundary between basalt and Ringold clay toward Providence Coulee. The sinuous reach of a steep-walled coulee trending southerly at the east side of the Othello Channels and joining Esquatzel Coulee in the SE$\frac{1}{4}$ sec. 24, T. 13 N., R. 30 E., is in alignment with this third coulee. This sinuous reach probably represents the pre-Ringold lower course of the coulee, exhumed from its cover of Ringold clay by glacial melt water in late Pleistocene time.

Three abandoned channelways breach the western rim of the Quincy Basin. From north to south, they are Crater Coulee, Potholes Coulee, and Frenchman Springs Coulee (fig. 4). They cannot be traced toward any particular coulee along the eastern margin of the basin. Brown tuffaceous sand assigned to the Ringold Formation lies unconformably over several members of the Yakima Basalt along the borders of these channelways. This erosional unconformity between the Ringold sand and the underlying basalt indicates that erosion of the channelways was active in pre-Ringold time. Very likely, some of these channelways in the project area had been excavated to their present depths when deposition of Ringold laminated clay started.

On the basis of physiographic evidence east of the project area, Bretz and others (1956, p. 1000) thought that the Washtucna and Esquatzel Coulees represent the former course of the Palouse River. The occurrence of a remnant of Ringold clay partially filling Hardesty Coulee, a tributary of Washtucna

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1/ This report contains a change in the age designation of the Ringold Formation as shown on the geologic map and sections of Grolier and Bingham (1971); it is reassigned to Pliocene and Pleistocene age (C. A. Repenning, written communication, 1970).

Coulee, indicates that at least part of Hardesty Coulee was eroded before deposition of the clay.

Sand Hollow, at the west end of the Royal Slope, is a pre-Ringold tributary of the Columbia River. The fill of Ringold clay there has been partially removed by erosion.

Several wind gaps across the crest of the Frenchman Hills and the Saddle Mountains are underlain by caliche-capped basalt gravel resting on brown tuffaceous sand assigned to the Ringold Formation. These wind gaps represent the courses of streams that were defeated by the rising anticlines before or during deposition of the Ringold laminated clay.

LATE PLIOCENE AND EARLY PLEISTOCENE EPOCHS

The main period of orogeny in Washington (uplift of the Cascade Range), which began in late Pliocene time, continued into early Pleistocene time. Eastern Washington became increasingly arid after the late Pliocene because the rising Cascade Range was an increasingly effective barrier to the eastward movement of moisture-laden air masses from the Pacific Ocean. As this rain shadow increased, the streams, except the ancestral Columbia and Snake Rivers, diminished in flow and finally disappeared.

Only the Columbia has maintained a course across the rising anticlines since the time of the Beverly deposition. This supports the contention that a change in climate was the principal factor in the disappearance of the streams in the project area; the Columbia and Snake Rivers are the only ones whose watersheds extend outside east-central Washington.

Though the Columbia River always has been a perennial stream, even it was unable to maintain a graded channel across the rising anticlines during at least two periods in the Pleistocene. As a result, lakes and(or) extensive flood plains formed in the intervening and deepening basins. The vast area of Ringold lacustrine deposits gives evidence of a very large, shallow lake that was an outstanding feature of central Washington. Although named Lake Lewis by Symons (1882, p. 108), in this report the lake will be labeled Lake Ringold to differentiate it from the late Pleistocene Lake Lewis II (herein called Lake Lewis). Lake Ringold either may have comprised several separate bodies of water, with each level controlled by a rising anticline, or it may have been associated with the uplift of the Cascades, which would have created a single body of water. The cause for the impoundment of Lake Ringold is the uplift of Horse Heaven Ridge outside the project area (Newcomb, 1958, p. 340). The uppermost level of Lake Ringold is now only relative because of continued deformation of the Yakima Basalt and Ringold Formation in post-Ringold time. The lacustrine deposits occur below an altitude of 1,000 feet in the Pasco Basin, but to the north they lap up over the warped surface of the basalt to altitudes of as much as 1,200 feet. The extent of the lake is inferred from the distribution of the Ringold laminated lacustrine clay and silt deposits.

In the project area, Lake Ringold covered the Pasco Basin, and was connected to the Quincy and Othello Basins by the straits at Sentinel Gap, by the Othello and Drumheller Channels, and possibly by several other saddles that remain as present-day wind gaps on the Frenchman Hills and Saddle Mountains. Arms of the lake extended up the canyon of the Columbia River into Sand Hollow, and eastward up Lind and Washtucna Coulees.

During the existence of Lake Ringold, volcanic ash was ejected from eruptive centers probably located west of the project area, near the present site of the Cascade Range. The debris, blown eastward by the wind, settled and accumulated on land as eolian sand and silt, and fell or was washed into the lake.

The rising anticlinal ridges bounding parts of the lake were eroding rapidly with the result that alluvial fans in places reached or were flooded by the lake waters. Thus, the fans built a piedmont slope of basaltic fanglomerate, with the brown tuffaceous sand and silt as interstitial material. These sediments interfinger with the lacustrine material along the slopes of the ridges.

The structural activity controlling sedimentation in the project area during late Pliocene and early Pleistocene time remained similar to that during deposition of the Beverly Member of the Ellensburg Formation; the basins were deepening and the anticlinal ridges were rising. The downwarping and uplift apparently continued throughout, and probably after, the time of deposition of the laminated clay in Lake Ringold. For example, in the eastern part of the project area, the Ringold Formation abuts disconformably against the basalt of the rising upland, and on the Taunton Terrace the beds of Ringold laminated clay are upturned about 20°. Also, basaltic fanglomerate accumulated along the flanks of rising anticlines and on piedmont slopes, where it interfingers with Ringold laminated clay deposited in the basin centers. On the other hand, there is no evidence to show that the laminated clay ever extended over the areas that are now the Frenchman Hills and Saddle Mountains; this indicates that deformation there could have been well advanced prior to deposition of the clay.

During the middle of the Ringold deposition, a change in the barrier forming Lake Ringold permitted the lake to overflow and the Columbia River to deposit a swath of sorted gravel across the Pasco Basin. According to Newcomb (1958, p. 336), the river shifted laterally in its course between Sentinel Gap and Wallula Gap (south of project area). The subsurface extent of the Ringold conglomerate in the project area (fig. 21) shows that just north of Pasco the Columbia River was at one time nearly 9 miles east of its present position.

Following the deposition of the conglomerate (fig. 25), during which time the river flowed across the shallow deposits it had brought to Lake Ringold, the basin again held a lake long enough for an additional 400 to 500 feet of silt, clay, and fine sand to be deposited. Final breaching of the enclosing rim permitted the Columbia River to deeply erode the soft deposits through which it now flows between Sentinel and Wallula Gaps.

Periodic climatic fluctuations during accumulation of the Ringold Formation are indicated by both the numerous caliche layers intercalated within the brown sand of the Ringold and the caliche layers that overlie the conglomerate at the White Bluffs. The massive caliche that caps the Ringold Formation also shows that a long interval of aridity followed the Ringold deposition.

PLEISTOCENE EPOCH

Middle Pleistocene

The arid climatic regime during the early Pleistocene probably was followed by a cycle of greater rainfall, as shown by the remnants of a once prominent local drainage network that cut through the caliche caprock. The climatic trend toward increased rainfall (probably contemporaneous with renewed deformation of basalt and a lowering of the regional base level, or possibly a combination of these factors) started a new cycle of erosion on the caliche-capped Ringold Formation. Remnants of the old drainage are strikingly preserved on the Royal Slope and in parts of the Pasco Basin.

Remnants of a Ringold fill, once continuous in the Othello Basin, underlie the northeastern part of the Royal Slope and the Smyrna Bench. The massive caliche that caps the Ringold clay there is breached by dry gulches that were once occupied by deeply incised streams. The lower ends of the gulches are abruptly truncated; near the valley of the master stream which they joined, erosion has removed the Ringold fill where large proglacial streams flowed westward along the Othello Basin in late Pleistocene time.

In the Pasco Basin, the principal valleys that originated during the same cycle of erosion are: (1) the V-shaped valley trending westward from sec. 3, T. 14 N., R. 28 E., to sec. 2, T. 14 N., R. 27 E., at the east end of the Wahluke Slope (McKnight, 1927, p. 459; Bretz, 1927, p. 461); (2) the northwestward-trending coulees along the White Bluffs—Parsons Canyon (fig. 7), Baxter Canyon and Rankin Canyon; and (3) the gulches that dissect the southward-facing bluffs underlain by the Ringold Formation in the NW¼ T. 10 N., R. 29 E., and the SW¼ T. 11 N., R. 29 E.

FIGURE 7.—Aerial view of Parsons Canyon, looking southeast from White Bluffs. (U.S. Bureau of Reclamation photo, 1954.)

Late Pleistocene

Great Flooding and Runoff

The events that ultimately led nearly to destruction of the drainage system that had developed on the Ringold Formation were related to great floods of melt water from regional glaciation in western North America. Several times late in the Pleistocene Epoch large Cordilleran ice sheets advanced southward from Canada to the Columbia River north of the project area. The glacier was split into lobes by the north-south ranges of the Okanogan Highlands and Selkirk Mountains. Some of these lobes blocked the Columbia River and its tributaries to form extensive glacial lakes. Two of these glacial lakes, Columbia and Spokane, overflowed their divides and spilled southward, cutting the Grand Coulee and the Wilson Creek-Telford scabland tract, respectively. The greatest quantity of water was released by failure of the ice dam that formed glacial Lake Missoula (Pardee, 1918, 1948; Bretz, 1930, 1959; Bretz and others, 1956). Tremendous volumes of water flowed down the channels in the regional slope of east-central Washington to the Snake and Columbia Rivers. The water combined with melt water from the west side of the most westerly lobe (Okanogan lobe) and from all the alpine glaciers on the eastern slope of the Cascade Mountains before funneling through the Horse Heaven Hills at Wallula Gap.

In addition to melt-water discharge from the north, the Snake River also received a sudden influx of water in late Pleistocene time. This occurred when Lake Bonneville overflowed at Red Rock Pass, near Preston, Idaho, and discharged into the Portneuf River, which joins the Snake at Pocatello, Idaho (Gilbert, 1890, p. 173; Malde, 1960, p. 295; Trimble and Carr, 1961, p. B164-B166; Sterns, 1962, p. 386).

Overtaxed by the influx of glacial melt water from so many sources, the preflood drainage system was largely destroyed, the coulees and canyons in and adjacent to the project area were cut, and much of the upland was cleaned of its loess mantle. The spectacular landforms that remain today, long known by their pioneer name of scabland and renamed even more appropriately "channeled scabland" by Bretz (1919), are attributable to the effects of over-

FIGURE 8.—Relief map of Columbia Plateau of eastern Washington, showing pattern of coulees of channeled scabland resulting from rapid erosion by the Spokane Flood of late Pleistocene time. Relief base from landform map of Washington, courtesy of Molenaar (1976).

fit streams on the land. Figure 8 shows the pattern in eastern Washington of the coulees resulting from the great floods.

The Columbia River, swollen by glacial melt water, first overflowed its banks and entered the Quincy Basin from the west through Crater and Potholes Coulees. The flow in this direction is indicated by caliche-capped gravel near the head of the two coulees with foreset beds dipping southeastward away from the river (Bretz and others, 1956, p. 985).

During a much later period, torrential streams entered the Quincy Basin through Grand Coulee, Dry Coulee, the valley of Crab Creek, Black Rock Coulee, Rocky Coulee, Bowers Coulee, and Lind Coulee.

The water entering the Quincy Basin from the north and east was temporarily ponded but subsequently flowed westward out through the three preexisting channelways to the west (Crater Coulee, Potholes Coulee, and Frenchman Springs Coulee) in a direction opposite to that of the earlier flow in these coulees. The floodwater also escaped, possibly later, through a gap at the east end of the Frenchman Hills at the present sites of the Warden and Drumheller Channels; this was thereafter the main exit from the Quincy Basin.

Probably the reason that the Drumheller Channels became the dominant outlet of the water temporarily ponded in the Quincy Basin is because the floor of the gap stood at an altitude lower than those of the three western channelways. The high-gradient torrential streams, which flowed down the south slope of the Frenchman Hills at the present site of the Drumheller Channels, removed the Ringold cover and excavated deep channels by plucking the columnar, vertically jointed basalt. The resulting waterfalls migrated upstream across the low end of the Frenchman Hills anticline and other subordinate anticlinal flexures north of it.

The direction of flow from the Drumheller Channels at first was mostly southeastward to the Othello Channels, over and around the east end of the Saddle Mountains. Later, after spillover had breached the Ringold fill across the lower part of the Royal Slope, the flow of melt water was westward down Lower Crab Creek valley. A small wind gap across the crest of the Saddle Mountains, with a sill at an altitude of 1,157 feet in the SW¼NE¼ sec. 5, T. 14 N., R. 29 E., was used as a spillway by ponded glacial melt water, probably before Othello Channels and Lower Crab Creek valley were fully excavated. A small waterfall with closed depressions was eroded in the basalt on the south slope of the Saddle Mountains, below the water gap in the SE¼SW¼ sec. 5.

South of the Othello Channels, the flow of melt water was split into several distributaries. Melt-water streams removed most of the Ringold fill overlying the basalt, and the preexisting drainage system that had developed on the Ringold Formation was largely removed. The melt water flowing southwestward toward the Columbia River enlarged and deepened preexisting stream channels now represented by Koontz Coulee, Ringold Coulee, and three smaller unnamed coulees across Columbia Flat. Melt water from the Othello Channels at times merged with that entering the project area through Washtucna Coulee between Connell and Mesa. The combined flow eventually discharged through the lower part of Esquatzel Coulee and the unnamed coulee east of Esquatzel Coulee, then into the proglacial Snake and Columbia Rivers.

The Snake River, likewise swollen by melt water and with a gravel-loaded bed above its present level, overflowed its banks a short distance upstream from the project area and swept across Snake River Flat. It removed loess and some Ringold lacustrine clay that overlie basalt there, and left a bar of gravel in its wake. The residual hills of loess and the gravel bar have been described and illustrated by Bretz (1925a, p. 108, fig. 5).

The melt water emptying into the basins brought in great volumes of basaltic gravel and sand. Gravel bars in the northern and eastern coulees are fluvial forms that were built from the bedload of these streams (Bretz and others, 1956, p. 977, pl. 8). An excellent example in Lind Coulee is provided by the bar near Roxboro. This deposit is the coarse bedload dropped because of a decrease in velocity caused by ponded water in the Quincy basin. Another example is the Priest Rapids bar, at the west end of Wahluke Slope, which is considered the remnant of a deltaic deposit in Lake Lewis during a stillstand at an altitude of about 860 feet.

The presence of ponded water in the Quincy Basin also is suggested at many places by the occurrence of two deposits of lacustrine silt, resembling the Touchet Beds of Flint (1938). One deposit of this lacustrine silt and associated erratics occurs up to an altitude of 1,400 feet (based on aerial-photo interpretation) in Bowers Coulee near the project boundary, and the other deposit reaches a common altitude of about 1,150 feet. Ice-rafted boulders occur on the north slope of the Frenchman Hills at altitudes as high as 1,360 feet in sec. 21, T. 17 N., R. 26 E. (Bretz and others, 1956, p. 999). That altitude is more than 200 feet above the floor of Quincy Basin. Similarly, faint shorelines southwest of the mouth of Lower Grand Coulee in sec. 23 and the SE$\frac{1}{4}$ sec. 14, T. 22 N., R. 26 E., are littered with ice-rafted debris at an altitude of 1,400 feet, more than 250 feet above the present level of Soap Lake. The highest scabland along the south slope of Pinto Ridge, north of the Crab Creek valley, also stands at much higher altitudes than any of the outlets used by glacial melt water in the Quincy Basin.

The highest recognizable spillways definitely attributable to the glacial outwash streams active in late Pleistocene time are: (1) a narrow channel in the SE$\frac{1}{4}$ sec. 15, T. 17 N., R. 30 E., 1 mile east of the eastern wall of the Warden Channel, whose floor stands at 1,315 feet and is underlain by channeled caliche; (2) a small scabland channel that is a part of the large plexus of channels at Frenchman Springs Coulee near the center of sec. 17, T. 18 N., R. 23 E., and stands at an altitude of 1,350 feet (Bretz and others, 1956, p. 999); and (3) a narrow channel that crosses Evergreen Ridge near the center of sec. 5, T. 18 N., R. 23 E., with a threshold of channeled caliche and basalt at an altitude of 1,383 feet.

Origin of, and Sedimentation in, Lake Lewis

The first runoff of large volumes of glacial melt water met with considerable resistance because of the limited capacity of the preglacial valleys and water gaps. The Columbia Gorge, deeply cutting the Cascade Range between Oregon and Washington, was the main water gap where obstruction occurred. The lake that was impounded upstream from this obstruction extended over the valley of the Columbia River and its tributaries to a level of 1,150 feet in the Pasco Basin (Bretz, 1919, p. 498-499, fig. 2; 1928b, p. 325-328; 1928c, pl. 5; Bretz and others, 1956, p. 983; Flint, 1938, p. 496-498, fig. 5; Bryan, 1927, p. 27; Lupher, 1944; Newcomb, 1962, p. 70). This is the submergence to which the name Lake Lewis applies.

The lacustrine silt deposited in Lake Lewis in the Pasco Basin is the Touchet Beds of Flint (1938). The time of sedimentation in Lake Lewis is uncertain relative to the time of channeling by glaciofluvial streams. At places, the lacustrine silt occurs well above the principal channelways that carried glaciofluvial discharge in late Pleistocene time. Thus, deposition of the lacustrine silt could have taken place in a lake extending over the area before the first torrential streams came into existence, or between two episodes of glaciofluvial discharge. However, Royal Slope, Wahluke Slope, Owens Flat, Columbia Flat, and the sites of Drumheller and Othello Channels were not submerged when the Ringold fill was channeled at these localities. Also, sedimentation in Lake Lewis preceded the last channeling episode on the Royal Slope and upper part of the Pasco Basin, because the lacustrine silt is not found in the scabland tracts or other melt-water channels there. The last settling of lacustrine silt in Lake Lewis took place after major use of the Snake River by glacial melt water, when the lake level was about 550 feet. The lacustrine silt now found on poorly defined terraces within the canyon of the Snake River would have been washed away by the stream currents had the Snake River discharged large volumes at a later time. Likewise, the presence of Touchet silt deposits at an altitude of only about 400 feet at old Wallula (just south of project area) beside the Columbia River indicates that its deposition postdated any great flows of water through Wallula Gap.

HOLOCENE EPOCH

The advent of the Holocene Epoch was marked by (1) a change in climate, (2) recession of continental ice from northern Washington, and (3) cessation of glaciofluvial discharge entering the project area. According to Hansen (1941, p. 503), the Holocene Epoch included "an initial cool and subhumid period, followed by desiccation and warming, and a final period of cooling and possible increase in moisture."

Present-day streamflow, small in comparison to that of the Pleistocene glaciofluvial discharge, has been unable to further modify the scabland channels; this has resulted in a general lack of stream integration within the project area. The lack of integration is indicated by the many short tributary gullies, the damming of streams by sand dunes, and the presence of many closed depressions in the abandoned glaciofluvial channelways that are occupied by water-table lakes. These lakes rarely overflow because of the lack of surface runoff within the individual basin, the large evaporation rate, and the presence of gravel-filled subsurface outlet channels in some places. Some of the lakes that have the least surface or subsurface outlets have become saline because of the concentration of solutes due to evaporation of the incoming ground water.

In the Holocene Epoch the processes of degradation and aggradation are active, as in the Pleistocene, but to a considerably lesser extent. Talus slopes, landslides, and gully erosion are degrading parts of the project area while stream deposition, sand dunes, and loess accumulation are aggrading other parts.

Wind deflation and redeposition of sand and silt are the processes that have most intensely modified the land surface in the project area during the Holocene Epoch. In the Quincy Basin and on the Pasco Slope, the glaciofluvial sand and lacustrine silt have been the source material for the dunes. In the Quincy Basin the deposits of sand have dammed the course of Crab Creek to form Moses Lake.

In the present climate, mechanical weathering is more dominant than chemical weathering, and the principal change in the basaltic cliffs has been the accumulation of talus. The valleys that were not actively eroded in late Pleistocene time have few, if any cliffs; they generally have smooth slopes because the deposition of loess on top of the talus has provided a protective cover to slow mechanical weathering.

Landslides occurred during and immediately after the last glaciofluvial flooding. They are found in several places in the project area, largely where the glacial streams had undercut steep slopes. The largest landslide (about 4 square miles) is on the north slope of the Saddle Mountains between Taunton and Corfu. Other large landslides have occurred on the slopes of the Saddle Mountains and the White Bluffs. No slides are known to have occurred late in Holocene time.

Gully erosion is active along the mountain slopes. Elsewhere in the project area, ephemeral streams are entrenched 3 to 10 feet into Holocene soil or alluvium. The start of this entrenchment by surface runoff apparently coincides with or has been accelerated by man's settlement with its consequent cultivation and grazing.

STRATIGRAPHY

MIOCENE AND PLIOCENE ROCKS

Yakima Basalt

The Yakima Basalt is the younger of two formations of basalt included in the Columbia River Basalt

System	Series and Subseries	Group	Nomenclature proposed by Mackin (1961): Formation	Nomenclature proposed by Mackin (1961): Member and unit	Nomenclature used in this report[1/]: Formation	Nomenclature used in this report[1/]: Member or unit	Thickness in feet
QUATERNARY	Holocene				Fluvial, lacustrine, and eolian deposits	Loess	0-25
						Dune sand	0-75
						Alluvium	0-20
						Colluvium	0-35
						Landslide debris	0-200
	Pleistocene					Lacustrine fine sand and silt	0-75
						Glaciofluvial gravel and sand	0-225
						Older alluvium	0-30
					Ringold Formation	Basaltic fanglomerate	0-260 (0-800)
						Buff laminated clay	0-500
TERTIARY	Pliocene					Quartzitic conglomerate	0-200
						Brown tuffaceous sand	0-100
						Gravel deposits in wind gaps	0-30
			Ellensburg Formation	Saddle Mountains Basalt Member	Ellensburg Formation	Beverly Member / Saddle Mountains Member	0-300 (0-400)
				Beverly Member Huntzinger Flow			
	Late Miocene	Columbia River Basalt Group[1/]	Yakima Basalt	Priest Rapids Member: Flow No. 4, Flow No. 3, Flow No. 2, Flow No. 1	Yakima Basalts	Priest Rapids Member	0-200
				Priest Rapids Member: Quincy Diatomite		Quincy Diatomite Bed	0-35
				Roza Basalt Member		Roza Member	0-200
				Frenchman Springs Basalt Member: Squaw Creek Diatomite Bed		Squaw Creek Diatomite Bed	0-5
				Frenchman Springs Basalt Member: Sentinel Gap Flow, Sand Hollow Flow, Gingko Flow		Frenchman Springs Member	250-375
				Vantage Sandstone Member		Vantage Sandstone Member	0-35
				Museum Basalt Member		Lower basalt flows	1000+
				Rocky Coulee Basalt Member			
				Unnamed flows			
	Middle Miocene				Picture Gorge Basalt	Not exposed if present	

FIGURE 9.—Stratigraphic sequence and relationship of principal rock units described in this report, and comparison with nomenclature of Mackin (1961).

1/ With the exception of the term Columbia River Basalt Group and the age of the Ringold Formation, the nomenclature and age assignments used herein follow that of the geologic map and sections by Grolier and Bingham (1971). However, the authors recognize that subsequent work by others has since revised the dating of some formations, members, and units.

Group (Griggs, 1976) which is of middle Miocene through Pliocene age. The older formation, termed Picture Gorge Basalt by Waters (1961, p. 591) is of middle Miocene age; the Yakima Basalt (see footnote to fig. 9, p. 20) is of late Miocene and early Pliocene age (Fiske and others, 1963, p. 63). Only the Yakima Basalt is known to occur in the project area. It underlies the entire area and forms the "bedrock" upon which rest sedimentary rocks of younger age.

Mackin (1961, p. 8, fig. 2) has described and named the units in the uppermost part of the Yakima Basalt. Later, Bingham and Grolier (1966) adopted a slightly modified version of Mackin's nomenclature, as shown in figure 9. The modified terminology is used in the present report.

The Yakima Basalt is a dark-gray, fine-textured, well-jointed, extrusive rock that occurs in individual lava flows, many of which are of great lateral extent. The overall appearance is monotonous, but many flows have distinctive features such as color, texture, degree of crystallization, hardness, porosity, vesicularity, thickness, breakage habits, kind of jointing, alteration, and permeability. The porphyritic nature of the basalt in some flows and the presence of a few sedimentary beds interstratified in the flows greatly facilitate mapping. Taken separately, any single feature is not adequate to trace individual flows laterally, but the constant association of several of these characteristics in one flow, when considered in relation to the characteristics of the overlying and underlying flows, is sufficient for stratigraphic identification of some flows over distances of tens of miles. For example, on the basis of its distinguishing characteristics, the Roza Member was traced by Bingham and Walters (1965, C87) to the vicinity of Pullman, 175 miles east of its type section in the Yakima Canyon. The geologic map by Grolier and Bingham (1971) shows that members of the Yakima Basalt are traceable across the length and breadth of the project area.

Field observations of the basalt are supplemented by subsurface data obtained from selected drillers' logs (Appendixes II and III). Four deep wells, drilled in search for oil within and outside the project area, have penetrated thicknesses of Yakima Basalt much greater than 1,700 feet. One of the wells, 21/31-10M1 (Appendix I for well-numbering system) drilled in 1960 at an approximate altitude of 1,610 feet, 5 miles east of the project boundary, penetrated 4,465 feet of basalt before entering 202 feet of shale and sand atop a granitic basement.

The two deepest wells in the project area were drilled in the Frenchman Hills. According to well drillers' logs and other reports, they encountered only basalt with occasional beds of tuff, shale, and sandstone. One well (17/24-23H1) at an altitude of 1,238 feet was drilled to a depth of 2,280 feet. The other (17/28-19D1) at an altitude of 1,443 feet was drilled in 1934 to a depth of 4,570 feet.

Another deep well was drilled in 1957 on the Rattlesnake Hills in Benton County at an altitude of about 2,875 feet in the SE$\frac{1}{4}$SE$\frac{1}{4}$ sec. 15, T. 11 N., R. 24 E., about 25 miles west of the Columbia River at Ringold. The well is reported to have penetrated 10,650 feet of basalt and associated volcanic sediments (M. T. Huntting, written communication, 1961).

During the field investigation, efforts were made to locate the sources of the basalt flows. Although no plug or feeder dike was observed that could be described with certainty, a circular depression 300 feet in diameter in the Priest Rapids Member could be the surface expression of a spiracle or a flooded feeder vent. It is located north of Upper Goose Lake, near the western end of the Drumheller Channels in SE$\frac{1}{4}$ sec. 23, T. 17 N., R. 28 E. The outer margin of the ringlike depression consists of columnar basalt (fig. 10), but within the concavity, the basalt is platy and rises toward the center in slabby swirls (fig. 11). This feature was first recognized by Bureau of Reclamation geologists (G. E. Neff, oral communication, 1960) and is similar to much larger rings (locally known as "craters"), which occur in the Roza Member near Odessa, outside the project area and 39 miles

FIGURE 10.—Outer rim of ringlike depression in basalt flow of Priest Rapids Member in SE¼ sec. 23, T. 17 N., R. 28 E. (U.S. Bureau of Reclamation photo.)

FIGURE 11.—Platy swirls near center of ringlike depression in basalt flow of Priest Rapids Member in SE¼ sec. 23, T. 17 N., R. 28 E. (U.S. Bureau of Reclamation photo.)

northeast of the ring in Drumheller Channels.

Another feature is the linear exposure of porphyritic Saddle Mountains basalt in the NE¼ T. 13 N., R. 29 E., which was shown on the geologic map of Grolier and Bingham (1971); this feature is in alignment with the regional structure of eastern Washington. Although it may be simply a valley fill, it may be related to a deep-seated structure such as a dike. The trend of N. 15° W., which is transverse to the regional slope, is not likely to be that of a valley fill. This outcrop also is in a straight alignment with several other exposures of porphyritic Saddle Mountains basalt to the southeast; this further suggests a structural rather than an erosional relationship.

Lower Basalt

All Yakima Basalt flows underlying the Vantage Sandstone Member are considered together and are referred to in this report as the "lower basalt." The color and texture of the lower basalt reflect its mineralogic composition. The basalt is dark gray to black, even when weathered. It is very fine grained to aphanitic, and differs from the overlying members of the formation in its tendency toward small splintery columnar jointing and, in many places, by spectacular fan jointing. The bases of some of the flows consist of thick zones of coarse rubble or agglomerate, which are rare or absent at the bases of most upper members of the Yakima Basalt. Waters (1961, p. 598) has determined the mineral composition of basalt flows exposed east of Crescent Bar which are representative of the lower basalt. Waters believes that the most distinctive mineralogical characteristics of the basalt are a high percentage of glass—nearly 30 percent—and an almost complete absence of olivine. The flows there average 31 percent plagioclase, 28 percent pyroxene, and nearly 11 percent opaque and alteration minerals.

The exposed top of the lower basalt is a surface of low relief in most places, developed by the easy erosion and removal of the soft Vantage Sandstone Member that overlies the lower basalt. Under normal differential erosion, which controls the retreat of basalt layers in upland areas, this stripped structural surface is the most prominent of many such surfaces on the basalt. In the coulees, the lateral removal of the overlying basalt has been accelerated by stream erosion of the Vantage Sandstone Member. Some examples of this stripped surface are the Babcock Bench (including its southward extension to Frenchman Springs Coulee), the surface followed by Interstate 90 for 4 miles north of the Vantage bridge, the bench above Brook Lake and Long Reservoir (renamed Billy

Clapp Lake), and the floor of the coulee tributary to Lower Grand Coulee east of the Great Blade.

The lower basalt is exposed where erosive processes have been most active. For example, outcrops are restricted mainly to the canyon of the Columbia River, to Lynch Coulee, the west end of Crater Coulee, the slopes of the Beezley Hills, Lower Grand Coulee, Dry Coulee, low benches in the eastern part of upper Crab Creek valley, and the north slope of the western half of the Saddle Mountains.

Vantage Sandstone Member

The Vantage Sandstone Member of late Miocene age overlies the lower basalt and underlies the Frenchman Springs Member. Roadcuts along old U.S. Highway 10, near the mouth of Schnebly Canyon west of the project area, were chosen by Mackin (1961, p. 12) as the type locality of the Vantage Sandstone Member. The sedimentary deposits that form the member are oxidized to pale yellow where exposed, but are reported as blue or green by drillers. The Vantage Sandstone Member consists of medium-grained, friable quartz-feldspar-mica sand, or of a weakly cemented tuffaceous sand containing hornblende. This distinctive tuffaceous sand is generally overlain by tuffaceous silt and clay mixed with quartz and mica. Poorly preserved carbonized twigs and plant impressions commonly occur in a black earthy zone near the top of the tuffaceous silt and clay.

The member is about 35 feet thick in the Vantage-Priest Rapids area, but thins to the northeast. In a railroad cut near Bacon siding in SE$\frac{1}{4}$ sec. 13, T. 23 N., R. 27 E., the Vantage Sandstone Member is no more than 1 foot thick.

The Vantage Sandstone Member is laterally continuous in the northern part of the project area, where it constitutes an important marker bed in the logs of many wells. The member rarely crops out along Coulee walls because it is concealed by talus that has fallen from overlying flows onto the margin of the lower basalt bench. The sandstone is completely stripped from most of this bench, and natural exposures, therefore, are few. The main exposures are in the creek bed of Lynch Coulee in the NW$\frac{1}{4}$ sec. 27, T. 21 N., R. 23 E.; at the bottom of gulleys dissecting the southeast slope of the Beezley Hills, southwest and northeast of Ephrata; in Lower Grand Coulee, about 1 mile north of Soap Lake, where a monoclinal flexure causes the basalt to rise sharply toward the northern upland; at the northwest corner of Frenchman Springs Coulee, in the SE$\frac{1}{4}$ sec. 19, T. 18 N., R. 23 E.; along Interstate 90, in the NW$\frac{1}{4}$ sec. 5, T. 17 N., R. 23 E., where the Yakima Basalt is thrust faulted along the north flank of the Frenchman Hills anticline; and in a gully at the west end of Smyrna Bench, in the SW$\frac{1}{4}$ sec. 34, T. 16 N., R. 25 E., where the rocks are overturned and faulted along the north limb of the Saddle Mountains anticline. Also, the Vantage Sandstone Member crops out on the deeply dissected southern slope of the Saddle Mountains, near the W$\frac{1}{4}$ corner sec. 7, T. 15 N., R. 26 E., and on the north slope in SW$\frac{1}{4}$NE$\frac{1}{4}$ sec. 3, T. 15 N., R. 25 E. Excellent manmade exposures occur in railroad cuts and roadcuts along State Highway 28 in the lower part of Lynch Coulee, and at the north end of Dry Coulee.

Frenchman Springs Member

The Frenchman Springs Member of late Miocene age overlies the Vantage Sandstone Member and underlies the Roza Member. Near Sentinel Gap it consists of the Gingko flow (lowest), Sand Hollow Flow, Sentinel Gap Flow, and the Squaw Creek Diatomite Bed (highest). The type locality selected by Mackin (1961, p. 13) is in Frenchman Springs Coulee (fig. 6). However, there the Sentinel Gap Flow, the stratigraphically highest flow, is missing as it did not extend that far north. The three flows, and possibly additional ones that are part of the member in Washtucna Coulee, are not differentiated and are mapped together on the geologic map of Grolier and Bingham (1971).

The basalt of the Frenchman Springs Member is dark gray to black, medium grained to aphanitic, and sparsely porphyritic. The mineral composition of a sample from Babcock Ridge is 34.6 percent plagioclase, 21.4 percent monoclinic pyroxene, 8.0 percent olivine, 6.9 percent opaque minerals, and 27.7 percent glass (Waters, 1961, p. 602, sample 10).

At its type locality in Frenchman Springs Coulee, the member is about 250 feet thick, but where the Sentinel Gap Flow is present, its thickness increases to about 375 feet. The member thins northward and northeastward from Soap Lake.

The areal distribution of the Frenchman Springs Member is very similar to that of the lower basalt. Exposures of Frenchman Springs basalt are restricted mostly to the Columbia River canyon walls, the Beezley Hills, Lind Coulee, the Drumheller Channels at the east end of the Frenchman Hills, and the Saddle Mountains west of Corfu.

The two characteristics useful in identifying the Frenchman Springs basalt are (1) the character and density of phenocrysts and (2) the existence of pillows and palagonite in the basal flow. Jointing habits of the Frenchman Springs basalt do not differ sufficiently from those of the overlying basalt members to be used independently as reliable guides in identification.

The Frenchman Springs basalt is identified by the presence of large, internally shattered plagioclase phenocrysts that are partly resorbed and weakly zoned. The phenocrysts occur as single crystals or in clusters and are roughly equidimensional, from one-half to 1 inch in diameter. They are transparent and clear at some localities, but are more commonly slightly stained by limonite. In places, the dense, columnar part of a flow may show only one large phenocryst per 1,000 square feet of exposure and, as Mackin (1961, p. 13) points out, "Clusters may occur near the top, near the base, or anywhere in the flow . . . ." Generally, pillows at the base of a flow and the hackly tabular jointed lava of some flow units are extremely porphyritic, but the typically large shattered phenocrysts may be rare even there.

Where a Frenchman Springs basalt flow is extremely porphyritic, and where the phenocrysts cannot be readily distinguished from those in other porphyritic members of the Yakima Basalt, definite stratigraphic identification of the Frenchman Springs basalt is dependent upon the position of the flow relative to the underlying and overlying members. The upper Frenchman Springs flow, exposed in the coulees of the northeastern part of the project area, is a case in point. The irregular distribution of phenocrysts in the Frenchman Springs Member is most obvious in the Sentinel Gap Flow (not separately mapped), which is almost nonporphyritic where it is columnar, and sparsely porphyritic where hackly or altered to palagonite.

The basal part of the Ginkgo flow, lowest flow of the Frenchman Springs Member, consists of a zone of pillows and palagonite that is 50 to 75 feet thick in the Vantage-Priest Rapids area. It is best exposed in a highway cut at the junction of Sand Hollow and the valley of the Columbia River (fig. 12).

FIGURE 12.—Pillow-palagonite zone in basal part of Frenchman Springs Member (Ginkgo flow) at mouth of Sand Hollow.

This zone has been traced to the northeastern part of the project area, where it is 20 to 30 feet thick. In the Vantage-Saddle Mountains area, it contains many petrified logs that lie partly or entirely imbedded between pillows or in palagonite. This zone is hidden under talus in parts of the project area.

Mackin (1961, p. 21) thought that in the Squaw Creek drainage basin, about 30 miles west of the project area, a bed of diatomite resting on Frenchman Springs basalt was "deposited in a lake impounded by the Sentinel Gap flow . . . ." This unit, named the Squaw Creek Diatomite Bed (Bingham and Grolier, 1966, p. 8), is thought to be part of the Frenchman Springs Member on the basis of its lateral equivalence and genetic relationship to the Sentinel Gap Flow.

East of the Columbia River, within the project area, the only observed exposure of the diatomite is an outcrop about 15 feet long and 5 feet thick in Frenchman Springs Coulee, in NE$\frac{1}{4}$NE$\frac{1}{4}$ sec. 29, T. 18 N., R. 23 E. Elsewhere in the project area, a thin bed of clay or diatomaceous siltstone (at most places less than 1 foot thick), silicified wood occurring as stumps, and odd-shaped concretions of silica mark the contact between the Frenchman Springs Member and the overlying Roza Member.

Roza Member

The Roza Member overlies the Frenchman Springs Member and underlies the Priest Rapids Member of the Yakima Basalt. The type locality of the Roza Member was selected by Mackin (1961, p. 21) as "a scarp on the . . . . east side of the Yakima River opposite Roza station." (Roza station is a railroad siding in the Yakima River canyon between Ellensburg and Yakima, approximately 25 miles southwest of the Vantage bridge over the Columbia River.) The Roza Member lies above the Frenchman Springs Member and its associated Squaw Creek Diatomite Bed, and is overlain in turn by the Quincy Diatomite Bed of the Priest Rapids Member.

The Roza basalt is dark bluish gray and weathers to a deep red-brown color. It is medium to coarse grained, porphyritic, and contains plagioclase, monoclinic pyroxene, iron-rich olivine, opaque minerals, and glass (Waters, 1961, p. 602, sample 8). The phenocrysts are plagioclase crystals that are light yellow or amber and transparent; they are lath shaped and average about 1 centimeter in length and one-half a centimeter in width.

The average thickness of the Roza Member is 100 feet, on the east wall of the Columbia River canyon. The basalt pinches out north of Potholes Coulee and thins north of the Crab Creek valley east of Adrian. It thickens eastward along the axis of the Frenchman Hills anticline, and is over 200 feet thick in the Drumheller Channels, where it consists of two flows.

The Roza basalt has a columnar jointing system through most of its thickness. Columns are 5 to 10 feet in diameter and tend to break into slabs and chips along platy joints normal to the column axis. In many places, these horizontal platy zones grade upward into a zone of swirling plates that characterize the upper parts of the columns. Wherever platy parting is well developed, the Roza basalt is coarser grained than average, and plagioclase phenocrysts look dark in the groundmass. In some places uniform pinches and swells in the column faces give the appearance of helical fluting irregularly along the columns. They are spectacularly displayed in the north wall of the south alcove at Frenchman Springs Coulee (fig. 6; Mackin, 1961, pl. 6A, p. 20).

The Roza Member is exposed high along the east wall of the Columbia River canyon from Sentinel Gap to a point 2$\frac{1}{2}$ miles north of Potholes Coulee, where it ends abruptly. It is exposed largely in the eastern part of the Quincy Basin, high along coulee walls. In Lower Grand Coulee it composes part of the Great Blade (fig. 13), and the scabland tract to the southeast. Roza basalt underlies most of the Royal Slope west of Low Gap and north of Sand Hollow. It

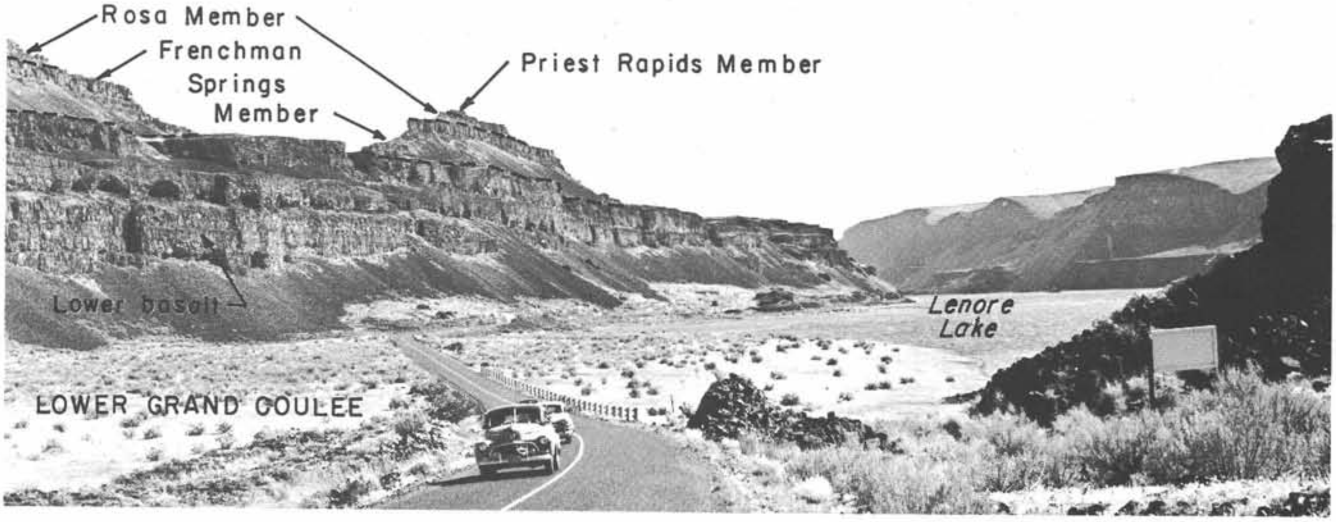


FIGURE 13.—Lower Grand Coulee, looking south, showing Roza basalt flow contacts exposed near the Great Blade (narrow ridge on skyline above automobiles). Note hanging valleys along wall of coulee in right background. (U.S. Bureau of Reclamation photo.)

crops out in the northern half of the Drumheller Channels, along the north and south slopes of the Saddle Mountains west of Corfu and, prior to submergence there in 1952, it was observed on the shore of Eagle Lakes in the Othello Channels (G. E. Neff, oral communication, 1962).

The Roza is the member of the Yakima Basalt most easily recognized in the field. It is the uppermost of several porphyritic flows (including those of the Frenchman Springs basalt) that are collectively referred to as "feldspar flows" by Bureau of Reclamation geologists, who have been using them for many years as a key to the stratigraphy and structure in their work on damsites in the northern half of the project area (W. H. Irwin 1941 and F. O. Jones 1945, unpublished U.S. Bureau of Reclamation reports).

The upper part of the Roza is easily recognized in the area bounded by Black Rock Coulee, Spring Coulee north of Pinto Ridge, and the course of Crab Creek between Adrian and the mouth of Black Rock Coulee.

Porphyritic basalt of the Roza Member is exposed east of Connell, high along the walls of Washtucna Coulee and its tributaries. Where the section is best exposed near Sulphur Lake, the stratigraphic sequence includes a coarse-grained gabbroic basalt amidst other flows of the Frenchman Springs Member, and also a nonporphyritic flow (possibly equivalent to Mackin's Sentinel Gap Flow) underlying Roza basalt. Basalt of the Roza Member is also exposed at the mouth of Providence Coulee and between Connell and Mesa.

In the eastern part of the project area, the lower flows of the Yakima Basalt and, rarely, the Frenchman Springs Member, which are necessary to make a positive identification of the Roza in the field, are not exposed. Also, in the upland and along the upper part of coulee walls, bedrock is generally concealed by loess or covered with talus. The Roza Member probably underlies the largest part of the loess-mantled upland from Hatton Coulee to Rocky Coulee, but there exposures are few, and identification of the basalt is based on lithology only.

The Roza basalt cannot be recognized with assurance in drillers' logs; the porphyritic nature of the flow generally is not recognizable in the fine cuttings of a churn drill.

Priest Rapids Member

The Priest Rapids Member overlies the Roza Member and in different places is overlain by the Saddle Mountains Member of the Yakima Basalt, the Beverly Member of the Ellensburg Formation, and sedimentary deposits of Pleistocene age; elsewhere, it is exposed at land surface. The type locality is upstream from the Priest Rapids Dam on the west side of the Columbia River, where the flows are exposed "in a south-dipping homocline on the west side of the reservoir" (Mackin, 1961, p. 23). There, the Priest Rapids Member consists of four lava flows and the Quincy Diatomite Bed. The diatomite bed may also contain layers of silt and clay.

The Priest Rapids basalt is grayish black, medium to coarse grained, and nonporphyritic, although some flows contain small phenocrysts easily noted in thin sections. Owing to a slightly diktytaxitic texture (Mackin, 1961, p. 23), the Priest Rapids basalt weathers easily. On a freshly broken sample the surface has a distinctive greenish cast, caused by saponite and chlorophaeite. Where weathered, the basalt is mottled brown, and the outer surface shows a distinctive red-brown color.

According to Waters (1961, p. 602, table 9, sample 9), the Priest Rapids basalt contains plagioclase, monoclinic pyroxene, olivine, glass, opaque minerals, and various decomposition products. The olivine-saponite content (12.5 percent) is the highest of the members of Yakima Basalt sampled by Waters. The percentage of glass (5.5 percent) is less than a third that of the Roza basalt, and less than a fifth that of the Frenchman Springs basalt.

The four flows of the Priest Rapids Member at the Priest Rapids damsite have a total thickness of 220±25 feet (Mackin, 1961, p. 23). From the central part of the project area, the Priest Rapids appears to thin rapidly northward and eastward. In the extreme northeastern part of the area, the columnar part of some of the Priest Rapids flows thins, with a proportionate thickening of the hackly part. It is uncertain whether the hackly basalt, which is black and finer grained than typical Priest Rapids basalt, is merely the entablature overlying the colonnade of a single Priest Rapids flow, or whether it represents another thin flow without colonnades.

Columns in the Priest Rapids basalt are as much as 10 feet in diameter. Platy parting normal to the column axis occurs to a lesser degree than in Roza basalt. In the Drumheller Channels, platy parting in Priest Rapids basalt occurs either in the upper or the lower part of the columns (fig. 14). The columns may break into large prismatic sections with spherical fracture surfaces of a roughly cup-and-ball character, also known as "ball-and-socket" joints. The shallow concavity of the upper surface of the blocks is more pronounced then in similar slabs from other members of the Yakima Basalt, and helps to identify the Priest Rapids basalt (fig. 14).

FIGURE 14.—Columnar jointing in basalt of Priest Rapids Member in SW¼NE¼ sec. 11, T. 16 N., R. 28 E. Platy parting is normal to axis of columns. The three columnar sections in foreground display typical shallow concavity.

As mapped in the northeastern part of the study area by Grolier and Bingham (1971), the Priest Rapids basalt constitutes the uppermost flows exposed along the coulee walls. East and north of the Crab Creek valley, it also is exposed in outliers resting on the Roza basalt. North of Pinto Ridge, four flows of the Priest Rapids Member are exposed and are reported in the logs of Bureau of Reclamation test holes (W. E. Walcott and G. E. Neff, unpublished U.S. Bureau of Reclamation report, 1950, p. 4).

The Priest Rapids Member is absent in most of the western part of the Frenchman Hills north of Sand Hollow and west of Low Gap Pass, although the basal part of a Priest Rapids flow, 1 or 2 feet thick, overlies the Quincy Diatomite Bed at places in the quarry in the SW$\frac{1}{4}$NE$\frac{1}{4}$ sec. 8, T. 17 N., R. 24 E. Priest Rapids basalt underlies Wahatis Peak and many of the hogbacks along the north and south slopes of the Saddle Mountains. East of the Saddle Mountains, the basalt crops out along the shorelines of Eagle Lakes and Scooteney Reservoir. In most of the scabland in and south of the Othello Channels the Priest Rapids flows are concealed by the overlying Saddle Mountains Member. Between Hatton Coulee and Bowers Coulee, the Priest Rapids basalt is nowhere exposed, and may not have extended there.

The lack of phenocrysts in the Priest Rapids Member is the most obvious identifying criterion in the field. The distinctive red-brown color of weathered surfaces also provides a clue.

In most of the project area, basalt of the Priest Rapids Member lies directly on the Quincy Diatomite Bed rather than on the Roza Member. The type locality of the Quincy Diatomite Bed is in the southwestern part of the Quincy Basin near Frenchman Springs Coulee. The writers agree with the statement by Mackin (1961, p. 26) that "On the assumption that the . . . lake [in which the Quincy Diatomite Bed was deposited] was impounded by one Priest Rapids flow and destroyed by another, the Quincy diatomite is considered to be a part of the Priest Rapids Basalt Member." The Quincy Diatomite Bed was shown on the geologic map of Grolier and Bingham (1971) wherever exposures warranted it, but none of the four basalt flows of the Priest Rapids Member were mapped separately.

The Quincy Diatomite Bed is as much as 25 feet thick in the southwestern part of the Quincy Basin, and as much as 35 feet thick on the south slope of the Frenchman Hills (fig. 15), where it is quarried in the E$\frac{1}{2}$ sec. 8, T. 17 N., R. 24 E. The thickness of the diatomite varies greatly, as does the occurrence of silt and clay layers, layers of banded opal, and nodules of opal within the diatomite. In many places, discontinuous layers or isolated nodules of opal are in or under the diatomite.

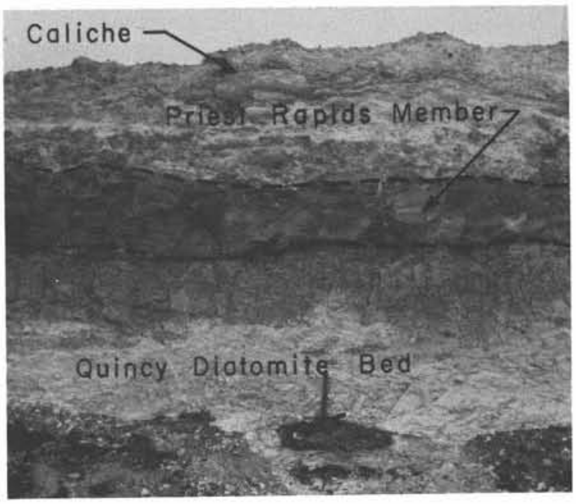


FIGURE 15.—Quincy Diatomite Bed in pit in SW$\frac{1}{4}$ NE$\frac{1}{4}$ sec. 8, T. 17 N., R. 24 E. Part of one flow of Priest Rapids Member overlies the diatomite. Pick rests on lens of opal within diatomite. Layer of massive caliche caps the basalt.

The effect of contact baking immediately beneath Priest Rapids basalt is shown by the development of columnar jointing in clay overlying diatomite. The columnar jointing in baked clay is best displayed in the northeastern part of the project area where a 1-foot layer of clay is the lateral equivalent of the diatomite.

The Quincy Diatomite Bed within the project area thins away from the southwestern part of the Quincy Basin and the western part of the Frenchman Hills. To the east it is still present in the Drumheller Channels where it ranges in thickness from 1 to 4 inches. It crops out along an east-facing cliff north of Goose Lake in the SW$\frac{1}{4}$NW$\frac{1}{4}$ sec. 23, T. 17 N., R. 28 E., and also on the north side of a Priest Rapids basalt outlier in the NW$\frac{1}{4}$NW$\frac{1}{4}$ sec. 36, T. 17 N., R. 23 E. An exposure in a drainage ditch along the Moses Lake-Stratford Road in NW$\frac{1}{4}$SW$\frac{1}{4}$ sec. 11, T. 20 N., R. 28 E. (Grolier and Foxworthy, 1961) shows the diatomite baked by the overlying Priest Rapids basalt and lying on top of the Roza basalt.

The Quincy Diatomite Bed also occurs at shallow depth in Adams County in the E$\frac{1}{2}$ sec. 25, T. 16 N., R. 30 E., where it was mined 50 or 60 years ago. In 1941, Mackin (written communication, 1946) examined this locality, now destroyed. He reported that the diatomite was at least 6 feet thick and was overlain by Priest Rapids basalt. The diatomite is reported in many well logs as "silica." Both the diatomite and its lateral equivalent of clay are useful markers in the subsurface correlation of members of the Yakima Basalt.

The lake or lakes in which the Quincy Diatomite Bed accumulated just prior to or during extrusion of the Priest Rapids basalt must have been extensive. Even where no diatomite occurs between Roza and Priest Rapids basalt, a zone of pillows and palagonite breccia as much as 10 feet thick characterizes the base of the Priest Rapids basalt. Many trees were engulfed by the advancing Priest Rapids flows. Petrified logs or stumps, some of them upright, occur at the base of the Priest Rapids basalt. Much of the petrified wood found on the Saddle Mountains occurs at this horizon.

Saddle Mountains Member

In the project area, the Saddle Mountains Member may overlie or be intercalated in the Beverly Member of the Ellensburg Formation, or it may overlie the Priest Rapids or Roza basalts. In turn, it is overlain by sedimentary deposits of late Pliocene and Pleistocene age in some parts of the area, or locally it is exposed at land surface. The Saddle Mountains basalt was named by Mackin (1961, p. 26), based upon one flow exposed at a type section near the east side of Sentinel Gap on the south limb of the Saddle Mountains, in the NW$\frac{1}{4}$ sec. 23, T. 15 N., R. 23 E. However, all the basalt flows above the Priest Rapids Member are now considered part of the Saddle Mountains Member (Bingham and Grolier, 1966, p. 12), regardless of whether they overlie or are intercalated with the Beverly Member of the Ellensburg Formation. Also mapped as part of the Saddle Mountains Member are individual flow tongues that were confined to shallow valleys in the underlying lavas and sedimentary rocks.

Field observations indicate that many flows, rather than a single flow, overlie the Beverly Member south and east from Sentinel Gap. Well drillers' logs show that the Beverly Member is discontinuous laterally, so that the Saddle Mountains basalt rests directly on the Priest Rapids Member in the southern part of the project area. The Saddle Mountains basalt is petrochemically similar to the Priest Rapids basalt (A. C. Waters, written communication, 1963), "and there is no doubt that it represents a continuation of the same great outpouring of basaltic lava as that represented by the rest of the Yakima Basalt."

At its type locality, the Saddle Mountains basalt is black, dense, very fine grained, and sparsely porphyritic. The basalt part of the flow consists of pillows with little or no palagonite. The pillow zone is crowned by cavernous basalt 2 to 3 feet thick.

Away from the type locality, much of the basalt is vesicular, regardless of whether it is black, aphanitic and hackly, or light to dark gray and columnar, or whether it consists of an agglomerate of reddish, semiangular pebbles and cobbles. The vesicles are commonly small—1 to 2 millimeters in diameter—and spherical. One flow of the Saddle Mountains basalt, which underlies some of the lower hogbacks on the south slope of the Saddle Mountains, is light gray, slightly diktytaxitic, and contains small phenocrysts of olivine. Where the porphyritic Saddle Mountains basalt was identified, it was distinguished from the nonporphyritic flows of the member on the geologic map of Grolier and Bingham (1971).

At the type area, the single Saddle Mountains flow is about 90 feet thick. Elsewhere, however, individual flows of Saddle Mountains basalt generally are thin and probably do not exceed 50 to 75 feet in thickness. The total number of flows probably increases toward the center of the Pasco Basin. The aggregate thickness of the Saddle Mountains basalt increases rapidly southward from the Othello Channels to the Snake River. The Saddle Mountains Member probably is thickest in the southern part of the Pasco Basin, where it may be 400 feet thick.

The Saddle Mountains basalt at its type area is characterized by narrow splintery columns and hackly jointing in the entablature of the flow. These features are characteristic of at least three of the flows identified elsewhere.

The Saddle Mountains Member extends southward from the southern limb of the Frenchman Hills anticline and from the southern half of the Drumheller Channels. The presence of only older basalts in isolated outcrops east of the Drumheller and Othello Channels indicates that the Saddle Mountains basalt does not extend eastward under the clay of the Ringold Formation that underlies Paradise Flats.

Basalt of the Saddle Mountains Member underlies the eastern two-thirds of the Royal Slope, and both sides of the Saddle Mountains. From the southern half of the Drumheller Channels south to the Snake River, most of the basalt exposed along coulee walls or in scabland channels is the Saddle Mountains basalt. Porphyritic Saddle Mountains basalt may be the rock exposed along the Snake River east of Martindale, in secs. 20, 21, 22, 15, T. 9 N., R. 31 E., and also 4 miles to the north, in secs. 31 and 32, T. 10 N., R. 31 E. At those places, the basalt contains large, shattered, partly resorbed plagioclase phenocrysts similar in appearance to those in the Frenchman Springs basalt, and small transparent lath-shaped plagioclase phenocrysts averaging 2 millimeters in length. Porphyritic basalt of similar lithology is exposed in sec. 25, T. 13 N., R. 29 E., and in the southern part of a linear ridge trending N. 15° W. across scabland from the NE$\frac{1}{4}$ sec. 14 into the NW$\frac{1}{4}$ sec. 24, T. 13 N., R. 29 E.

Porphyritic Saddle Mountains basalt of another type occurs in a borrow pit in the NW$\frac{1}{4}$ sec. 12, T. 11 N., R. 30 E., a quarter of a mile east of Eltopia. At this locality, small subangular fragments of unaltered gray dense basalt are engulfed in a matrix of light-gray, slightly porphyritic basalt that is partly granulated and altered to clay. The phenocrysts consist of lath-shaped plagioclase and are about 1 centimeter long.

Saddle Mountains basalt is intertongued with sediments of the Beverly Member at several places in the west-central part of the project area. A flow resting on pumicite is exposed in NW$\frac{1}{4}$NE$\frac{1}{4}$ sec. 23, T. 15 N., R. 23 E., on the east side of the north-trending ridge on which the type section of the Saddle Mountains and Beverly Members are located. The northern front of the flow abuts pumicite. The flow is overlapped by additional layers of pumicite. This flow is not exposed in the pumicite quarry 0.25 mile to the southwest, and probably does not extend there. Also, the Saddle Mountains basalt interstratified with the Beverly Member is exposed in two northeastward-trending synclines near the top of the Saddle Mountains escarpment, east of Beverly (figs. 16 and 17).

In addition to the flood basalt flows, just described, the Saddle Mountains Member also occurs as narrow linear flow tongues. In the project area, three flow tongues, the Warden-Othello, Cactus, and Jericho flow tongues, have been recognized as consisting of Saddle Mountains basalt. They all rest on older members of the Yakima Basalt rather than being

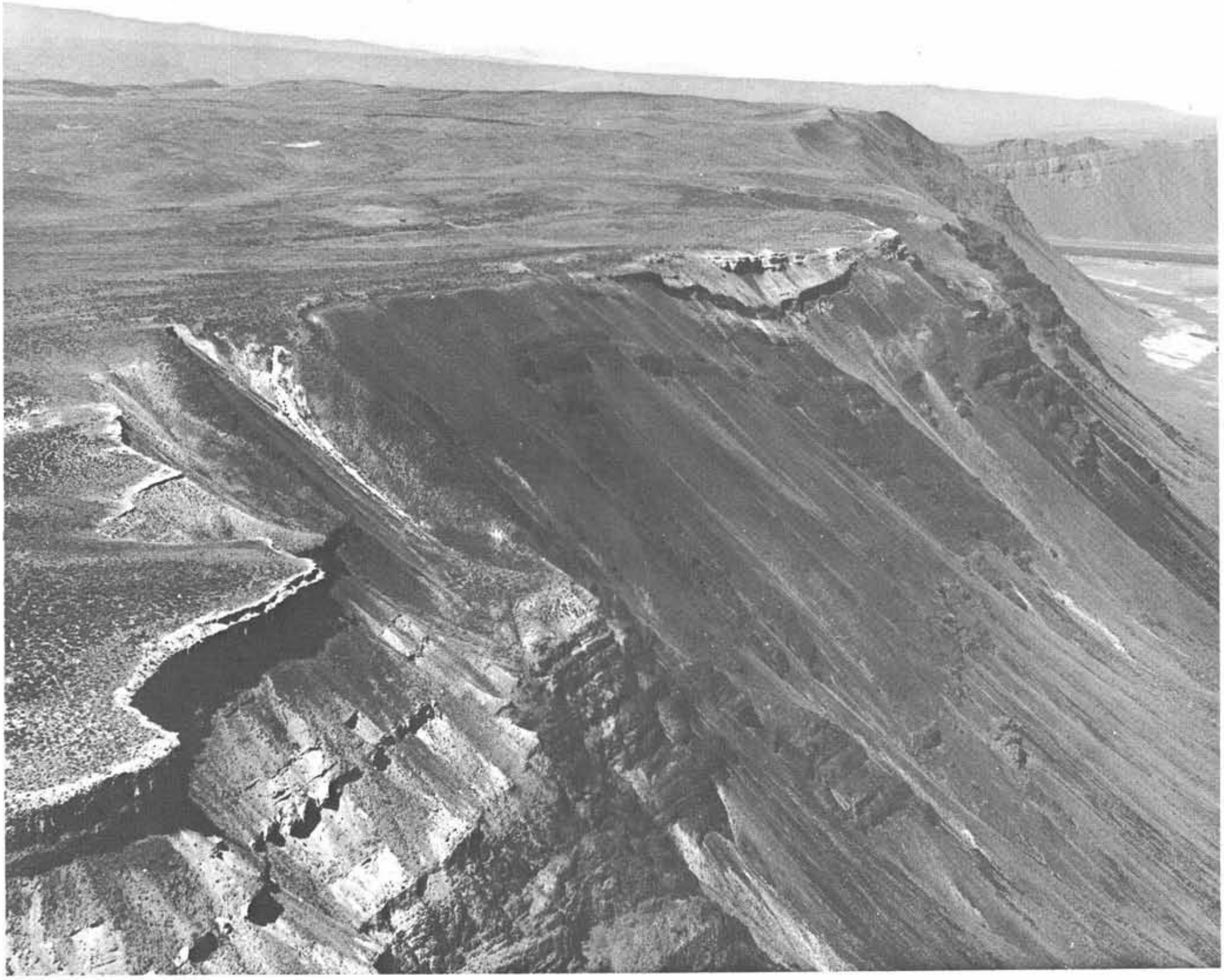


FIGURE 16.—Aerial view of north face of Saddle Mountains, looking southwest, showing two synclines cut by the fault scarp. Sharp syncline of figure 17 is in left foreground. (Grant County PUD photo, 1960).

interstratified with the Beverly Member. The tongues occur near the northern or eastern limits of the Saddle Mountains basalt, where the lava that flowed up the valleys eroded during the time interval between the extrusion of Priest Rapids and Saddle Mountains basalts.

The basalt in the flow tongues is black, fine grained to aphanitic, predominantly hackly, and similar to the basalt that forms the entablature of the Saddle Mountains flow at the type section.

The Warden-Othello flow tongue is a prominent sinuous ridge 20 to 40 feet above the surrounding scabland that extends for 12 miles from the NW$\frac{1}{4}$ sec. 29, T. 16 N., R. 29 E., near Owl Lake 2 miles northwest of Othello (fig. 18), to the SW$\frac{1}{4}$ sec. 21, T. 17 N., R. 30 E., 1-3/4 miles southwest of Warden. Between the Drumheller Channels and the Warden Channel, the Warden-Othello flow tongue is concealed by clay of the Ringold Formation, which under-

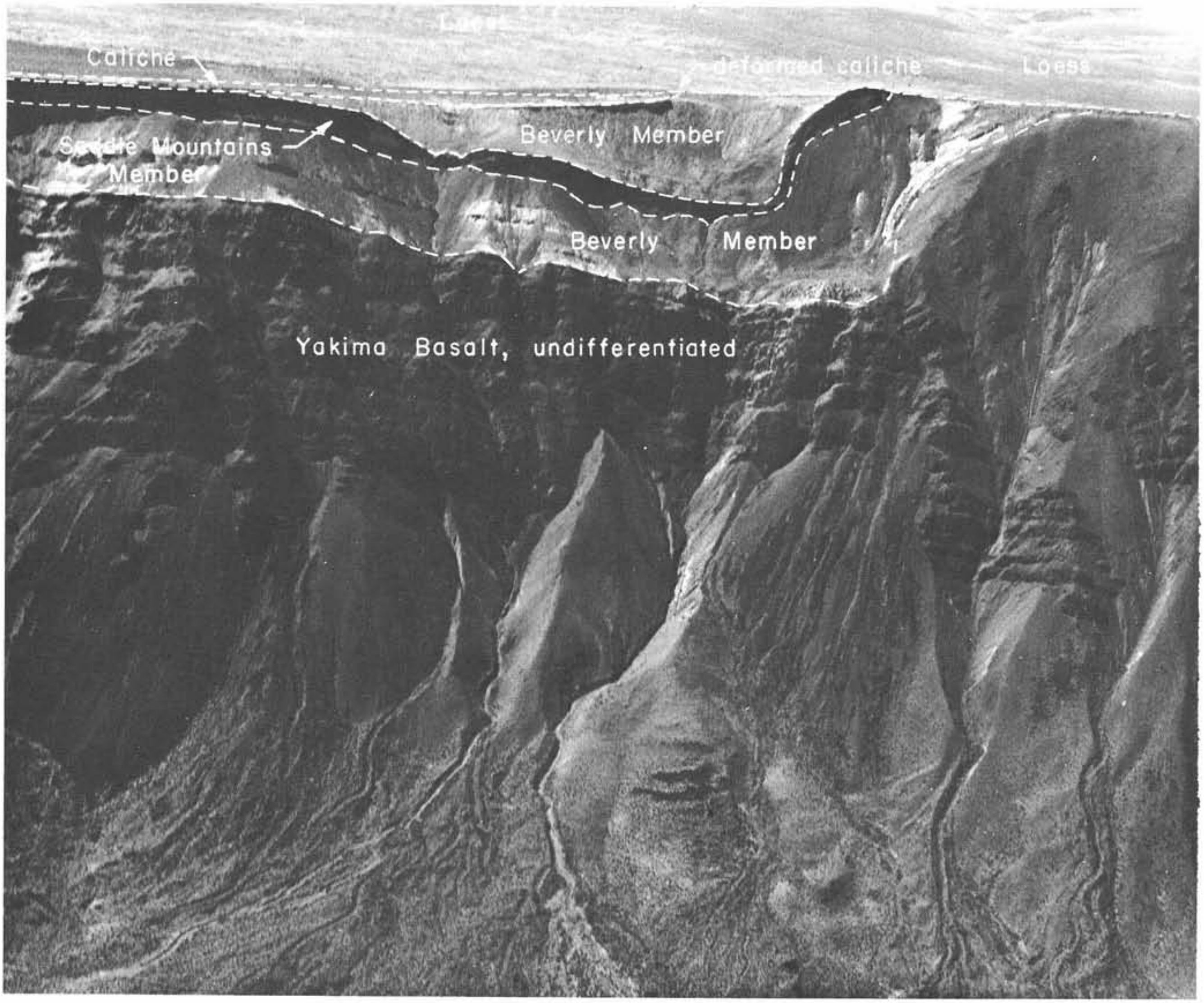


FIGURE 17.—Aerial view of north face of Saddle Mountains, showing sharp syncline in Yakima Basalt and interbedded Beverly Member. Extreme right end of caliche is deformed upward, similar to the right limb of the syncline in the underlying sediments and basalt. (Grant County PUD photo, 1959.)

lies Jackass Mountain, and by glaciofluvial gravel of late Pleistocene age farther east. South of Warden, where this tongue disappears eastward under ashy silt, it is approximately 500 feet wide. Because it is obscured by talus or swamps, the basalt contact of the tongue is nowhere exposed. The tongue is underlain by Priest Rapids basalt northwest of Othello and Roza basalt in the Warden Channel. Irregular marginal grooves, now partially filled with water, have been etched out by late Pleistocene fluvial erosion.

This sinuous ridge of basalt also has been described as possibly a dike (Neff, 1962, p. 20-21).

The Cactus flow tongue is named after Cactus, a Burlington Northern railway siding in Esquatzel Coulee. The tongue is exposed on the north and south walls of Esquatzel Coulee in sec. 16, T. 13 N., R. 31 E. Basalt of the Cactus flow tongue is black and

FIGURE 18.—Aerial view looking north, showing Warden-Othello flow tongue trending northeast from north end of Owl Lake, 2 miles northwest of Othello.

dense. The tongue forms a protrusion in the north wall of Esquatzel Coulee, where it has been extensively quarried. The columnar jointing displayed in the tongue was described and illustrated by Campbell and others (1916, p. 166, pl. 22-B), before quarrying began. Fluvial erosion has accentuated the topograhic expression of the tongue across the surrounding scabland, where it trends N. 15° W. for 4 miles from Old Maid Coulee in the SW$\frac{1}{4}$ sec. 22, T. 13 N., R. 31 E., to a point 2 miles north of Esquatzel Coulee, in the NE$\frac{1}{4}$ sec. 5, T. 13 N., R. 31 E. At the latter place, the tongue disappears under slope wash and the Ringold Formation. The Cactus flow tongue fills a channel, 500 feet wide and 150 feet deep, cut through one Priest Rapids flow and into Roza basalt. The lateral contact of the Cactus flow tongue against the Roza basalt along the walls of the ancient valley cannot be observed in cross section from Esquatzel Coulee, because the contact is obscured by slope wash. However, the basal contact between the tongue and the Roza basalt at the bottom of the ancient channel is exposed in sec. 16, along the south wall of Esquatzel Coulee a few feet above the coulee floor.

The Jericho flow tongue is exposed only in cross section along the north wall of Jericho Coulee, in the N$\frac{1}{2}$ sec. 25, T. 16. N., R. 23 E., 2 miles northeast of Beverly. It is three-quarters of a mile wide, and 100 feet thick. Its base is concealed by drifting sand and talus. The tongue consists of black fine-grained basalt with hackly jointings. At the eastern end of the tongue it is in contact with columnar basalt, probably of another Saddle Mountains flow. The contact at the western end is covered by fluvial gravel.

Beverly Member of the Ellensburg Formation

In the project area, the Ellensburg Formation is represented only by the Beverly Member, which is named for the town of Beverly.

The type locality of the Beverly Member is on the east side of Sentinel Gap on the south slope of the Saddle Mountains, in secs. 23 and 24, T. 15 N., R. 23 E. There, the member consists of beds of quartzite-bearing conglomerate, pumicite, and tuffaceous sand, silt, and clay. The conglomerate and pumicite make up more than 50 percent of the member at its type locality. Elsewhere, the sedimentary rocks in the Beverly Member may include basaltic fanglomerate and grit, as well as conglomerate, pumicite, and tuffaceous sand, silt, and clay. The individual facies referred to above were not mappable at the scale of the geologic map of Grolier and Bingham (1971).

The age of the Beverly Member is uncertain, because only a few fossil fresh-water mollusks and fossil leaves were found in 1958 during excavation work at the site of Priest Rapids Dam, about 9 miles south of the type locality. The mollusks occurred in a light-gray poorly consolidated siltstone. According to D. W. Taylor of the U.S. Geological Survey (written communication, 1962), the mollusks consisted of the fresh-water clam, Anodonta and the fresh-water snail Bellamya n. sp. Taylor states that "the collection at locality 22683 includes the same extinct species of Bellamya [found in faunal assemblages at four other localities in eastern Washington]. The occurrence of Coretus [in one of the other four collections] tends to suggest an age no younger than middle Pliocene. I do not believe a precise age assignment is warranted, nor can one even date the collections as Miocene vs. Pliocene. Probably they are all from the same time interval—early Pliocene, for example, or late Miocene, or very late Miocene to earliest Pliocene."

Conglomerate

Conglomerate of the Beverly Member is restricted to the vicinity of the type locality. It thins to extinction in less than 3 miles east of the Sentinel Bluffs, on the south slope of the Saddle Mountains. The conglomerate consists of beds of well-rounded, commonly elongate pebbles and cobbles. Quartzite and felsite porphyry are the most distinctive components; these, along with basalt, agate, and granitic and metamorphic rocks, are commonly embedded in a tuffaceous sand matrix. Two of the lower conglomerate beds at Sentinel Bluffs are composed predominantly of basalt pebbles.

A thin bed of very coarse, stratified sand, consisting of milky white and yellow quartz grains, between the Priest Rapids Member and the Saddle Mountains Member in the central and eastern part of the Saddle Mountains probably is the lateral equivalent of the conglomerate. This sand is exposed at the mouth of a northward-trending gulley in the SW$\frac{1}{4}$ sec. 22, T. 15 N., R. 24 E.; in a notch on the crest of the ridge at the western end of the Corfu landslide, in the SE$\frac{1}{4}$NW$\frac{1}{4}$ sec. 16, T. 15 N., R. 27 E.; on a small peak in the NE$\frac{1}{4}$SW$\frac{1}{4}$ sec. 12, T. 15 N., R. 26 E.; and on the crest east of the W$\frac{1}{4}$ corner sec. 14, T. 15 N., R. 27 E.

Two small exposures of the conglomerate are also found in the western part of the Othello Basin. Open-work quartzite gravel resting on the Priest Rapids basalt underlies two small knobs on a scabland bench in Jericho Coulee, south of the W$\frac{1}{4}$ corner sec. 28, T. 16 N., R. 24 E. The gravel there consists of quartzite, basalt, and agate pebbles cemented with a crystalline carbonate that partially fills the voids between the pebbles. Foreset beds in the gravel dip eastward. Another exposure of felsite porphyry pebbles and a very few quartzite and agate pebbles of smaller size, embedded in a fine tuffaceous sandstone, occurs on the west side of the road at the base of a cliff of basalt at the junction of Red Rock Coulee and Natural Corral, in the SW$\frac{1}{4}$NW$\frac{1}{4}$ sec. 17, T. 16 N., R. 26 E. A thickness of about 4$\frac{1}{2}$ feet is exposed, and the base of the conglomerate is hidden by talus. The overlying basalt appears similar to Priest Rapids basalt, and the conglomerate thus may be stratigraphically lower than that of Sentinel Gap.

A loosely consolidated gravel consisting of highly polished, red-stained, typically elongate pebbles and cobbles of quartzite and lamprophyre porphyry in a friable quartzose and micaceous sand matrix is exposed in the NE$\frac{1}{4}$SE$\frac{1}{4}$ sec. 11, T. 14 N., R. 29 E. The outcrop lies in the Othello Channels, in a trench that contains a stream between two of the Eagle Lakes. This conglomerate is about 10 feet thick, and contains few or no basalt pebbles. Because of its position between the Priest Rapids and Saddle Mountains Members, it is correlated with the Beverly Member.

Logs of a few wells at scattered points in Franklin County suggest that a conglomerate equivalent to that of the Beverly Member may occur beneath basalt, where it was deposited by ancient shifting streams. The extent or direction of the channels recorded by this conglomerate cannot be determined with present data.

Pumicite

Pumicite is the most distinctive and perhaps the most laterally persistent lithologic unit of the Beverly Member. It consists almost entirely of volcanic glass shards, arranged in layers, many of which contain ripple marks. Its occurrence at Sentinel Bluffs in the upper part of the Beverly Member has been described by Twiss (1933), Carithers (1946), Mason (1953), and Mackin (1961).

Throughout its known area extent, the pumicite in the Beverly Member contains layers of spherical accretionary lapilli, termed chalazoidites. Such lapiili from pumicite sampled at the Sentinel Bluffs have been described by Moore and Peck (1962, p. 184-185, 187, fig. 1, pl. 2, and table 1).

Perfect sorting of the pumicite and the gradual decrease in the size of the glass shards eastward from Sentinel Bluffs support many observers' conclusions that the pumicite was airborne and originated from the west. The lateral gradation in grain size is not easily detectable over a short distance in the field, but is conspicuous when pumicite sampled at Sentinel Bluffs is compared with that obtained from hogbacks south of the Smyrna Bench.

The effects of contact metamorphism are more pronounced at the contact of the pumicite and overlying Saddle Mountains flows than at any other stratigraphic level in the project area where sedimentary rocks underlie basalt. In many places, the pumicite underlying the Saddle Mountains Member is baked to the color and hardness of a red brick for several feet below the contact, and columnar jointing in it is well developed.

Pumicite is exposed at many localities on the slopes of the Saddle Mountains. In the small north-trending synclines in secs. 3 and 4, T. 15 N., R. 24 E., pumicite in the Beverly Member is directly overlain by basalt (figs. 16 and 17). In the larger northeast-trending syncline in sec. 2, T. 15 N., R. 24 E., pumicite is underlain by a grit and fanglomerate of angular to subangular basalt pebbles and grains that are of Roza and Priest Rapids lithology.

Pumicite and tuffaceous silt and clay interstratified with basalt crop out discontinuously along some of the lower hogbacks on the north and south flanks of the Saddle Mountains as far east as R. 27 E. They also occur within large slumped blocks in the Corfu landslide.

Pumicite, probably at the stratigraphic level of the Beverly Member, also occurs in the Snake River canyon. Prior to the filling of Sacajawea Lake, pumicite was exposed in railroad cuts along the old right-of-way of the Spokane, Portland, and Seattle Railway.

In the $SE\frac{1}{4}$ sec. 5, the $NW\frac{1}{4}$ sec. 8, and in secs. 7 and 18, T. 9 N., R. 32 E., layers of pumicite 4 to 5 feet thick are interbedded with a coarse tuffaceous sand and clay 20 to 30 feet thick.

Tuffaceous Sand, Silt, and Clay

Layers of tuffaceous sand and white to buff silt and clay are interbedded with the conglomerate and pumicite at Sentinel Bluffs. Brown tuffaceous sand, silt, and clay occur principally in the deepest part of the two northeast-trending synclines at the top of the Saddle Mountains escarpment, in secs. 2 and 4, T. 15 N., R. 24 E. Immediately east of a sharp syncline that terminates the western limb of the eastern and larger syncline, extremely slickensided black clay, lying on the Saddle Mountains Member, is itself overlain by a brown tuffaceous sand containing scattered yellowish-orange quartzite pebbles. In the largest northeast-trending syncline, the sedimentary rocks overlying the Saddle Mountains Member are about 200 feet thick and are slightly discordant with the syncline. The presence of Roza and Priest Rapids basalt pebbles in the grit underlying the tuffaceous sand indicates that the sand unconformably overlies older units around the margins of the synclinal basin.

The Beverly Member is considered to be intertongued with the Saddle Mountains Member in the two northeast-trending synclines on the north scarp of the Saddle Mountains, so that a part of the Beverly Member appears to overlie basalt. The validity of this interpretation depends upon the eventual correlation of the Saddle Mountains basalt flow on the north scarp with the flow on the southern limb of the Saddle Mountains anticline in sec. 24, T. 15 N., R. 23 E. If, conversely, the flow in the north scarp were eventually correlated with the flow that overlies the Beverly Member at the type section, the question would arise as to whether the uppermost sedimentary rocks in the north scarp belong to the Ringold Formation rather than to the Beverly Member. As a general rule, both the Beverly Member and the Ringold Formation are disconformable over basalt near the troughs of the synclinal basins and their distinction could not always be made satisfactorily in this vicinity.

Beds of brown tuffaceous sand or buff silt and clay occur in the approximate stratigraphic position of the Beverly Member at other localities in the project area. For example, tuffaceous silt and clay underlie a flow tongue of Saddle Mountains basalt along the east bank of the Columbia River about 2,000 feet downstream from Priest Rapids Dam, in the $SE\frac{1}{4}$ sec. 2, T. 13 N., R. 23 E. Discontinuous outcrops of tuffaceous silty clay of the Beverly Member also occur along the loess-mantled western scarp of the Drumheller Channels. One of the outcrops is in a roadcut in the $SE\frac{1}{4}$ sec. 4, T. 16 N., R. 28 E., where the uppermost foot of a bed of tuffaceous clay, baked at the contact with the overlying basalt, is exposed at road level. The base of the clay is not exposed. The basal part of the overlying basalt there consists of a spectacular 30-foot exposure of pillows and palagonite. Another exposure is on a small pinnacle near the west margin of the Drumheller Channels, in the $SE\frac{1}{4}SW\frac{1}{4}$ sec. 9, T. 16 N., R. 28 E. The sedimentary rocks at this locality are about 30 feet thick and are capped by basalt. They consist of three layers of sand, which exhibit an upward gradation from festooned crossbedding to horizontal lamination, and a gradual upward decrease in grain size. The mineral constituents include quartz, feldspar, glass shards, basalt, and heavy mineral grains mixed with well-rounded granules of pumice.

The two outcrops in the Drumheller Channels described above are the surface exposures of a sedimentary bed that is interstratified with basalt under the eastern half of the Royal Slope, where it is reported at depth in many drillers' logs as far west as Royal City and as far north as the line of subdued hogbacks in the $SE\frac{1}{4}$ T. 17 N., R. 27 E., and the $SW\frac{1}{4}$

T. 17 N., R. 28 E. These silt-mantled hogbacks are underlain by Saddle Mountains basalt that in turn overlies the Beverly Member in the Drumheller Channels exposures described above. The uppermost sedimentary bed within the basalt, as reported in logs of wells located on the Taunton Bench and under the terrace at Othello, probably is a part of the Beverly Member.

Offlap Relationship of the Flows

An offlap relationship exists away from the plateau margin, each succeeding flow not generally extending beyond the limits of previous flows. Several factors that provide evidence for the offlap are as follows:

1. The basalt flows that are stratigraphically the highest occur south of the Frenchman Hills. Their outcrop pattern is due mainly to the centrally directed dip toward the Pasco Basin.

2. The number of flows that make up the Saddle Mountains Member diminishes northward; two are thought to extend onto the Royal Slope and one to the crest of the Frenchman Hills.

3. A decrease in the number of flows also occurs in the Priest Rapids Member in the northwestern part of the project area. However, the presence of four Priest Rapids flows in the Bacon syncline, at the extreme northern part of the project area (W. E. Walcott and G. E. Neff, 1950, p. 4, unpublished U.S. Bureau of Reclamation report), indicates that the margin of the subsiding basin was still north of this area during Priest Rapids time.

The primary cause of the offlap (fig. 19) between flows is presumably the contemporaneous subsidence of the lava field and extrusion of the flows. However, it also is possible that the exhaustion of the lava supplying a particular flow limited its extent, or the uneven subsidence of the lava field caused thinning or nondeposition of flows locally. This may explain conditions along the east side of the project area, where one Priest Rapids flow is present near Connell, none are exposed near Lind Coulee, and four are present in the Bacon syncline. Another explanation is that the flows may merely have decreased in extent upward and the subsidence of the Pasco Basin came later.

Variation in Thickness of the Flows

The impression gained from field observation is that the thicknesses of some flows decrease away from the central part of the project area. For example, the thickness of the Priest Rapids Member at Wahatis Peak on the Saddle Mountains is estimated to be about 300 feet, whereas the measured thickness at Priest

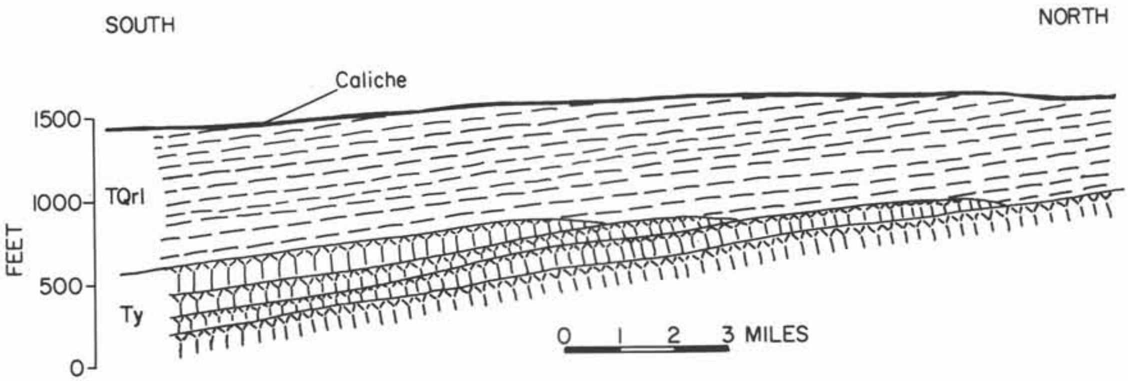


FIGURE 19.—Diagrammatic section showing the apparent regional unconformity due to offlap of individual basalt flows. Ty, Yakima Basalt, undifferentiated; TQrl, Ringold lacustrine clay and silt.

Rapids Dam is 220 feet (Mackin, 1961, p. 23). A decrease in overall thickness of the Priest Rapids, Roza, and Frenchman Springs Members is noticeable between Drumheller Channels and the extreme northern part of the project area. This change may be due either to the absence of one or more flows that never extended that far, or to a decreasing thickness in the individual flows themselves, or to both.

Directions of Lava Movement

One of the criteria used to determine the direction of flow in basalt is primary foreset bedding developed in pillow-palagonite complexes in the basal part of the flow (Waters, 1960, p. 361, pl. 2). In the project area, pillow-palagonite complexes are best developed in the basal part of at least two flows within the lower basalt, two flows in the Frenchman Springs Member, and several flows in the Priest Rapids Member.

The pillows and palagonite stringers in the basal part of the Sentinel Gap flow (Frenchman Springs Member), in the NW$\frac{1}{4}$ sec. 28, T. 17 N., R. 23 E., are arranged in foresets that dip northward when viewed normal to the strike of the flow on the east wall of the Columbia River canyon (Mackin, 1961, p. 16, pl. 3). The same northerly dip in foreset layers of pillows occurs in the pillow-palagonite complexes of the other basalt flows mentioned above, even in the extreme northeastern part of the area. This observation indicates that many, if not all, of the flows originated to the south. Obviously, then, the regional component of dip of the flows toward the Pasco Basin is not primary, but is due to post-extrusion subsidence centered in the Pasco Basin.

PLIOCENE AND PLEISTOCENE RINGOLD FORMATION

The term Ringold Formation was first used by Merriam and Buwalda (1917, p. 260) to name strata exposed in the White Bluffs, which extend for 30 miles along the east bank of the Columbia River north of Pasco. The formation, of Pliocene and Pleistocene age, consists of four principal lithologic units or lithofacies. These are brown tuffaceous sand and silt; a quartzite-bearing conglomerate; buff laminated clay, silt, and fine sand; and a basaltic fanglomerate. All four lithofacies are interbedded or intertongued.

Only three of the principal lithofacies occur in the White Bluffs, and they have been described there by Newcomb (1958, p. 328-340). The basaltic fanglomerate occurs only around the margins of synclinal basins, where uplift and erosion of the basalt substratum were greatest. The north-facing cliffs of the Smyrna Bench, where typical exposures of the basaltic fanglomerate occur, are here proposed as the type locality of the fanglomerate.

The fossils collected by Merriam and Buwalda (1917, p. 257 and 261) indicated a Pleistocene age for sediments exposed in the White Bluffs, based on a mammalian fossil assemblage. The fossils collected by Strand and Hough (1952, p. 154) were determined as middle or late Pleistocene in age.

Fossil mollusks and ostracods have been collected at three localities and identified as Pliocene to late Pleistocene by D. W. Taylor and I. G. Sohn of the U.S. Geological Survey (written communication, 1962 and 1959, respectively).

The nature of the contact between the Ringold Formation and the underlying basalt is of considerable interest in determining the relative ages of the Ellensburg and Ringold Formations. As suggested by Brown and McConiga (1960), "The indicated composition, sequence, continuity, and thickness of the beds do not indicate an unconformity of consequence or other significant change to distinguish the post-basalt sediments from the interbasalt sediments (the Ellensburg(?) Formation)." Also, the writers have found that the lower parts of both formations are nearly concordant over basalt in the center and flanks of the structural basalt basins in eastern Washington.

At the type locality along the White Bluffs, the basal part of the Ringold Formation is nowhere exposed, so that the nature of the contact can only be deduced from indirect evidence. Drillers' logs of two wells (12/28-24N1 and 13/27-14R2) on sloping terraces near the base of the White Bluffs—between the bluffs and the Columbia River—indicate that clay, silt, sand, and gravel extend downward 200 to 300 feet to basalt. The grain size and the lithology of these subsurface sedimentary rocks are shown by drill cuttings to be similar to those of the Ringold Formation exposed in the bluffs. This similarity, as well as an inferred stratigraphic continuity, has led Newcomb (1958, p. 330, 335) to extend the Ringold Formation at the type locality downward to basalt—a view shared by Brown and McConiga (1960, p. 51).

The caliche capping the Ringold atop the cliffs stands at an altitude of about 900 feet, whereas basalt in well 13/27-14R2 was penetrated at an altitude of 95 feet; thus, the total thickness of the Ringold Formation resting on basalt near this well is about 800 feet.

Saddle Mountains basalt is exposed in the floor of Ringold Coulee, about 5 miles northeast of its mouth, in the SW$\frac{1}{4}$SE$\frac{1}{4}$ sec. 21, and also in the S$\frac{1}{2}$ sec. 28, T. 13 N., R. 29 E., at altitudes of 650 and 580 feet, respectively. There, the contact between the basalt and the overlying Ringold Formation is concealed by talus and fluvial gravel. Nevertheless the dip of the Yakima Basalt surface in Ringold Coulee is southwesterly (Walters and Grolier, 1960, pl. 3), and a slight disconformity exists between the Yakima Basalt and the Ringold Formation (fig. 20). An 8-foot bed of red silt that occurs high in the Ringold Formation near the top of the White Bluffs has been traced 5$\frac{1}{2}$ miles up Ringold Coulee by Brown and McConiga (1960, p. 44-45, 47, figs. 1-3). The red marker bed dips southwestward at 10 feet per mile, whereas the underlying basalt slopes in the same direction at 74 feet per mile (Appendix IV), the difference of less than 1° is imperceptible. Likewise along the flanks of anticlinal folds where the Ringold Formation dips as much as 20°, the slight discordance is not noticeable.

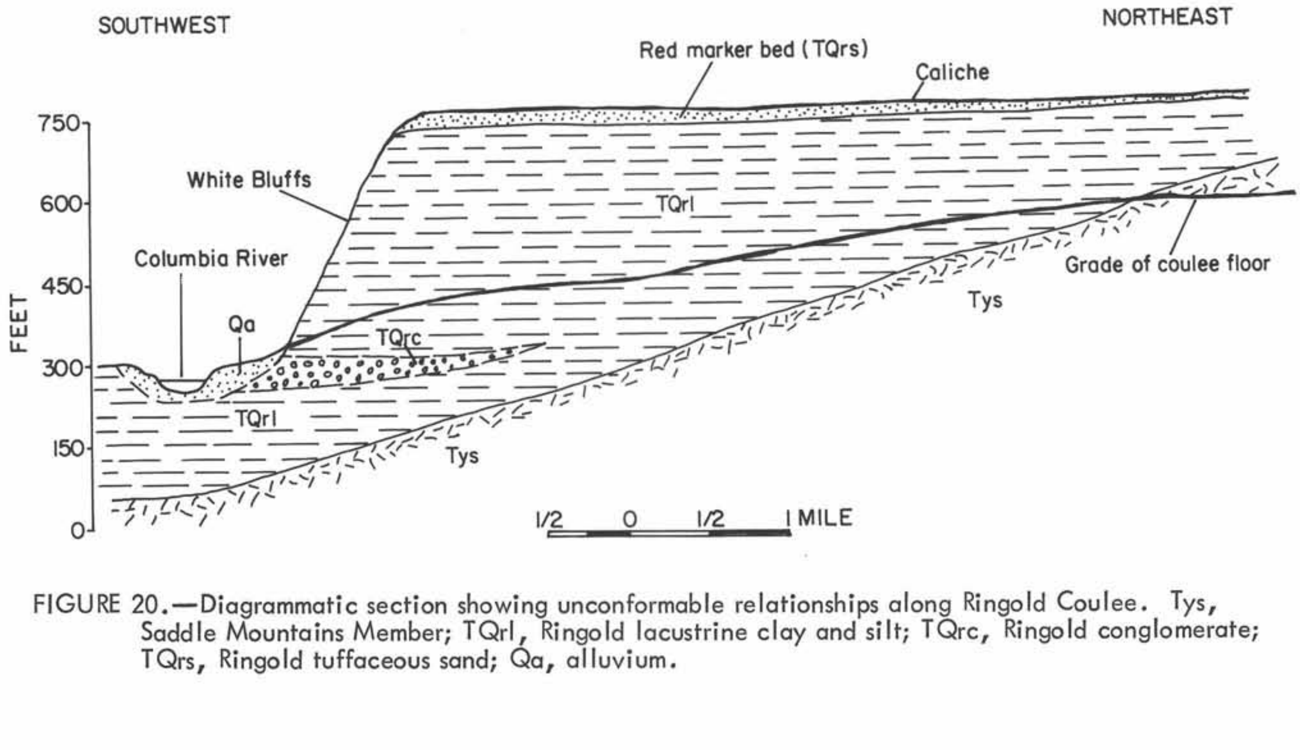


FIGURE 20.—Diagrammatic section showing unconformable relationships along Ringold Coulee. Tys, Saddle Mountains Member; TQrl, Ringold lacustrine clay and silt; TQrc, Ringold conglomerate; TQrs, Ringold tuffaceous sand; Qa, alluvium.

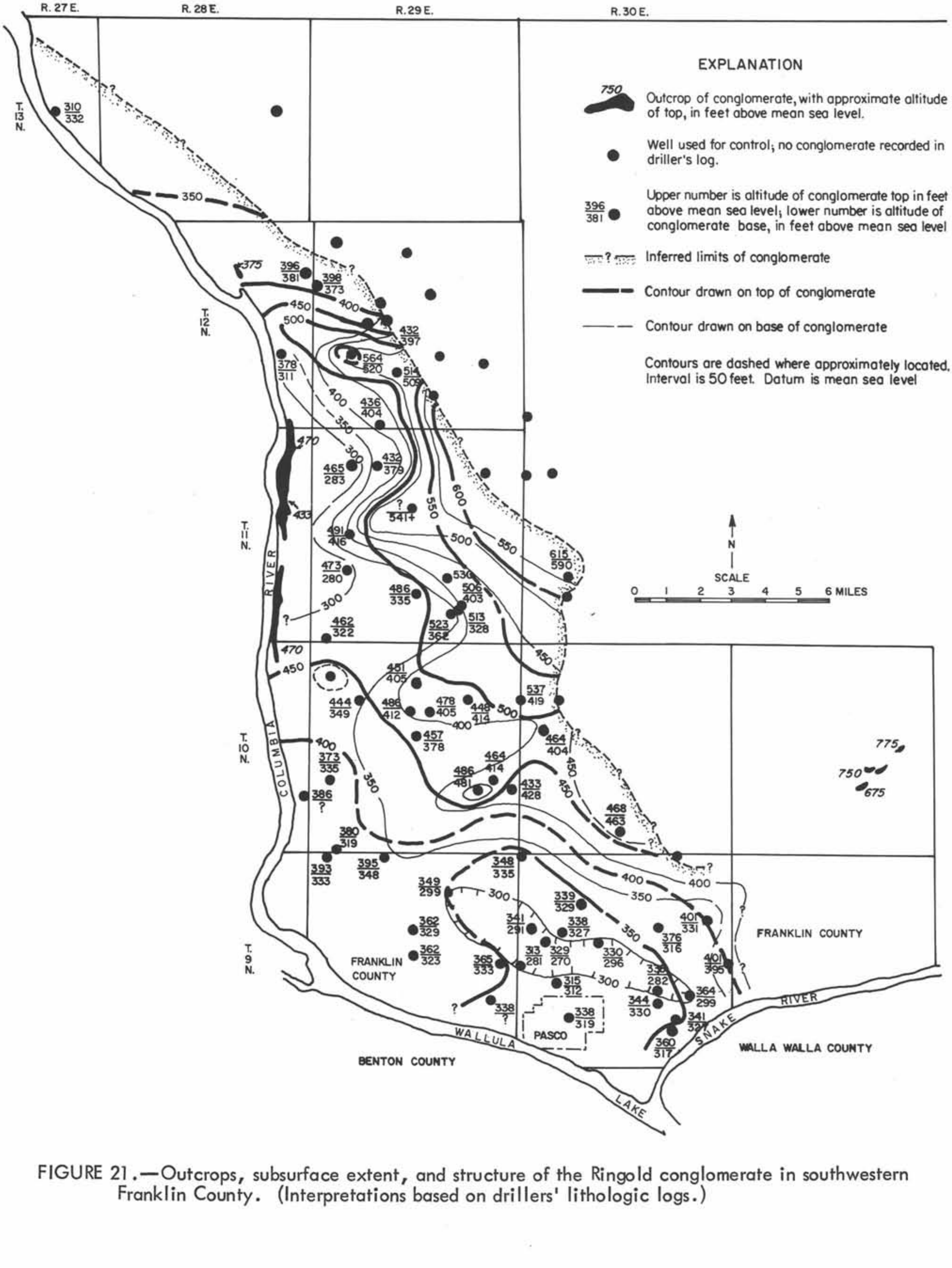


FIGURE 21.—Outcrops, subsurface extent, and structure of the Ringold conglomerate in southwestern Franklin County. (Interpretations based on drillers' lithologic logs.)

The lack of a widespread erosional interval either between the basalt and the Ringold beds or between the individual Ringold beds themselves at or near the type locality suggests that the disconformity mentioned above actually is a structural discordance, rather than erosional. Support for structural control is offered by Brown and McConiga (1960, p. 53) who state that "The gradual general decrease in dips upward in the Ringold section favors essentially continuous deformation . . . during the period of deposition."

The Ringold conglomerate facies probably was originally a part of an alluvial plain that had a gradient southward along the White Bluffs. With this assumption and those offered in the preceding paragraph, the contours on figure 21 provide additional evidence of structural deformation of the Ringold Formation. The contours on the top and bottom of the conglomerate facies show that it was deformed primarily during deposition but also, to a lesser degree, after deposition. The contours on the bottom of the conglomerate, as represented by the (fine-line) 400-foot contour on figure 21, indicate that deformation in the form of four westward-plunging noses occurred largely before the final deposition of the conglomerate. Deformation continued after the last deposition of the unit, by the uplift of a broad area between the mouth of Ringold Coulee and the south end of the White Bluffs, as well as by amplification and extension of the small tight structure just east of Ringold. The contours

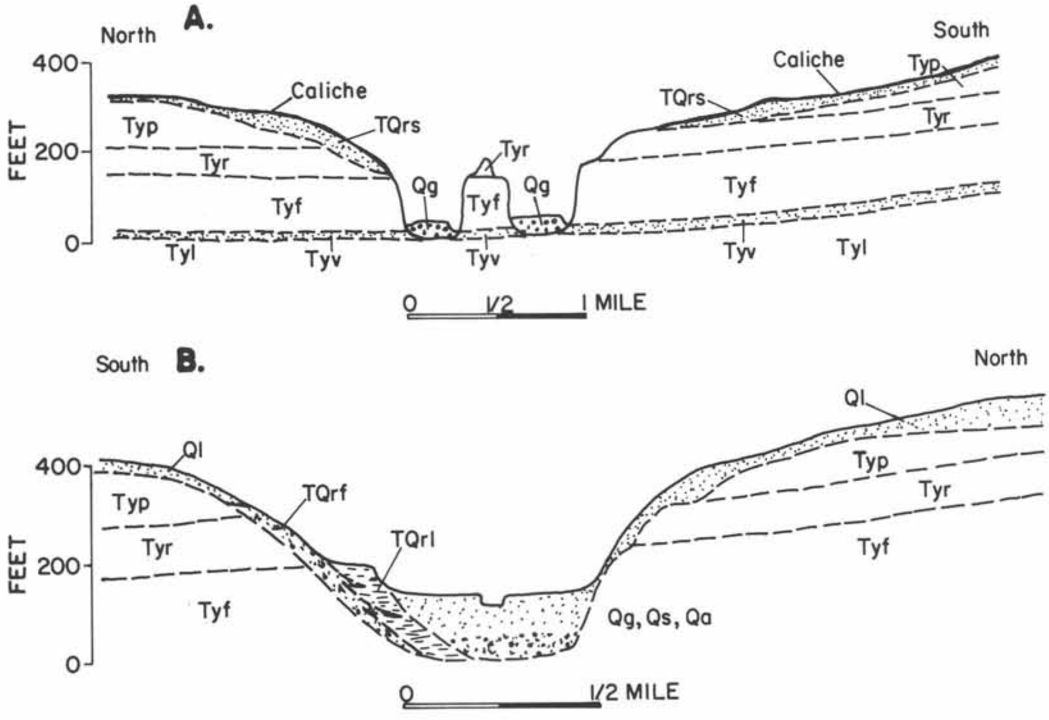


FIGURE 22.—Diagrammatic sections across (A) Potholes Coulee and (B) Lind Coulee near Providence, showing unconformable relationships. Tyl, lower Yakima Basalt (flows below the Vantage Sandstone Member); Tyv, Vantage Sandstone Member; Tyf, Frenchman Springs Member; Tyr, Roza Member; Typ, Priest Rapids Member; TQrf, Ringold fanglomerate; TQrl, Ringold lacustrine clay and silt; TQrs, Ringold tuffaceous sand; Ql, loess; Qg, Qs, Qa, fluvial gravel and sand (flood deposits) and alluvium.

show that, at present, the overall structure has an amplitude of more than 120 feet and a plunge of more than 150 feet westward.

The nature of the contact between the Ringold and underlying basalt is far different from this structural discordance at many locations away from the central parts of the Pasco Basin. One situation is an apparent angular unconformity between the Ringold and the underlying basalt (fig. 19). The contact of the Yakima Basalt and the Ringold Formation (regionally, not at the scale of a small basin) is a gap in rock sequence due to missing flows.

The second type of Ringold-Yakima Basalt contact away from the central Pasco Basin is indicated by an erosional unconformity. Deep valleys (represented by coulees like Sand Hollow, Lind, and Washtucna) that were cut through several members of the Yakima Basalt were ponded and became the sites of lacustrine deposition, with the result that the buff laminated clay or the brown tuffaceous sand facies of the Ringold Formation, in places, unconformably overlies successive members of the Yakima Basalt (fig. 22). An erosional unconformity also exists at the contact between the brown tuffaceous Ringold sand and the Yakima Basalt.

Many stratigraphic sections of the Ringold Formation have been measured or estimated along the White Bluffs by previous workers (Calkins, 1905, p. 35-36; Merriam and Buwalda, 1917, p. 261; Glover, 1941, p. 90-92; Newcomb, 1958, p. 331-333, and written communication, 1963). In some of these sections, the total thickness of sand, silt, and clay exceeds 500 feet. The present writers believe that the gravel, sand, and silt of late Pleistocene age that overlies the Ringold Formation in places along the White Bluffs has been erroneously identified as part of the Ringold in a few of these sections. This is because many beds of late Pleistocene age that are concealed by soil, slope wash, or alluvium have been mistakenly included in the measured sections. Therefore, an exposed thickness of the Ringold Formation as great as 500 feet at any one locality along the bluffs is doubtful.

Merriam and Buwalda (1917, p. 263) thought that "the Ringold Formation was deposited in a basin the walls of which were essentially the Yakima Range on the west, the Saddle Mountains on the north, and the lava plateaus on the east and south." They also said (p. 264), "It is probable that this formation originally extended up through the gap which the Columbia has cut across the Saddle Mountains, and that it was deposited over areas north of that range." Later, Culver (1937, p. 57-58) traced the Ringold "to Esquatzel Coulee in the vicinity of Eltopia and northward near Connell." He also stated (p. 58) that deposition of the Ringold Formation "extended northward past Othello and beyond Moses Lake . . . ." The postulations and observations of these early workers have been supported by Taylor (1948, pl. 1), and by our field data.

The buff laminated clay and silt facies in the project area is largely limited to the Pasco, Quincy, and Othello Basins. Other basins on the basalt surface outside the project area, both to the north and to the east, probably received lacustrine deposits also, but not necessarily from the same source.

The brown tuffaceous sand facies is widespread within the project area, and beyond the project area in all directions. As a very fine sand and silt, it extends well beyond the project area to the east. At the White Bluffs, the massive beds of brown tuffaceous sand are, in places, interstratified with lacustrine deposits consisting predominantly of buff laminated clay. The Ringold sections exposed on the Smyrna and Corfu Benches show that basinward interfingering of the brown tuffaceous sand, buff laminated clay, and basaltic fanglomerate facies occurred primarily in the southern part of the Othello Basin. To the north, on the Royal Slope, 4 or 5 miles north of the benches, only buff laminated clay is exposed.

The quartzite-bearing fluvial conglomerate facies apparently is limited to the Pasco Basin. The basaltic fanglomerate facies is associated with piedmont slopes. For example, it occurs along some of the steep slopes of the Saddle Mountains, the Beezley Hills, and in one place along the valley wall of Lind Coulee near Providence.

Where the Ringold Formation was channeled by melt-water streams in late Pleistocene time, or otherwise dissected, its upper boundary is an erosional unconformity beneath glaciofluvial and Holocene eolian deposits. Where the Ringold Formation was not channeled, it is capped by a layer of massive caliche. This caliche layer is the youngest of many such layers that occur repeatedly, although less prominently, throughout the Ringold section.

The caliche is chiefly a mixture of calcium carbonate and silica and contains minor amounts of iron and manganese oxides. It accumulates over a long time interval at land surface or a few feet below it, under an arid or semiarid climatic regime. It occurs either as a white thin coating on the underside of, or around, rock particles of any size, or as a firm cement that binds particles of silt, pebbles, and even boulders into hard rock. In places, the two types of caliche grade into one another. Price (1940, p. 1939), who has studied caliches in the southwestern United States, has named the first type an "outcrop-cementation" caliche, and the second a "plateau-blanket" caliche.

Outcrop-cementation caliche is of relatively minor importance in the project area. It occurs principally in fluvial gravel of late Pleistocene age and is briefly described in a later section of this report. Plateau-blanket caliche, on the other hand, forms the caprock, as thick as 20 feet, that overlies the brown tuffaceous sand, the buff laminated clay, the conglomerate, or the fanglomerate facies of the Ringold Formation. The caprock is an important marker horizon, because it can be traced and identified easily throughout much of the area.

Hypotheses for the origin of caliche fall within three main groups (Bretz and Horberg, 1949, p. 507): (1) Deposition with a surface-water body through chemical or biochemical processes; (2) deposition near the ground surface due to the ascending movement and evaporation of ground water; and (3) deposition at the base of the B soil horizon through leaching of pedocal[1/] soils by descending water (illuviation[2/]). In the central part of the three project-area basins, where the Ringold Formation consists predominantly of laminated silty sand and clay, the massive plateau-blanket caliche caprock represents an evaporite or duricrust[3/]. In the upland parts of the project area, plateau-blanket caliche forms many layers and caps the fanglomerate and brown tuffaceous sand of the Ringold Formation. The third hypothetical process cited above probably was dominant in the accumulation of caliche on the upland. The varying thickness of sand between any two layers of caliche shows that the balance between soil aggradation, weathering, and soil erosion shifted many times for long and short intervals. At places, soil aggradation at land surface accelerated enough so that the corresponding time interval within the caliche zone is represented by uncemented material. In contrast, the presence of small solution cavities in caliche, filled with sand, may be evidence of a degrading soil profile at the time of caliche deposition.

1/ Soil of arid or semiarid region, enriched in lime, which accumulates in regions of low temperature and rainfall and prairie vegetation.

2/ The deposition in an underlying layer of soil of colloids, soluble salts, and small mineral particles which have been leached out of an overlying soil layer. The action occurrs in humid climates.

3/ The case-hardened crust of soil formed in semiarid climates by the precipitation of salts at the ground surface as the ground water evaporates.

The caliche caprock represents a time interval during which all processes of aggradation were greatly reduced. After the caliche had accumulated, sufficient time elapsed for the caliche cap to become dissected and discontinuous, and for a drainage network to develop on the Ringold Formation. The now discontinuous caliche caprock of the Ringold Formation represents a long time break—specifically, the time required for its accumulation, plus that required to lower base levels, permitting dissection of the caliche. In this situation, the apparent conformity of Ringold beds, caliche caprock, and overlying glaciofluvial deposits or loess on the interfluves is coincidental, and cannot be taken as positive evidence of the lack of a time break in deposition or an erosional unconformity.

Although impervious when first exposed to water, the caliche is readily dissolved by circulating water (G. E. Neff, written communication, 1959). Karst topography, which is known to develop on caliche (Price, 1940, p. 1938), has not been definitely identified in the project area. Nevertheless, a few closed depressions on Paradise Flats, in T. 15 N., Rs. 30 and 31 E., and on the Royal Slope north of Natural Corral, in sec. 9, T. 16 N., R. 26 E., may owe their origin not only to deflation of silt and sand overlying caliche, but to solution sinks in the caliche as well.

Brown Tuffaceous Sand

At a place in the vertical cliff along the north bank of the Columbia River, in sec. 2, T. 14 N., R. 26 E., Calkins (1905, p. 35-36) measured a stratigraphic section that contained beds of "red sandstone and terra cotta silt." These and rocks of similar appearance and lithology are referred to as "brown tuffaceous sand" throughout the present report. Bryan (1927, p. 26) concluded that the sand (which he described as "finely laminated gray-brown silt") was water-laid, based on exposures along the old state highway between Lind and Connell, in the SE$\frac{1}{4}$ sec. 20, T. 14 N., R. 32 E. The sand occurs there in the upper part of the bluffs and is capped by caliche. The lower contact is hidden by drifting sand. The brown tuffaceous sand probably overlies the buff laminated clay exposed in the roadcut a quarter of a mile west and in a new roadcut along U.S. Highway 395 in the SE$\frac{1}{4}$ sec. 19, T. 14 N., R. 32 E., away from the base of the bluffs.

The brown tuffaceous sand facies in the project area is similar to the soil substratum described by Bryan (1927, p. 28) in a roadcut immediately west of the railroad siding at Fishtrap, 25 miles southwest of Spokane (east of the project area). Bryan (1927, p. 44) called this substratum the basal part of the Palouse loess. The substratum lies beneath the light-colored loess that forms the so-called inner core of the Palouse Hills. A. C. Waters (written communication, 1961), who has studied the Palouse loess sheets petrographically, recognized several stratigraphic units. The basal unit is reddish brown and locally indurated, resembling a soft brick.

In this report, where the field identification of rocks was based solely on megascopic characteristics, a layer-by-layer correlation of the brown tuffaceous sand underlying the uplands with the sand intercalated in the buff laminated clay at or near the White Bluffs was not possible.

The sand facies consists of compact to loosely consolidated tuffaceous debris wherever exposed in the eastern and northern uplands as well as at the type locality along the White Bluffs. The sand is massively bedded at most localities, though massive and thin-bedded layers alternate in the upland east of lower Esquatzel Coulee, in Franklin County.

The vertical, somewhat prismatic parting in the layers of the sand, and the massive bedding, both of which are the structural characteristics of loess, suggest a primary eolian deposition. Root casts of caliche and calcareous nodules and tubules occur in many exposures.

The brown tuffaceous sand contains a paleosol[1/] or a succession of paleosols, as indicated by the many caliche layers (some of which may have developed at the base of the B horizon), by calcareous concretions—interpreted as root casts—and by burrows of small rodents.

At many places, the brown tuffaceous sand facies lies unconformably over the Yakima Basalt. In the western part of the Quincy Basin, at the head of Crater Coulee, the sand unconformably overlies the Roza Member. In the scabland south of the Potholes Coulee, and in the northeastern part of the Drumheller Channels, it is unconformable over both the Roza and the Priest Rapids Members (fig. 22A). In the vicinity of Warden, it is unconformable over the Frenchman Springs Member.

Reworked brown tuffaceous sand and silt was interfingered with the buff laminated clay facies in a nearshore environment as seen in a railroad cut 5 miles southwest of Warden, in the NW¼ sec. 1, T. 16 N., R. 29 E., and in the north-facing bluff of the Smyrna Bench.

Reddish-brown massively bedded sand similar to that in the White Bluffs is overlain by typical buff laminated clay in a roadcut 2 miles north of Othello, in the NW¼ sec. 27, T. 16 N., R. 29 E. The base of the sand is not exposed, but probably rests on the Saddle Mountains Basalt Member.

Brown tuffaceous sand is overlain by buff laminated clay in the walls of Lind Coulee, a short distance west and east (fig. 23) of the State Highway 17 bridge. Two miles downcoulee from the bridge, in the SW¼ sec. 3, T. 17 N., R. 29 E., the sand rests unconformably atop the Roza Member.

The brown tuffaceous sand facies also underlies extensive areas along the western and southwestern margins of the Quincy Basin. East of the Drumheller and Othello Channels, the sand crops out in gullies east of the areas underlain by buff laminated clay. The sand also occurs high along the slopes of ridges like the Frenchman Hills and the Saddle Mountains and even along the crests where not removed by erosion. Examples of this type of occurrence are the roadcut exposures south of Low Gap Pass on the Frenchman Hills, in the NW¼NE¼ sec. 13, T. 17 N., R. 24 E., and those in the S½ sec. 32, T. 15 N., R. 29 E., near the eastern end of the Saddle Mountains. The sand also underlies the upland in the eastern and northern parts of the project area, where it has not been removed by erosion. Exposures are rare because it is covered by younger loess. Most of the exposures were too small to be mapped at the scale of the geologic map of Grolier and Bingham (1971). They are restricted mostly to roadcuts and, more rarely, to the steep slopes of melt-water channelways and gullies. A few of the best localities are: (1) the northern edge of an alcove a quarter of a mile southeast of Black Lake in Black Rock Coulee, in the SW¼NE¼ sec. 3, T. 20 N., R. 29 E. (at this locality, the brown sand, several feet thick and capped with caliche, overlies the Priest Rapids Member); (2) the roadcut along State

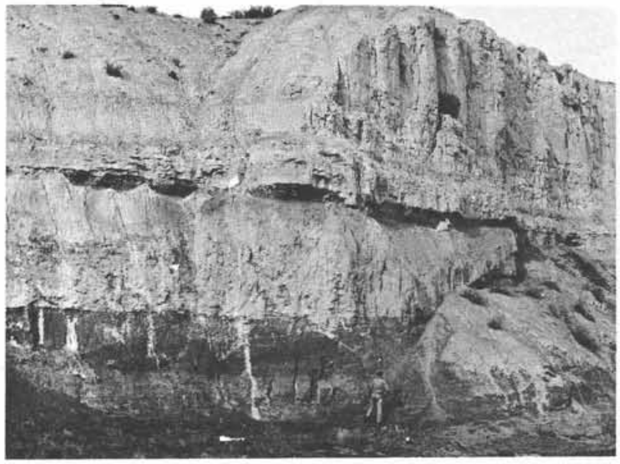


FIGURE 23.—Brown tuffaceous sand (darker sediments near base of wall) and overlying lacustrine clay of Ringold Formation, in south wall of Lind Coulee, 500 feet east of Highway 17 in NE¼NE¼ sec. 35, T. 18 N., R. 29 E. (U.S. Bureau of Reclamation photo.)

1/ Buried soil horizon of the geologic past.

Highway 26 a quarter of a mile west of Providence Coulee; (3) roadcuts along the north slope of Dunnigan Coulee, in secs. 23 and 24, T. 13 N., R. 33 E.; (4) a roadcut on the Moses Lake-Stratford Road, in the NW¼ sec. 11, T. 20 N., R. 28 E. (Grolier and Foxworthy, 1961); and (5) a roadcut in the southwestern part of the Quincy Basin, at the head of a gully in the SW¼NW¼ sec. 33, T. 18 N., R. 23 E. Except for its being coarser grained, the brown sand in these outcrops is similar to the loess rock at Fishtrap (in eastern Lincoln County) described by Bryan (1927, p. 28).

The accumulation of the brown tuffaceous sand facies in eastern Washington doubtless occurred at repeated intervals of geologic time. Because of its widespread geographic distribution, the eolian sand, which grades eastward to a silt, is not associated with a particular topographic feature in the project area as it is in the hills in the core of the Palouse country, the extension of the Palouse country proposed by Caldwell (1961, p. 117, fig. 1), or the linear hills described by Lewis (1960, p. 98, fig. 1).

Large amounts of the tuffaceous sediment accumulated in lake basins where the laminated clay was settling. The brown tuffaceous sand is the dominant constituent of many beds at various stratigraphic levels of the Ringold Formation; it may underlie, overlie, or be interstratified with lacustrine deposits at the White Bluffs. Each stratum of the tuffaceous sand probably records an eruption of pyroclastic debris.

Quartzitic Conglomerate

The conglomerate facies exposed at White Bluffs (fig. 21) is an aggregate of quartzite, other metamorphic rock, granitic rock, porphyry, and basalt pebbles and cobbles, most of which are well rounded with some flat and elongate; the interstitial material is largely subangular quartzose sand. It is loosely coherent but has strongly cemented zones; it was called "conglomerate" by Newcomb (1958) to distinguish it from the glaciofluvial gravel.

The constituents of the Ringold conglomerate at the White Bluffs, their elongation, and their degree of roundness are similar to those of the conglomerate in the Beverly Member of the Ellensburg Formation. Lithologically, they also resemble coarse flood-plain deposits of the present-day Columbia and Snake Rivers, although the Ringold constituents are more elongate, with a lower percentage of basalt pebbles. When stratigraphic relationships are obscured by cover, it is difficult to assign with assurance isolated outcrops of quartzitic conglomerate to the proper unit. Quartzitic conglomerate may have been deposited at any time before or after the deposition of scabland gravel.

Well drillers' logs show that the conglomerate facies occurs as a lens in the silt and clay of the Ringold Formation (fig. 21). The maximum reported thicknesses of the conglomerate within the project area are 182 feet in well 11/29-5N1 and 193 feet in well 11/29-20N1 (where it rests on basalt); the unit thins rapidly to its eastern edge.

Conglomerate similar to that exposed in the southern part of the White Bluffs crops out at three other places in the project area: (1) A conglomerate overlain by a layer of caliche is poorly exposed under drifting sand in secs. 13, 23, and 27, T. 10 N., R. 31 E., about 11 miles northeast of Pasco (fig. 21). It consists of elongate quartzite, other metamorphic rock, and porphyry pebbles, but it differs from the Ringold conglomerate at the type locality in having a higher percentage of greenstone pebbles. The base is not exposed. (2) A gravel consisting of pebbles and cobbles of quartzite, greenstone, and basalt is exposed in a pit 4½ miles northeast of Pasco, in the NE¼NE¼ sec. 23, T. 9 N., R. 30 E. (fig. 21). Its age is not known but it may be Ringold on the basis of lithologic similarity to the conglomerate exposed along the White Bluffs. Foreset bedding in this gravel dips northward and away from the Snake River (fig.

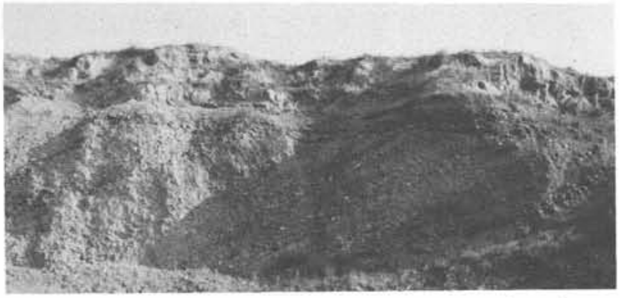


FIGURE 24.—Quartzite- and greenstone-bearing gravel possibly of Ringold age in gravel pit in NE$\frac{1}{4}$NE$\frac{1}{4}$ sec. 23, T. 9 N., R. 30 E. Foreset bedding in gravel dips northward away from Snake River; bedding of the glaciofluvial basaltic gravel dips gently southward. Windblown silty sand overlies the basaltic gravel.

FIGURE 25.—Quartzite-bearing conglomerate of the Ringold Formation in roadcut at mouth of Parsons Canyon, in the NE$\frac{1}{4}$SW$\frac{1}{4}$ sec. 1, T. 11 N., R. 28 E. Crossbedding of conglomerate here dips eastward away from Columbia River.

24; Bretz, 1928c, p. 681). (3) A somewhat indurated quartzitic gravel similar to the Ringold conglomerate, but lacking the caliche cap, is exposed in pits in the NW$\frac{1}{4}$SW$\frac{1}{4}$ sec. 3, T. 16 N., R. 23 E., and the SW$\frac{1}{4}$ SW$\frac{1}{4}$ sec. 10, T. 16 N., R. 23 E. The base is not exposed at either locality, but field relationships suggest that it unconformably overlies the Roza Member at the first locality and the Priest Rapids Member at the second. Gravel of similar lithology occurs as lag after wind erosion of the sand along the top of the gravel terrace and along the gravel scarp north of the mouth of Jericho Coulee, 2 miles northeast of Beverly.

In the northernmost outcrops of the conglomerate facies along the White Bluffs, the imbrication of the pebbles indicates that deposition took place in a stream flowing southward or southeastward (fig. 25). Farther south, crossbedding and shingling point to repeated shifting in current direction.

Buff Laminated Clay

The buff laminated clay facies includes strata of clay with subordinate silt and fine sand, which were deposited in a lacustrine environment in the Pasco Basin. The unit is presently best exposed in the White Bluffs, where beds of sand, silt, and clay progressively grade upward into one another, with the superposition of beds recurring many times in the section. This recurrence suggests cycles of deposition during which environmental conditions, probably those of a large lake or river flood plain, repeated themselves. The clay and silt commonly are indurated to the consistency of a claystone and siltstone, and are either laminated or massively bedded, commonly breaking into blocks along vertical joint planes (fig. 26). Where the clay lies at considerable depth below the regional water table and therefore has not been oxidized since its deposition, it is blue or green, and is known to drillers as "blue clay" (Newcomb, 1958, p. 335-336).

Sand and silt layers within the clay facies are predominantly quartzose and micaceous. In many of the thinner layers, the sand is loose and exhibits false cross-lamination associated with current ripples; in other layers, the sand is indurated.

Buff laminated clay extends away from the White Bluffs beneath a cover of glacial melt-water deposits, rarely thicker than 100 feet, or under a very thin cover of Holocene loess or dune sand. The clay

FIGURE 26.—Laminated, thinly bedded siltstone and claystone of the Ringold Formation in borrow pit in SW$\frac{1}{4}$SW$\frac{1}{4}$ sec. 20, T. 15 N., R. 30 E.

extends as far north as the upper part of Wahluke Slope, where it crops out in a roadcut in the NW$\frac{1}{4}$SE$\frac{1}{4}$ sec. 24, T. 15 N., R. 23 E., at an altitude of 1,160 feet, and also in a gully in the NW$\frac{1}{4}$ sec. 27, T. 15 N., R. 24 E., at an approximate altitude of 1,100 feet. To the northeast, beyond the Othello Channels, the clay facies is exposed in the bluffs at the western and southern margins of Paradise Flats (fig. 26), and in bluffs 1$\frac{1}{2}$ miles northeast of Connell.

North of the Saddle Mountains and Othello Channels, buff laminated clay underlies the high part of the Royal Slope from the western margin of the Drumheller Channels to the vicinity of Royal City. It is exposed in the north-facing bluffs of the Taunton Terrace, and occurs sporadically near the top of the bluffs along the north-facing cliffs of the Corfu and the Smyrna Benches. The clay also partly underlies Sand Hollow at the western end of the Royal Slope, where it is exposed in borrow pits, resting unconformably on both the Frenchman Springs and Roza Members of the Yakima Basalt.

The only exposures of the clay facies in the Quincy Basin are in the SW$\frac{1}{4}$SW$\frac{1}{4}$ sec. 6, T. 17 N., R. 28 E., along the southern part of the bluffs, east of Pelican Horn of Moses Lake (Grolier and Foxworthy, 1961), and along the vertical banks of Lind Coulee (fig. 23), in the western part of the coulee's course toward Potholes Reservoir. Data from well drillers' logs indicate that the clay fills the deepest part of the Quincy Basin under a thick cover of younger deposits.

The areas presently underlain by buff laminated clay are discontinuous. Many hills within meltwater channels represent outliers composed mostly of buff clay. Among them are: (1) The large crescent-shaped hill mostly in T. 10 N., R. 30 E., where buff clay is exposed in shallow dug pits; (2) Jackass Mountain, 3$\frac{1}{2}$ miles southwest of Eltopia, in secs. 17, 20, and 29, T. 11 N., R. 30 E.; (3) a second Jackass Mountain, 5$\frac{1}{2}$ miles north of Othello mostly in sec. 3, T. 16 N., R. 29 E.; and (4) Cash Butte, 3 miles southwest of Royal City, in the SW$\frac{1}{4}$ sec. 10, T. 16 N., R. 25 E.

Buff laminated clay also is exposed in isolated outcrops far to the east of the previously mentioned areas. For example, it crops out at road level along the northwest side of the Pasco-Kahlotus highway, in the SE$\frac{1}{4}$ sec. 3 and in the NE$\frac{1}{4}$ sec. 15, T. 10 N., R. 32 E., at an approximate altitude of 900 feet.

Erosional remnants of a once-extensive valley fill of buff laminated clay occur in two of the coulees that trend westward from the eastern upland. One of the remnants is near the junction of Hardesty and Washtucna Coulees at an altitude of about 1,020 feet in the SW$\frac{1}{4}$ sec. 32, T. 14 N., R. 33 E., where it was first reported by Bretz and others (1956, p. 1004). The other remnant is an outcrop of massively bedded clay, unconformable against the Frenchman Springs Member on the south wall of Lind Coulee, at a maximum elevation of 1,340 feet in the NE$\frac{1}{4}$ sec. 33 and the NW$\frac{1}{4}$ sec. 34, T. 17 N., R. 32 E. Clay 120 feet thick also is reported overlying basalt in well 16/32-3B1, 1$\frac{1}{2}$ miles to the southeast.

Clay and coarse sand that may be equivalent in age to the buff laminated clay facies but spatially unconnected with the principal area of Ringold deposition in the central and southern parts of the project area occur in the Bacon syncline, in the extreme northern part of the project area. The sediments are exposed in a railroad cut near the center of sec. 9, T. 23 N., R. 28 E., where they rest on the Priest Rapids Member and are overlain by fluvial gravel.

Basaltic Fanglomerate

A part of the basaltic fanglomerate facies of the Ringold Formation is best exposed along the steep cliffs of the Smyrna Bench, a truncated remnant of the piedmont slope into the Othello Basin from the central part of the Saddle Mountains. There, the fanglomerate is about 400 feet thick and includes several lenses of buff laminated clay.

The basaltic fanglomerate facies consists of angular to subangular, poorly sorted pebbles, cobbles, and boulders of basalt in a matrix of brown tuffaceous sand and silt. Layers of fanglomerate tens of feet thick locally alternate with, or grade laterally into, layers composed almost entirely of brown tuffaceous sand containing root casts and tubules coated with caliche. Basaltic rock fragments in the sand are from flows of the Priest Rapids, Roza, and Frenchman Springs Members that were undergoing erosion on the nearby mountain slopes at the time of fanglomerate deposition. The nonbasaltic rock fragments are opal, petrified wood, and sporadic, small, subangular, wind-faceted pebbles of black volcanic glass that occur in place near the top of the Beverly Member.

The fanglomerate facies at this locality consists of two units: a lower one with the basalt cobbles and boulders stained brown, and a thin upper unit with fresher unstained basalt cobbles and boulders. The lower unit rests on a sequence of brown tuffaceous sand, bluish-black clay 10 to 20 feet thick, and the Saddle Mountains Member of the Yakima Basalt. The upper unit is capped by the massive caliche.

Immediately southeast of Sentinel Bluffs, the basaltic fanglomerate facies overlies hogbacks of the youngest basalt on the south slope of the Saddle Mountains, and forms northwest-trending bluffs across secs. 23 and 26, T. 15 N., R. 23 E. The fanglomerate there differs in lithology from that at any other locality because, in addition to angular basalt debris, it contains well-rounded pebbles and cobbles of quartzite, felsite porphyry, and other rock types reworked from the Beverly Member. Because the Beverly conglomerate crops out north and upslope from the hogbacks referred to above, it probably was the source of the rounded non-basalt debris in the Ringold fanglomerate. At the southwestern end of this area, glaciofluvial erosion has stripped the fanglomerate from the underlying Saddle Mountains Member. In the bluffs and the gullies in secs. 23 and 26, the exposed fanglomerate consists predominantly of massive brown tuffaceous sand with thin layers of angular basalt fragments. The exposed section there is capped by caliche.

Channeling by glaciofluvial streams has produced many good exposures of the basaltic fanglomerate facies at other places in the project area. An exposure nearly 3 miles long occurs atop a monocline west of Corfu, where the fanglomerate is associated with lacustrine sand, silt, and clay, and unconformably rests on the Saddle Mountains Member near the top of the monocline.

Basaltic fanglomerate about 130 feet thick and capped with massive caliche is exposed at the top of the cliff in secs. 31 and 32, T. 16 N., R. 25 E. It rests disconformably over the Priest Rapids Member. Badlands-type pinnacles, resulting from wind and water erosion of the fanglomerate, stand out a few feet below and north of the cliff. The altitude of the base of the fanglomerate there is about 1,300 feet.

Basaltic fanglomerate underlies the higher part of the Wahluke Slope along the south flank of the

Saddle Mountains anticline and the piedmont slope at the base of the Beezley Hills, north of Winchester and Quincy in the Quincy Basin. The fanglomerate at those localities is exposed in only a very few places. Away from the mountain front, the fanglomerate is overlain by a thin veneer of unconsolidated alluvial-fan debris and eolian silt accumulated in Holocene time.

The basaltic fanglomerate facies crops out for a distance of 2 miles along the west wall of the Ephrata Channel where the piedmont slope of the Beezley Hills has been truncated. There, the fanglomerate rests unconformably over the Frenchman Springs and Roza Members and is made up of debris eroded from the Beezley Hills monocline.

In the Columbia River water gap through the Frenchman Hills anticline, angular basalt debris rests on the bench atop the lower flows of Yakima Basalt, and against the scarp in the SE$\frac{1}{4}$ sec. 8, T. 17 N., R. 23 E. There, the basalt debris is mixed with the brown tuffaceous sand of the Ringold Formation. A thick caliche layer caps the talus, and is parallel to the talus slope. This deposit is a remnant of talus that probably extended continuously along the Frenchman Hills scarp prior to its almost complete removal by glacial melt-water streams. The talus also crops out under gravel deposits almost as far north as Trinidad. This deposit, in secs. 17, 18, and 20, T. 20 N., R. 23 E., has been described by Bretz and others (1956, p. 987-988).

The presence of the caliche-capped fanglomerate of apparent Ringold age in the Frenchman Hills water gap means that part of the uplift of the Frenchman Hills and the entrenchment of the ancestral Columbia River had occurred before this part of Ringold deposition. Also, this relation would indicate that the Columbia River occupied this part of its course before the accumulation of the brown tuffaceous sand in talus.

<u>Gravel Deposits in Wind Gaps</u>

Gravel and sand deposits occur in wind gaps within both the Saddle Mountains and Frenchman Hills. The deposits lie well above the shorelines of lakes that developed as a result of submergence by glaciofluvial currents in late Pleistocene time. Three deposits are tentatively considered of Ringold age and mapped as "undifferentiated Ringold Formation."

In the Saddle Mountains, two gravel deposits are referred to the Ringold. One is a gravel and sand deposit that occurs as a bar in an abandoned water gap in secs. 30 and 31, T. 15 N., R. 29 E. It ranges from 1,210 to 1,330 feet in altitude (Bretz and others, 1956, p. 999). The deposit is composed of rounded pebbles that are predominantly basaltic and of quartz and feldspar sand. The deposit rests on Saddle Mountains basalt, and is capped with a layer of massive caliche that is several feet thick. The maximum thickness of the deposit is about 30 feet. The other deposit consists of basaltic gravel with a caliche cap and occurs at the east end of the Saddle Mountains, mostly in sec. 10, T. 14 N., R. 29 E. The deposit ranges in altitude from about 1,150 to 1,270 feet. The best exposure is at the top of the scarp in the SE$\frac{1}{4}$NE$\frac{1}{4}$ sec. 10.

In the Frenchman Hills, a gravel deposit occurs in the SW$\frac{1}{4}$NE$\frac{1}{4}$ sec. 19, T. 17 N., R. 27 E., in a pit at the southeast end of an emergency airstrip. The deposit is at an altitude of 1,410 feet. It is composed of rounded pebbles of basalt that are cemented by caliche in the upper few feet. The gravel is about 8 feet thick, and overlies caliche-impregnated brown tuffaceous sand of the Ringold Formation. The deposit extends northward down the slope of the Frenchman Hills and is exposed again in a roadcut in the NW$\frac{1}{4}$ NW$\frac{1}{4}$ sec. 20, T. 17 N., R. 27 E. At this locality, the brown tuffaceous sand is interstitial to the basalt pebbles; however, it is labeled Qg on the geologic map of Grolier and Bingham (1971).

PLEISTOCENE DEPOSITS

Alluvium Older Than Glaciofluvial Deposits

Several alluvial deposits, seemingly unrelated to one another or to any other time-stratigraphic unit in the area, are known from isolated exposures. Some deposits are too small to be mapped, and others were not differentiated from fluvial gravel. The chief deposits are listed below because of their stratigraphic interest and their bearing on the geologic history. They are scattered in the project area.

Conglomerate in Esquatzel Coulee

An erosional remnant of an indurated conglomerate rests on a basalt bench about 50 feet above the valley floor of Esquatzel Coulee. It is exposed near the base of the north coulee wall, at an altitude of about 700 feet, 1 mile northeast of Mesa and half a mile east of State Highway 17, in the SW$\frac{1}{4}$SE$\frac{1}{4}$ sec. 24, T. 13 N., R. 30 E. It is not shown on the geologic map of Grolier and Bingham (1971).

The conglomerate consists of angular slabs of basalt, and small rounded pebbles of basalt and quartzite tightly cemented by carbonate. It is 25 to 30 feet thick and rests on the Roza Member. Based on its lithology and degree of cementation, the conglomerate is tentatively interpreted as an erosional remnant of alluvium in a prescabland valley. The alluvium may have been deposited when the ancestral Palouse River followed a southwesterly course down the sites of Washtucna and Esquatzel Coulees into the Pasco Basin as described by Bretz (1928c, p. 662).

Gravel Deposit in Old Maid Coulee

On the north side of Old Maid Coulee, a gravel deposit is exposed in a pit in the NE$\frac{1}{4}$NE$\frac{1}{4}$ sec. 24, T. 13 N., R. 31 E., on the west side of U.S. Highway 395, and in the NE$\frac{1}{4}$NW$\frac{1}{4}$ sec. 19, T. 13 N., R. 32 E., east of the highway. The exposure is at an altitude of 1,050 feet. It is shown as fluvial gravel on the geologic map of Grolier and Bingham (1971).

Quartzite and caliche pebbles, and micaceous quartz and feldspar sand are mixed with gravel composed of weathered basalt ranging from granule to boulder size. The deposit is capped with caliche; the base is not exposed. Foreset bedding dips southwestward, down coulee. Bryan (1927, p. 27 and pl. 5A) first described this deposit and interpreted it as "waterworn gravel, evidently deposited by a shifting, but not necessarily turbulent stream." This gravel also has been described in detail by Flint (1938, p. 518), and later by Bretz and others (1956, p. 1006) who thought that the gravel probably was deposited by a glaciofluvial stream.

Gravel Deposits in the Quincy Basin

Several deposits of fluvial gravel in the Quincy Basin are older than most of the glaciofluvial gravel. The deposits, consisting of cobbles and boulders of basalt, caliche, and granite, underlie the area east of Potholes Coulee and the channel floor at the head of Crater Coulee. They were included with the glaciofluvial gravel on the geologic map of Grolier and Bingham (1971). In a pit where the deposit in Crater Coulee is well exposed, it consists of coarse basalt cobbles with some boulders of the Vantage Sandstone Member (Bretz and others, 1956, p. 985, 988). Materials of similar composition underlie parts of secs. 30, 31, and 32, T. 19 N., R. 24 E. (fig. 27).

A striking feature of these deposits is the dip of their bedding surfaces, which is to the east, southeast, or south, away from the Columbia River. Thus, the sediments must have been deposited by a large river flowing in a direction opposite to that indicated by the present longitudinal profile of Crater and Potholes Coulees. This conclusion is supported by the presence of the soft Vantage sandstone boulders

FIGURE 27.—Caliche-capped basaltic gravel in pit in NW$\frac{1}{4}$ sec. 31, T. 19 N., R. 24 E. Foreset bedding in gravel dips eastward away from the Columbia River.

that could have been transported only a few miles from nearby outcrops in Lynch Coulee, and by the decrease in average grain size of the deposit eastward from the coulees. Possibly this is a record of an early flow of glacial runoff into the Quincy Basin and resulted in the high erratics and silt deposits that preceded the erosion of the low outlet through Drumheller Channels.

Gravel Deposit in Lynch Coulee

A small deposit of gravel is exposed on the north slope of Lynch Coulee, in a railroad cut along the abandoned right-of-way of the Burlington Northern railroad, near the center of sec. 32, T. 21 N., R. 23 E. The gravel, about 15 feet thick, is deeply weathered, and consists of angular cobbles and boulders of basalt. The base is not exposed. The degree of weathering, the crumbling aspect of the rock fragments, and the till-like appearance make this deposit unique among those observed in the project area. The deposit is not shown on the geologic map of Grolier and Bingham (1971).

Glaciofluvial Deposits

The glaciofluvial deposits consist of the unconsolidated silt, sand, and gravel unconformably overlying the older rocks and unconsolidated materials. These materials record deposition in temporary lakes and in or near the channels used by the catastrophic floods during late Pleistocene time. In this report, these outwash deposits are subdivided and referred to as "lacustrine fine sand and silt," and "glaciofluvial gravel and sand."

Lacustrine Fine Sand and Silt

The lacustrine deposits were first recognized and described from exposures in Walla Walla Valley by Bretz (1928b, p. 325-328). Bretz (1928c, pl. 5) also showed their areal extent to the area northeast of Pasco, and thence to Providence Coulee, Wahluke Slope, Paradise Flats, the Royal Slope, and Sand Hollow. He also showed the occurrence of the deposits in the area northeast of Warden, and in a later report (Bretz and others, 1956, p. 983) they were described in a part of the Quincy Basin.

The name "Wahluke" was originally suggested by Beck (1936, p. 12) for the thin layer of sediments lying above "lime rock" (plateau-blanket caliche), in a "typical exposure . . . above undoubted Ringold at the head of the Beverly highway grade a mile or so . . . [north] of Wahluke." He also pointed out the occurrence of similar deposits in the valley cut and fill along the lower part of Lind Coulee. These deposits, which include Holocene silt and sand, he termed "the Tiflis member of the Wahluke sediments."

Flint (1938, p. 493-499, fig. 5) named the fine-grained deposits the "Touchet beds," in an area extending from the Horse Heaven Hills south of the project area northward to the Saddle Mountains and eastward to Walla Walla. The name is taken from particularly good exposures in the vicinity of Touchet in the Walla Walla Valley (outside the project area, 23 miles southeast of Pasco).

Two principal periods of deposition of lacustrine silt are recognized in the Pasco and the Quincy Basins, based on the areal distribution of the silt below two faintly defined shorelines, at altitudes of 1,150 feet and 850 feet in the Pasco Basin, and of 1,350 feet and 1,200 feet in the Quincy Basin. The older silt occurs above the lower shoreline and may be recognized by a slight discoloration from the original buff to a light brown. Both the older and younger silts occur below the lower altitude in the basins of deposition like the Quincy, Othello, and Pasco Basins and in valleys tributary to scabland channels.

Although the grain size is silt and fine sand, as many as five textural constituents were recognized by Bretz (1928b, p. 325-326):

1. Coarse, angular black (fresh) basaltic sand and gravel, occurring in lenses and pockets, or disseminated in grains throughout stratified silt.
2. Light-colored very fine sand, occurring as very irregular lenses within fine gravel in some parts of the deposits.
3. Grayish-buff to light-brown stratified and locally laminated silt.
4. Ice-rafted pebbles, cobbles, and boulders, consisting mostly of metamorphic and other nonbasaltic rock fragments, usually found in the silty phase of the deposits.
5. Yellowish sandy clay surrounding ice-rafted fragments (this constituent was not observed during the present investigation).

Up tributary valleys, away from the major glaciofluvial channelways, the average grain size of the lacustrine sediments decreases. Down valley, the average grain size grades into basaltic sand and gravel.

Stratification is at a minimum near the upper altitude limit of the silt. The deposits at lower altitudes are characterized by rhythmic bedding as a sand and silt alternation varying from a few inches to a few feet in thickness. This variation in stratification is best illustrated in exposures near the confluence of Weber and Lind Coulees, in the NW$\frac{1}{4}$ sec. 3, T. 18 N., R. 30 E., and also in a railroad cut in the NW$\frac{1}{4}$ sec. 4, T. 17 N., R. 30 E. The massive layers of silt stand out in relief on the vertical face of some of the exposures. They are etched by wind erosion, which preferentially removes the less compacted and finer material in the crenulated laminae of the current-bedded sand layers.

The lacustrine fine sand and silt is calcareous throughout. It is loosely compacted, slightly cohesive, and temporarily stands in steep walls in ravines and cuts.

The silt forms a thin veneer, only a few feet thick, along the valley and basin slopes and is as much as 75 feet thick at the bottom of and against the lower slopes of coulees. There, the silt forms nearly flat-topped graded terraces that stand as much as 50 feet above the present valley floors.

Perhaps the most obvious characteristic of the lacustrine fine sand and silt is the presence of innumerable clastic dikes. The dikes are generally vertical multiple bands of the same material in a maze of fissures that opened repeatedly after most of the deposit had been laid down. The dikes occur either singly or as multiple compound dikes, with a maximum thickness of several feet. The beds in the dikes are layered generally parallel to the fissure walls, and many dikes contain cross laminae of sand in one or more layers. Fissure walls are coated with paper-thin layers of clay. In some places, the dikes extend downward through the massive caliche layer (fig. 28A), upon which the lacustrine fine sand and silt may rest, into the brown tuffaceous sand and lacustrine clay of the Ringold Formation—for instance, along the bluffs in the SW$\frac{1}{4}$ sec. 3, T. 14 N., R. 26 E. (fig. 28B), and in isolated exposures of Ringold Clay along the Pasco-Kahlotus Highway, in sec. 3, T. 10 N., R.

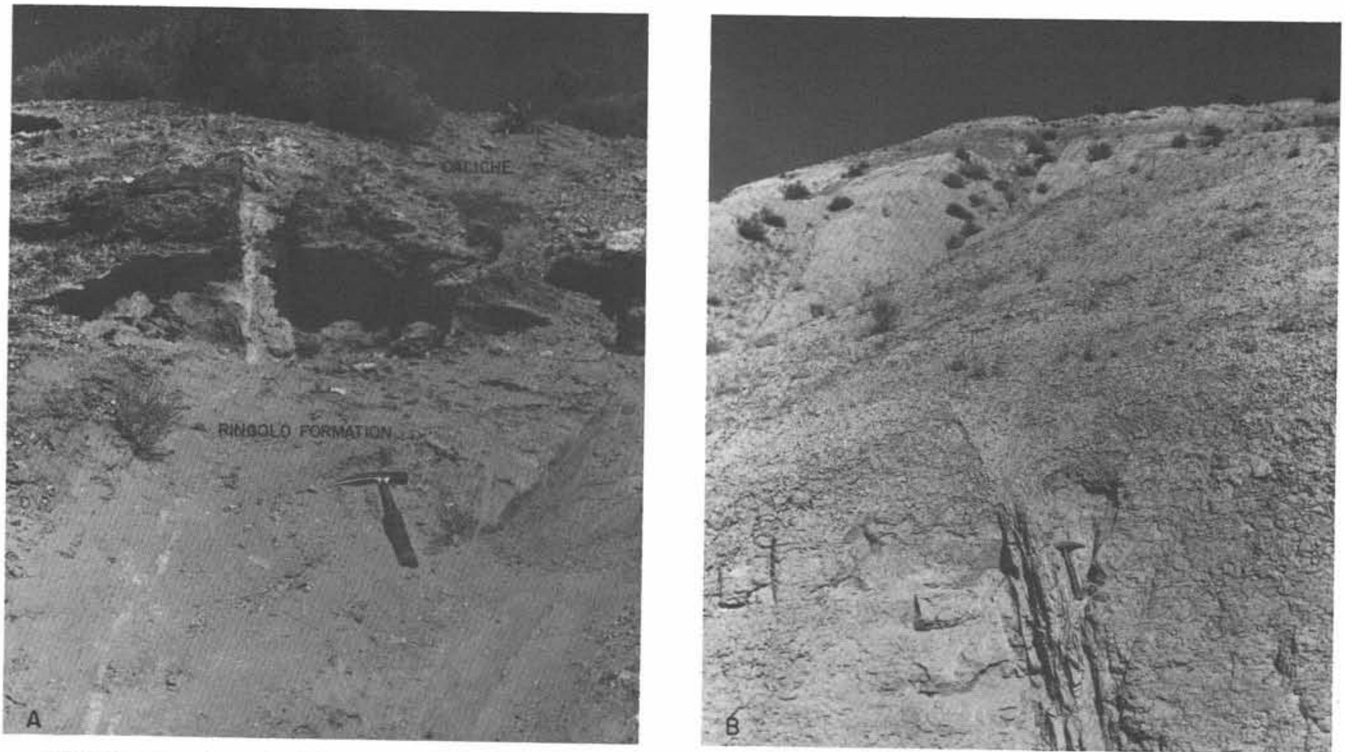


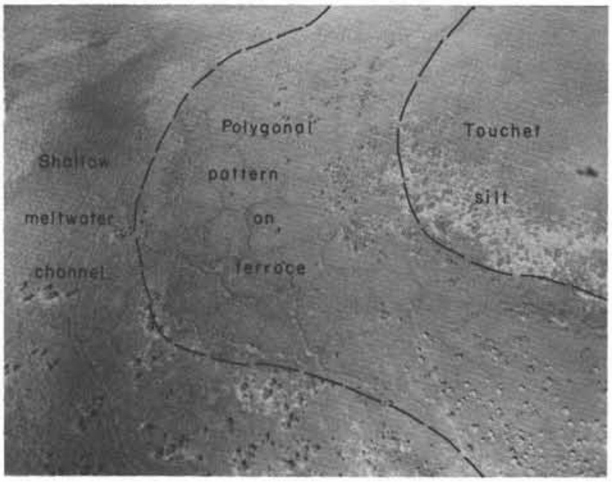
FIGURE 28.—Clastic dikes cutting through Ringold Formation and overlying caliche in SW¼SW¼ sec. 3, T. 14 N., R. 26 E. A. Dike cutting caliche that caps Ringold Formation. B. Dike cutting Ringold laminated clay. Caliche capping shown in figure 28A is at top here, about 150 feet above hammer. (General Electric Hanford Operations photo.)

32 E. The lithologic characteristics of the dikes in the area southeast of this project area have been extensively described by Lupher (1944).

In plan, the dikes form a polygonal network (fig. 29A) "with a distance across the cells ranging from 50 to 400 feet . . . ." (Newcomb, 1962, p. 70). This polygonal network is most easily seen in the Pasco Basin in sand-dune blowouts, or where the dikes have penetrated the coarse sand phase of the glaciofluvial sediments, and the overlying silts have been removed by deflation, as they were in the N½ T. 9 N., and the S½ T. 10 N., R. 30 E. In the latter place, the polygonal pattern is emphasized by vegetation (fig. 29A and B). The growth of ragweed and Russian thistle is more vigorous along the dikes than is the cheat grass over the permeable sand and gravel, probably because the high silt content of the dikes favors a high moisture retention.

Examination of the dikes by Newcomb has lead him to interpret (1962, p. 70) that " . . . the clastic dikes resulted from upward injections of ground water. . . . caused by bank-storage effluent when a pressure difference was produced by a large lowering of Lake Lewis. . . . later injections . . . produced the many laminae of the dikes."

The authors infer that most of the dikes resulted from the filling of desiccation cracks or deep frost cracks that occurred in the bottom sediments of the intermittent Lake Lewis during glacial periods. Lake Lewis is thought to have been short lived but recurrent because of the lack of definite shorelines and the absence of clay-size particles in the cyclic bedding of the lacustrine facies. Periodic drying of the lake permitted either one or both of the crack types to occur along the same zone of weakness as previous cracks, producing the vertically bedded dikes.



A. Aerial view of pattern.

B. Closeup of pattern.

FIGURE 29.—Polygonal pattern of clastic dikes in coarse sand in NE$\frac{1}{4}$ sec. 36, T. 10 N., R. 30 E. At this locality, polygonal pattern is restricted to terrace east of shallow melt-water channel. Southwesterly winds removed lacustrine fine sand and silt that mantled terrace, exposing plan view of dike pattern.

The upper limit of deposition of the glaciofluvial deposits is commonly marked by the presence of erratic pebbles and boulders which are scattered or in nests. These were interpreted long ago by Russell (1893, p. 26-27) and Bretz (1919, p. 498-499, fig. 2) as debris which was ice rafted and then stranded along the shoreline of the lakes in which deposition of the silt is assumed to have taken place. These "berg dumps" occur, for example, on a terrace along Highway 26, 2 miles east of Taunton, in the NE$\frac{1}{4}$NE$\frac{1}{4}$ sec. 9, T. 15 N., R. 28 E., at an elevation of about 780 feet, and along the Wahluke Slope in the NE$\frac{1}{4}$ NW$\frac{1}{4}$ sec. 27, T. 15 N., R. 24 E., at an altitude of about 1,150 feet.

Inconspicuous strandlines, marked by almost imperceptible scarplets bordering narrow wave-cut terraces, are also useful in recognizing the upper limit of the lacustrine silt deposits. They are most easily seen on air photographs, but in some places may be easily recognized in the field as on the loess-mantled north slope of the Saddle Mountains immediately above the Taunton Terrace.

Glaciofluvial Gravel or Sand

Glaciofluvial gravel or sand was deposited at points where scabland channels emptied into basins or large river valleys, or where the channels widened; in such places, a diminished water velocity resulted in deposition of part of the floodwater sediment load. The scabland channels, having been scoured by rivers and floods of great capacity, are today remarkably free of gravel deposits.

The pebbles, cobbles, and boulders in the glaciofluvial gravel characteristically are well rounded to subrounded, with many percussion marks. Weathering rinds on the basalt particles are very thin or absent. The deposits are well sorted, have an open-work texture, and consist almost entirely of basalt, with less than 1 percent granitic and metamorphic material. In contrast, glaciofluvial gravel near the levels of the Columbia and Snake Rivers is as much as 60 percent nonbasaltic (Roald Fryxell, oral communication, 1963). The gravel deposits near these rivers were included with the alluvial deposits on the geologic map of Grolier and Bingham (1971) because of their similar lithology, and because the exact termination of the last melt-water flow in these rivers is indistinguishable.

Locally, caliche and Ringold pebbles and blocks occur in the glaciofluvial gravels. Gravel containing these materials exists marginal to scabland tracts, where development of the scabland consisted of little more than channeling of caliche. Typical examples are the deposits of caliche gravel that are spread over Paradise Flat north and south of Othello and the mounds of rounded caliche pebbles in the southwestern part of the Quincy Basin in the NE$\frac{1}{4}$ sec. 8, T. 18 N., R. 23 E.

A thin coating of calcium carbonate occurs on the underside of fluvial gravel particles to a depth of several feet below land surface in most places, though the amount is generally insufficient to cement the gravel.

The deposits of fluvial gravel and sand can be divided into four general geomorphic shapes: deltas; bars, natural levees, and terraces; valley fill; and discontinuous gravel masses in abandoned plunge pools.

The deltas extend outward like fans from the mouths of scabland channels that open into structural basins or onto prescabland piedmont slopes. An example of this type of deposit is the Quincy Basin fill, which is shaped like a cone with its apex near the mouth of Lower Grand Coulee. The deltaic fill exhibits a striking southward decrease in grain size and change in stratification. Near the mouth of Lower Grand Coulee in T. 21 N., Rs. 26 and 27 E., the delta consists of torrentially bedded, very coarse basaltic boulder gravel, which is well exposed in a gravel pit in Ephrata in the NW$\frac{1}{4}$ sec. 22, T. 21 N., R. 26 E. In the central part of the basin, the topset beds of the delta comprise well-stratified sand and granule gravel. This phase is well exposed in a roadcut on Interstate 90, immediately west of the east side of Morrison Flat, in the SE$\frac{1}{4}$ sec. 30, T. 19 N., R. 27 E. In the southern part of the Quincy Basin, between Morrison Flat and the Frenchman Hills, most of the bottomset beds of the delta consist of fine sand and silt. Some of this sand subsequently migrated eastward, as dunes blown by the prevailing westerly wind, and dammed Crab Creek and Rocky Ford Creek, forming Moses Lake. The maximum known thickness of the deltaic deposits in the Quincy Basin is 143 feet (well 19/26-11J1), about 11 miles west of the city of Moses Lake.

Deposits of gravel, similar to but smaller in area than those of the Quincy Basin, flare out south of the Drumheller and Othello Channels; they have been partly removed by subsequent flood erosion. Also, the huge gravel and sand deposit trending southeast from Sentinel Gap and underlying most of T. 14 N., Rs. 23 and 24 E., is part of a delta deposited by the proglacial Columbia River as it entered Lake Lewis. At Mattawa, the delta is 225 feet thick.

The gravel deposits with giant subfluvial stream forms—the bars, natural levees, and terraces—are common along all channels used by the floodwaters. The deposits contain basaltic material of all grain sizes but are predominantly gravel. Bars are well developed in Crab Creek between Wilson Creek and Brook Lake, where they have been described by Bretz and others (1956, p. 977-980, pl. 8). The natural levees are best exemplified by the long curving ridge that passes through the east side of the city of Moses Lake and extends southward along the Potholes Reservoir. Terraces are formed along the sides of channels that contained perennial streams, or they may be remnants of older deposits at places where they were reworked as a result of lowering base levels. Deposits along the north side of Crab Creek valley northeast of Adco and along Rocky Ford channel are examples of the two terrace types. Similar deposits also occur in the Columbia and Snake River valleys. Two outstanding examples are the gravel terraces and bars underlying parts of T. 9 N., Rs. 29 and 30 E., and T. 10 N.,

R. 29 E.; and the terrace known as Snake River Flat northeast of Ice Harbor Dam.

The glaciofluvial material remaining in the scabland valleys is largely gravel, which can be considered the bedload of the glaciofluvial floods. Deposits of this type occur in Dry Coulee, in the valley of Crab Creek east of Adrian, in Washtucna and Esquatzel Coulees, and along the channel of Lower Crab Creek between Corfu and Jericho. In Crab Creek valley, the thickness of the gravel averages about 100 feet, whereas it is 180 feet thick at Connell (Calkins, 1905, p. 43), about 120 feet thick near Mesa, and 160 feet thick at Eltopia.

Discontinuous gravel deposits occur around abandoned plunge pools, as in the small tributary of Esquatzel Coulee in sec. 18, T. 13 N., R. 31 E., on the bottom of the Potholes and Frenchman Springs Coulees, and in most of the small plunge pools in the Drumheller and Othello Channels. Other discontinuous gravel masses occur on basalt benches (for example, the gravel underlying a terrace in the SE$\frac{1}{4}$ sec. 17, T. 17 N., R. 23 E., 1 mile north of the Vantage Bridge) and in the cliffed recesses of coulees (such as in Washtucna Coulee at Sulphur Lake and in Esquatzel Coulee a mile north of Mesa).

HOLOCENE DEPOSITS

Eolian Deposits

In the project area, eolian materials exposed at the surface consist of young loess and dune sand deposited during the Holocene Epoch. Like the older loess deposited throughout the Pleistocene Epoch, it is composed of silt-sized material that settled to the land surface after being carried aloft and widely distributed by the wind. It includes some sand and clay-size particles intermixed, derived largely from beds of the Pleistocene lacustrine silt. The dune sand, on the other hand, is blown along the ground surface, largely by creep and saltation (bouncing).

Loess

In any particular area, the deposit is thickest on the leeward side of hills, such as the north-facing slopes of Frenchman Hills and Saddle Mountains. The loess is easily distinguished from the undisturbed lacustrine silt at higher altitudes by the lack of stratification and by the development and depth of the soil profile, even though both deposits are of similar light gray-brown color. Where it overlies the older loessal deposits, such as the brown tuffaceous sand of the Ringold Formation, the Holocene loess generally is separated from it by a layer of plateau-blanket caliche.

As shown on the geologic map of Grolier and Bingham (1971), loess of Holocene age extends onto Babcock and Pinto Ridges, Frenchman Hills, Saddle Mountains, and the uplands east of the Quincy and Pasco Basins (the loess has been mapped only where it is more than about 5 feet thick). In the project area, the areal extent of the Holocene loess corresponds approximately to that of the Shano and Ritzville soil series (Gilkeson, 1958; 1965, p. 3-5).

Regularly spaced surficial mounds of Holocene silt with sporadic pebbles are well developed in the project area. They occur in two physiographic environments: on gravel and stripped basalt surfaces in channels formed by early glaciofluvial erosion; and on the wind-swept slopes of the Beezley Hills, Pinto Ridge, and other high slopes, where the silt is less than 10 feet thick. The mounds average about 25 feet in diameter, 3 feet in height, and are spaced 10 to 50 feet apart.

The silt has been removed from the intermound areas, probably by deflation and soil creep. On sloping surfaces, the intermound areas approach a desert-pavement type of surface, whereas on flat surfaces, they are covered by a thin layer of silt. The intermound areas may exhibit a poorly developed polygonal stone net, bare of silt, surrounding the mounds. These areas of patterned ground are probably analogous to those described by Malde (1964, p. 202, pls. 1 and 4) on the Snake River plain and by Brunnschweiler (1964, p. 227, fig. 5) and Kaatz (1959) on the Columbia Plateau.

A layer of white volcanic ash, from a few inches to a foot thick, occurs in the upper part of the Holocene loess, as it does in all other undisturbed Holocene deposits in the area. The layer is an important marker, and can be observed in many roadcuts. The widespread ash bed has been interpreted as originating from Mount Mazama (Crater Lake) and its age is approximately 6600 years (Powers and Wilcox, 1964). Locally, an older ash layer is bedded in lower parts of the loess. The older ash, from Glacier Peak, has been identified in the northern part of the project area (Quincy Basin and Crab Lake depression in secs. 7, 8, and 9, T. 21 N., R. 30 E.). Its deposition is now thought to have occurred about 12,000 years ago (Fryxell, 1965); hence it dates this part of the loess as of late Pleistocene age.

Dune Sand

In the project area active sand dunes occur in the central and southern Quincy Basin, in the areas northeast of Pasco, in the Lower Crab Creek valley, and along the Columbia River.

In the Quincy Basin, the dunes that constitute the natural dam impounding the water of Moses Lake contain a mixture of basaltic, opal, and quartz-feldspar sand, which has drifted from the lacustrine fine sand and silt underlying the southern and southwestern parts of the Quincy Basin. In the area north and northeast of Pasco, dunes consist of sand reworked and blown from a thin veneer of lacustrine sand and silt. The barchan (crescent shaped) dunes along the Columbia River and in the Lower Crab Creek valley east of Beverly consist of sand derived from the floodplain deposits of the Columbia River. The geologic map of Grolier and Bingham (1971) shows only the larger areas of active dunes, and does not show the innumerable established dunes in the southern parts of the Quincy and Pasco Basins, nor the many dune islands in Potholes Reservoir. There, the sharp crests of the dunes rising above the water surface are being slowly leveled by the wind which blows the dry sand off the saturated bases.

The parabolic and transverse dunes in the southern parts of the Quincy and Pasco Basins have left "tails," longitudinal dunes many miles long. These tails (long low ridges of sand) are now largely stabilized; they extend from the source area of the sand and show the direction of past dune migration.

A notable exception to the dunes of sand are the dunes of silt at the top of the Ringold scarp, which have been blown upward from the White Bluffs in the area from sec. 11, T. 13 N., R. 27 E., to sec. 33, T. 13 N., R. 28 E. Only those dunes near the mouth of Ringold Coulee, in secs. 13 and 14, T. 12 N., R. 28 E., are shown on the geologic map of Grolier and Bingham (1971).

Alluvium

Holocene alluvium consists of channel-lag gravel and channel-fill material of silt to gravel size. The channel-lag gravel on the narrow flood plains of the Columbia and Snake Rivers is only slightly modified glaciofluvial gravel because the capacity of either river during Holocene times, even during peak floods, does not approach that of their ancestors in the late Pleistocene. Channel-fill material of sand and silt is exposed in discontinuous low terraces along the course of Crab Creek and Lower Crab Creek and in most coulees (Daugherty, 1956, p. 234-235, fig. 8).

In the small channels of intermittent streams that flow in coulees at times of flash flood or during spring snowmelt, most of the alluvium is tan or gray silt and sand, reworked from earlier deposits, mostly from the lacustrine silt of Pleistocene age or from loess. Some of the alluvial silt is deposited when the water is temporarily ponded behind gravel bars which form natural dams or obstructions in the coulees. Many of these deposits contain Holocene fresh-water gastropods and pelecypods.

A well-formed alluvial fan made up of reworked clay and sand of the Ringold Formation has been deposited at the mouth of an unnamed coulee northeast of the Othello Channels, mostly in secs. 3, 4, 9, and 10, T. 14 N., R. 30 E. The coulee carries the discharge of several tributary coulees heading in Paradise Flats. Prior to irrigated farming, discharge from this coulee flowed 4 miles westward through dunes and blowouts to the depression now occupied by Eagle Lake.

Coarser basaltic alluvium, some of which is subangular, also is transported at times of floods by intermittent streams; this alluvium is either reworked from bars of Pleistocene gravels, or eroded from the coulee bed. For example, the flood of 1956 in Providence Coulee carried runoff and alluvium through Connell, and down the entire length of Esquatzel Coulee to the Pasco sump north of the Pasco airport (Anderson and Bodhaine, 1956, p. 2).

One of the most perfectly shaped alluvial fans in the project area is that upon which the city of Ephrata is built. The fan consists of semiangular basaltic debris brought down from the Beezley Hills by the intermittent creek in Two Springs Canyon. The debris has accumulated within an abandoned meltwater channel. Spring snowmelt or storm runoff from Two Springs Canyon no longer is free to flow across the fan; it is now channeled off to the north where the silt and sand fraction of its load accumulates in an elongate closed depression nearly 2 miles long north of the city.

White marl occurs in the Soap Lake depression. Isolated outcrops of it are also found in the abandoned plunge pools of Potholes and Crater Coulees.

Peat beds about 20 feet thick lie in a depression formerly occupied by Crab Lake in the Crab Creek valley. There the streamflow was impounded by alluvium from Wilson Creek, a tributary from the north beyond the project area. The lake has been subsequently drained by ditching for cultivation. A small part of the depression lies within the extreme northeastern part of the project area, in sec. 18, T. 22 N., R. 30 E. Also, peat and muck lie in a low area near the north end of Moses Lake at the mouth of Rocky Ford Creek (Rigg, 1958, p. 52-53), and in a depression just south of the divide between Soap Lake and its post-Pleistocene outlet toward Rocky Ford Creek, in secs. 5, 6, and 8, T. 21 N., R. 27 E.

Colluvium

Subangular basaltic debris, carried down from the slopes of the Beezley Hills and the Saddle Mountains, accumulates near the heads of the alluvial fans. In one part of the Wahluke Slope, basaltic debris is building up at the base of a small dissected scarplet that marks one of the levels of ancient Lake Lewis, at an altitude of about 850 feet across T. 15 N., R. 25 E. Along the north escarpment of the Saddle Mountains, east of Beverly, giant debris cones of angular basalt rise from the base of the mountains and extend to the mouths of truncated gullies 1,000 feet or more above; at their upper ends they merge laterally with talus.

Talus accumulates at the base of most basalt cliffs in the coulees. It is prominent in the Columbia and Snake River canyons and in the scabland channels, but it is best developed along the fault scarps on the north face of the Saddle Mountains, and in Dry Coulee, in the SE$\frac{1}{4}$ sec. 22, T. 23 N., R. 27 E.

FIGURE 30.—Aerial view looking southeast, showing multiple landslides and debris flows on north slope of Saddle Mountains between Taunton and Corfu. Flood plain of Lower Crab Creek in middleground.

Landslide Debris

Landslides have affected both the Yakima Basalt and the Ringold Formation. Parts of the large landslide west of Taunton have slumped at different times, so that several ages of sliding are apparent. Several of the most recent rock flowages have built debris flows that extend as much as 1,200 feet into Lower Crab Creek valley. These debris flows grade upslope into typical backward rotated landslide blocks (fig. 30). Most landslides in the basalt include interbedded sedimentary materials that probably served as weak planes on which sliding started. Fragments and blocks of all these materials can be found in the landslide debris.

Along the White Bluffs, many landslides have been caused by undercutting of the weak Ringold Formation by the Columbia River. There, slices of the bluffs more than a mile long have moved toward the river with a backward rotation motion.

STRUCTURE

Geologic structures in the project area are only parts of the larger features that characterize the related structural evolution of the entire Columbia Plateau and of the Cascade Range provinces. The events that produced these structural features—subsidence, folding, and faulting—occurred on a regional scale.

REGIONAL SUBSIDENCE

The evidence for broad regional subsidence in the project area during extrusion of the basalt is based on the observations that (1) the oldest flows of the Yakima Basalt are exposed farthest away from the Pasco Basin, and (2) there is a primary offlap of many younger flows relative to the older ones. Although the foregoing conditions might also be explained as being due to a decrease in quantities of lava during the later extrusions, the writers concur with the belief that either continual or periodic subsidence caused these characteristics of the basalt flows.

The direction of movement of most lava in the project area was apparently northward or northeastward, based on the direction of foreset bedding in pillow zones, the direction of bending of pipe vesicles, and several other criteria (Waters, 1960, p. 350-366). The regional thinning of the flows in a northerly direction in the project area is an additional clue that the flows spread to the north. All of these features imply that the area of extrusion was south or southeast of the project area, and that it may have roughly coincided with the center of the structural basin. Features and criteria pertaining to regional subsidence are described in more detail below.

FOLDING AND FAULTING

In the project area the Yakima Basalt is complexly folded and faulted. The structures are parts of tectonic deformations of the entire region; consequently, many of the main anticlines, synclines, and monoclines extend beyond the project boundaries, especially westward. Three fold and fault systems are present within the study area. The most prominent is a system that trends almost due east; upon this are imposed a north-northeast-trending system, and a more subdued northwest-trending system.

East-Trending Folds and Faults

The most prominent folds and faults occur in the west-central part of the area and trend easterly. Two long anticlines are expressed topographically by the Saddle Mountains and the Frenchman Hills. These two uplifts are separated by an asymmetric syncline which is referred to herein as the Othello Basin. South of, and parallel to, the Saddle Mountains, another deep east-trending asymmetric syncline (Wahluke syncline) is indicated by data from several deep wells. North of the Frenchman Hills lies the saucer-shaped Quincy Basin, which in turn is bounded on the north by the Beezley Hills monocline, the Pinto Ridge anticline, and the Soap Lake anticline (W. E. Walcott and G. E. Neff, unpublished U.S. Bureau of Reclamation administrative report, 1950).

Saddle Mountains Anticline

The elongate and asymmetrical Saddle Mountains anticline extends 70 miles eastward from near Ellensburg to the Othello Channels; the 45-mile-long segment east of Sentinel Gap lies within the project area and is the most intensely faulted in the area. The degree of deformation increases westward along the axis. For descriptive purposes, the ridge within the project area is subdivided into an eastern part that extends from the Othello Channels to the west side of the Corfu landslide, a central part that extends westward to the west end of the Smyrna Bench, and a western part that extends thence westward to Sentinel Gap.

The overall structural geology of the Saddle Mountains is that of a broad (1 to 4 miles wide) asymmetrical anticline. In its eastern part, the structure has a steeper northern limb (fig. 31A), and locally in the central part it has a steeply dipping south limb and an overturned north limb. The western part of the Saddle Mountains (fig. 31C) consists of both an asymmetrical anticline and northward, a syncline, the north limb of which terminates in an 1,800-foot-

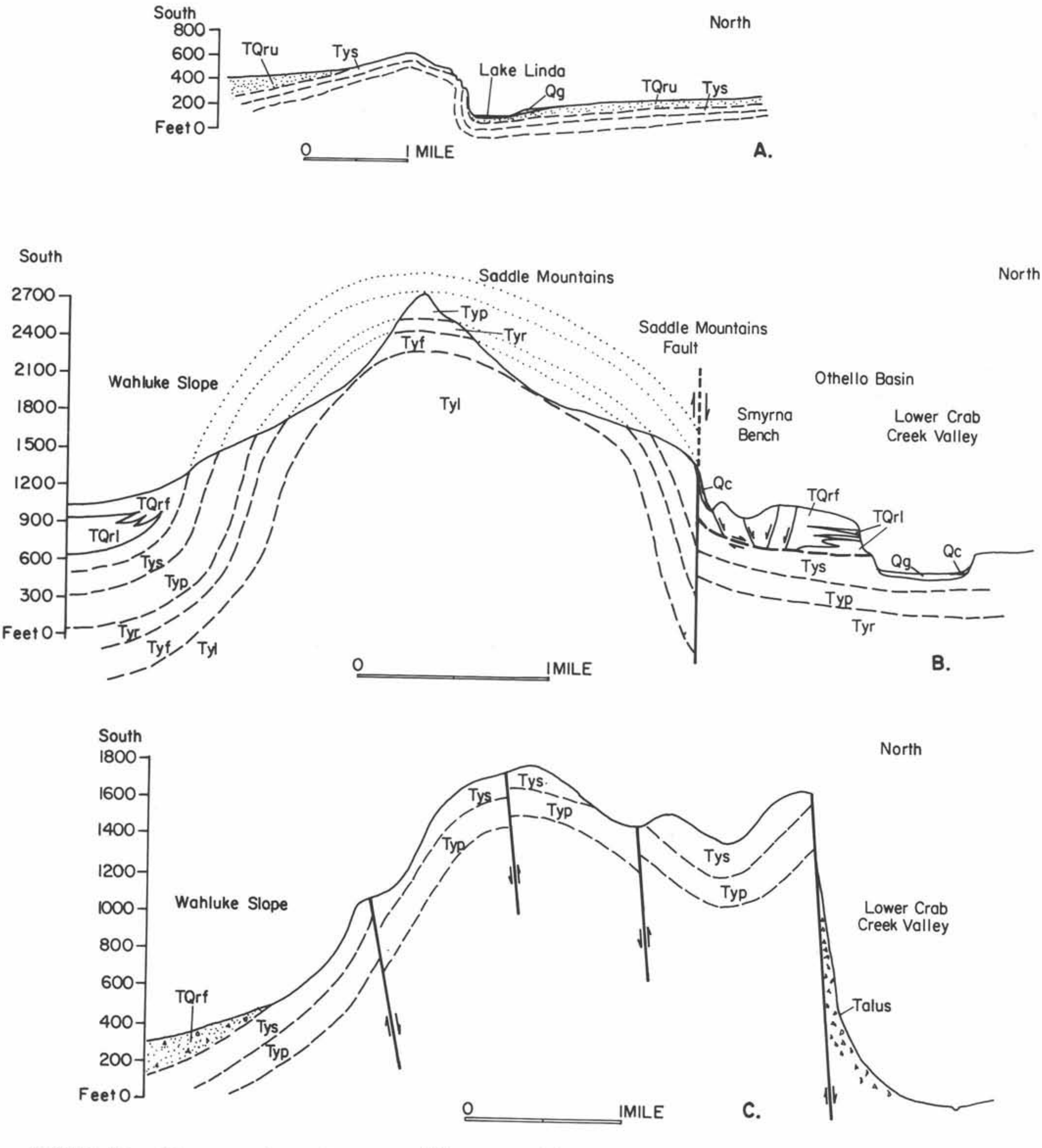


FIGURE 31.—Diagrammatic sections across (A) eastern, (B) central, and (C) western part of the Saddle Mountains, showing generalized structural configurations. Tyl, lower basalt, undifferentiated; Tyf, Frenchman Springs Member; Tyr, Roza Member; Typ, Priest Rapids Member; Tys, Saddle Mountains Member; TQrf, Ringold fanglomerate; TQrl, Ringold lacustrine clay and silt; TQru, Ringold Formation, undifferentiated; Qg, fluvial gravel; and Qc, colluvium.

high fault-line escarpment east of Beverly.

The gently undulating crest of the anticline in the eastern part declines eastward from an altitude of 1,800 feet at the Grant-Adams County line to about 1,150 feet, where it plunges eastward into the floor of the Othello Channels. About 2 miles west of Eagle Lake, a subordinate monocline separates from the main anticlinal axis. The monocline is a sharp flexure on the north limb of the anticline that causes the basalt to stand vertically (fig. 31A) near Linda Lake in the SE$\frac{1}{4}$ sec. 30, T. 15 N., R. 29 E. The monocline cannot be traced westward from the east end of the Taunton Terrace because of a cover of Ringold clay and loess.

In the central part of the Saddle Mountains, the anticlinal arch is 2 to 2$\frac{1}{2}$ miles wide. The south limb there dips as much as 65°, while the north limb is also steeply dipping but partly faulted (fig. 31B) and partly overturned. Prior to landsliding and faulting west of the Taunton Terrace, this north limb probably was continuous with that described at the very eastern end of the range.

In a zone just south of the Smyrna Bench, the west-trending flexure is represented by the steeply dipping flows in the faceted spurs along the fault-line scarp. The attitude of these flows contrasts with the flat-lying flows exposed at the north side of the bench and at the crest of the mountain.

The fault along the north limb of the anticline is a nearly vertical shear zone that parallels the eastward trend of the Saddle Mountains (Grolier and Bingham, 1971). The fault zone extends westward from the bench near Corfu along the fault-line scarps through a gully in W$\frac{1}{2}$SW$\frac{1}{4}$ sec. 34, T. 16 N., R. 25 E., at the west end of the Smyrna Bench (Grolier and Bingham, 1971). From there it continues westward, under the talus of the north-face fault-line scarp of the western part and across the Columbia River just north of Sentinel Gap. The vertical displacement along this fault reaches several hundred feet in secs. 3, 4, and 5, T. 15 N., R. 26 E., where the spur ridges of the upland are sharply truncated and the small valleys above the fault are "hanging." Along this fault, the Saddle Mountains have been uplifted sufficiently to permit shallow slumping from the spurs, along the steeply dipping flows, onto the downdropped Smyrna Bench in secs. 1, 2, 3, 4, and 5, T. 15 N., R. 26 E., along the north limb of the anticline.

The multiple scarps which bound the Smyrna Bench on the south side are from one-half to less than a quarter of a mile north of the main fault. The vertical displacement along these scarps is opposite to that of the main fault, so that a narrow graben (fig. 31B) separates the Smyrna Bench from the fault-line scarp along the north slope of the Saddle Mountains.

At the extreme west end of the Smyrna Bench, the Yakima Basalt and the Vantage Sandstone Member are overturned and dip about 60° southward (Grolier and Bingham, 1971) on the north side of the main fault zone. Immediately south of the fault zone there, the Vantage Sandstone is vertical and unbroken. Farther south, (uphill) the lower basalt flows are overturned. The overturned flows north of the fault zone are considered to be slump blocks, which came from the overturned fold to the south and which have slid about 500 feet over the sedimentary rocks of the Ringold Formation underlying the Smyrna Bench (Grolier and Bingham, 1971, and Bingham and others, 1970).

The western part of the Saddle Mountains consists of the main anticlinal arch and about 1$\frac{1}{2}$ miles to the north, a subordinate syncline (fig. 31C). The anticline is intensely fractured by high-angle faults that trend nearly eastward along the summit and along the limbs. The south limb, especially throughout T. 15 N., R. 24 E., is cut into blocks by several small faults, and is further complicated by slumping. A transverse fault trending approximately northward cuts through the anticline in secs. 15 and 22, T. 15 N., R. 24 E. (Grolier and Bingham, 1971).

The subordinate syncline trends eastward and is subdivided by northward- and northeastward-trending cross folds into three small synclinal basins. These three small structural basins are underlain by sedimentary rocks of the Beverly Member of the Ellensburg Formation, Saddle Mountains Member of the Yakima Basalt, fanglomerate assigned to the Ringold Formation, and loess.

A steep escarpment, rising as much as 1,800 feet above the level of Lower Crab Creek at Sentinel Mountain, truncates the north limb of the east-trending syncline from Sentinel Mountain to the west end of the Smyrna Bench. The stratigraphic displacement along the escarpment at Sentinel Mountain exceeds 1,000 feet. The escarpment was interpreted by Russell (1893, p. 96) as due to a fault and by Calkins (1905, p. 37-38, 42) as due to the erosional removal of one limb of a sharp flexure. Laval (1956, p. 126-128) stated that, at least on the west side of the Columbia River near Sentinel Gap, the lower flows have been thrust northward over the Priest Rapids Member. However, the critical relationships are hidden under talus and sand dunes, and the scarp is interpreted as a near-vertical fault by Grolier and Bingham (1971).

The top of the escarpment is marked by an erosional surface that truncates the basalt and sedimentary rocks folded in the synclines and the subdividing anticlinal flexures (fig. 16). The erosional surfaces are underlain in places by a layer of massive caliche or elsewhere by a 10- to 20-foot thickness of fanglomerate and composed of angular basalt fragments cemented with caliche, which caps the truncated basalt and overlying sedimentary rocks. The caliche cap also is deformed and folded by the north-trending flexures (fig. 17).

Away from Sentinel Gap the Saddle Mountains Member underlies the crest of the anticline and most of the east-trending syncline to the north, in secs. 2, 3, and 4, T. 15 N., R. 24 E. Near Sentinel Gap many basalt flows have been removed by erosion from both limbs of the anticline. The Roza Member also crops out on the north limb, where it is exposed in small gullies cutting through the cover of loess in the E$\frac{1}{2}$SE$\frac{1}{4}$ sec. 1, and the N$\frac{1}{2}$NE$\frac{1}{4}$ sec. 12, T. 15 N., R. 24 E. The south limb of the Saddle Mountains anticline is deeply dissected by gully erosion, and the Roza Member also crops out in the SE$\frac{1}{4}$ sec. 8 and the NW$\frac{1}{4}$ sec. 16, T. 15 N., R. 25 E. For 2$\frac{1}{2}$ miles east of Sentinel Gap, the Saddle Mountains Member and the underlying Beverly Member of the Ellensburg Formation are stripped from the crest of the anticlinal arch.

The rest of the Saddle Mountains to the east are unusual in that the uppermost members of the Yakima Basalt underlie the crest of the mountain and have been partly removed from its flanks by erosion. This structural and erosional anomaly is due not only to the broad arch and the steep limbs of the Saddle Mountains anticline, but to cross folds of low amplitude and long wavelength that produce undulations along the crest. The undulations partly control the erosion pattern, and the basalt that is stratigraphically the highest occurs in shallow synclines or sags along the crest.

Frenchman Hills Anticline

Within the project area, the Frenchman Hills anticline extends from the Columbia River to the northern part of the Drumheller Channels. The ridge is slightly arcuate southward and is formed by a slightly asymmetric anticline that is characterized by a short, steep north limb, where the flows dip as much as 60° northward toward the Quincy Basin, and by a long gently sloping south limb that represents the north flank of the Othello synclinal basin.

The crest of the anticline is gently undulating. The crest is offset about 1$\frac{1}{2}$ miles in the SW$\frac{1}{4}$ T. 17 N., R. 26 E., as shown on the geologic map of Grolier and Bingham (1971); the nature of the structural complication responsible for this offset is hidden beneath sedimentary cover.

At the western end of the Frenchman Hills within the project area, 31 miles west of the Drumheller Channels, the north limb of the Frenchman Hills anticline apparently has been thrust a few hundred feet northward in the SE¼ sec. 32, T. 18 N., R. 23 E. (geologic map of Grolier and Bingham, 1971). The thrust plane is exposed in a roadcut along the eastbound lane of Interstate 90, and shows the Frenchman Springs Member thrust over the brown tuffaceous sand of the Ringold Formation.

The Frenchman Hills anticline continues eastward into the eastern part of the Drumheller Channels. In the Drumheller Channels, the anticline is succeeded northward by subordinate shallow synclines and anticlines, the northernmost of which is an anticline called the Lind Coulee flexure by Bureau of Reclamation geologists (F. O. Jones, unpublished U.S. Bureau of Reclamation administrative report, 1945). The Lind Coulee flexure was not traced east of the Drumheller Channels, but it probably continues eastward, and merges with another fold, thus accounting for the high ridge immediately south of Lind Coulee east of Warden.

Lind Coulee Fault

The Lind Coulee fault is one-quarter to three-quarters of a mile north of the Lind Coulee flexure in secs. 3, 4, 5, and 6, T. 17 N., R. 29 E. (geologic map of Grolier and Bingham, 1971). This low-angle thrust fault crops out east of, and lies under, the Potholes Reservoir. The thrust plane is poorly exposed except along the south bank of Lind Coulee in the SW¼ sec. 5, T. 17 N., R. 29 E. At this locality, the Roza basalt is intensely sheared and thrust over brown tuffaceous sand of the Ringold Formation; the thrust plane there dips about 10° southward.

Pinto Ridge Anticline

Two asymmetric anticlines lie in the extreme northern part of the project area. One of these flexures, named the Pinto Ridge anticline by Bretz (1932, p. 3) extends eastward past Long Lake beyond the project boundary.

The Pinto Ridge anticline is faulted along parts of its north limb. The Pinto Fault (fig. 32) partly truncates the north limb on the east, just inside

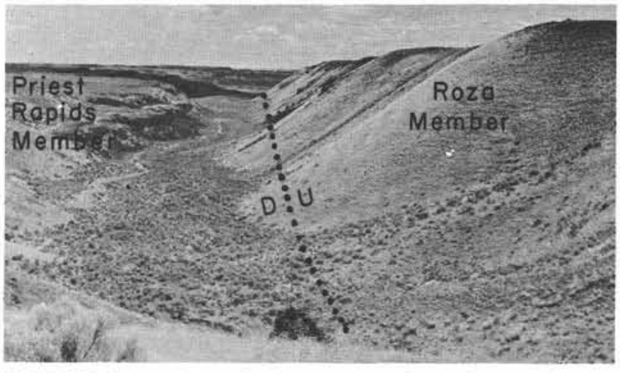


FIGURE 32.—View looking east along Spring Coulee, showing scarp of the Pinto Fault. (U.S. Bureau of Reclamation photo, 1949.)

the project boundary, and the Dry Coulee Fault (fig. 33) partly truncates it at its western end, as shown on the geologic map of Grolier and Bingham (1971). The upthrown block is on the south side of both faults.

Soap Lake Anticline

Also in the extreme northern part of the project area is the Soap Lake anticline, whose axial trend is normal to Lower Grand Coulee at Soap Lake. This anticline extends 6 miles eastward from the Beezley Hills monocline to Dry Coulee. The anticline is asymmetric with a steeper south limb.

Minor Sags and Anticlinal Flexures

East-trending structural sags and associated small anticlinal flexures are best displayed in the cliffs above the Columbia River gorge, west of the Quincy Basin. Prescabland and scabland stream channels in that area were formed subsequent to this folding; the locations of Potholes Coulee and Frenchman Springs

FIGURE 33.—View looking southwest across Dry Coulee, showing trace of the Dry Coulee Fault. (U.S. Bureau of Reclamation photo, 1949.)

Coulee are controlled by east-trending sags in the underlying basalt. At least one additional sag lies between these two coulees, which explains the outlier of Roza basalt on a bench otherwise underlain by the Frenchman Springs Member, in the NW¼ sec. 31, T. 19 N., R. 23 E. The highest parts of Babcock and Evergreen Ridges coincide with the east-trending anticlinal flexures between the sags; the east wall of the Columbia River gorge is eroded convexly eastward across sags, and convexly westward across anticlinal flexures.

In the eastern part of the project area, the course of many coulees now deeply entrenched into basalt possibly was controlled by gentle eastward-trending folds in basalt. Some of these coulees are, from north to south, Crab Creek valley, the lower part of Rocky, Bowers, Lind, Washtucna, Old Maid, and Dunnigan Coulees.

Northeast-Trending Folds

Folds that trend northeastward in the project area include the Lynch Coulee anticline and syncline, Beezley Hills monocline, an unnamed monocline across the Lower Grand Coulee, and minor sags and intermediate anticlinal flexures in the eastern part of the area.

Lynch Coulee Anticline and Syncline

The Lynch Coulee anticline and syncline, northwest of the Quincy Basin, are asymmetric, tight folds that trend approximately N. 30° E. They are in alignment with the upper and lower bends of a monocline west of the Columbia River (Don Walcott, written communication, 1956).

The southeastern limb of the anticline dips gently into the Quincy Basin, whereas the northwestern limb dips up to 60° to the adjacent syncline. The syncline, asymmetric in the opposite direction, has a gentle northwestern limb which grades into the long slope that extends southeastward from the south wall of Moses Coulee (outside the project area). The plunge of the syncline is southwest within the project area.

Beezley Hills Monocline

The Beezley Hills monocline, along the southeast- and south-facing slope of the Beezley Hills, extends northeastward and becomes the Coulee monocline north of the mapped area. Approximately 2½ miles southwest of Ephrata, the monocline bends abruptly westward and continues in that direction just outside the project area.

The Coulee monocline, first recognized by Calkins (1905, p. 38), illustrated by Bretz (1932, p. 60-68), and mapped by W. E. Walcott and G. E. Neff (unpublished U.S. Bureau of Reclamation administrative report, 1950) lies mostly outside the project area. It is a continuation of the Beezley Hills monocline starting northwest of Little Soap Lake, but it has been almost completely obliterated by the up-coulee migration of recessional waterfalls of the Dry Falls cataract. The islands in Lake Lenore and northward to Blue Lake (out of the project area) are the remnants of the dipping flows. On those islands the basalt flows dip about 50° eastward, in sharp contrast with the near-horizontal altitude of the flows exposed along both walls of Lower Grand Coulee.

The Beezley Hills-Coulee monocline is a principal tectonic structure of the northern part of the Columbia Plateau. The basalt in it is several hundred feet below its general position on the high plateaus to the north. The monocline is deeply dissected by gully erosion. Erosional removal of basalt along the slope of the monocline amounts to several hundred feet.

Unnamed Monocline Across Lower Grand Coulee

An unnamed monocline trends northeast from the site of its merger with the Beezley Hills-Coulee monocline about half a mile south of Little Soap Lake. The monocline passes into the high-angle Dry Coulee Fault about $5\frac{1}{2}$ miles northeast of the Coulee monocline (fig. 33). The Dry Coulee Fault in turn continues for another 4 miles into the south limb of the Bacon syncline. The Bacon syncline, recognized and named by Bretz (1932, p. 3, fig. 1), also trends northeast at the northern border of the project area, in T. 23 N., R. 28 E., and continues northeast for many miles along the south side of the Coulee monocline.

The unnamed monocline in effect lowers the altitude of the Yakima Basalt several hundred feet from the High Hill anticline, the next structural feature to the north. The flexure is sharpest at its western end, in the east wall of Lower Grand Coulee (W. E. Walcott and G. E. Neff, unpublished U.S. Bureau of Reclamation administrative report, 1950). There, erosional truncation of the Frenchman Springs Member and many flows of the lower basalt reveals flows steeply dipping to the southeast.

Minor Sags and Intermediate Anticlinal Flexures

Northeast-trending sags in the eastern part of the project area can be recognized in the field where loess is absent as in the scabland channels. The best examples of these are in a scabland tract in T. 22 N., R. 29 E., and also in parts of Black Rock Coulee. The structural sags provided the lower divides through which the Lake Missoula floodwater overflowed out of the Crab Creek valley. The anticlinal flexures were not crossed by the floodwater, and are still overlain by brown tuffaceous sand assigned to the Ringold Formation. Farther southeast, the parallel southwestward trend of many coulee segments suggests that a part of the drainage was controlled by these minor sags.

Northwestward-Trending Folds

The northwest-trending fold system is the least pronounced of the three systems observed in the area. These structures with northwest trends were observed in the project area, and are described as follows:

1. In the northwestern corner of the Quincy Basin, Crater Coulee trends northwest, normal to the synclinal valley occupied by Lynch Coulee. Apparently, Crater Coulee was formed along a structural sag in the basalt.

2. About $2\frac{1}{2}$ miles north of Beverly, the basalt flows dip 3° to 4° SW. in the bluff. At the

south end of the outcrop, the basalt, at the northern edge of the gravel cover, dips to the northeast. This sag in the basalt may be a continuation of the northwestward-trending Rye Grass syncline on the west side of the Columbia River.

3. A low northwesterly-trending anticlinal hump may be present in the basalt exposed south of Ephrata, largely in secs. 3, 4, and 10, T. 20 N., R. 26 E.

4. The main axis of the Pasco syncline trends northwest, based on the altitude of the basalt surface reported in well logs. This trend also is reflected in the structure contours on the Ringold conglomerate (fig. 21), indicating that the conglomerate also was involved in the regional deformation.

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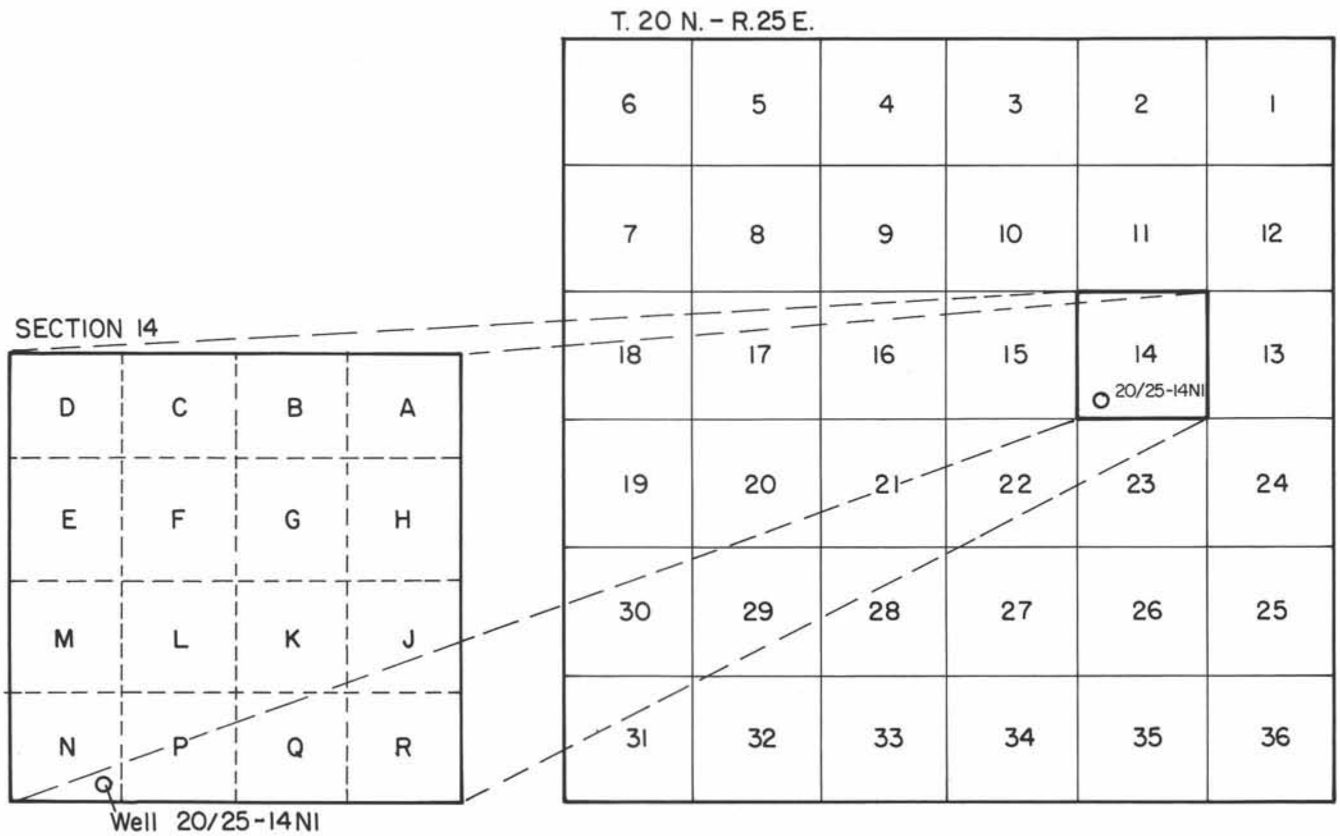
APPENDIX I

WELL-NUMBERING SYSTEM

Wells and test holes used in this report are designated by Geological Survey numbers that indicate their location according to the rectangular system for subdivision of public land. The numbering system is illustrated below, using well 20/25-14N1 as an example. For that well, the numerals preceding the hyphen indicate the township and range north and east of the Willamette meridian and base line, respectively (T. 20 N., R. 25 E.). The first numeral after the hyphen indicates the square-mile section in which the well is located (sec. 14), and the capital letter indicates the 40-acre tract within that section. Within each 40-acre tract, the wells are numbered serially according to the order of well inventory.



APPENDIX II

SELECTED DRILLERS' LOGS OF WELLS, WITH STRATIGRAPHIC INTERPRETATIONS

	Thickness (ft)	Depth (ft)
9/30-18B1. U.S. Government Naval Air Station. Altitude, 416 ft. Drilled by Durand & Son, 1943. Casing, 20-in. to 1,045 ft.		
Pleistocene:		
Fluvial and lacustrine deposits:		
Fluvial gravel:		
Gravel, fine	23	23
Sand, fine	8	31
Sand, fine with gravel	22	53
Sand, fine	17	70
Sand, medium	5	75
Ringold Formation:		
Conglomerate:		
Gravel	5	80
Sand and gravel, cemented	11	91
Sand, gravel and boulders	25	116
Sand and gravel, loose	4	120
Sand and gravel, partly cemented	6	126
Lacustrine clay:		
Shale, sandy, gray	26	152
Clay, yellow	8	160
Shale, gray	15	175
Shale, blue	7	182
Shale, hard	3	185
Pliocene-Miocene:		
Ellensburg Formation and Yakima Basalt; undifferentiated:		
Basalt, black	15	200
Basalt, gray	38	238
Basalt, black	42	280
Basalt, broken, with gray shale	21	301
Basalt, creviced	33	334
Basalt, gray	77	411
Shale, green	19	430
Basalt, gray	126	556
Rock, rotten, and blue shale	13	569
Rock "shell"	4	573
Shale and rock	2	575
Shale, green	5	580
Basalt, broken, black, with shale seams	30	610
Basalt, gray	30	640
Basalt, black	24	664
Basalt, gray	38	702
Shale, dark	13	715
Shale, with shells	5	720
Basalt, broken, gray	20	740
Basalt, dark, gray	5	745
Basalt, gray	60	805
Basalt, broken, sandy	5	810

Continued

APPENDIX II—Continued

	Thickness (ft)	Depth (ft)
9/30-18B1 —Continued		
Shale, green	10	820
Basalt, gray	3	823
Shale, green	22	845
Basalt, gray	39	884
Basalt, broken	6	890
Shale, green	15	905
Basalt, gray	5	910
Shale, green	20	930
Basalt, broken	5	935
Basalt, gray	2	937
Basalt, gray and green shale	8	945
Basalt, broken, dark	22	967
Basalt, broken, and gray shale	10	977
Basalt, broken	14	991
Shale, dark	4	995
"Shell," hard	1	996
Basalt, hard, gray	10	1,006
Basalt, dark	46	1,052
Basalt, gray	39	1,091
Basalt, black	57	1,148
Basalt, gray	156	1,304
Shale, green	9	1,313
Basalt, broken, with gray shale	4	1,317
Shale, gray	7	1,324
"Shell," hard	1	1,325
Shale, green	6	1,331
"Shell," hard, green shale and rock in thin layers	11	1,342

11/29-5N1. West 15 Domestic Water Co.
Altitude, 915 ft. Drilled by V. E. Dilley, 1957.
Casing, 8- to 6-in. to 804 ft.; perforations, 576-640 ft.

	Thickness (ft)	Depth (ft)
Holocene and Pleistocene:		
Loess and fluvial and lacustrine deposits, undifferentiated:		
Topsoil	18	18
Pleistocene:		
Ringold Formation:		
Lacustrine clay and silt:		
Caliche	4	22
Clay	18	40
"Sandrock"	90	130
Sand	6	136
Clay, brown	16	152
Clay, gray	21	173
Clay, yellow	13	186
Sand, brown	61	247
Clay, gray, contains sand	53	300
Clay, brown	40	340
Clay, black, with a little gravel	20	360
Clay, gray, and sand	15	375
Sand, brown	25	400
Clay, brown	12	412
"Sandrock," brown	28	440

Continued

APPENDIX II—Continued

	Thickness (ft)	Depth (ft)
11/29-5N1—Continued		
"Sandrock," blue	10	450
Conglomerate:		
Sand and gravel, water-bearing at 576 ft.	135	585
Gravel and brown clay	5	590
Sand and gravel, water-bearing	42	632
Lacustrine clay:		
Clay, contains gravel	8	640
Pliocene-Miocene:		
Yakima Basalt:		
Saddle Mountains Member:		
Basalt, black	25	665
Basalt, porous, brown	15	680
Basalt, black	55	735
Shale, green	25	760
Clay, gray	15	775
Basalt, black	55	830
Basalt, gray	30	860
Basalt, black	76	936
Basalt, black, some shale	34	970
Basalt, porous, black	10	980
Basalt, black	140	1,120

14/23-2A1. Priest Rapids Development Co.
Altitude, 776 ft. Drilled by Barnett & Barnett, 1956.
Casing, 12- to 10-in. to 455 ft.

	Thickness (ft)	Depth (ft)
Pleistocene:		
Fluvial and lacustrine deposits:		
Fluvial gravel:		
Sand and gravel	154	154
Ringold Formation:		
Lacustrine clay and silt:		
Clay	27	181
Conglomerate:		
Gravel	44	225
Gravel, sandy, cemented	84	309
Shale, sandy, water-bearing at 318 ft.	53	362
Sand and gravel, cemented	60	422
Lacustrine clay:		
Clay, brown	7	429
Pliocene-Miocene:		
Yakima Basalt:		
Saddle Mountains Member:		
Basalt, broken	30	459
Ellensburg Formation:		
Beverly Member:		
Clay, blue	27	486
Sand and gravel, cemented	48	534
Clay and shale	36	570
Yakima Basalt:		
Priest Rapids Member:		
Basalt, gray, water-bearing at 613 ft.	133	703
Basalt, hard, gray	5	708

Continued

APPENDIX II—Continued

	Thickness (ft)	Depth (ft)
14/23-2A1—Continued		
Basalt, soft, water-bearing	9	717
Basalt, hard, gray	19	736
Clay, blue	9	745
Basalt, gray, water-bearing	4	749
Clay, blue	11	760
Basalt, gray	4	764
16/29-34D1. Chef Reddy Food Plant. Altitude, 1,078 ft. Drilled by O. F. Zinkgraf, 1962 Casing, 20- to 16-in. to 374 ft.		
Pleistocene:		
Fluvial and lacustrine deposits:		
Fluvial gravel:		
Gravel and boulders	17	17
Basalt, broken, and clay	9	26
Ringold Formation:		
Lacustrine clay:		
Clay, hard, light tan	54	80
Clay, hard, dark brown	34	114
Clay, yellow, contains sand	11	125
Clay, yellow	10	135
Clay, gray	21	156
Clay, blue	4	160
Clay, gray	6	166
Clay, blue, contains sand	9	175
Clay, blue, traces of basalt at 179 ft.	21	196
Pliocene-Miocene:		
Yakima Basalt:		
Saddle Mountains Member:		
Basalt, brown	33	229
Basalt, red	4	233
Basalt, medium hard, brown	15	248
Shale, brown	1	249
Basalt, broken, and blue clay	9	258
Basalt, black, contains green shale	30	288
Basalt, black, contains reddish-brown shale	17	305
Basalt, firm	10	315
Basalt, with blue shale	6	321
Priest Rapids Member:		
Basalt, medium hard to hard, gray	15	336
Basalt, medium hard, light tan	4	340
Basalt, soft to medium hard (caving)	34	374
Basalt, soft, and caving badly	3	377
Basalt, medium hard, brown	4	381
Basalt, black and gray	48	429
Basalt, hard, gray	26	455
Basalt, softer, black	48	503
Basalt, hard, gray	13	516
Basalt, soft, black, contains layers of green and yellow shale	27	543
Roza Member:		
Basalt, hard, gray	17	560
Basalt, "honeycombed," black	45	605

Continued

APPENDIX II—Continued

	Thickness (ft)	Depth (ft)
16/29-34D1.—Continued		
Basalt, firm, black	2	607
Basalt, loose and broken, black	18	625
Basalt, firm, black	5	630
Basalt, soft to firm, black	12	642
Basalt, medium hard, gray	8	650
Basalt, hard	9	659
Basalt, medium hard, black	22	681
Basalt, hard, black	3	684
Basalt, medium hard, black	19	703
Basalt, gray	4	707
Basalt, hard, gray	26	733
Basalt, medium hard, caving, black	13	746
Frenchman Springs Member:		
Basalt, softer, black	44	790
Basalt, hard, black	53	843
Basalt, soft, black	3	846
Basalt, hard, black (caving slightly at 891 ft.)	75	921
Basalt, loose	1	922
Basalt, hard	22	944
Basalt, soft, caving badly	55	999
Vantage Sandstone Member:		
Sand and clay	3	1,002
Shale, green	1	1,003
Lower basalt, undifferentiated:		
Basalt, medium hard, black	40	1,043

17/25-23K1. U.S. Bureau of Reclamation
Altitude, 1,227 ft. Drilled by B. L. Price, 1955-57.
Casing, 12-in. to 80 ft.; 6-in. liner, 818 to 839 ft.

	Thickness (ft)	Depth (ft)
Artificial fill:		
Caliche and basalt gravel (fill from Frenchman Hills tunnel excavation)	6	6
Pleistocene:		
Ringold Formation:		
Lacustrine clay and tuffaceous sand:		
Caliche and basaltic gravel	14	20
Sand, fine, calcareous, chiefly quartz, brown	30	50
Silt, sandy, light brown	10	60
Clay, compact, green to light tan	8	68
Pliocene-Miocene:		
Yakima Basalt:		
Priest Rapids Member:		
Basalt, slightly weathered, dark brown to black	9	77
Basalt, dense, hard, black	13	90
Basalt, medium hard, slightly iron-stained, dark gray	30	120
Basalt, vesicular, black	30	150
Basalt, medium hard, dark gray to black	40	190
Basalt, weathered, broken, brown	6	196
Basalt, scoriaceous, iron-stained	44	240
Roza Member:		
Basalt, medium hard, with broken zones, black	65	305
Basalt, vesicular, dark brown	4	309
Basalt, medium hard, dark gray	34	343

Continued

APPENDIX II—Continued

	Thickness (ft)	Depth (ft)
17/25-23K1 —Continued		
Basalt, broken, vesicular, black to brown	3	346
Basalt, brown to black	11	357
Frenchman Springs Member:		
Basalt, scoriaceous, soft, broken, brown	13	370
Basalt, slightly iron-stained, medium hard, black	20	390
Basalt, black to brown	25	415
Basalt, hard, gray	34	449
Basalt, jointed, gray	6	455
Basalt, hard, gray	29	484
Basalt, jointed, brown to black	26	510
Basalt, very hard, gray	20	530
Basalt, jointed, black	5	535
Basalt, hard, dense, black	64	599
Basalt, vesicular, hard, black	3	602
Basalt, jointed, medium hard, brown	6	608
Basalt, dense, hard, gray to black	40	648
Basalt, vesicular, broken, brown to black	11	659
Basalt, hard, dense, black	26	685
Vantage Sandstone Member:		
Basalt, interflow, vesicular and opaline minerals	28	713
Lower basalt, undifferentiated:		
Basalt, hard, black	35	748
Basalt, jointed, black	10	758
Basalt, hard to very hard, black	44	802
Basalt, vesicular, black	10	812
Basalt, medium hard, black	20	832
Interflow, chiefly opal	3	835
Basalt, hard, gray	10	845
Basalt, weathered, dark brown	12	857
Basalt, gray to black	51	908
Basalt, hard, brown	36	944
Basalt, dense, very hard, black	13	957

19/26-11J1. A. E. Lund. Altitude, 1,254 ft. Drilled by A. Lund, 1949. Casing, 12-in. to 400 ft.; perforations at 185, 240, and 270 ft.

	Thickness (ft)	Depth (ft)
Holocene:		
Soil	4	4
Pleistocene:		
Fluvial and lacustrine deposits:		
Fluvial and lacustrine sand:		
Sand	143	147
Ringold Formation:		
Tuffaceous sand and silt:		
Sandstone, water-bearing	143	290
Lacustrine clay:		
Clay, blue	110	400
Pliocene-Miocene:		
Yakima Basalt:		
Priest Rapids Member:		
Basalt	60	460
Basalt, reddish-brown	20	480
Basalt, hard, tested 20 gpm at 486 ft	20	500

Continued

APPENDIX II—Continued

	Thickness (ft)	Depth (ft)
19/26-11J1—Continued		
Basalt, porous	10	510
Basalt, soft, red	9	519
Basalt, hard, black	21	540
Roza Member:		
Basalt and clay	15	555
Basalt, gray	10	565
Basalt, porous	32	597
Basalt, water-bearing	17	614
Basalt, hard	33	647
Basalt, broken, and blue clay	15	662
Frenchman Springs Member:		
Basalt, hard, blue	33	695
Basalt, creviced, water-bearing	10	705

19/28-15Q1. City of Moses Lake. Altitude, 1,070 ft. Drilled 1950. Casing, 16-in. to 132 ft.

	Thickness (ft)	Depth (ft)
Holocene:		
Soil	2	2
Pleistocene:		
Fluvial and lacustrine deposits:		
Fluvial gravel:		
Gravel, coarse	8	10
Gravel and boulders	4	14
Boulders	4	18
Boulders and coarse gravel	3	21
Gravel, coarse	3	24
Gravel, dry	8	32
Sand and gravel	10	42
Boulders	11	53
Ringold Formation:		
Lacustrine clay:		
Clay, sandy, yellow	11	64
Tuffaceous sand and conglomerate:		
Gravel, loose	1	65
Sand and gravel	9	74
Gravel and broken basalt	6	80
Pliocene-Miocene:		
Yakima Basalt:		
Roza Member:		
Basalt, brown (caving)	22	102
Basalt, broken, blue	8	110
Basalt, blue	5	115
Basalt, blue (caving)	7	122
Basalt, creviced, blue	10	132
Basalt, hard, blue	11	143
Basalt, black	15	158
Basalt, creviced, dark	28	186
Basalt, blue	14	200
Squaw Creek Diatomite Bed:		
Shale, blue	5	205
Frenchman Springs Member:		
Basalt, creviced, blue	20	225
Basalt, hard, blue	36	261

Continued

APPENDIX II—Continued

	Thickness (ft)	Depth (ft)
19/28-15Q1.—Continued		
Shale, blue	4	265
Basalt, broken, blue	27	292
Basalt, hard, blue	9	301
Basalt, gray	47	348
Basalt, broken, blue	16	364
Basalt, gray	23	387
Basalt, blue	56	443
Vantage Sandstone Member:		
Clay, blue	7	450
Shale, brown	8	458
Lower basalt, undifferentiated:		
Basalt, broken, blue	18	476
Basalt, hard, gray	122	598
Basalt, broken, blue	7	605
Basalt, black	51	656
Basalt, creviced, black	14	670
Basalt, broken, black	84	754
Basalt, hard, blue	8	762
Basalt, hard, dark	61	823
Basalt, dark (caving)	4	827
Basalt, hard, dark	19	846
Basalt, dark (caving)	4	850
Basalt, hard, dark	23	873
Basalt, hard, gray	36	909

19/32-31C1. Henry Gering. Altitude, 1,469 ft.
Drilled by Joy Drilling Co. Casing, 12-in. to 20 ft.

	Thickness (ft)	Depth (ft)
Holocene and Pleistocene:		
Loess:		
Topsoil	10	10
Pliocene-Miocene:		
Yakima Basalt:		
Roza Member:		
Basalt, black, water-bearing	53	63
Basalt, gray	67	130
Basalt, brown, water-bearing	30	160
Basalt, gray	50	210
Basalt, broken	45	255
Basalt, gray	95	350
Basalt, broken	48	398
Basalt, gray	48	446
Basalt, broken	6	452
Vantage Sandstone Member:		
Shale, black	40	492
Lower basalt, undifferentiated:		
Basalt, broken, water-bearing	22	514
Basalt, gray	6	520

20/25-8P1. R. M. Anderson. Altitude, 1,242 ft.
Drilled by G. C. Hoff, 1918. Casing, 10-in. to 150 ft.

	Thickness (ft)	Depth (ft)
Holocene:		
Soil	3	3

Continued

APPENDIX II—Continued

20/25-8P1—Continued	Thickness (ft)	Depth (ft)
Pleistocene:		
Fluvial and lacustrine deposits:		
Fluvial and lacustrine sand:		
Sand, coarse, black	20	23
Ringold Formation:		
Tuffaceous sand and silt:		
Silt and "limerock"	130	153
Pliocene-Miocene:		
Yakima Basalt:		
Roza Member:		
Basalt, black	40	193
Basalt, hard, gray	32	225
Squaw Creek Diatomite Bed:		
Diatomite	5	230
Frenchman Springs Member:		
Basalt, "honeycombed," black, water-bearing	10	240
"Petrified wood," agatized	10	250
Peat	20	270
Basalt, black	50	320
Basalt, "honeycombed," black, water-bearing	10	330
Basalt, black	60	390
Basalt, "honeycombed," water-bearing	20	410
Basalt, black	20	430

23/28-4D1. U.S. Bureau of Reclamation (Long Lake Damsite Test Hole 25). Altitude, 1,538 ft.
Drilled by U.S. Bureau of Reclamation, 1946.

	Thickness (ft)	Depth (ft)
Measuring point above land-surface datum	3	3
Pleistocene:		
Fluvial and lacustrine deposits:		
Fluvial gravel:		
Sand and gravel	27	30
Pliocene-Miocene:		
Yakima Basalt:		
Priest Rapids Member:		
Basalt, vesicular	8	38
Basalt, dense	13	51
Basalt, vesicular; flow contact at 56 ft.	5	56
Basalt, very vesicular	14	70
Basalt, vesicular	13	83
Basalt, slightly vesicular	6	89
Basalt, dense; flow contact at 115 ft.	26	115
Basalt, vesicular	32	147
Quincy Diatomite Bed:		
Ash, white and palagonite	3	150
Roza Member:		
Basalt, vesicular	33	183
Basalt, dense	17	200
Basalt, dense	36	236

Continued

APPENDIX II—Continued

	Thickness (ft)	Depth (ft)
23/28-4D1.—Continued		
Frenchman Springs Member:		
Basalt, vesicular	10	246
Basalt, dense; flow contact at 267 ft.	21	267
Basalt, vesicular	9	276
Basalt, dense	14	290
Basalt, vesicular	11	301
Basalt, dense	7	308
Sand, black, and green mud	6	314
Basalt, vesicular	17	331
Basalt, dense	7	338
Basalt, vesicular	4	342
Basalt, dense	10	352
Ash and vesicular fragments	1	353
Basalt, dense	3	356
Basalt, vesicular	2	358
Lower basalt, undifferentiated:		
Basalt, fragmental, with brown crystalline material	7	365
Basalt, decomposed	3	368
Basalt, vesicular	19	387
Basalt, slightly vesicular	12	399
Basalt, vesicular, calcite in seams; flow contact at 427 ft.	51	450
Basalt, dense	17	467
Basalt, vesicular	2	469
Ash, tan, flow contact	4	473
Basalt, very vesicular	4	477
No core	2	479
Basalt, vesicular	4	483
Basalt, dense	4	487
Basalt. vesicular	8	495
Basalt, dense, pyrite in seams	14	509
Basalt, amygdaloidal	6	515
Basalt, dense	2	517
Basalt, vesicular	21	538
Basalt, slightly vesicular	16	554
Basalt, dense	70	624
Basalt, vesicular	4	628
Basalt, dense; flow contact at 635 ft.	7	635
Basalt, vesicular	5	640
Mud seam	2	642
Basalt, vesicular and cemented breccia	4	646
Basalt, vesicular	10	656

APPENDIX III

DEPTH INTERVALS OF STRATIGRAPHIC UNITS AS PENETRATED BY SELECTED WELLS

Well number	Altitude (ft)	Depth intervals, in feet below land surface, interpreted from drillers' logs and surficial geology							
		Glacio-fluvial sand and gravel	Ringold conglomerate	Saddle Mountains Member	Priest Rapids Member	Roza Member	Frenchman Springs Member	Vantage Sandstone Member	Lower basalt flows
9/29-21M1	360	0-28	...	...	...	...	...	...	...
9/30-18H1	410	0-65	65-140	220-1043?	...	...	...	...	...
20F1	420	0-65	65-200	200-1030?	...	...	...	...	...
9/32-8G1	448	0-120	...	...	120-210	...	...	...	...
10/29-8K1	645	...	...	299-463	...	...	...	...	...
9R1	614	...	...	249-388	...	...	...	...	...
10D1	686	...	195-290	290-618	...	...	...	...	...
11K1	660	...	...	245-397	...	...	...	...	...
15M1	566	...	...	195-352	...	...	...	...	...
19O1	508	4-173	210-238	...	...	...	...	...	...
10/30-18G1	539	0-50	75-135	135-716	...	...	...	...	...
11/29-2L1	927	...	...	300-459	...	...	...	...	...
16A1	916	...	...	375-960	...	...	...	...	...
17N1	891	...	400-475	475-565	...	...	...	...	...
20N1	890	...	417-610	610-936	...	...	...	...	...
26P1	707	...	201-304	309-426	...	...	...	...	...
31O1	862	...	400-540	540-725	...	...	...	...	...
11/30-3L1	770	...	...	45-105	...	...	...	...	...
6B1	780	...	...	142-196	...	...	...	...	...
7D1	848	...	...	281-427	...	...	...	...	...
10B1	765	...	...	5-115	...	...	...	...	...
34L1	626	7-15	...	15-400	400-526	...	...	...	...
12/28-24N1	396	...	18-85,127-155	205-693?	720?-755?	...	...	...	...
12/29-3O1	822	...	...	258-317	...	...	...	...	...
4O1	856	...	...	298-463	...	...	...	...	...
7N1	588	...	190-215	267-465	...	...	...	...	...
12N1	935	...	...	328-430	...	...	...	...	...
20N1	860	...	296-340,356-380	380-540	...	...	...	...	...
12/30-5E1	966	...	...	245-451	...	...	...	...	...
15B1	820	...	...	20-210	...	...	...	...	...
18C1	965	...	...	343-600	...	...	...	...	...
18R1	970	...	...	260-519	...	...	...	...	...
19A1	816	...	...	34-106	...	...	...	...	...
21G1	803	7-15	...	15-211	...	...	...	...	...
34J1	755	24-87	...	87-276	...	...	...	...	...
13/28-13N1	952	...	...	561-1111	1117-1119	...	...	...	...
13/29-20N1	982	...	...	520-870	...	...	...	...	...
23M1	695	...	...	42-310	...	...	...	...	...
13/30-29A1	888	1-43	...	65-275	287-540	...	...	...	...
14/23-2A1	776	0-153	225-309,362-422	429-459	570-764	...	...	...	...
36L1	550	3-142	...	142-195?	199-350	...	...	...	...
14/25-1D1	660	...	...	355-798,905-924	...	...	...	...	...
21B1	640	0-25	343-407	520-522	...	...	...	...	...
14/27-24C1	862	...	...	589-1261	1291-1396	...	...	...	...
14/30-2A1	1170	...	...	133-387	...	...	...	...	...
27J1	942	...	...	1-340	360-381	...	...	...	...
15/26-2801	770	...	...	275-660	788-892	...	...	...	...
15/27-5J1	635	...	...	21-100	...	...	...	...	...
5R1	726	...	...	0-128	133-358	...	...	...	...
32E1	730	...	...	277-765	846-1034	1034-1123	...	...	...
34L2	697	...	...	208-350	350-604	604-636	...	...	...
15/28-6M1	632	9-13	...	13-131	172-176	...	...	...	...
8E1	855	...	...	252-371	414-415	...	...	...	...
15/29-3D1	1091	...	...	212-363	363-574	574-693	...	...	...
4P1	992	...	...	102-163	197-335	...	...	...	...
9E1	967	8-57	...	160-259	...	...	...	...	...
18B1	919	3-41	...	132-330	360-495	...	...	...	...
27E1	959	2-45	...	135-222	232-400	...	...	...	...

APPENDIX III—Continued

Well number	Altitude (ft)	Depth intervals, in feet below land surface, interpreted from drillers' logs and surficial geology							
		Glacio-fluvial sand and gravel	Ringold conglomerate	Saddle Mountains Member	Priest Rapids Member	Roza Member	Frenchman Springs Member	Vantage Sandstone Member	Lower basalt flows
15/30-4M1	1149	...	...	...	138-338	...	...	...	...
10P1	1227	...	...	...	160-300	306-518	...	...	...
23H1	1195	...	...	51-145	170-408	408-497	497-748	748-763	763-850
33A1	1173	...	...	266-343	360-480	...	...	...	...
15/32-13D1	1216	...	...	...	...	...	8-172	172-175	175-602
16/24-1G1	1215	...	...	...	25-205	205-315	340-710	710-748	748-800
16/25-1Q1	1030	...	...	40-179	254-475	481-575	575-917	...	...
8M1	1194	...	...	...	41-112	...	...	...	...
16/27-12M1	980	...	...	92-191	210-300	...	...	...	...
16/28-4B1	1020	...	...	3-160	167-179	...	...	...	...
5H1	1076	...	...	75-149	...	...	...	...	...
5N1	1000	...	...	66-180	214-290	...	...	...	...
6P1	1048	...	...	124-262	...	...	...	...	...
7K1	960	...	...	70-200	...	...	...	...	...
16/29-34D1	1077	...	...	196-315	321-516	543-732	732-999	999-1003	1003-1043
34R1	1116	...	...	178-311	339-561	571-725	725-900	...	...
16/30-16L1	1226	...	...	...	25-80	80-229	...	...	...
18A1	1190	...	...	...	47-277	281-392	...	...	...
16/30-31R1	1142	...	...	...	148-250	...	...	...	...
16/31-35J1	1450	...	...	...	...	55-157	157-403	403-478	478-901
17/25-31N1	1192	...	...	...	26-110	...	...	...	...
17/27-31D1	1170	...	...	132-215	270-470	470-580	580-810	...	...
17/28-31L1	1139	...	...	70-163	195-230	...	...	...	...
33C1	1102	5-24	...	25-168	...	...	...	...	...
17/30-2Q2	1179	...	...	...	...	...	10-190	190-191	191-255
9N1	1238	25-48	...	...	...	48-159	179-319	...	...
12H1	1253	...	...	...	...	...	44-170	170-178	178-350
33K1	1344	...	...	...	...	37-246	246-628	628-634	634-1002
17/31-8R1	1249	...	...	...	...	...	62-155	...	...
18/23-36H1	1302	...	...	...	102-209	228-375	375-498	498-505	505-670
18/24-32D1	1239	...	...	...	30-160	180-220	...	...	...
32N2	1320	...	...	...	135-255	280-374	374-425	...	...
33Q1	1239	...	...	...	60-168	...	...	...	...
18/28-2E1	1142	3-42	...	...	41-90	90-187	226-426	426-465	...
4D2	1067	...	...	...	43-100	100-190	...	...	...
4D3	1077	...	...	...	57-115	...	...	...	...
24G2	1138	...	...	...	146-194	194-350	350-546	...	...
24K1	1142	...	...	...	110-219	219-335	335-453	...	...
26F1	1110	0-70	...	...	141-312	312-455	455-?	...	?-801
26J1	1154	0-19	...	...	112-160	...	...	...	...
18/29-1A2	1274	...	...	...	30-238	238-348	348-510	...	...
1B1	1243	...	...	...	15-181	181-320	320-542	...	...
1L1	1260	...	...	...	12-288	288-390	390-616	...	...
6R1	1164	...	...	...	32-156	156-283	283-525	...	...
11A2	1251	...	...	...	28-159	159-270	297-527	527-547	547-648
17P1	1170	0-42	...	...	90-294	300-342	...	...	...
33H1	1138	0-95?	...	...	130-175	190-210	...	...	...
18/30-3A2	1140	...	...	...	...	...	109-266	...	...
16R1	1206	0-60	...	...	...	113-185	...	...	...
18/31-4G1	1190	...	...	...	...	...	46-267	267-282	282-293
13R2	1462	...	...	...	...	29-148	148-391	391-428	428-613
18/32-28C1	1484	...	...	...	55-143	143-265	265-553	553-587	587-616
19/24-3B1	1235	3-54	...	...	54-186	...	...	...	...
7J1	1256	...	...	...	167-267?	267-?	?-502	...	...
28N1	1225	5-8	...	...	80-210	...	...	...	...
28N2	1224	5-11	...	...	80-140	...	...	...	...
19/26-11J1	1254	0-147	...	...	400-540	540-647	647-705	...	...
19/27-4M3	1126	...	...	...	139-270	270-378	...	...	...
7A1	1182	0-100	...	...	236-302	302-392	392-570	...	...
7J1	1166	2-65	...	...	203-256	...	...	...	...
7K1	1165	3-70	...	...	205-242	...	...	...	...
8C1	1150	0-30	...	...	140-180	190-210	...	...	...
12A1	1124	...	...	...	121-205	...	...	...	...
18R2	1136	2-63	...	...	208-250	...	...	...	...

APPENDIX III—Continued

Well number	Altitude (ft)	Depth intervals, in feet below land surface, interpreted from drillers' logs and surficial geology							
		Glacio-fluvial sand and gravel	Ringold conglomerate	Saddle Mountains Member	Priest Rapids Member	Roza Member	Frenchman Springs Member	Vantage Sandstone Member	Lower basalt flows
19/27-19R1	1122	2-20	...	...	166-225	...	...	...	...
20F1	1077	0-40	...	...	99-207	...	...	...	...
26B1	1108	0-85	...	...	150-237	...	...	...	...
19/28-4P1	1179	0-106	...	...	200-260	271-415?	415?-748	...	...
6M4	1082	0-40	...	...	115-197	197-263	...	...	...
7E1	1122	2-102	...	...	...	155-208	...	...	...
12K1	1192	2-68	...	...	82-118	118-218	228-329	...	...
13R1	1201	2-25	...	...	71-137	150-260	260-475	475-488	488-568
14H3	1082	...	...	...	7-47	95-112	...	...	...
15Q1	1070	2-53	...	...	80-102	102-200	205-442	442-458	458-909
22B2	1072	0-41	...	...	40-154	154-270	275-528	528-545	545-763
24H1	1200	26-29	...	...	126-144	144-257	257-345	...	...
24Q1	1202	2-70	...	...	91-140	140-260	...	...	...
25C1	1196	2-64	...	...	110-145	145-270	270-293	...	...
27H1	1090	...	...	...	...	97-336?	336?-650	650-689	689-1045
28K4	1073	0-12	...	...	...	44-275?	275-665	665-690	690-1000
29N2	1056	0-74	...	...	48-145	145-330	342-688	688-696	696-800
34G1	1152	2-40	...	...	90-162	162-381?	381?-419	...	...
19/29-1G1	1240	...	...	...	...	...	24-227	227-240	...
7A1	1200	...	...	...	13-113	153-170	...	...	...
9H1	1300	...	...	...	47-223	232-239	242-588	588-605	605-637
22E1	1294	...	...	...	15-225	245-320	325-525	525-560	560-585
22R1	1282	...	...	...	50-157	205-290?	290-560	...	...
29A1	1188	...	...	...	30-80	88-208	235-410	410-432	432-486
19/30-32N1	1278	...	...	...	...	23-125	125-332	332-345	345-351
35K1	1170	...	...	...	...	...	35-239	239-264	264-360
19/31-16H1	1414	...	...	...	...	12-142?	142?-307?	...	307?-565
19/32-19N1	1290	...	...	...	...	28-53	53-290	290-302	302-411
19/32-31C1	1469	...	...	...	...	10-160	160-453	453-492	492-520
31G1	1438	...	...	...	...	48-140	140-403	403-438	438-445
20/23-1Q1	1358	...	...	...	...	60-170?	170?-404	404-430	430-505
11H2	1335	...	...	...	...	45-228?	228?-442	442-475	475-1000
12A1	1336	...	...	...	...	25-105?	105?-345	345-355	355-370
12J1	1324	...	...	...	...	...	?-388	388-391	...
22M2	1338	...	...	...	...	54-159?	159?-412	412-420	...
25E1	1300	...	...	...	...	48-168	175-447	447-452	452-480
35A1	1290	...	...	...	...	...	?-405	405-411	411-414
20/24-6E1	1376	...	...	...	...	...	?-408	408-420	...
20/25-3R1	1272	...	...	...	...	15-160	...	...	...
5N1	1292	...	...	...	...	38-138	138-310	...	...
6D1	1408	...	...	...	...	31-91	91-245	...	...
8P1	1242	3-23	...	...	...	153-225	270-430	...	...
10E1	1202	...	...	...	...	16-62	65-92	...	...
16J1	1228	2-55	...	...	80-100	155-228	...	...	...
16K1	1226	0-40	...	...	160-212	...	...	...	...
20A1	1184	2-15	...	...	120-?	?-260	275-415	...	...
21A2	1226	0-40	...	...	...	219-324	339-555	...	555-652
20/26-21A1	1242	0-95	...	...	256-290	290-390	392-416	...	...
22P1	1248	0-70	...	...	350-387	...	...	...	...
20/27-28D1	1120	0-25	...	...	...	100-170	...	...	...
20/28-2N1	1237	...	...	...	72-88	...	...	...	...
13B1	1274	0-59	...	...	59-97	97-203	...	...	...
13Q1	1275	...	...	...	103-126	...	...	...	...
16R1	1152	0-78	...	...	...	92-113	...	...	...
17G1	1155	3-64	...	...	...	89-153	153-212	...	...
26N2	1106	0-30	...	...	30-46	49-80	...	...	...
32C1	1195	0-123	...	...	176-185	200-310?	310?-480?	...	480-725
32J1	1187	0-88	...	...	177-223	225-293	294-496	496-532	532-712
33E1	1169	0-85	...	...	...	142-230?	230?-418	418?-452	452-791
20/29-18J1	1279	0-50	...	...	50-80	...	...	...	...
18N1	1278	0-52	...	...	52-149	...	...	...	...
18Q1	1272	0-70	...	...	70-153	...	...	...	...
19C1	1276	0-51	...	...	51-168	...	...	...	...
20D1	1277	0-50	...	...	50-172	...	...	...	...

APPENDIX III—Continued

Well number	Altitude (ft)	Depth intervals, in feet below land surface, interpreted from drillers' logs and surficial geology							
		Glacio-fluvial sand and gravel	Ringold conglomerate	Saddle Mountains Member	Priest Rapids Member	Roza Member	Frenchman Springs Member	Vantage Sandstone Member	Lower basalt flows
20/29-27J1	1230	0-73	...	...	...	...	73-155	155-182	182-394
28C1	1288	...	...	...	15-80	80-190	190-385	385-407	407-416
29D1	1270	0-?	...	...	...	190-215	...	...	...
20/29-29P1	1278	0-50?	...	...	50-62	...	...	...	...
29P2	1277	...	...	...	12-72	...	...	...	...
30A1	1274	0-?	...	...	...	160-216	...	...	...
20/30-35R1	1340	...	...	...	...	...	3-117	...	...
21/23-26H1	1700±	...	...	...	...	...	4-88	88-133	133-310
21/25-27J1	1780	...	...	...	...	...	51-190	190-240	240-520
33H1	1509	...	...	...	...	40-140	140-390	390-396	...
35P1	1396	...	...	...	...	35-85	85-288	288-293	293-312
21/26-10C1	1271	0-49	...	...	...	...	49-146	...	...
10G2	1290	0-67	...	...	...	67-81	85-127	...	...
14L1	1288	0-80	...	...	...	210-320	320-568	568-650	650-1025
15F1	1272	...	...	...	...	283-327	327-500	...	...
16B1	1345	...	...	...	...	...	...	...	0-252
16B3	1345	...	...	...	...	...	...	...	1-260
21E1	1490	...	...	...	...	...	100-220	220-299	259-618
28K1	1230	...	...	...	...	34-149?	149?-426	426-483	483-580
29F1	1542	...	...	...	...	15-85	85-335	335-345	...
21/28-2D2	1314	0-35	...	...	...	35-218	218-308	...	...
35M1	1262	0-100	...	...	...	100-163	...	...	...
21/30-3E1	1684	...	...	...	19-176	176-270	270-451	...	...
22/26-13M1	1374	...	...	...	...	...	...	...	27-330
24L1	1157	3-81	...	...	...	...	96-345	345-400	400-435
25L1	1202	0-90	...	...	...	...	190-434	434-473	...
22/27-19N1	1118	0-54	...	...	...	...	54-270	270-300	300-466
22/28-5C1	1366	5-225	...	...	...	...	...	...	225-303
34H2	1339	0-60	...	...	...	...	60-283	...	...
22/29-17D1	1633	...	...	...	24-142	142-217	217-384	384-421	421-625
23/27-14Q1	1270	...	...	...	...	...	...	...	25-400

APPENDIX IV

GRADIENTS OF THE RED MARKER BED AND BASALT SURFACE, AND CHANGE IN THE THICKNESS OF THE RINGOLD FORMATION (BELOW THE RED MARKER BED) IN RINGOLD COULEE

Location or well number[1/]	Red marker bed[2/]		Top of basalt		Ringold Formation below red marker bed	
	Altitude (ft above sea level)	Gradient (ft per mile)	Altitude (ft above sea level)	Gradient (ft per mile)	Thickness (ft)	Decrease in thickness (ft per mile)
Well 12/28-24N1 (White Bluffs)	850+	10	205	74	645	64
Location 13/29-21 SE$\frac{1}{4}$ (north wall, Ringold Coulee)	910		650		260	

1/ Distance between sites, 6 miles.
2/ After Brown and McConiga, 1960, p. 44, fig. 1.