GEOHYDROLOGY AND DIGITAL SIMULATION OF THE GROUND-WATER FLOW SYSTEM IN THE UMATILLA PLATEAU AND HORSE HEAVEN HILLS AREA, OREGON AND WASHINGTON

By A. Davies-Smith, E.L. Bolke, and C.A. Collins

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CONVERSION FACTORS

For the convenience of readers who may prefer to use metric (International System) units rather than the inch-pound units used in this report, values may be converted by using the following factors.

Factors for converting English units to metric units are shown to four significant figures.

Multiply inch-pound unit	By	To obtain metric unit
acre-foot (acre-ft)	1,233	cubic meter (m^3)
acre-foot (acre-ft)	0.001233	cubic hectometer (hm^3)
foot (ft)	0.3048	meter (m)
inch (in.)	25.40	millimeter (mm)
mile (mi)	1.609	kilometer (km)
square foot (ft^2)	0.09294	square meter (m^2)
square mile (mi^2)	2.59	square kilometer (km^2)
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
foot per second (ft/s)	0.3048	meter per second (m/s)

<u>Sea level</u>: In this report "sea level: refers to the National Geodetic Vertical Datum of 1929 (NGVD of 1929)--a geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, formerly called "Sea Level Datum of 1929."

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ABSTRACT

The Columbia Plateau is underlain by massive basalt flows of the Columbia River Basalt Group, with a composite thickness of about 10,000 feet. The oldest and most extensive basalt unit, the Grande Ronde Basalt, is estimated to be about 8,000 feet thick beneath the Columbia River. Overlying the Grande Ronde is the Wanapum Basalt, which in the study area reaches a thickness of about 1,000 feet near the Columbia River. The Saddle Mountains Basalt, overlying the Wanapum Basalt, is about 800 feet thick. Sedimentary deposits overlie the basalt, principally near the Columbia River, and reach a thickness of about 150 feet.

Individual basalt flows within the massive outpourings range from a few feet to as much as 300 feet thick. Characteristically, the basalt flows have a dense center with vertical jointings and a scoriaceous or brecciated zone at the top and bottom. Porous zones between basalt flows are called interflow zones; they consist of weathered basalt flow-top breccia and scoria and may also contain sedimentary interbeds.

Structural features in the study area include a series of anticline-syncline pairs. An arcuate topographic feature that extends east-west through the area is called the Willow Creek monocline. A head gradient as steep as 400 feet per mile across the feature indicates that it may be a hydrologic barrier.

The main avenues of ground-water movement in the Columbia River Basalt Group are the interflow zones between basalt layers. These complex, poorly-interconnected zones may be rather extensive in the lateral direction but largely isolated from overlaying and underlying interflows by poorly permeable basalt flow centers. Wells in the Columbia River Basalt Group usually intercept more than one interflow zone. Another control on ground-water movement is the "barrier" effect of stratigraphic pinch-outs and offsets. Although these are not entirely impermeable to ground water, they do retard ground-water movement and may effectively isolate parts of the units.

Geohydrologic units (aquifers) in the study area were delineated by vertical grouping of wells with similar head, separated by zones of low permeability. Four such units were defined for this report. The uppermost unit (layer 1) consists of unconsolidated deposits of gravel, sand, silt, and clay that overlie the basalt; layer 2 consists of the Saddle Mountains Basalt, layer 3 the Wanapum Basalt, and layer 4 consists of the upper thousand feet of Grande Ronde Basalt.

A three-dimensional finite-difference model to simulate ground-water flow was calibrated for steady-state and transient-flow conditions. The steady-state model was used to simulate ground-water conditions prior to major ground-water development, which began about 1950. The transient model was used to simulate ground-water conditions from 1950 to 1982. Additionally, the transient model was used to predict possible future ground-water development alternatives.

On the basis of model analysis, the major components of the water budget for inflow before ground-water development were recharge from precipitation--about 146 ft^3/s (cubic feet per second) or 106,000 acre-ft/yr (acre-feet per year), ground-water leakage from streams--about 31 ft^3/s (22,400 acre-ft/yr), and subsurface flow entering the boundaries of the study area--about 15 ft^3/s (10,900 acre-ft/yr). The major outflow component of the budget was leakage to streams--about 185 ft^3/s (134,000 acre-ft/yr).

During the period 1950-82, maximum water-level declines of about 300 feet occurred in parts of the Grande Ronde Basalt unit (layer 4). During the same period, ground-water withdrawal increased to about 182 ft^3/s (132,000 acre-ft/yr). On the basis of model analysis, return flow from surface irrigation increased 36 ft^3/s (26,100 acre-ft/yr), ground-water storage decreased by about 87 ft^3/s (63,000 acre-ft/yr), ground-water leakage to streams decreased by about 42 ft^3/s (30,400 acre-ft/yr), and leakage from streams increased by about 17 ft^3/s (12,300 acre-ft/yr).

INTRODUCTION

Early water use in the Umatilla Basin consisted primarily of (1) municipal use, in towns such as Pendleton, Heppner, Hermiston, and Arlington; and (2) rural domestic and stock-watering use by dryland wheat farmers. Surface-water diversions, shallow wells, and sumps supplied water for flood and wheel-line irrigation systems in the valley bottoms. Irrigated crops commonly included pasture and alfalfa. In about 1964 higher prices for wheat encouraged additional irrigation and caused changes in cropping patterns.

An irrigation "boom" in the 1960's and 1970's centered around Hermiston, Oregon. Owners of large corporate farms close to the Columbia River installed pumping stations to lift Columbia River water to their arrays of center pivots. Away from the rivers, development spread to the uplands as irrigators turned for large supplies of ground water to the basalt bedrock aquifers underlying Umatilla, Morrow, and Gilliam Counties. Dryland wheat farming gave way to irrigated wheat and row crops. Multiple center-pivot systems and greater depth to water in the uplands contributed to the drilling of deeper wells and installation of pumps of higher capacity than those used in the valley bottoms. Since 1960, more than 300 irrigation wells have been drilled into the basalt in the three-county area (Smith, Collins, and Olson, 1983).

The development of this water source slowed in the late 1970's, due to increasing recognition of problems associated with large amounts of pumpage. These problems have included widespread water-level declines and well interference, increasingly expensive operation, and deteriorated performance of wells. Previous studies by U.S. Geological Survey and Oregon Water Resources Department (OWRD) indicated the need for OWRD to take regulatory action to control ground-water withdrawals in the area.

However, little was known about the magnitude of recharge, discharge, vertical movement of water, and hydraulic characteristics of the ground-water reservoir. In 1980 the Geological Survey entered into a cooperative agreement with OWRD to describe and quantify the ground-water resources of the Umatilla Basin.

Purpose and Scope

The major purposes of this report are to describe the aquifer system (its boundaries, hydrologic properties and flow system) and to quantify recharge and pumpage. In order to enhance understanding of the ground-water system, a digital-flow model was developed and calibrated. The model was used to simulate ground-water levels for the period 1950-1982 and to predict ground-water levels to the year 2000.

This report covers the period from 1950 to 1982 and is based partly on data collected in the field and partly on data available in the files of OWRD, U.S. Geological Survey, and various other state and federal agencies. Field data collection by OWRD and the Geological Survey included locating wells, making water-level measurements in wells, obtaining power-consumption and flow-meter readings from wells, and mapping to supplement existing geologic maps. Available data included drillers' reports, streamflow measurements, and precipitation, temperature, land-use, and soils information.

Previous Work and Acknowledgments

Previous work in the area includes a study of the stratigraphy of the Columbia River Basalt Group (Swanson and others, 1979); reconnaissance geologic maps of the Columbia River Basalt Group (Swanson and others, 1981); a map study of geology and structure of The Dalles 1:250,000 quadrangle by Bela (1982); geologic studies by Shannon and Wilson, Inc. of Portland for Portland General Electric (1973 and 1974); and Newcomb's (1967) discussion of The Dalles-Umatilla syncline.

Wagner's (1949) hydrologic study of the Umatilla River basin includes a well and spring inventory with drillers' logs. Hogenson's (1964) discussion of the geology and ground-water resources in the Umatilla Basin includes a later inventory of wells. Robison's (1971) atlas of the hydrology of the Hermiston-Ordnance area contains information on geology, hydraulic heads, water chemistry, and carbon-14 dates. Water quality in the basalt is further described and summarized in a report by Newcomb (1972). Gonthier and Harris (1977) made an analysis of the water resources of the Umatilla Indian Reservation on the eastern edge of the study area. Ground-water studies by McCall (1975) and Bartholomew (1975) of Oregon Water Resources Department were compiled at the time the State held hearings on the proposed Ordnance and Butter Creek critical ground-water areas; these studies include tabulations of water rights. The most recent published well inventory is in the report by Smith, Collins, and Olson (1983).

Land-use mapping has been done by several agencies, including the U.S. Army Corps of Engineers (Johnson and others, 1981), Battelle Pacific Northwest Laboratories for Rockwell Hanford Operations (Stephan and others, 1971), and Environmental Remote Sensing Applications Laboratory (ERSAL) in cooperation with the Soil Conservation Service and the U.S. Forest Service (Murray, 1981).

Numerous studies describe the hydrology of other parts of the Columbia Plateau; the more extensive of these, such as that by Gephart and others (1979), include discussions applicable to northeastern Oregon.

Special thanks are due to the many well owners in the study area who allowed access to their wells and power records; their cooperation has been indispensable. P. L. Oberlander, D. W. Miller, J. R. Bull, and R. Almy of OWRD shared the data they collected in the field and compiled from OWRD files. Mr. Oberlander (now at Battelle Pacific Northwest Laboratories) in particular shared his scientific observations and the results of his study of the area. Steve Applegate (then State Watermaster in Pendleton) and his assistant Tony Holcomb have been sources of practical, first-hand information on the wells in the area.

Location of the Study Area

The study area includes 5,800 square miles in northeastern Oregon and southeastern Washington (fig. 1). Data for Washington were provided by Pacific Northwest District Office of the Geological Survey in Tacoma, Washington, which is carrying out a similar study on the north side of the Columbia River in the Horse Heaven Hills area. The present study emphasizes the 3,800-square-mile area in Oregon that includes parts of Umatilla, Morrow, and Gilliam Counties.

The Oregon part of the study area has been called by many names. Walker (1977) includes the area in the larger Deschutes-Umatilla Plateau. Hogenson (1964) subdivided the area into the Blue Mountain upland, the Blue Mountain slope, the Pendleton plains, and the Umatilla lowlands; Robison (1971) also refers to the Umatilla lowlands. In this report the Oregon part of the study area will be called the Umatilla Plateau and the Washington part will be called the Horse Heaven Hills area.

From the Blue Mountains on the south edge of the area, the land slopes rather gently northward toward the Columbia River, which flows east to west and marks the Oregon-Washington State line. Similarly, the Horse Heaven Hills area slopes southward from the Horse Heaven Hills toward the Columbia River. Together, the north and south slopes form a broad trough of arable land.

Limited precipitation (as little as 8 inches per year along the Columbia River) precludes most agricultural land uses except range and dryland wheat farming, unless the land is irrigated. The Oregon side includes several major streams: the Umatilla River, with tributaries Butter Creek, Birch Creek, McKay Creek, and Wildhorse Creek; and Willow Creek, with tributary Rhea Creek. Rock Creek, a tributary of the John Day River, forms the west edge of the study area. Most of these streams have small summer flows and are subject to diversions for irrigation. The Columbia River, with an average discharge of 182,400 ft^3/s (cubic feet per second) at McNary Dam (U.S. Geological Survey, 1983), has diversions at a number of sites in the study area where pumping stations lift water to irrigate extensive lowland farms.

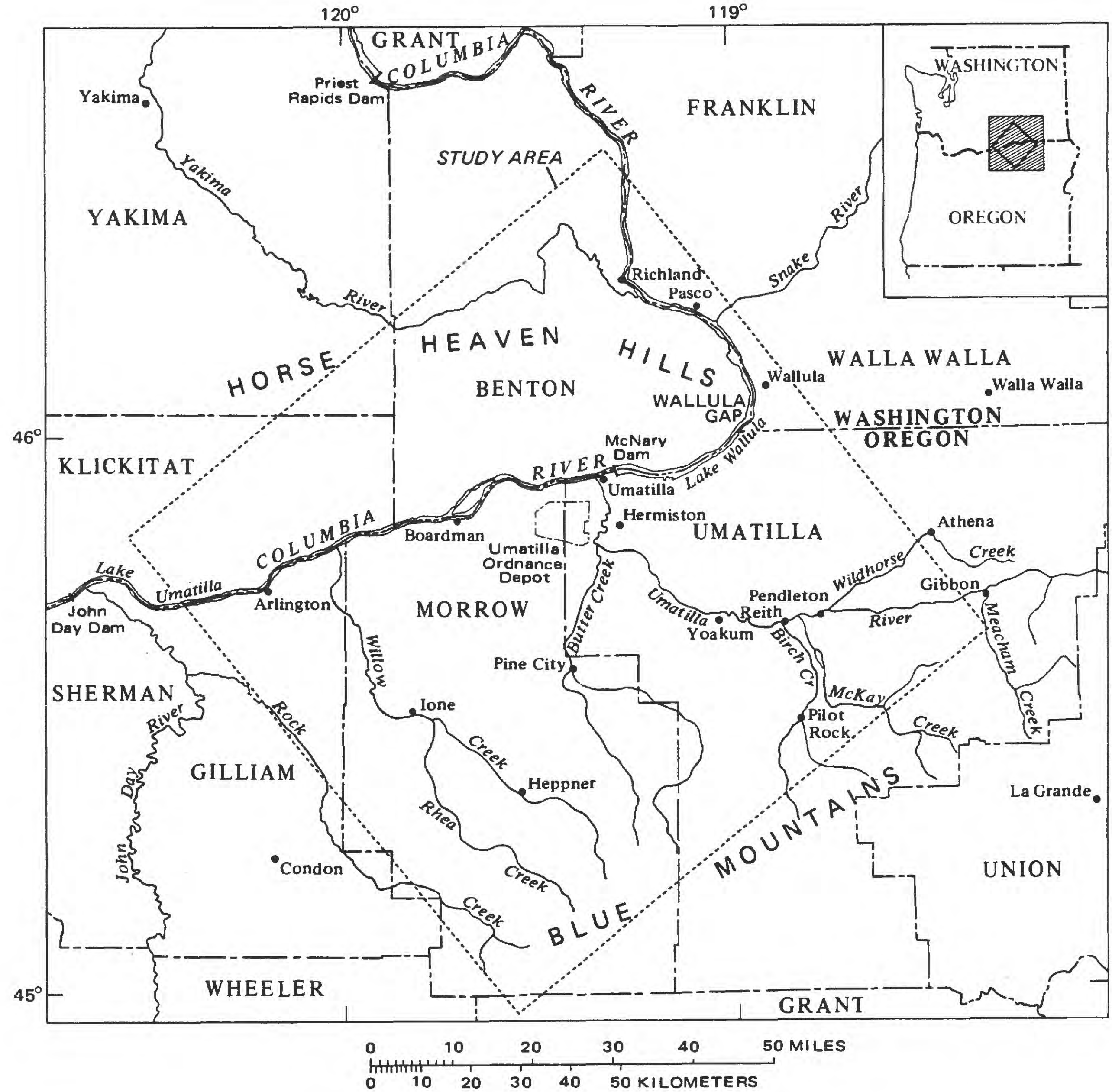


Figure 1.--Location of study area.

GEOLOGIC AND STRUCTURAL FRAMEWORK

The study area is located at the southern edge of the Columbia Plateau. The Plateau is underlain by flood basalts of the Miocene Columbia River Basalt Group and Holocene-to-Miocene sediments. These massive basalt outpourings, each of successively less volume, followed and filled ancient topographic lows and lapped onto the Blue Mountains in northern Oregon (figs. 2 and 3). Most of the flows originated to the east, in southeastern Washington, northeastern Oregon, and southern Idaho. Due to decreasing volume of the successive basalt outpourings and to the structural uplift of the Blue Mountains, each successive flow halted further to the north, resulting in thinning of the basalt sequence from north to south in Oregon. An oil test well near Condon, about 14 miles north of the edge of the basalt, penetrated over 2,400 feet of basalt.

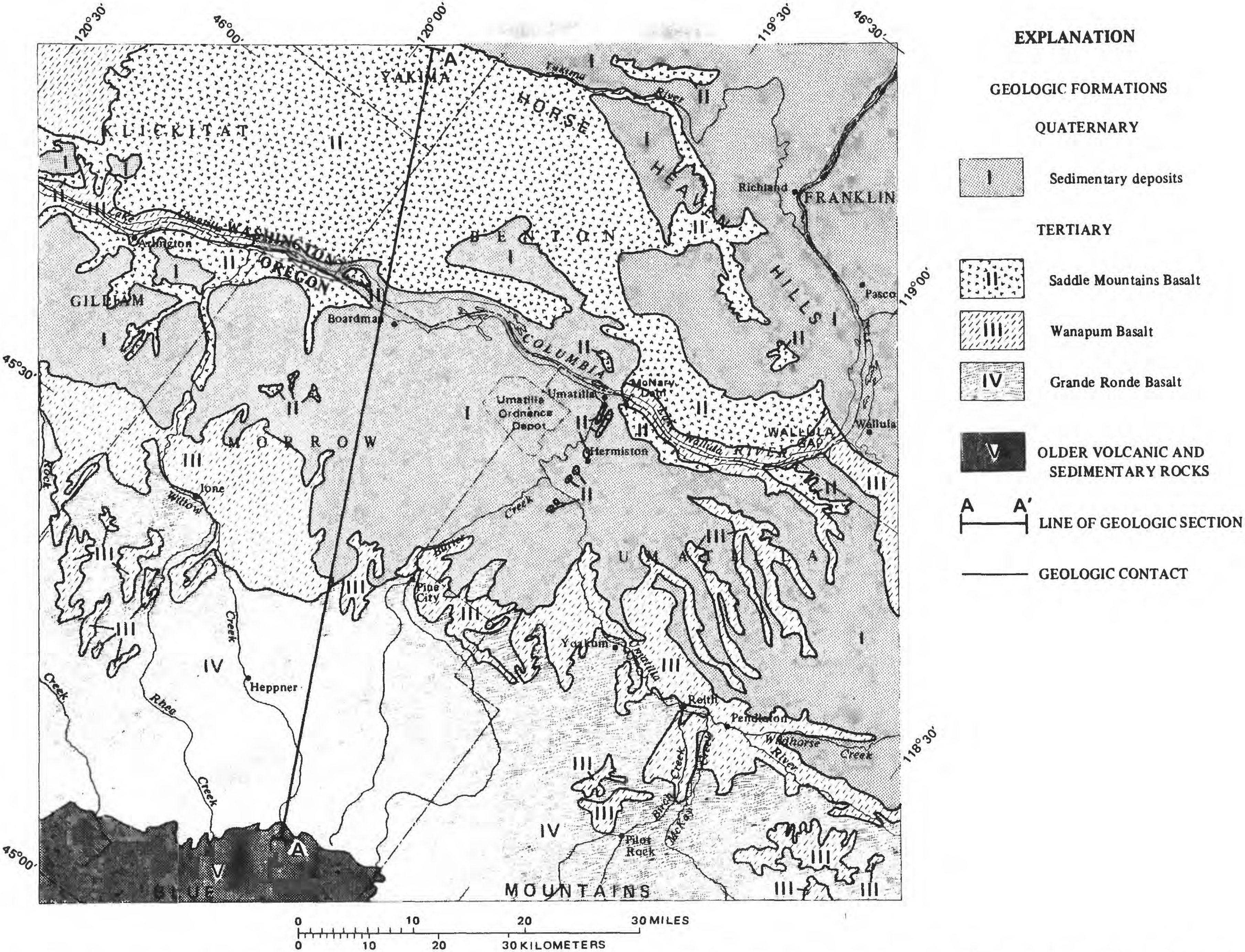


Figure 2.--Generalized geology of study area. Geology adapted from Swanson and others (1979 and 1981), Bela (1982), Shannon and Wilson (1973), and Newcomb (1967).

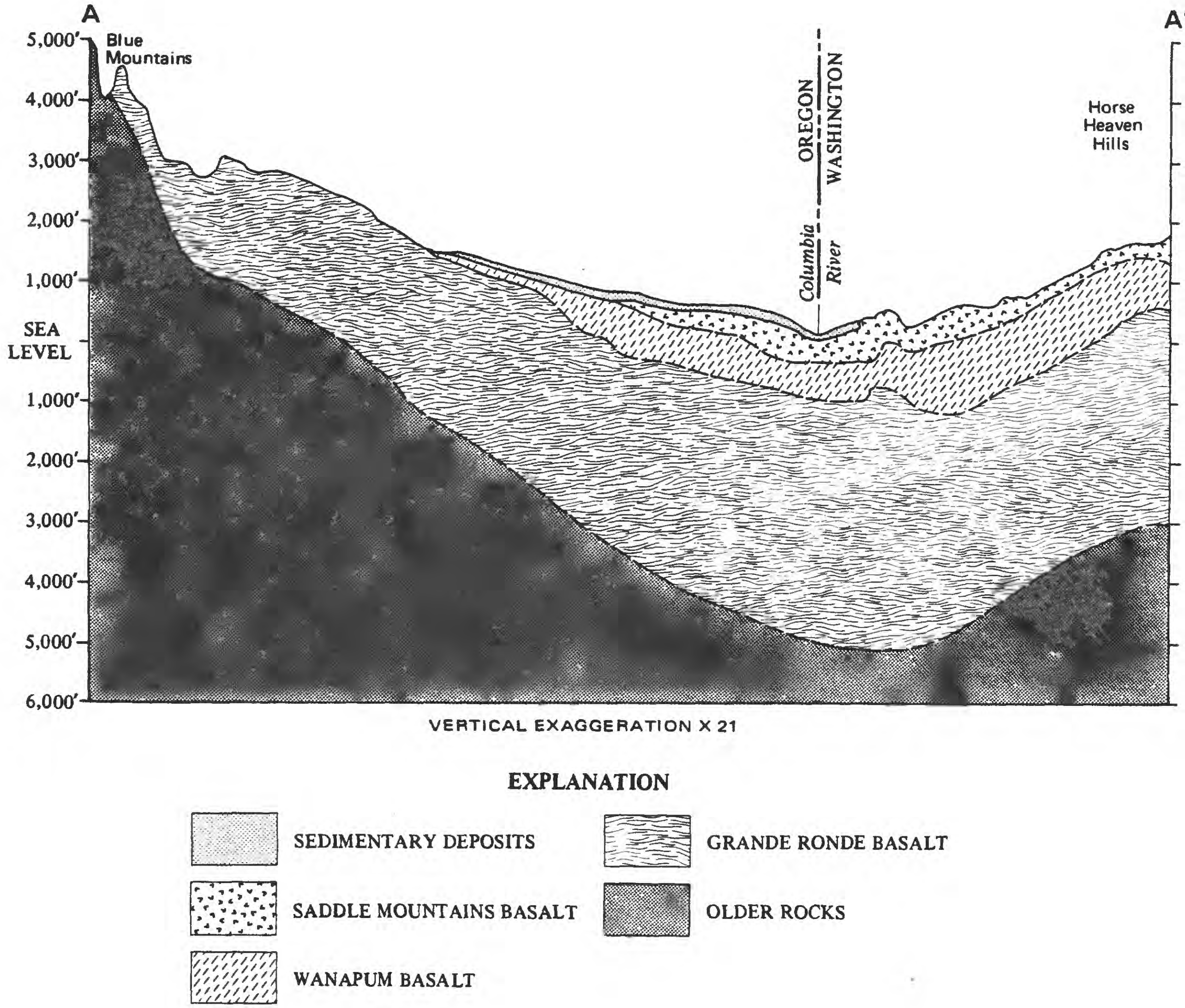


Figure 3.--Schematic diagram of geologic cross section A-A'.

According to proprietary magneto-telluric data, the basalt may be more than 10,000 feet thick in the Boardman area (Tom Hepner, Geotronics, Inc., Austin, Texas, oral commun., April 23, 1984). Sediments of glaciofluvial origin and alluvium overlie the basalt near and along the Columbia River. Loess covers much of the central part of the study area. The maximum thickness of the sediments is about 150 feet.

Individual basalt flows range from a few feet to as much as 300 feet thick. Characteristically, these flows have a dense center with vertical jointing and a scoriaceous or brecciated zone at the top and bottom (fig. 4). The upper surface may be weathered, and the base of the flow may show evidence of having been cooled in water. The porous zones between basalt flows are called interflow zones. They consist of weathered basalt, flow-top breccia, and scoria. Interflows may also contain sedimentary interbeds. Interbeds are sedimentary deposits, ranging from a few feet to more than 200 feet in thickness, that occur between two basalt flows. These interbeds are composed of clay, silt, and sand, with gravel layers in some places. The color of the interbeds ranges from red or brown to green or blue.

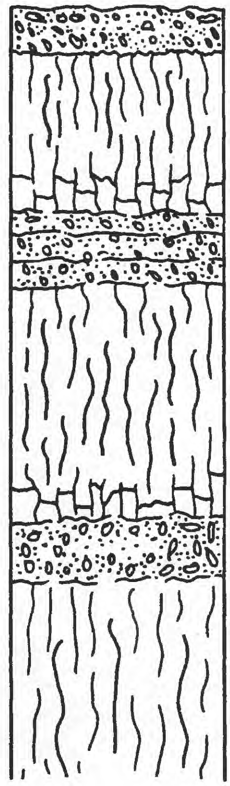


Figure 4.--Schematic diagram of basalt flows, interflow zones and interbeds.

During the period of deposition of the Columbia River Basalt Group, the volume of successive flows diminished and the time interval between flows increased. As a result, the thickness of sedimentary interbeds tends to be greater between younger basalt flows. Thus the interbeds are thickest and most significant in the Saddle Mountains Basalt and are least evident in the Grande Ronde Basalt. Sedimentary deposition was concentrated in the structural basins.

Stratigraphy

The oldest and most extensive basalt unit in the study area is the Grande Ronde Basalt (table 1). The upper two units of the Grande Ronde have been identified in surface mapping by Swanson and others (1981). More than 3,000 feet of the Grande Ronde are exposed in the Snake River Canyon. The thickness of the Grande Ronde in the study area is not known, but it thins southward to a feather edge against the Blue Mountains. It has recently been estimated to be more than 8,000 feet thick beneath the Columbia River.

For the purposes of this report, the Vantage member of the Miocene Ellensburg Formation is referred to as the Vantage interbed. It overlies the Grande Ronde Basalt and is a fairly continuous weathered interflow zone, associated in some places with a clay or sandstone interbed. Overlying the Vantage interbed is the Wanapum Basalt, represented in the study area by the Frenchman Springs Member, the Roza, and the Priest Rapids Member. The maximum thickness of the Frenchman Springs Member is approximately 600 feet, and it extends as far south as Pilot Rock, Oregon. The occurrence of Priest Rapids and Roza is limited to the northwestern part of the study area. Total thickness of Wanapum Basalt reaches about 1,000 feet near the Columbia River. Weathered interflows without sedimentary interbeds are common in the Wanapum and in the upper 200 to 400 feet of the Grande Ronde.

Table 1.--Stratigraphic column showing relation between formations and members of the Miocene Columbia River Basalt Group and major intercalated sedimentary units within the study area

	Formation	Members or Magneto-stratigraphic units	Sedimentary interbed
Holocene to Miocene	Loess, alluvial lacustrine, and glaciofluvitile sedimentary deposits		
Miocene	Saddle Mountains Basalt[1]		Rattlesnake Ridge
		Elephant Mountain	
		Pomona	
			Selah
		Umatilla	
			Mabton
	Wanapum Basalt[1]	Priest Rapids	
		Roza	
		Frenchman Springs	
			Vantage
	Grande Ronde Basalt[1]	N2	
		R2	

[1] The Saddle Mountains and Wanapum Basalts are included in the Yakima Basalt subgroup.

The Saddle Mountains Basalt overlies the Wanapum Basalt. Its maximum thickness in the study area is 800 feet in Washington at a location about 5 miles north of Boardman and the Columbia River. The Saddle Mountains Basalt is present only in the northern part of the study area, extending about 15 miles south of the Columbia River. From oldest to youngest, it consists of the Umatilla Member, the Pomona Member, and the Elephant Mountain Member (table 1).

In the study area, the Saddle Mountains Basalt has three major interbeds. The Rattlesnake Ridge is the interbed between the Elephant Mountain and Pomona Members; the Selah interbed is between the Pomona and Umatilla Members; and the Mabton interbed is between the Umatilla Member of the Saddle Mountains Basalt and the underlying Priest Rapids Member of the Wanapum Basalt. The sedimentary interbeds are members or informal units of the Ellensburg Formation (Schminke, 1964). The Rattlesnake Ridge is of limited extent in Oregon, but the Selah and Mabton occupy a 10- to 15-mile-wide strip south of the Columbia River. These major interbeds are commonly described by drillers as massive green or blue clay layers.

For a summary of the lithologic characteristics of the Columbia River Basalt Group units, see Swanson and others (1979).

Sedimentary deposits overlie the basalt along the Columbia River, their southern edge approximately coinciding with the southern edge of the Saddle Mountains Basalt (fig. 2). These deposits cover much of the eastern part of the Horse Heaven Hills area. Along the Columbia River, the sediments consist of recent alluvium and Pleistocene lacustrine silt and sand and of glaciofluvatile coarse sand and gravel (Hogenson, 1964). In addition, the higher ground throughout the area is blanketed with loess.

Structure

The basalt on both sides of the Columbia River dips gently toward the river (figs. 5, 6, and 7). This gentle downwarp is The Dalles-Umatilla syncline (Newcomb, 1967). Along most of the length of the syncline, the axis lies along the Columbia River and diverges toward the Umatilla River Valley near Hermiston (fig. 8).

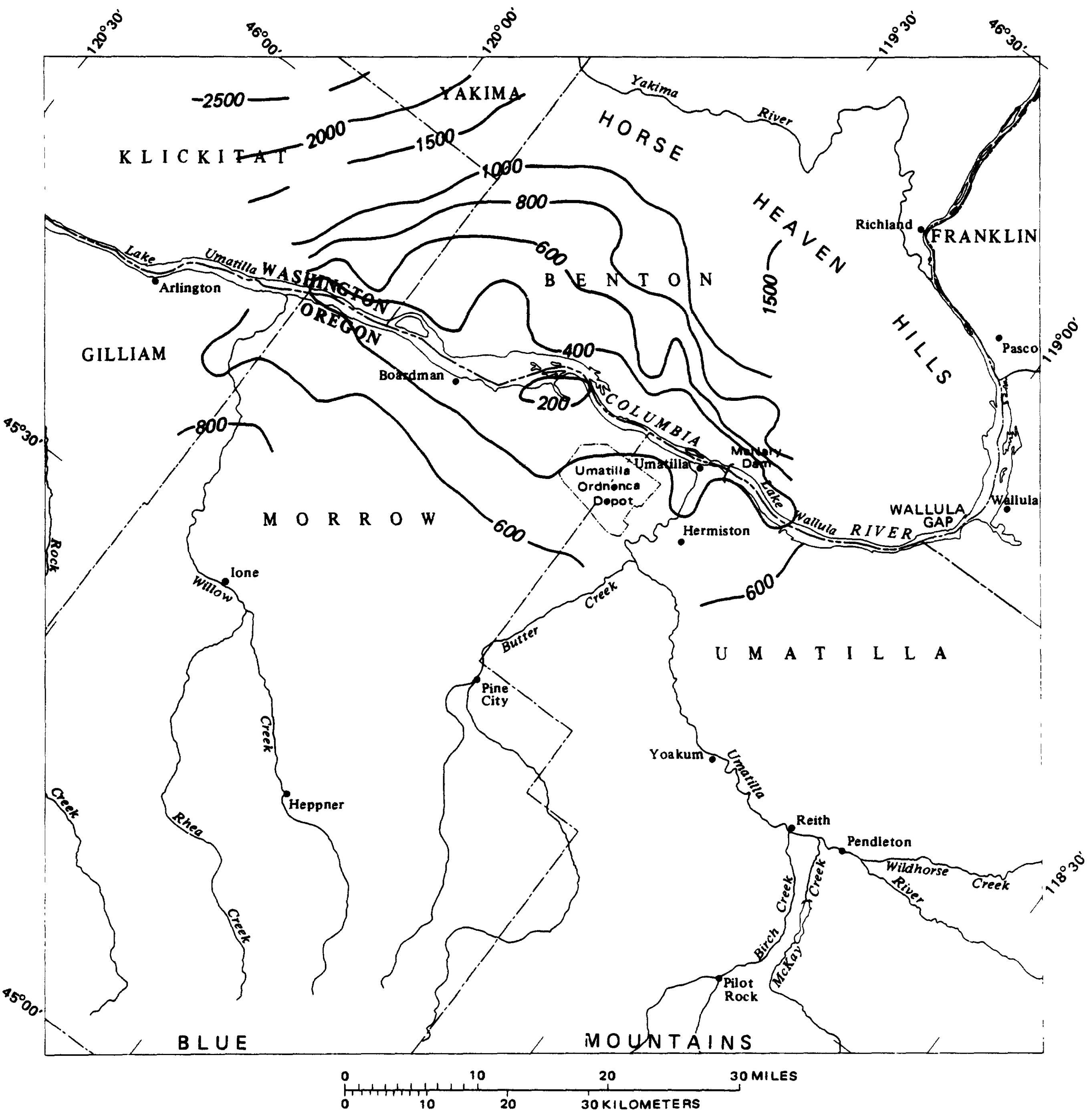
The study area is bounded on the south by the Blue Mountain anticline, which provides approximately 5,000 feet of total relief. The Blue Mountains have a core of Mesozoic intrusive rocks; these, along with Paleozoic and Tertiary sedimentary and volcanic rocks, are exposed along the crest of the Blue Mountains.

The Horse Heaven Hills lie along the north and east margins of the area. The Columbia River Basalt Group is continuous across the Horse Heaven Hills, but the rocks are steeply folded and faulted in the Horse Heaven Hills anticline. Near Richland, Washington, the structure has a sharp bend, which may actually be the intersection of two structural trends (Myers and others, 1979). The anticlinal trend continues southward across the Columbia River near Wallula Gap and into Oregon. The structure is flanked on the east by the Wallula Gap fault and the Walla Walla fault system.

Rock Creek, on the western edge of the area, coincides with a double bend in the Columbia River, perhaps indicating lateral right offset. Although short fault segments have been mapped along Rock Creek, no major structural feature has been defined.

Structural features within the study area include a series of anticline-syncline pairs. On the north side of the Columbia River, parallel to The Dalles-Umatilla syncline, a belt of anticlines forms the Columbia Hills and Patterson Ridge. In Oregon, the broad, arcuate folds of the Agency syncline and Rieth anticline follow the trend of Blue Mountains. Service anticline is a tight anticline-syncline pair, faulted at some points along its length. It swings westward, as marked by thrust faulting and some normal faults, and extends as far west as Rock Creek (Swanson and others, 1981). Northwest-southeast trending structures include the Arlington-Shutler Butte lineament.

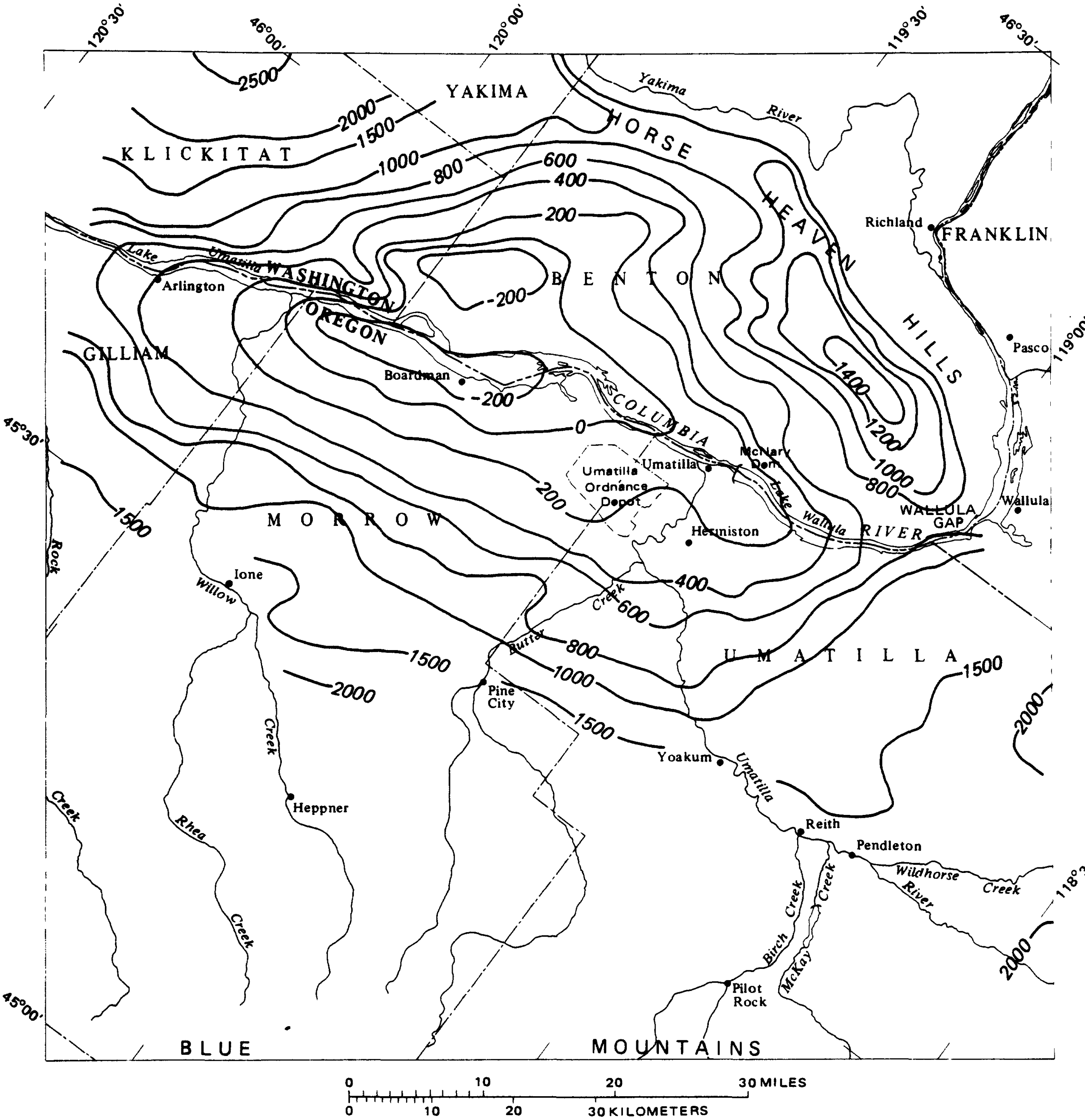
An arcuate topographic feature which extends east-west from south of Arlington, Oregon, to north of Pine City, Oregon, has been called the Willow Creek monocline (Bela, 1982; Shannon and Wilson, Inc., 1973) or the Willow Creek lineament (Oberlander, Oregon Water Resources Department, oral commun., 1980). This feature apparently is a hydrologic barrier and will be discussed in the next section.



EXPLANATION

——400—— STRUCTURE CONTOUR–Shows altitude of top of Saddle Mountains Basalt. Contour interval 200 and 500 feet. Datum is sea level.

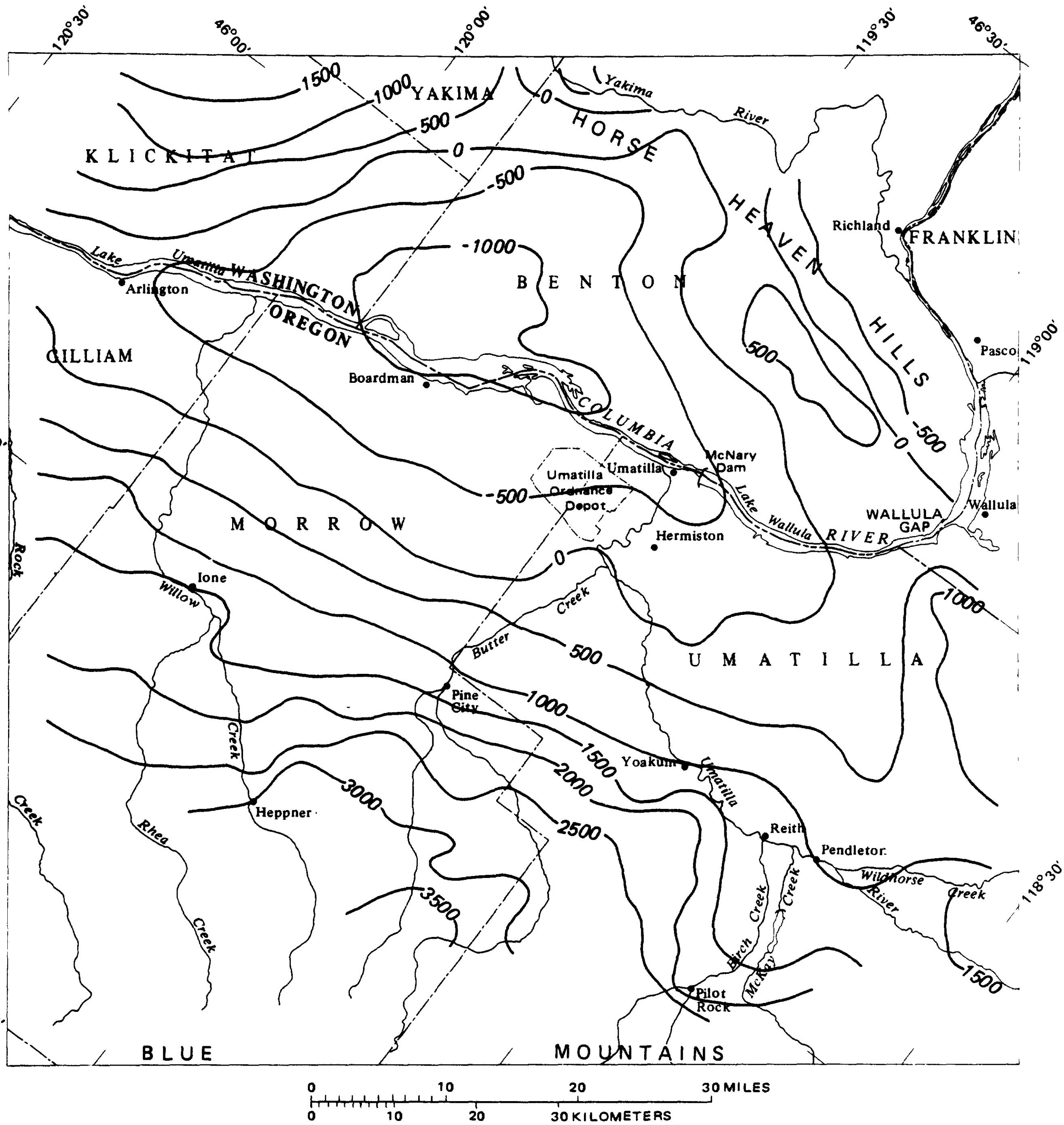
Figure 5.--Structure of top of Saddle Mountains Basalt.



EXPLANATION

—— 200 —— STRUCTURE CONTOUR–Shows altitude of top of Wanapum Basalt. Contour interval 200 and 500 feet. Datum is sea level.

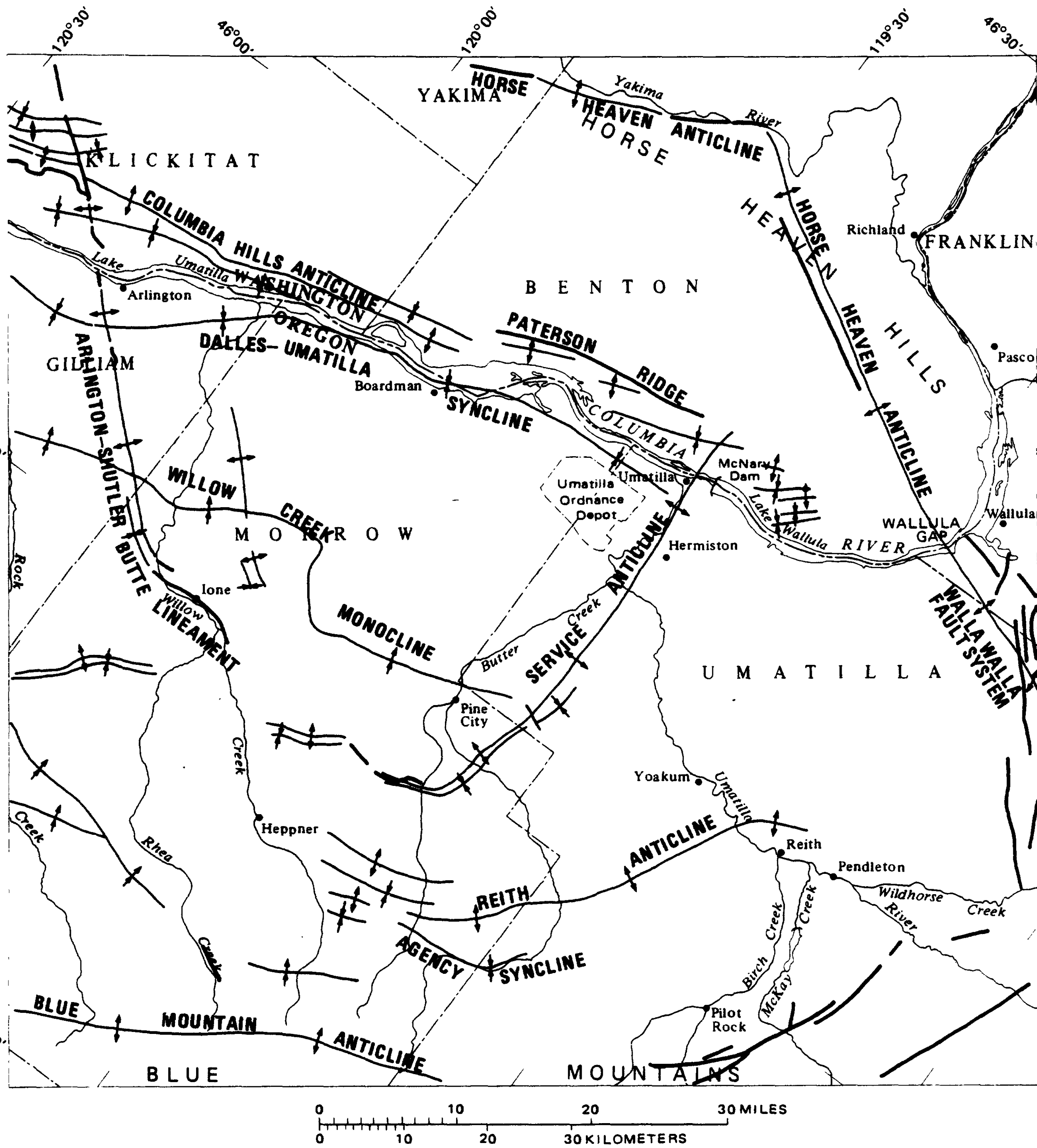
Figure 6.--Structure of top of Wanapum Basalt.



EXPLANATION

——1000—— STRUCTURE CONTOUR–Shows altitude of top of Grande Ronde Basalt. Contour interval 500 feet. Datum is sea level.

Figure 7.--Structure of top of Grande Ronde Basalt.



EXPLANATION

ANTICLINE — MONOCLINE

SYNCLINE — FAULT

Figure 8.--Major structural features. Geology adapted from Swanson and others (1979 and 1981), Bela (1982), Shannon and Wilson (1973), and Newcomb (1967).

Hydrologic evidence suggests that the monocline extends downward into the Grande Ronde Basalt. It is possible that the Willow Creek monocline is the surface expression of a fault with north side downthrown. Numerous basalt flows, notably the more recent flows of the Saddle Mountains Basalt, terminated or thinned out across this feature, resulting in disruption of the continuity of interflow zones across it. The nature of this disruption could be the actual truncation of interflow zones or a draping and stretching of flows across the structure. The progressive downwarping north of the lineament could have been caused by the subsidence of the central part of the Columbia Plateau, which began prior to Wanapum time and continued into late Saddle Mountains time (Swanson and others, 1979).

HYDROLOGY

Ground-water Reservoir

Ground water in the Columbia River Basalt Group moves through a complex system of geohydrologic units (aquifers) that are poorly connected in the vertical direction. For several reasons the system should be viewed as a whole. First, all of the basalt geohydrologic units are confined or semiconfined, allowing changes in pressure in the system to have rapid and far-reaching effects, as shown by the recent water-level declines caused by pumping. Second, although the vertical connections are poor, vertical movement of water is significant, particularly when induced by pumping. Third, numerous uncased wells have locally increased vertical movement of ground water between interflow zones and, over time, head differences between zones have decreased.

Short-term performance of wells is controlled by localized properties of the geohydrologic units and rates of flow through the units. However, it is important to view ground-water withdrawal in relation to natural recharge and discharge of the system. The ultimate amount of water that can be withdrawn depends not only on the storage and transmissive properties of the units and the natural recharge but also on the amount of natural discharge that can be diverted to the pumping wells and the amount of additional infiltration that can be induced from streams. Over a long-term period, ground-water withdrawals will affect surface-water supplies to some degree.

Geologic structure plays a major role in the ground-water system. Two aspects in particular affect ground-water development. One is the stratification of basalt flows; individual interflow zones may be rather extensive horizontally, but commonly are largely isolated from overlying and underlying interflows by poorly permeable basalt flow centers. In practical terms, this stratification can mean that adjacent wells penetrating different depths may have entirely different water levels and performance. A second important effect of structure is the "barrier" or flow impediment effect of stratigraphic pinchouts and offsets. These so-called barriers are not entirely impermeable to ground water, but they do isolate wells and retard ground-water movement to the downgradient side.

Boundaries

The natural geohydrologic boundaries of the flow system are the Blue Mountains on the southeast and the Horse Heaven anticline on the north. The southwestern boundary is less clearly defined, but Rock Creek, a tributary of the John Day River, is incised deeply into the Grande Ronde Basalt and was selected as a probable geohydrologic boundary for the purposes of this study.

For the most part, the Grande Ronde Basalt pinches out at the Blue Mountains anticline on the south edge of the study area (fig. 8). This broadly uplifted area is bounded on the north by the Blue Mountains fault and describes a sweeping curve parallel to and south of the Rieth anticline and the Agency syncline. Rocks predating the Columbia River Basalt Group are exposed in many places in its core but are considered to be insignificant for the purposes of this study.

The Horse Heaven Hills, which form both the northern and eastern boundaries of the study area, are a sharply folded and faulted anticline. High hydraulic heads and a reversal of gradient in the Horse Heaven Hills indicate a ground-water divide separating southward flow toward the Columbia River from northward and eastward flow into the Yakima Basin and the Pasco Basin (Frank Packard, U.S. Geological Survey, written commun., 1983; Gephart and others, 1979, plates 11-9 and 11-10). West of Pasco, Washington, the trend of the Horse Heaven anticline changes from approximately east-west to northwest-southeast. It continues as a well-defined structural and topographic feature across the Columbia River into Oregon. In Oregon, as in Washington, the high hydraulic heads along its fault-disrupted crest indicate that it is a major ground-water divide.

Ground-water flow directions in the basalts near Rock Creek are generally normal to the Columbia River. Rock Creek was chosen as the western boundary because (1) it is aligned nearly parallel to the general direction of ground-water flow thus forming a probable no-flow boundary, (2) it is distant from the major area of ground-water declines so pumping is not likely to induce flow across it, and (3) it has eroded deeply into the Grande Ronde Basalt, thus severing most interflow zones that transmit water. Additionally, Rock Creek is aligned with a sharp bend of the Columbia River; the bend shows evidence of strike-slip faulting (Swanson and others, 1981). The trend continues northwestward into Washington as a series of faults that offset the Columbia Hills. This faulting may alter the natural ground-water flow and has been included as a probable boundary.

Delineation and Hydrologic Properties of Geohydrologic Units

The interflow zones in the basalts are generally the more permeable units, whereas the basalt centers and interbeds generally represent less permeable units. A detailed representation of the ground-water flow system would require definition of the spatial distribution of each of these units--a definition that is not presently possible from the available data. For the purposes of this study, the basalts and interbeds were divided into three major geohydrologic units (aquifers) on the basis of geologic data and ground-water levels. Additionally, the sedimentary deposits that overlie the basalts near the Hermiston-Umatilla area comprise a fourth unit.

Apparent transmissivity values for each of the four units were computed from about 1,700 short-duration specific capacity tests, using Brown's (1963) extension of the Theis equation. These apparent values were divided by the total of saturated thickness open to the well, in order to obtain values of hydraulic conductivity. Most of the wells that are completed in the Wanapum Basalt also penetrate the Saddle Mountains Basalt; thus hydraulic conductivity for these units is a composite value. Mean values of hydraulic conductivity, as determined from well data where the well is completed only in each unit, are 0.28 ft/s (feet per second) for the sediments overlying the basalt, 0.00021 ft/s for the Saddle Mountain Basalt, 0.00197 ft/s for the Wanapum Basalt, and 0.00075 ft/s for the Grande Ronde Basalt. These values are comparable to those determined from aquifer tests conducted by OWRD (Oregon Water Resources Department, written commun., 1981).

Measurements of the vertical hydraulic conductivity of the geohydrologic units are not available. MacNish and Barker (1976) report that vertical hydraulic conductivity in the Walla Walla River basin is about 5×10^{-8} ft/s. Tanaka and others, in Gephart and others (1979), estimated a considerably smaller value of 2×10^{-10} ft/s. Where vertical hydraulic conductivity values are this low, it is possible that interconnection of aquifers by open boreholes could overshadow the effects of natural leakage between units. Such changes may have occurred in some heavily developed parts of the Umatilla Plateau.

For this study, vertical hydraulic conductivity could not be directly calculated. A relation developed from the Horse Heaven Hills study in Washington in conjunction with a cross-section model in Oregon was used for an initial estimate of vertical hydraulic conductivity. The relation between the vertical and lateral hydraulic conductivity, and the thicknesses of upper and lower units is described by McDonald and Harbaugh (1984) and modified for the Horse Heaven Hills study is as follows:

$$V_c = \left[\frac{2 \times FACT}{\frac{b_1}{K_{L_1}} + \frac{b_2}{K_{L_2}}} \right]$$

where

V_c = vertical hydraulic conductivity divided by thickness $(\frac{1}{T})$ between two units;

b_1, b_2 = thickness (L) of upper and lower units;

K_{L_1}, K_{L_2} = lateral hydraulic conductivity $(\frac{L}{T})$ of upper and lower units;

FACT = ratio of vertical to lateral hydraulic conductivity $(\frac{K_v}{K_L})$;

The empirical value developed for $\frac{K_v}{K_L}$ from the Horse Heaven Hills data for basalt was 0.003. This value, along with lateral hydraulic conductivity, and thickness, were used to derive values of V_c for basalt units.

In order to determine whether or not these values were reasonable, a separate determination of V_c for the low hydraulic conductivity material was calculated by constructing a cross-section flow model in the Oregon part of the study area. The model was constructed so that each unit (Saddle Mountain, Wanapum, and Grande Ronde) was simulated separately from intervening low hydraulic conductivity material (clay). The hydraulic conductivity of the three units was held constant, while the hydraulic conductivity of the intervening beds was varied during model calibration. The hydraulic head in the Saddle Mountains and Grande Ronde layers was held constant. The model calculated the head in the Wanapum layer and this calculation of head was compared with observed heads in the Wanapum. The hydraulic conductivity of the intervening beds was adjusted until the calculated heads in the Wanapum approximated the observed heads.

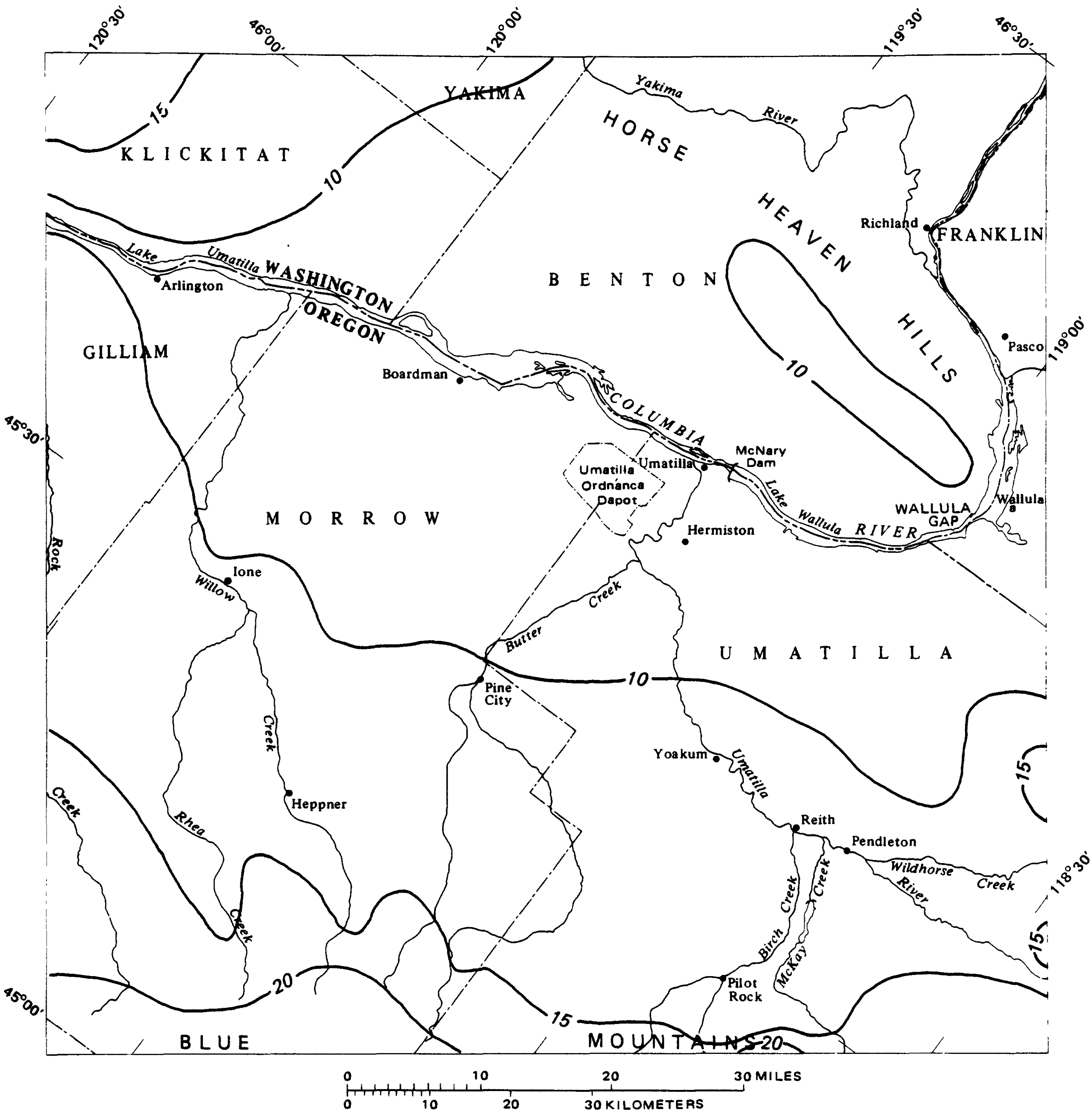
Results of this cross-section model analysis, although preliminary in nature, indicate that values of V_c were reasonable when compared with results of the Horse Heaven Hills study. The values of V_c, as calculated from the cross-section model, ranged from 1.0×10^{-13} to 6.0×10^{-15} Sec^{-1}, whereas V_c values from the Horse Heaven Hills study ranged from 1.0×10^{-10} to 1.0×10^{-15} Sec^{-1}.

Thus, V_c values as calculated from the empirical relation defined above was used as initial input data to the three dimensional model. The model will be discussed in another section of this report.

Storage coefficients calculated from aquifer tests in the study area ranged from 0.00001 to 0.0015 (Oregon Water Resources Department, written commun., 1981). Gephart and others (1979) report storage coefficients for the Columbia River Basalt Group that range from 0.00003 to 0.0012 for 10 test sites. Luzier and Burt (1974) computed storage coefficients of 0.0025 and 0.0065, typical of leaky artesian conditions, for the basalt aquifer in the Odessa area of Washington, using a volumetric analysis. The average storage coefficient derived from MacNish and Barker's (1976) model of the basalt aquifer in the Walla Walla River basin was 0.00046. In this study, the values derived during the Horse Heaven Hills study were used as estimates for specific yield and storage-coefficient: 0.01 for the Saddle Mountain Basalt and 0.003 for the Wanapum and Grande Ronde Basalts. A value of 0.15 was used for the sediments overlying the basalt.

Recharge

The major source of recharge to the geohydrologic units is precipitation that infiltrates the ground, especially in highland areas where interflows are exposed at land surface. Recharge rates in the study area are directly related to the amount of precipitation. Most of the recharge from precipitation occurs during the winter and spring. Precipitation on the Umatilla Plateau (fig. 9) ranges from about 8 inches per year near the Columbia River to more than 20 inches per year in the Blue Mountains (Johnsgaard, 1963). Potential evaporation rates greatly exceed annual precipitation. However, evapotranspiration is largest during the summer and precipitation occurs mainly during the winter. A study of historical weather data by Johnsgaard (1963) showed that stations in the study area that are above an altitude of about 1,400 feet have an average of 5 months in which precipitation exceeds evapotranspiration. This excess precipitation ranges from 4.4 inches to as much as 7.0 inches. Areas below 1,400 feet have excess precipitation for only 4 months, with surpluses ranging from 2.9 to 5.9 inches.



EXPLANATION

——10—— LINE OF EQUAL AVERAGE ANNUAL PRECIPITATION, 1930-57–Interval 5 inches. (Data from U.S. Weather Bureau, published by U.S. Soil Conservation Service, 1964).

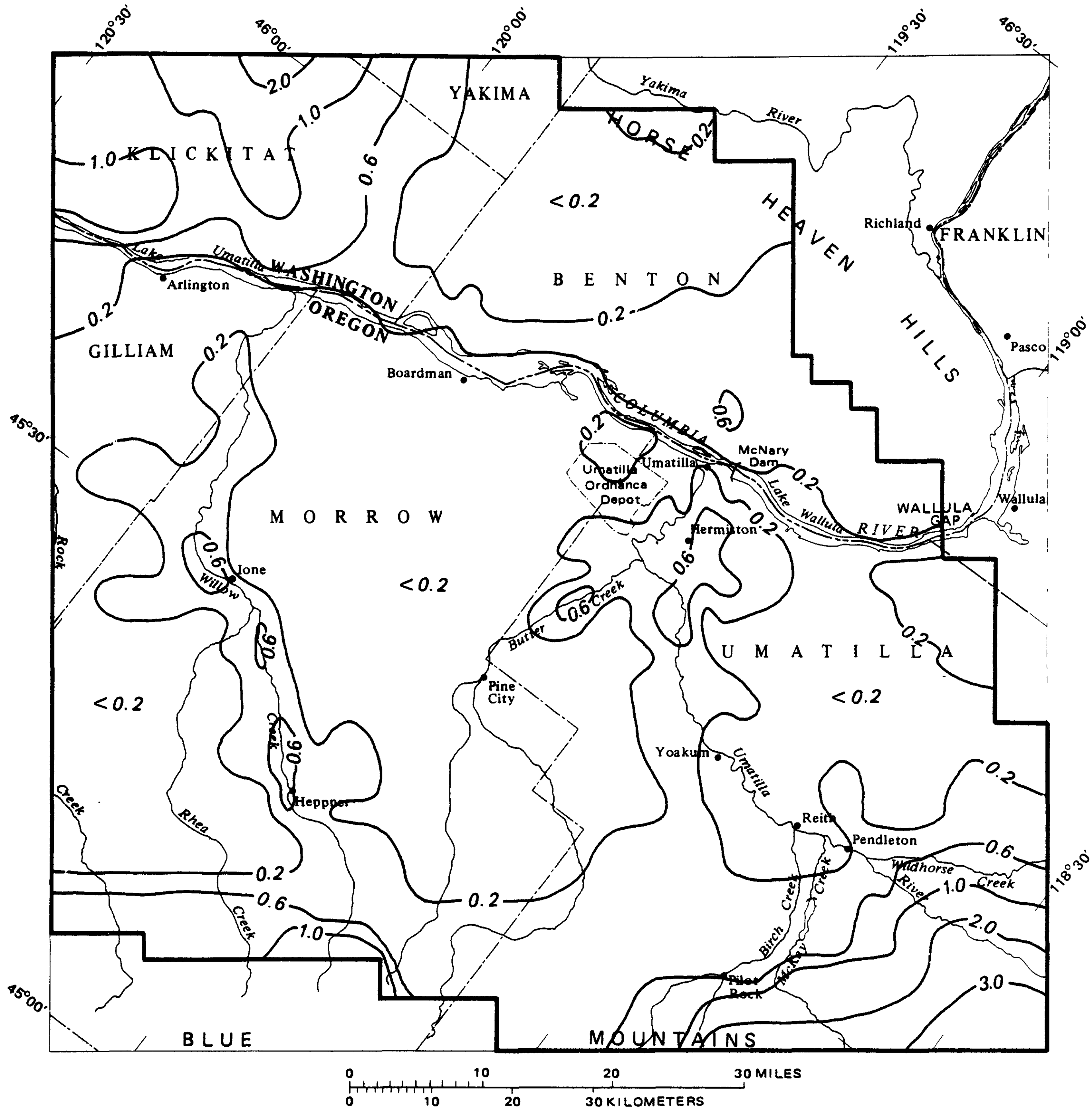
Figure 9.--Average annual precipitation.

Another source of recharge to the geohydrologic units is leakage from streams. Most streams in the region gain water from the ground-water system in their upper reaches and some lose water to the ground-water system farther downstream. Many small streams are ephemeral in their lower reaches. Some of their water evaporates and some recharges the sediments overlying the basalt. Because the entire basalt sequence is tilted and beveled off, the stream channels cross successively younger interflow zones in a downstream direction; water moves into and out of the basalt in these zones in accordance with the relation between aquifer heads and stream-surface altitude. Water levels in some wells close to the Columbia River have risen as a result of the filling of Lake Umatilla behind the John Day Dam. For example, the hydrograph for a City of Arlington well illustrated in Smith, Collins, and Olson (1983, p. 36), shows a rise of 62 feet over the 15-year period following the filling of the lake.

In some areas, irrigation water may percolate downward to the underlying basalt. Although head buildups from this source have not been observed in the basalt units in Oregon, head buildups in the basalt in the eastern part of the Horse Heaven Hills area in Washington have occurred.

Recharge from precipitation cannot be measured directly under normal field conditions because it is governed by a complex interplay of precipitation, temperature, evaporation, solar radiation, plant cover, soil characteristics, and soil-moisture conditions. However, recharge was estimated for the study area by use of a method developed by Bauer and Vaccaro (1987) for the Regional Aquifer System Analysis (RASA) of the Columbia Plateau. That method, according to Bauer and Vaccaro, "computes recharge from precipitation, evaporation, and stream runoff and consists of simplified physical-process submodels that allow a user to determine which components of the hydrologic system are important....The minimum data sets are daily precipitation, daily maximum and minimum temperatures, available soil-water capacities, soil thickness, soil texture, and land use." Daily stream discharge and monthly estimates of ground-water discharge to streams are used where the data are available. Model-computed surface-runoff values are used if data are not available, but with less reliable results. Although the accuracy of the method is unknown, its use is acceptable because it relies on empirical techniques for results and because most data, such as streamflow, precipitation, temperature, soil type and thickness, and land use, can be measured directly. For a complete description of the method the reader is referred to Bauer and Vaccaro (1987).

The above method was applied using the Umatilla Plateau and Horse Heavens Hills area data for the period 1956 to 1977. With this method, average recharge from precipitation and from application of irrigation water from surface sources varies from more than 3 inches annually in the Blue Mountains to less than 0.2 inches annually over large parts of the study area (fig. 10). The distribution of recharge shown in figure 10 includes the effects of long-term recharge from surface-water irrigation near the Umatilla River, Butter Creek, and Willow Creek. These areas have been irrigated annually since the mid-to-late 19th Century, and the recharge from this source is assumed to be nearly constant over a long period of time. Estimates of the amount of surface water applied to the land surface for irrigation were made from a report by Simons (1953), who calculated streamflow depletion due to irrigation for the streams noted above. If the application of surface water had not occurred, the long-term recharge from precipitation for areas near the Umatilla River and Butter and Willow Creeks would be less than 0.2 inches annually.



EXPLANATION

——*1.0*—— LINE OF EQUAL AVERAGE ANNUAL RECHARGE, 1956-77–Interval 0.4 and 1.0 inches.

▬▬▬ MODEL BOUNDARY

Figure 10.--Average annual recharge.

Ground-water Movement

Ground water in the Umatilla Plateau and the Horse Heaven Hills area flows toward the Columbia River from recharge areas in adjacent highlands--the Blue Mountains and the Horse Heaven Hills. On the basis of limited data, the shapes of the potentiometric surfaces for each geohydrologic unit subparallel the surface topography. Wells penetrating the four geohydrologic units are scattered throughout the area and do not provide adequate control for detailed mapping of the potentiometric surfaces. In addition, wells solely supplied by the Grande Ronde Basalt are nonexistent in the northern part of the study area. Because of these inadequacies and limitations, water-level contour maps are not shown.

At any given location, both horizontal and vertical movement of ground water occurs in the basalts. In general, horizontal movement predominates in the interflow zones, whereas vertical movement of water is believed to predominate in the basalt flow centers, because of vertical jointing, and also in the interbeds. Large differences in water levels across flow centers, combined with the great areal extent of the basalts, can result in a considerable quantity of water moving vertically. Information on head changes in wells comes from those drillers' logs that have notations of water levels at various drilling depths and, in some cases, from well-deepening records. Where there are no measurements of heads in the Grande Ronde Basalt, these vertical differences are used to make estimates of vertical head differences. The amount of vertical head change is added to or subtracted from the head in each unit as appropriate to provide points for interpolating the head surface. Lacking additional information, heads in the Grande Ronde Basalt are assumed to be the same as those in the Wanapum Basalt. This assumption has the effect of minimizing exchange of water between these two units.

The Willow Creek monocline forms an impediment to flow, especially in the Grande Ronde Basalt. The head gradient across Willow Creek monocline and its role as an impediment to ground-water flow were first noted by P. L. Oberlander and D. W. Miller (Oregon Water Resources Department, written commun., May 1981). Water passes through slowly, as indicated by a gradient as steep as 400 feet per mile. (Head gradient south of the monocline is about 100 to 150 feet per mile; to the north of the monocline it flattens to approximately 8 feet per mile.) It appears that either the continuity of interflow zones is interrupted across this feature or that the transmissivity of the zones is severely diminished. In either case, this type of feature--that impedes ground-water flow--is referred to in this report as a barrier; it should be understood that only a partial blocking of flow is implied by this term. Frank Packard (U.S. Geological Survey, oral commun., 1983) has suggested that the slow rate of flow through such barriers encourages the deposition of zeolites and other secondary minerals in the pore spaces of the rock, further diminishing the flow through the barrier.

The magnitude and direction of vertical head changes in the Grande Ronde Basalt are as much as 380 feet over a 56-foot vertical interval. In general, heads decrease with depth in the upland recharge areas and north of the Willow Creek monocline, indicating downward movement of water in these areas. Upward movement is indicated by increasing heads with depth and flowing wells. These are common in the major stream valleys near the Columbia River and immediately to the south of the Willow Creek monocline. Heads in wells penetrating the Saddle Mountains Basalt commonly decrease with depth.

The "damming" of water on the upgradient side of the Willow Creek monocline causes heads to increase with depth and causes ground water to move upward. This is an explanation for an unusually large number of flowing wells on the south side of the monocline. On the north side (or downgradient side) heads decrease with depth.

Carbon-14 dates of water from 20 wells in the study area range from approximately 1,500 years before present to almost 40,000 years before present. More than half of the samples were collected by Oregon Water Resources Department personnel and the remainder, with two exceptions, were collected and processed by the Geological Survey. The average age of the samples was 16,500 years. (Carbon-14 age indicates the length of time since the water was exposed to atmospheric carbon dioxide.) The dates showed no clear relation to well depth or unit, probably because of intermixing of older water with younger water within the well bore in some of the wells, or perhaps intermixing of well water with streamflow or seepage from surface water applied irrigation water.

Discharge

Natural

Natural discharge of the geohydrologic units is principally to streams. Discharge from the shallow basalt also occurs to tributary streams near their headwaters, although these streams lose water to the units in their lower reaches. The regional discharge area for all units is believed to be the Columbia River.

Discharge of ground water by evapotranspiration is small, except perhaps in some of the wide valley bottoms where ground water is close to land surface. In some areas the uppermost unit may also lose water to evapotranspiration. These evapotranspiration losses, however, were considered to have only negligible effects on ground-water discharge.

Pumpage

Pumpage data for the Washington part of the study area were obtained from results of the Horse Heaven Hills study (Frank A. Packard, U.S. Geological Survey, written commun., 1986). The sources of information on ground-water pumpage in the Oregon part of the study area are as follows:

(1) Flowmeters have been installed on most large capacity wells in the areas under study by OWRD. Cumulative totals are recorded annually by the State Watermaster and other OWRD and Geological Survey personnel who inspect the wells. Flowmeter data are available from the late 1970's to the present, but data for the late 1970's are sparse. The flowmeters are the best source of information. For the metered wells, a statistical relation was developed between power consumption and volume of water pumped. This relation was used to estimate pumpage from the power records of unmetered wells.

(2) Some municipal and industrial-commercial users provided data on their annual and (or) monthly water use. Other municipalities provided information on number of connections and growth trends.

(3) Power records obtained from Pacific Power and Light, Columbia Basin Electrical Cooperative, and Umatilla Electric Cooperative Association give monthly power usage. Well inspectors record power-meter readings and, when opportunity permits, the additional information needed to compute well efficiencies. Power records are available for 1976 through 1982 for approximately 30 percent of the irrigation wells in the area.

(4) Information on irrigated acreage was obtained from examination of LANDSAT imagery and Geological Survey 7-1/2-minute orthophoto quads, from the OWRD water-rights records, from information provided by the Corps of Engineers during their cooperative study with the Geological Survey EROS Data Center, the Columbia River and tributaries irrigation-withdrawals analysis project (Johnson and others, 1981), and from data provided by the ERSAL at Oregon State University. The irrigated-acreage information was used in conjunction with flowmeter and power records to obtain average application rates and kilowatt-hour per acre-foot factors to estimate water use for wells with minimal data.

Using the information as given above, pumpage was estimated for each of the four geohydrologic units for the period 1950 to 1982. The results in acre-feet are shown in figure 11. Pumpage for wells completed in more than one unit, where specific data was lacking, was considered proportional to the thickness of each unit penetrated by uncased well bores. Because flowmeter data are considered to be the most accurate data for calculating pumpage and because flowmeters have been installed on most large-capacity wells in the study area, the calculations of pumpage are believed to be accurate to ±15 percent.

Pumpage from industrial, irrigation, and municipal wells in the study area are included in figure 11; specifically, any site known to have a flowmeter (10 acres or more irrigated) or a pump of over 10 horsepower was included. Of the 332 wells with known horsepower ratings, 140 had pumps rated at 100 horsepower or greater. Domestic use of ground water from individual small-diameter wells is not included in figure 11. Figure 11a shows pumpage for each State (Oregon and Washington) and figure 11b shows pumpage delineated for each unit. The areal distribution of pumpage is discussed in the model section of this report.

Water-level Changes

The locations of wells are shown in figure 12, and figure 13 shows the hydrographs of water levels in the wells that were selected to depict typical water-level changes that have occurred in each of the four geohydrologic units of the study area during the period 1953 to 1982. Although the hydrographs show both seasonal and long-term changes, in this study only the long-term changes are discussed. The long-term changes in water levels are due principally to the effects of man's activities--ground-water pumping and recharge from surface water.

In the overlying sedimentary unit, well G1 (an irrigation well west of Hermiston) shows a decline of about 30 feet from 1961 to 1974 and then a rise of about 20 feet from 1976 to 1982. The decline is due to pumpage and the rise may be due to recharge from surface water that is diverted from the Umatilla River. These diversions began in 1976 and are routed by pipeline and finally by open channel to the vicinity of well G1.

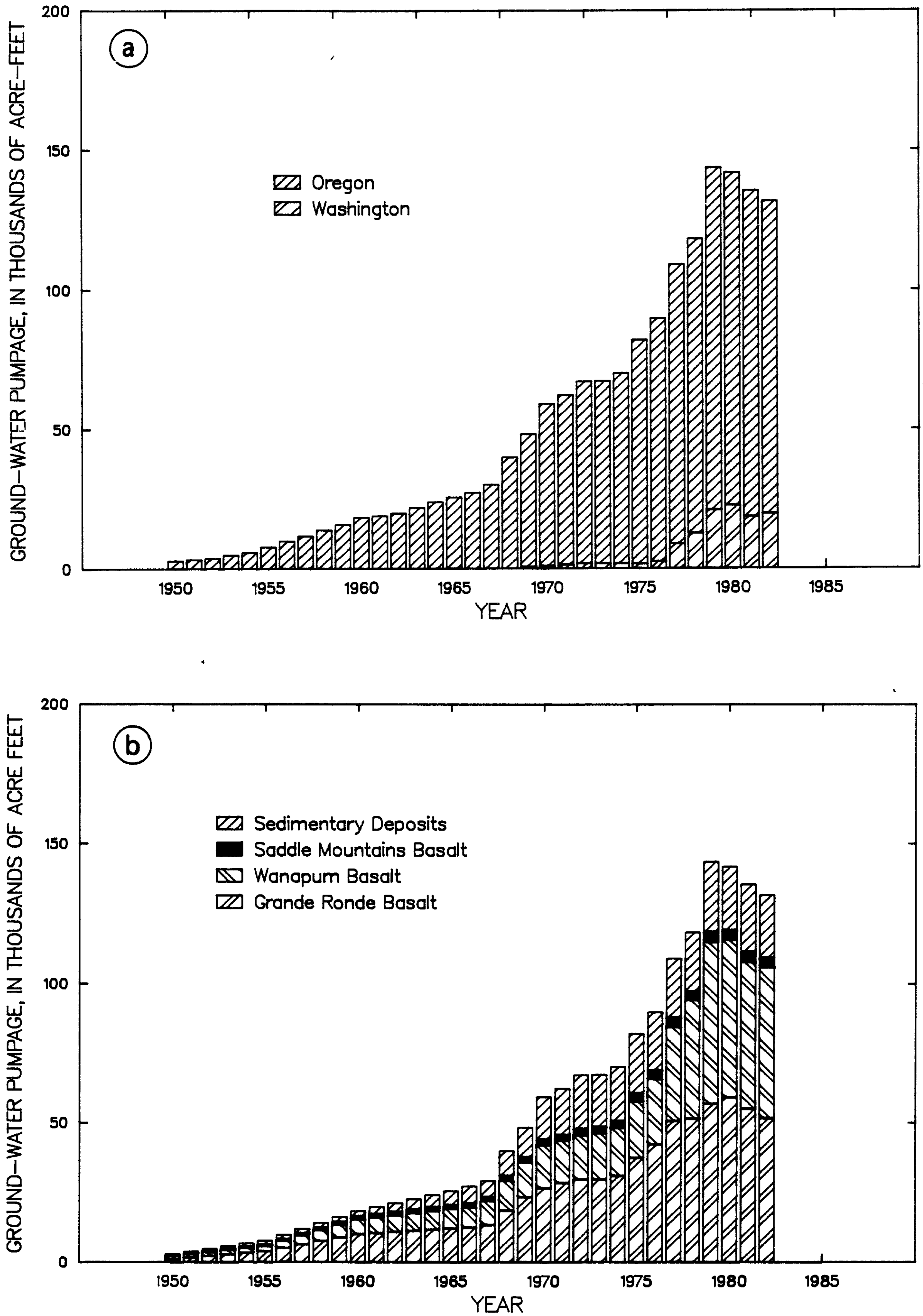


Figure 11.--Ground-water pumpage from irrigation, industrial, and municipal wells in the study area (a) delineated by state (b) delineated by geohydrologic unit.

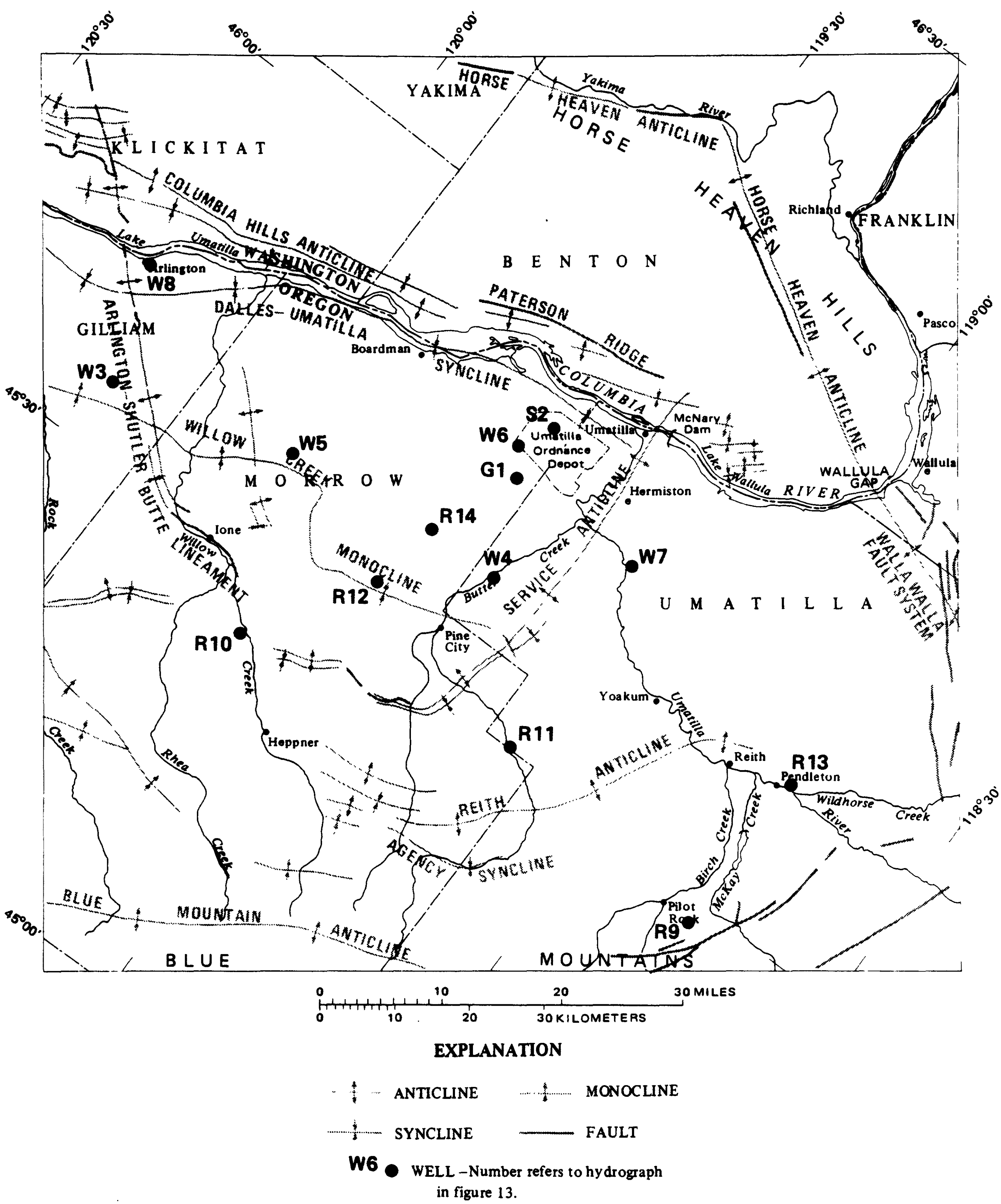


Figure 12.--Location of selected wells.

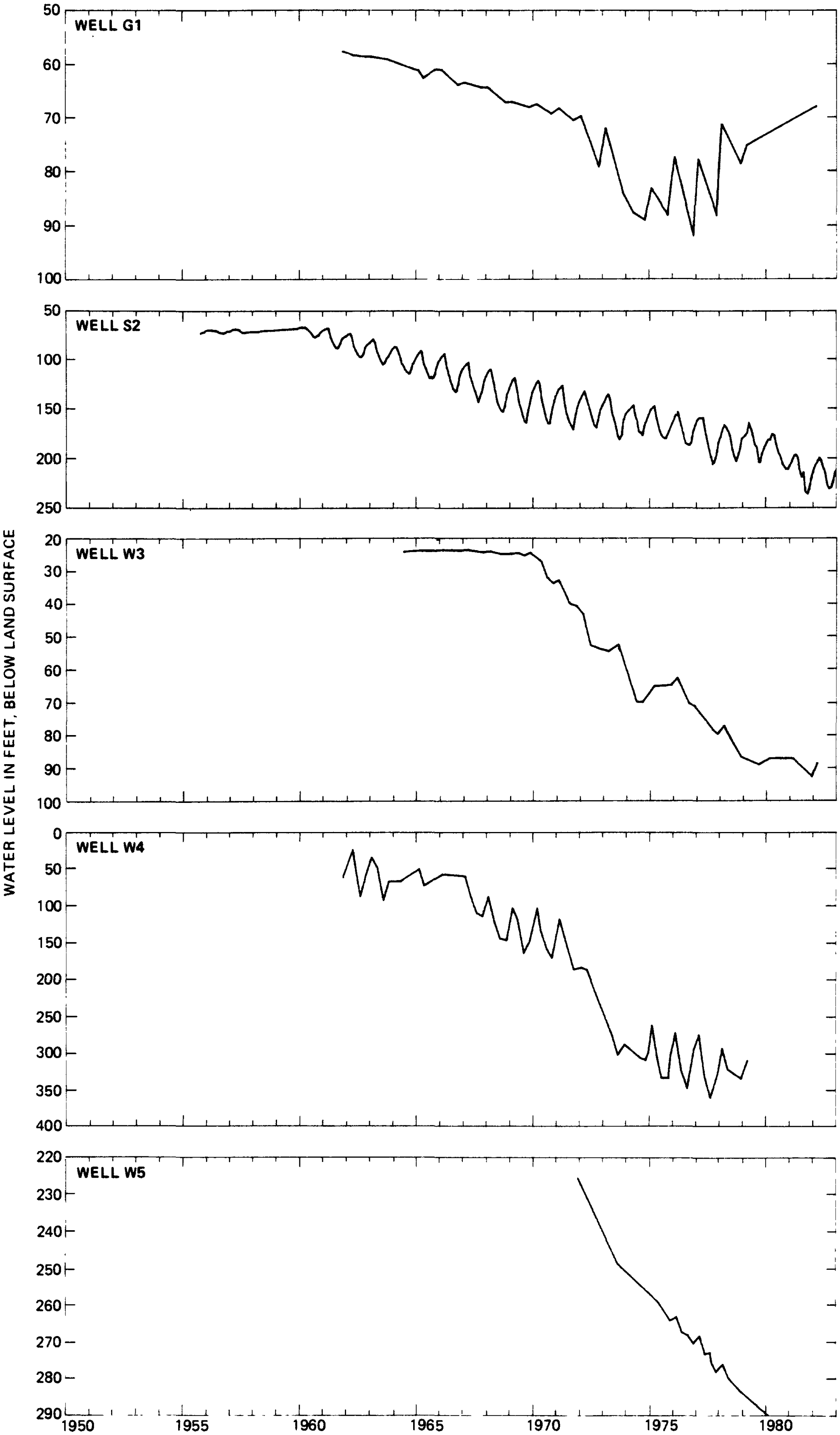


Figure 13.--Water levels in selected wells.

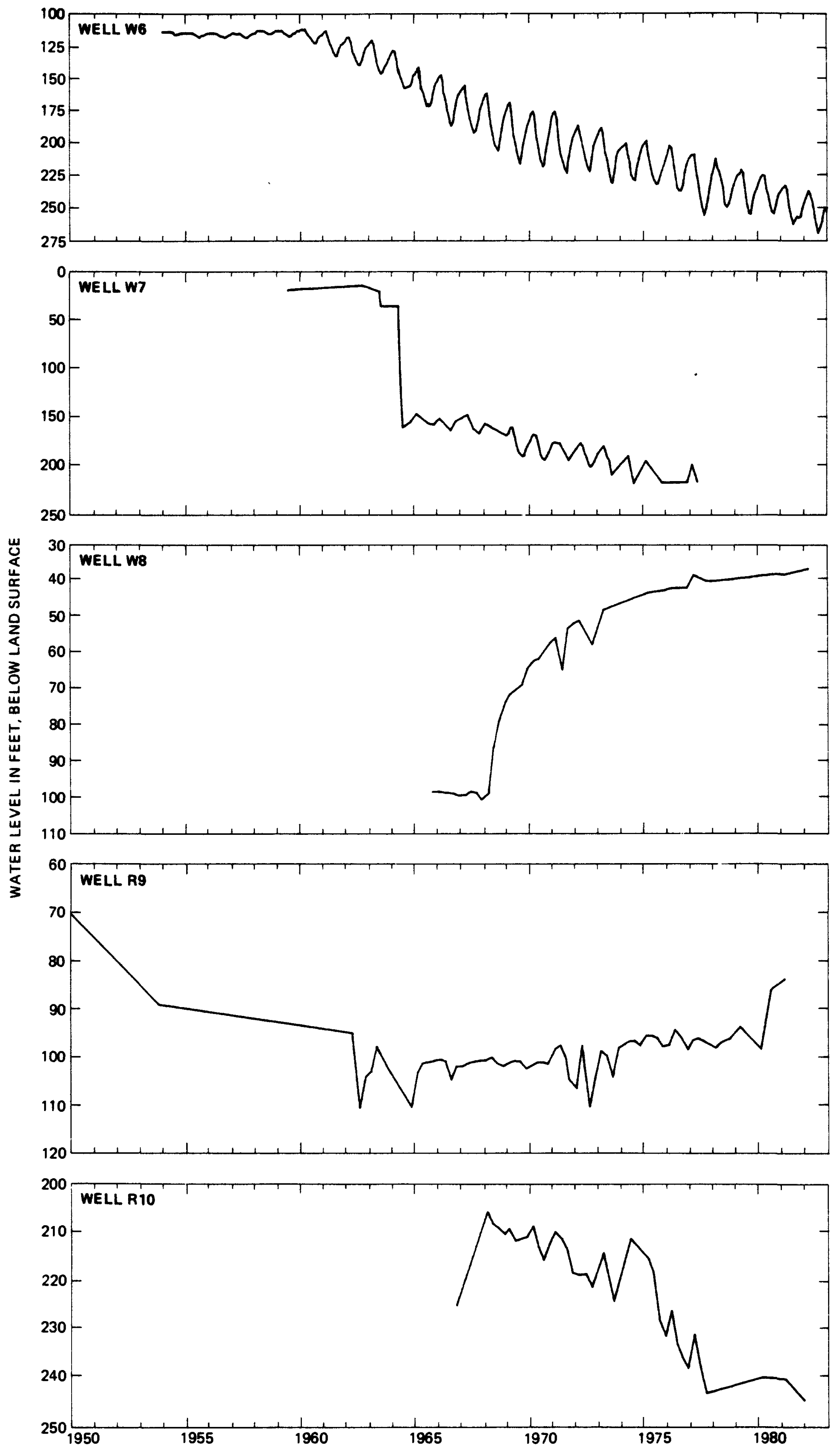


Figure 13.--Water levels in selected wells--continued.

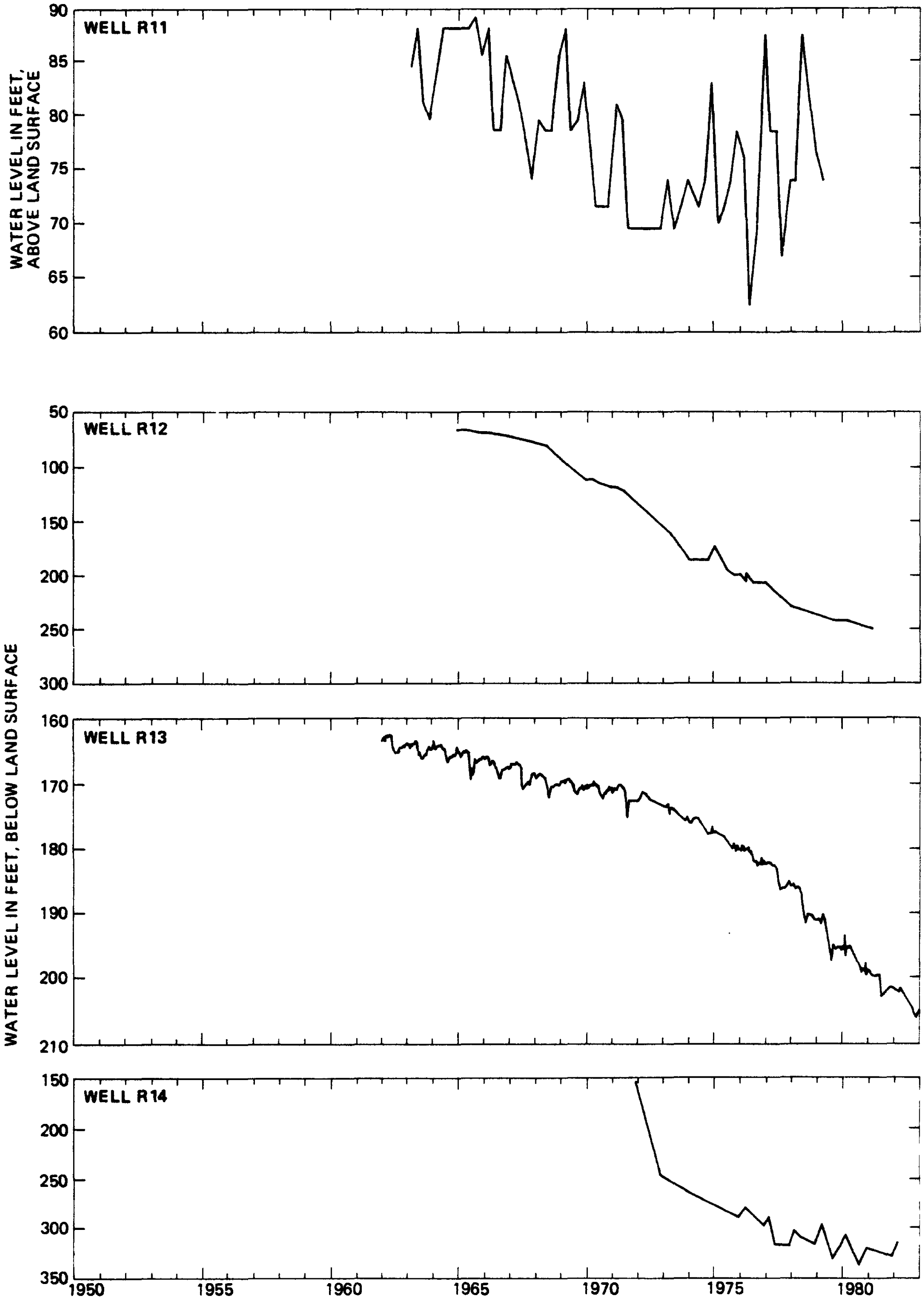


Figure 13.--Water levels in selected wells--continued.

Additionally, water pumped from the Columbia River and spread on lands that lie west and southwest of the Umatilla Ordnance depot may have an effect on shallow water levels in unit 1, but no data are available to evaluate this surface application of water--the water is either consumptively used or runs off from irrigated areas and is locally ponded.

Wells S2 and W3 are completed in both the Saddle Mountain and Wanapum Basalt. Well W3 is used for irrigation, and water levels declined about 75 feet from 1970 to 1978, but changed little thereafter. Well S2 was constructed for industrial use, and the water level declined about 100 feet from 1964 to 1982.

The greatest declines of water levels in the study area occur in the Wanapum Basalt, as noted in the hydrograph of well W4. The decline was about 300 feet from 1962 to 1982. Wells W5 and W6 show declines of about 65 and 130 feet respectively. Wells W4, W5, and W6 are all east of the Service anticline and north of the Willow Creek monocline, and all are used for irrigation. The water level in well W7 east of the Service anticline declined about 40 feet from 1964 to 1977, after the well was deepened. An exception to water-level declines in the Wanapum Basalt occurs near Arlington in the vicinity of the Columbia River. Here the water level in well W8 rose about 62 feet from 1968 to 1982, as a result of leakage from Lake Umatilla when the John Day Dam on the Columbia River was closed in 1968.

The largest declines in the Grande Ronde Basalt are in the area north of the Willow Creek monocline and are shown in the hydrographs of R12 and R14. These are both irrigation wells and their declines were about 200 and 170 feet respectively during the period 1964-82. Areas of lesser decline in the Grande Ronde Basalt are south of the Willow Creek monocline and east of the Service anticline. Wells R10 and R13 have declines of about 30 and 45 feet during the period from 1962 to 1982. The hydrographs of wells R9 and R11 south of the Willow Creek monocline show little change in water level for the 1953-82 period.

In summary, the largest declines in observed long-term water levels occur in the Wanapum and Grande Ronde Basalts north of the Willow Creek monocline. Long-term water-level rises occur in the vicinity of the Columbia River near Arlington in the Wanapum Basalt and in the shallow sediments west of Hermiston.

DIGITAL SIMULATION MODEL

The Umatilla Plateau ground-water flow system was simulated by using the Geological Survey's modular three-dimensional finite-difference model (McDonald and Harbaugh, 1984). The Strongly Implicit Procedure (SIP) solution algorithm is used to solve iteratively the sets of flow equations formulated by the program.

In order for the model to converge to an interim solution, the acceleration parameter and the seed were varied by trial and error until a solution was achieved. The optimum values for the acceleration parameter and the seed were 0.15 and 0.50 respectively.

Four layers were used to simulate the four geohydrologic units in the model. Layer 1 is simulated as an unconfined unit and represents the sediments overlying the Saddle Mountain Basalt. Layer 2 represents the Saddle Mountains Basalt, layer 3 represents the Wanapum Basalt, and layer 4 represents the Grande Ronde Basalt.

Because partial dewatering has occurred in some areas, layers 2 and 3 are simulated as semiconfined to allow for recomputation of the transmissivity based on changes in saturated thickness. Layer 4 is simulated as confined.

The finite-difference grid has 1,404 rectangular blocks that range in size from 1.5 to 4 miles on a side, with areas ranging from 2.25 to 16 square miles. The smallest blocks are in the area of greatest ground-water withdrawals. The grid is rotated 36 degrees counterclockwise from the landnet for alignment with the predominant direction of ground-water flow and to minimize the number of inactive blocks. The grid size is 39 rows and 36 columns. The grid and its orientation with respect to geographic features are shown in figure 14.

Steady-state Analysis

The steady-state analysis represents the period prior to about 1950, when no major ground-water development had taken place. The ground-water system was considered to be in a state of equilibrium, or steady state. Natural recharge to the layers over a long period of time equaled natural discharge and storage remained unchanged.

Boundaries

For the steady-state analysis, the boundary conditions are shown in figures 15a, b, c, and d. Layer 1 extends over a small area near the center of the model. Most of the sedimentary deposits overlying the basalt in the study area (see fig. 2) are unsaturated. The area within the boundary in figure 15a represents the sediments that are saturated and where ground water has been developed. The sedimentary deposits near and beneath streams were not simulated because they are relatively insignificant. The contacts of the saturated part of layer 1 with the underlying basalt at the edges of layer 1 were simulated as lateral zero-flow boundaries. Likewise, for layers 2, 3, and 4, wherever formation outcrops exist, no lateral flow occurs and the boundaries are simulated as zero flow.

In layer 2 at the Horse Heaven Hills anticline, lateral flow was simulated as zero flow because the hills form a natural drainage divide coincident with an elongated fault zone that runs near and along the crest of the Hills. In layers 3 and 4, the east boundary coincides with the natural drainage, and the area along this boundary parallels the Walla Walla fault system. This boundary was also simulated as zero flow.

In layer 4, the southeast boundary was simulated using a head-dependent flux condition because the model boundaries do not extend entirely to the drainage divide, and some lateral flow probably enters this boundary. In layers 3 and 4 at the Horse Heaven Hills anticline, the head-dependent flux boundary is also used, but because the geohydrologic unit does not crop out nor are wells drilled into the unit, the flux across this boundary in layers 3 and 4 is not known. These conditions were tested during the transient (induced stress) analysis of the model. Along the west boundary of the model, the flow lines of the ground-water system for layer 4 are generally parallel to the boundary and normal to the Columbia River. For steady-state simulation, no lateral flow is assumed to cross this boundary. A head-dependent flux boundary is included at a few nodes in layer 4 along the west boundary to test the zero-flow conditions.

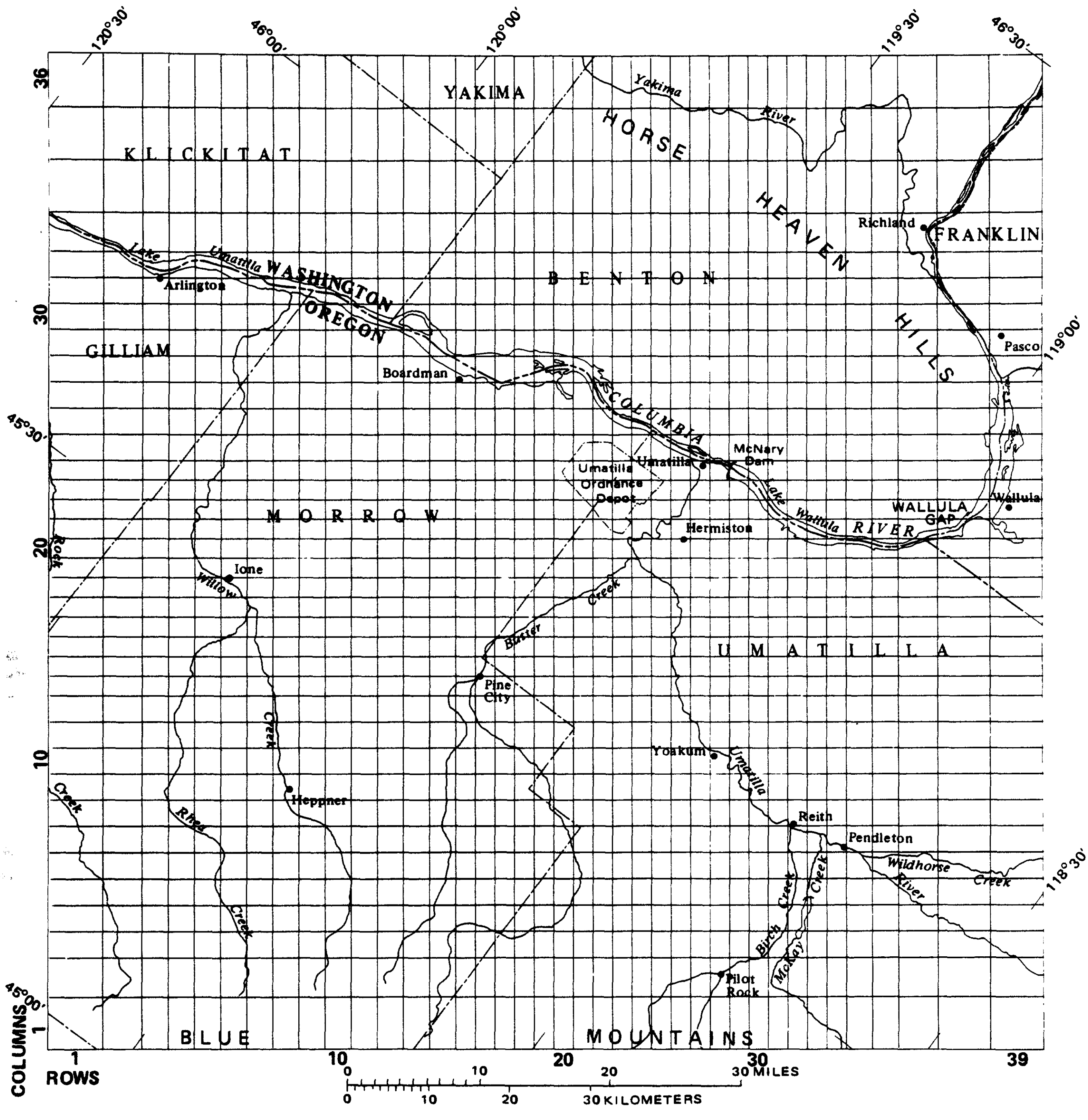
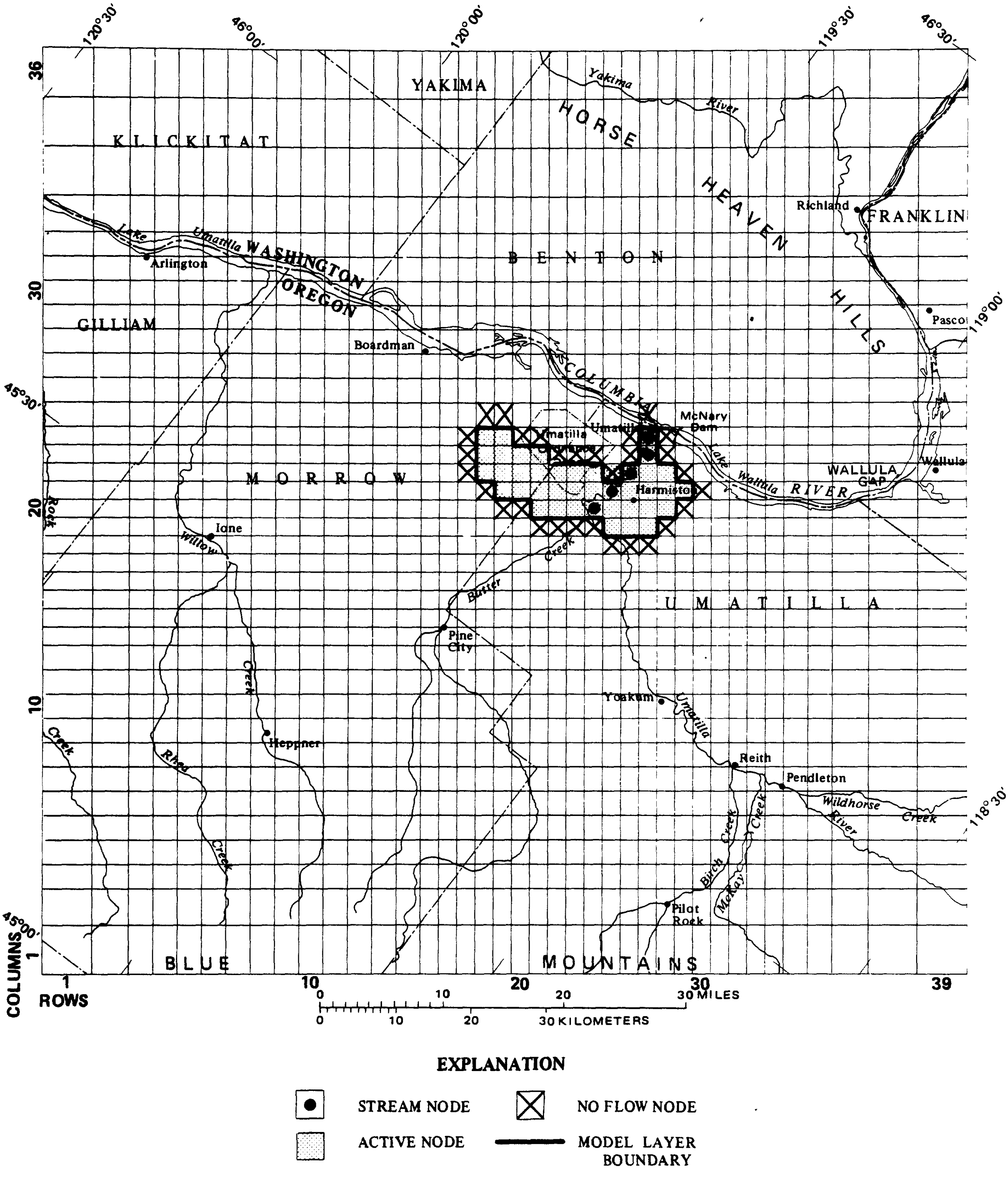
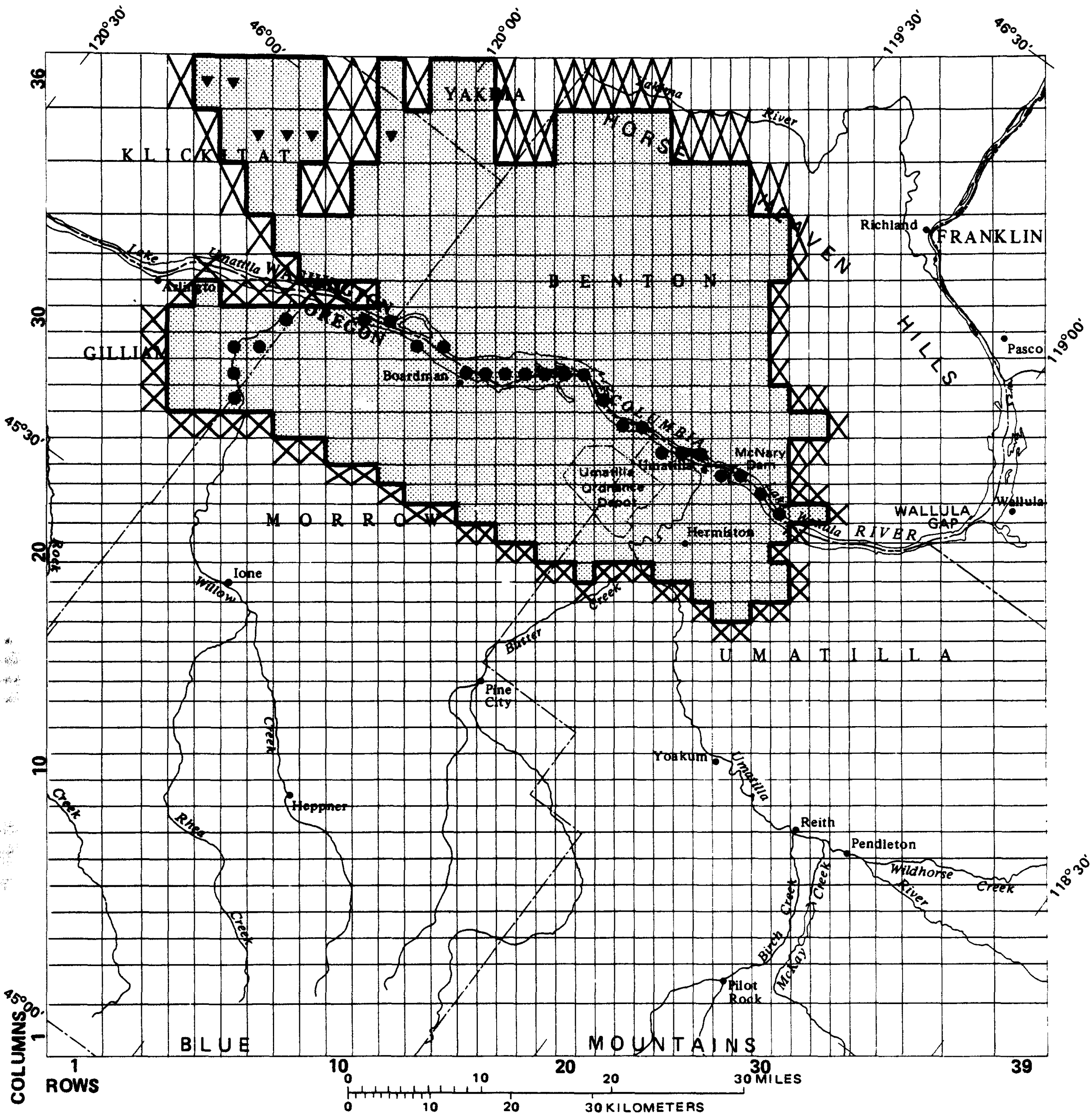


Figure 14.--Relation of model grid to geographic features.



EXPLANATION

Figure 15a.--Model grid of node simulation for layer 1.



EXPLANATION

● STREAM NODE	ACTIVE NODE	MODEL LAYER BOUNDARY
▼ DRAIN NODE	NO FLOW NODE	

Figure 15b.--Model grid of node simulation for layer 2.

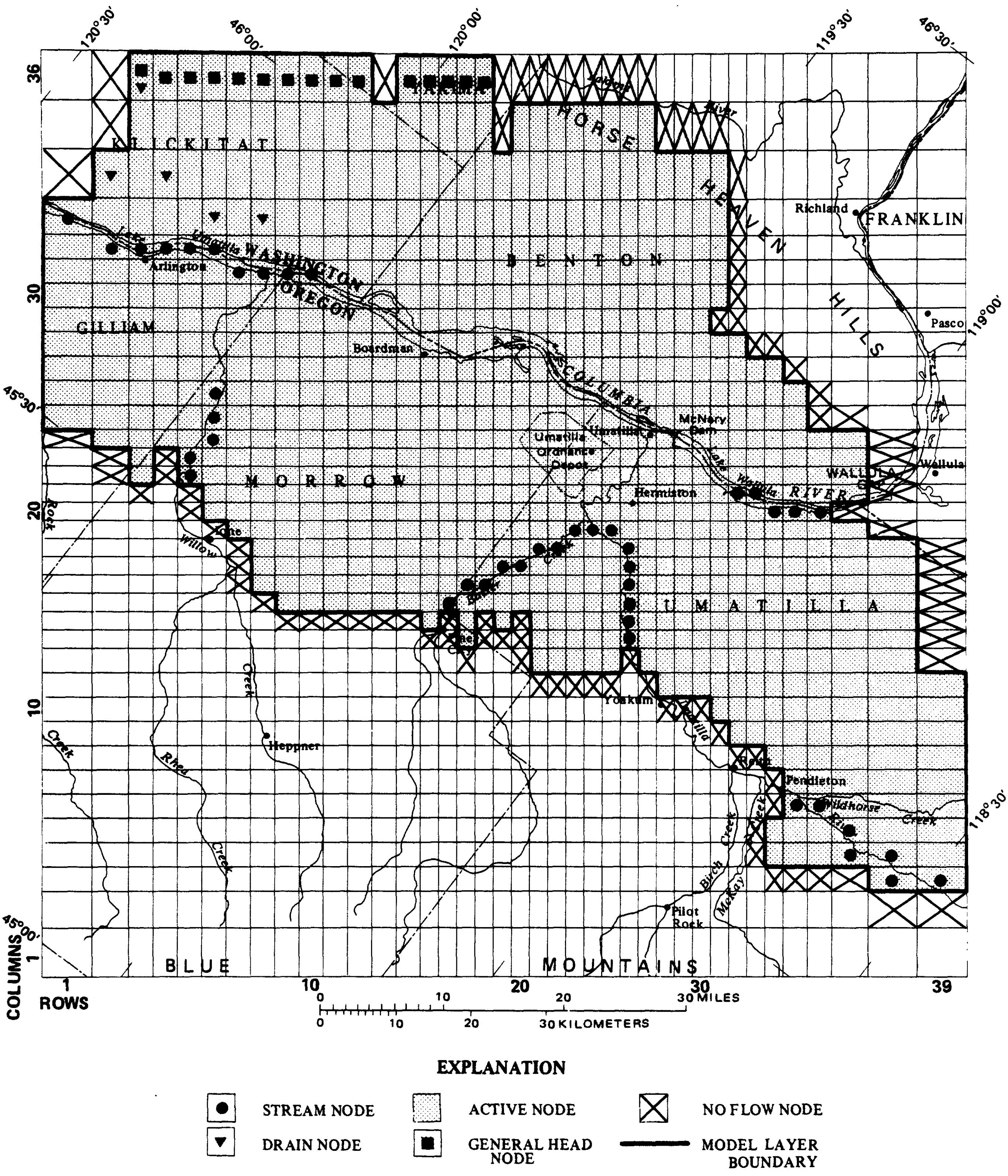
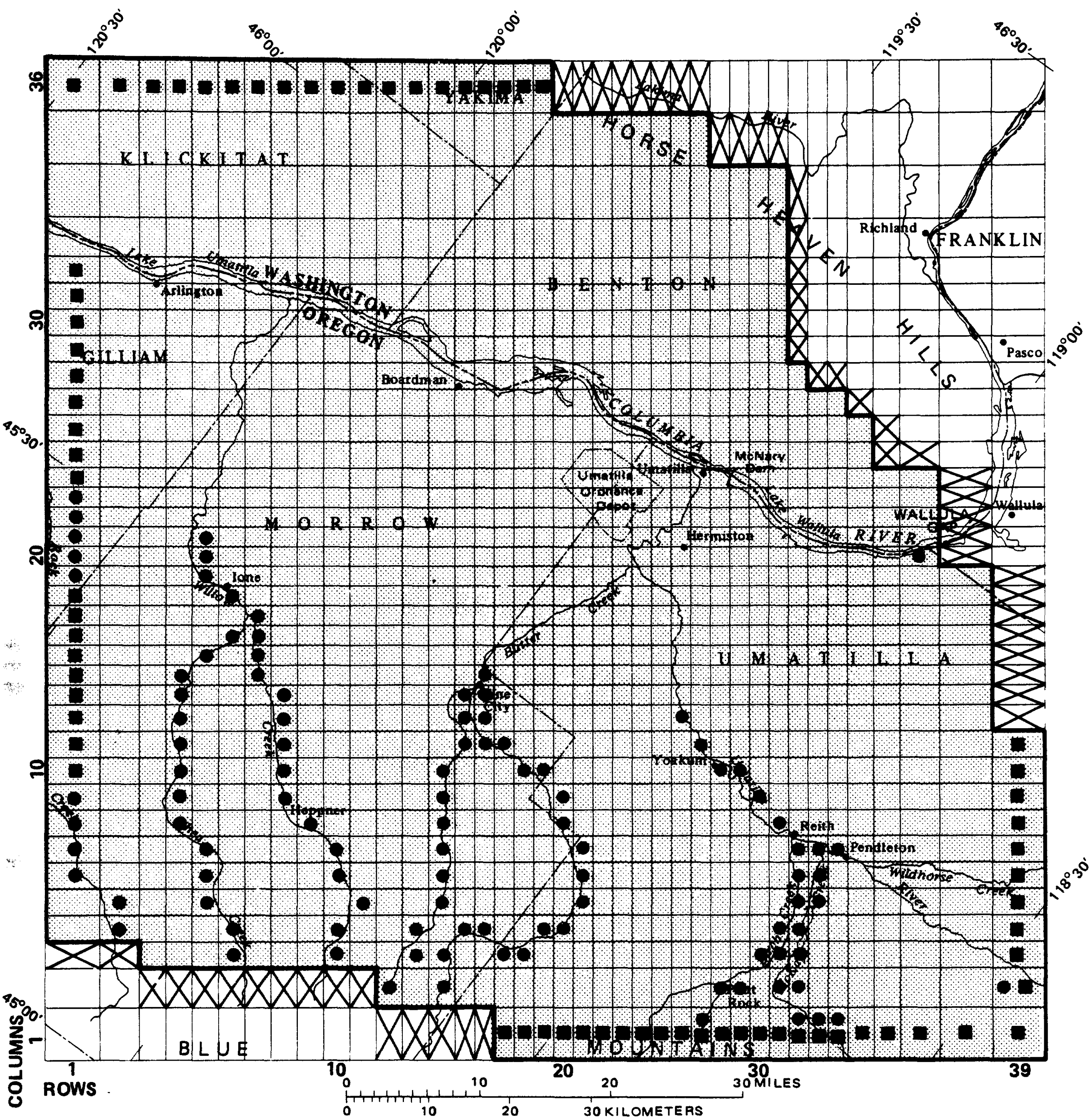


Figure 15c.--Model grid of node simulation for layer 3.



EXPLANATION

●	STREAM NODE	░	ACTIVE NODE	———	MODEL LAYER BOUNDARY
■	GENERAL HEAD NODE	☒	NO FLOW NODE		

Figure 15d.--Model grid of node simulation for layer 4.

For all layers where streams are hydraulically connected to the layers, the internal boundaries were simulated according to the Darcy relation:

$$Q = \frac{k}{m} (h_s - h_a)A,$$

where

Q = flow rate from stream to layer, L^3/T;
k = vertical hydraulic conductivity of the streambed, $\frac{L}{T}$;
m = thickness of the streambed, L;
h_s = elevation of the stream surface, L;
h_a = elevation of the aquifer head, L; or if h_a is below the streambed, then the elevation of the streambed is used; and
A = area of streambed, L^2.

The hydraulic conductivity of the streambed (k) was the same as that used for estimating the conductance between layers.

Using this relation for the river-layer connection, the model calculates the flow rate entering and leaving each node (figs. 15a-d) in the model.

Simulation of drains in the system is similar to that for streams, except that water is allowed to move only from the layer to the drain and not vice versa. Thus, if the the head in the layer is below the head in the drain, the flow is set equal to zero. Drain nodes are shown in figure 15a-d; they represent seepage from the ground-water system to the narrow, deeply-entrenched stream channels in the western part of the Horse Heaven Hills area in Washington.

Initial Model Input

The input data to the steady-state model is as follows:

(1) An initial estimate was made of hydraulic-head distribution prior to ground-water development. Although head data were sparse during the predevelopment period, the estimate of head distribution was constructed from water levels reported on drillers' logs and was supplemented with water levels from a few observation wells.

(2) Values for the altitude of the tops and bottoms for each node of layers 1, 2, and 3, and the top of layer 4 were interpreted principally from drillers' logs. The model computes thickness of each layer from the data. For layer 4, a thickness of 1,000 feet was used.

(3) Values of lateral hydraulic conductivity for each node in each layer were determined from aquifer tests, where available, and from specific capacity tests. Values of transmissivity were estimated for layer 4 from hydraulic conductivity and thickness.

(4) Values of hydraulic conductivity divided by thickness for each node in the intervals between the four layers were derived from cross-section modeling and an empirical relation between lateral hydraulic conductivity and thickness, as explained in a previous section of the report.

(5) An estimate of average recharge for each uppermost node in the model was calculated from the method of Bauer and Vaccaro (1987), as shown in figure 10.

(6) Values for river and drain heads, areas (derived from topographic maps), and conductances were coded for each river and drain node in the model.

Calibration

The procedure to calibrate the steady-state model consisted principally of adjusting the least known variable in the model. That variable is the vertical hydraulic conductivity divided by thickness (k/b) between layers 1 and 2, 2 and 3, and 3 and 4. Initially, the array multipliers for each layer were adjusted until a model solution was obtained. The array multipliers, thus obtained, were 0.1 for $(k/b)_{1-2}$, 0.08 for $(k/b)_{2-3}$, and 0.08 for $(k/b)_{3-4}$.

Secondly, the procedure consisted of adjusting the k/b on a node-by-node basis between layers 1 and 2, 2 and 3, and 3 and 4, until model-calculated heads in each layer closely approximated observed heads in control wells. The locations of the control wells are shown in figure 16. Additionally, where layer 4 is not overlain by layer 3 (no leakage), values of transmissivity in layer 4 were adjusted to match heads in the control wells. A few adjustments in lateral hydraulic conductivity were made in layers 2 and 3 near the axis of the Horse Heaven Hills anticline, where adjustments in vertical hydraulic conductivity were unsuccessful for calibration. Calibration required decreasing lateral hydraulic conductivity values in layer 1 by a factor of 0.01. Figures 17a, b, and c show the values for k/b, and figures 18a b, c, and d show the values of transmissivity as determined during steady-state calibration.

The average differences[1] between the observed heads at control wells and the calculated contoured heads were ±7, ±40, ±29, and ±52 feet for layers 1 through 4 respectively. The head distribution of each layer as calculated by the steady-state model is shown in figures 19a, b, c, and d.

A statistical comparison between measured stream gains and losses and model-calculated stream gains and losses was not done because adequate streamflow records are not available. Streamflow records at existing gaging sites are affected by numerous ungaged diversions for irrigation, both within and entering into the study area. Also, because of the large average flow of the Columbia River (about 1,000 times the ground-water component of the water budget), measured gains or losses to the Columbia River from ground water are indeterminate.

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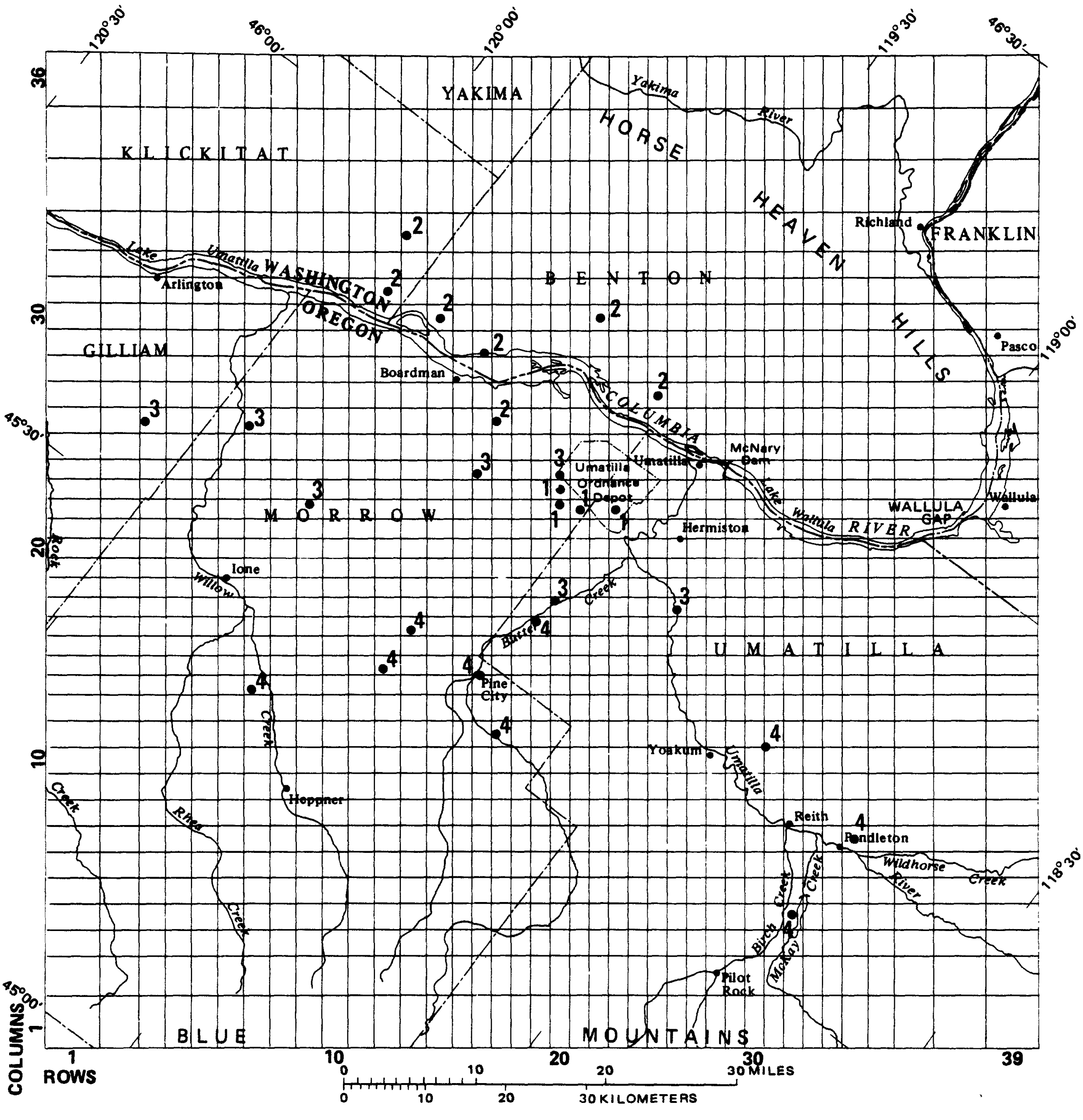
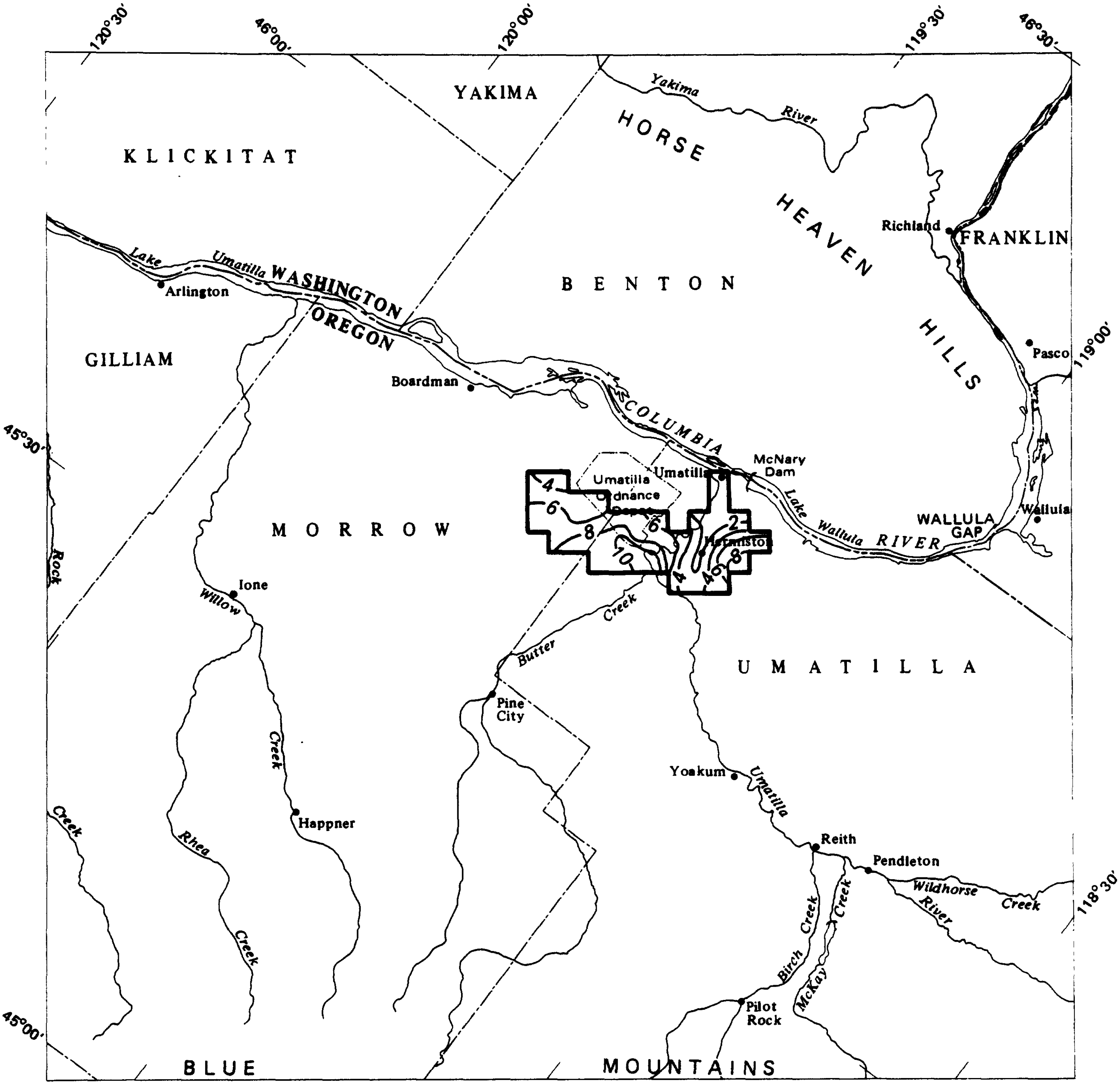
[1] The average difference is the sum of absolute differences divided by the number of wells.



EXPLANATION

•1 WELL USED FOR WATER-LEVEL CONTROL --Number refers to layer in which well is completed.

Figure 16.--Location of wells used for water-level control during model simulation.



0 10 20 30 MILES

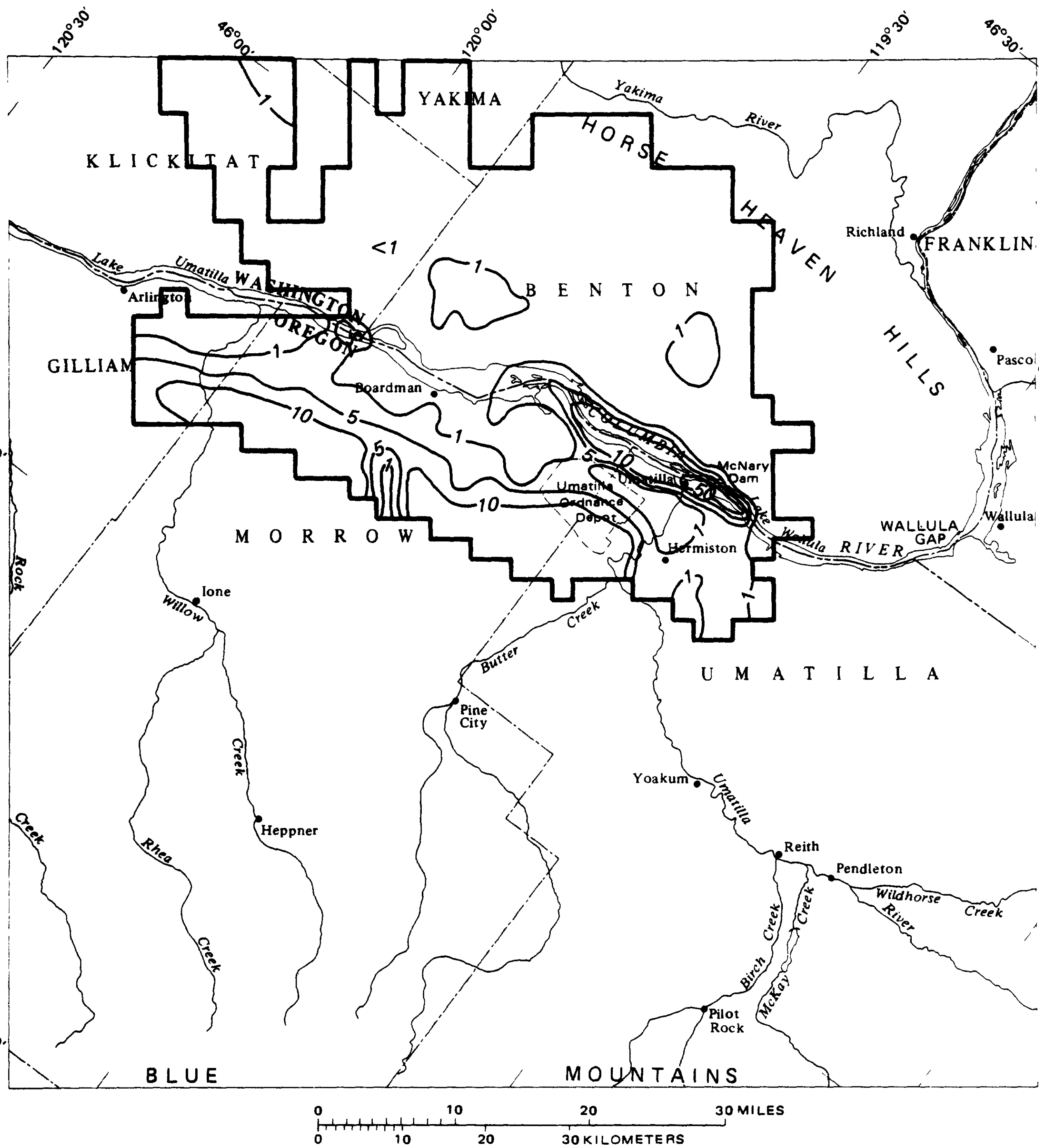
0 10 20 30 KILOMETERS

EXPLANATION

——2—— LINE OF EQUAL VERTICAL HYDRAULIC CONDUCTIVITY DIVIDED BY THICKNESS. Interval 2(ft/s)/ft X 10^{-11}.

——— MODEL LAYER BOUNDARY

Figure 17a.--Vertical hydraulic conductivity divided by thickness between layers 1 and 2.

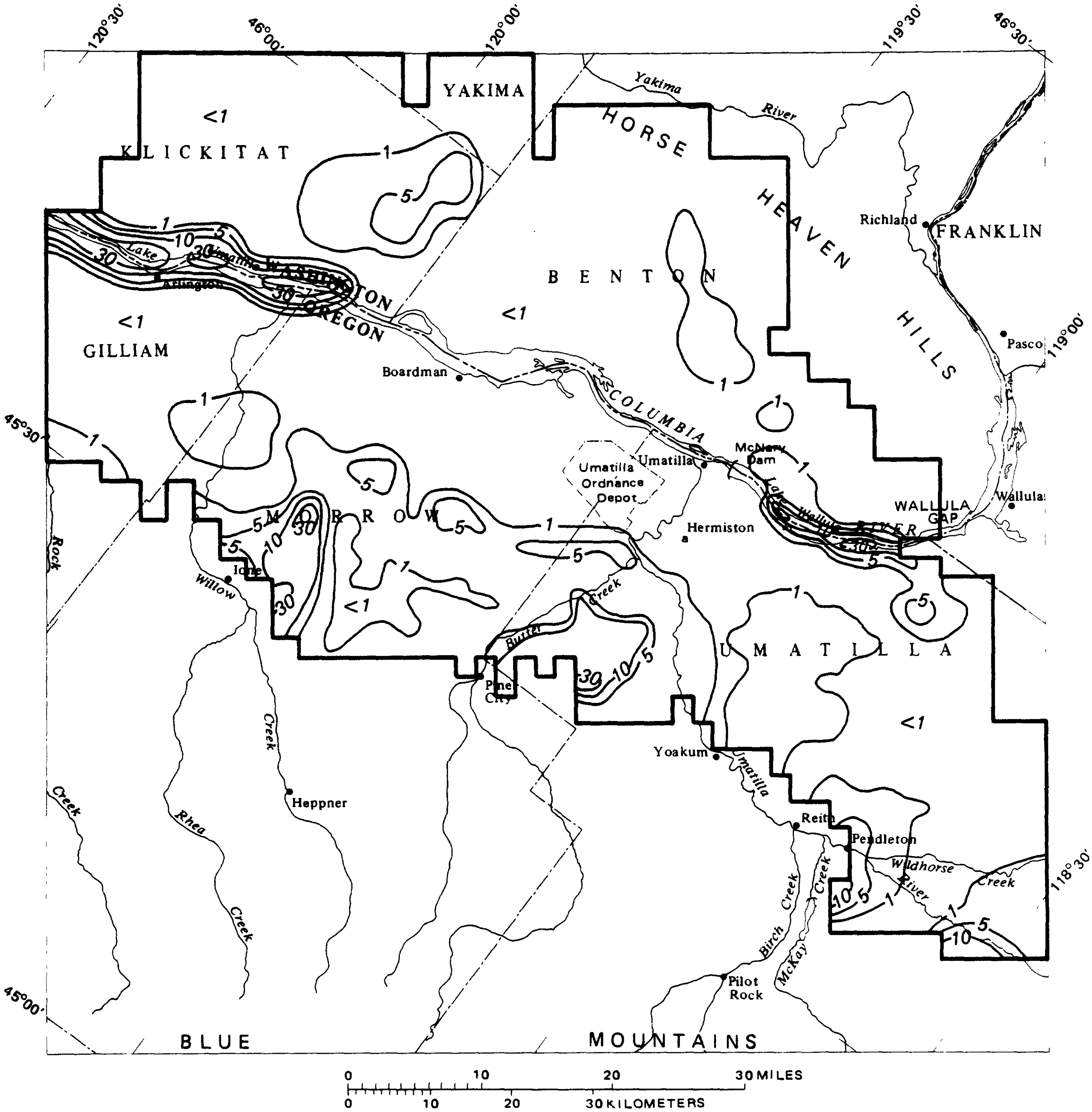


EXPLANATION

——1—— LINE OF EQUAL VERTICAL HYDRAULIC CONDUCTIVITY DIVIDED BY THICKNESS. Interval 4, 5, and 40 (ft/s)/ft X 10^{-11}.

——— MODEL LAYER BOUNDARY

Figure 17b.--Vertical hydraulic conductivity divided by thickness between layers 2 and 3.

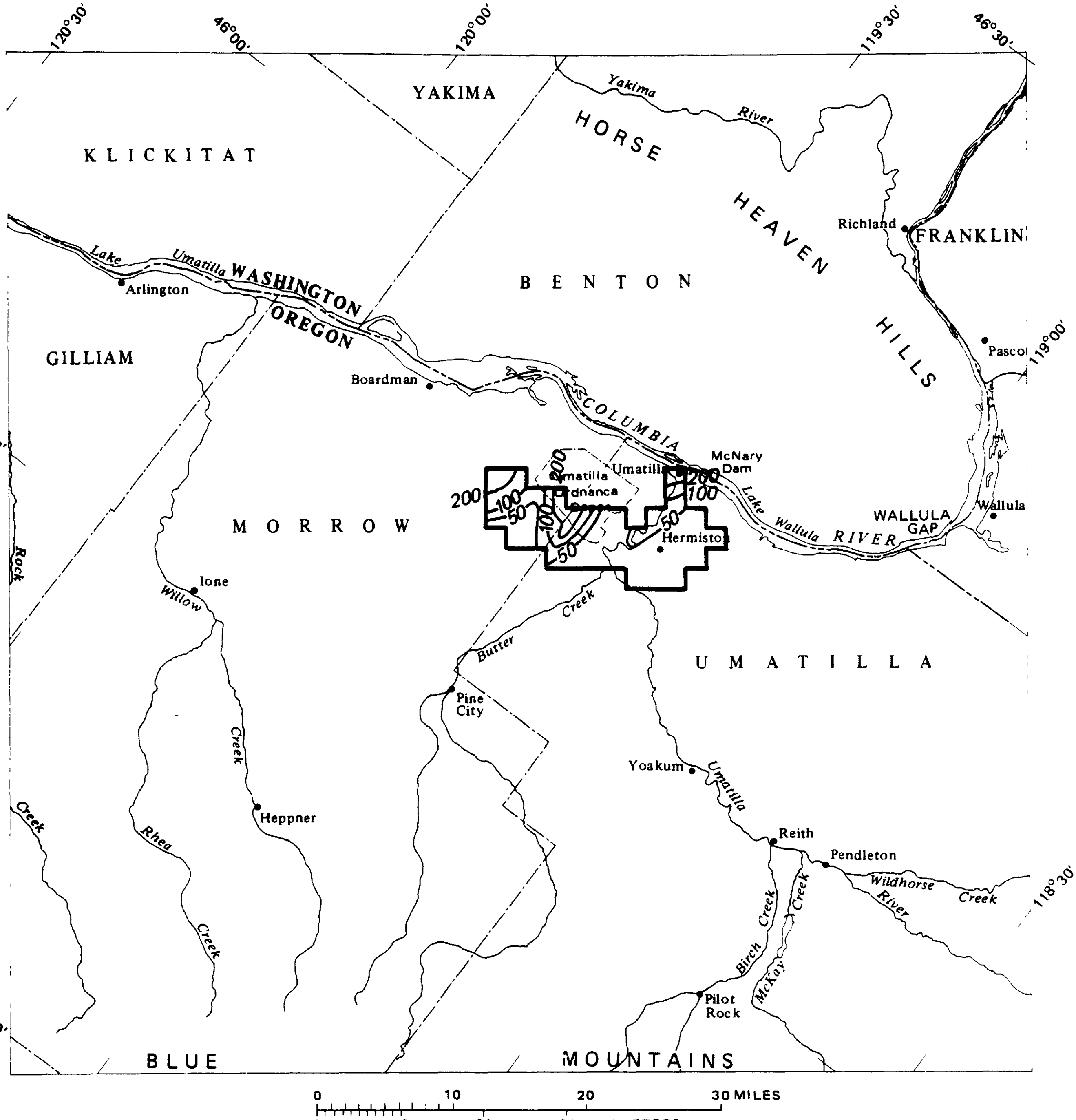


EXPLANATION

—30— LINE OF EQUAL VERTICAL HYDRAULIC CONDUCTIVITY DIVIDED BY THICKNESS. Interval 4, 5, and 20 (ft/s)/ft X 10^{-14}.

MODEL LAYER BOUNDARY

Figure 17c.--Vertical hydraulic conductivity divided by thickness between layers 3 and 4.

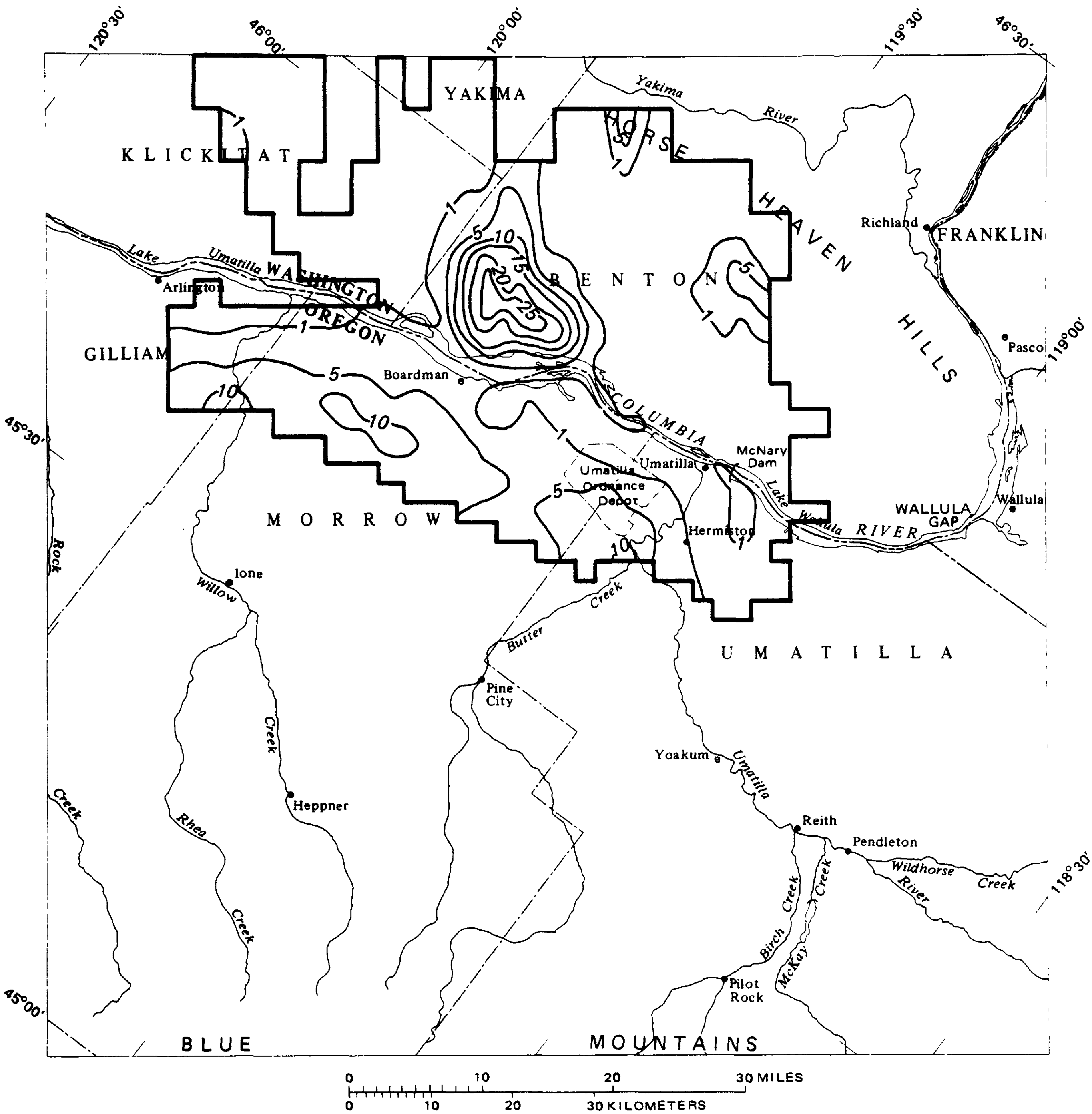


EXPLANATION

—— 50 —— LINE OF EQUAL TRANSMISSIVITY. Interval 50 and 100 X 10^{-2} ft^2/s.

———— MODEL LAYER BOUNDARY

Figure 18a.--Transmissivity for layer 1.

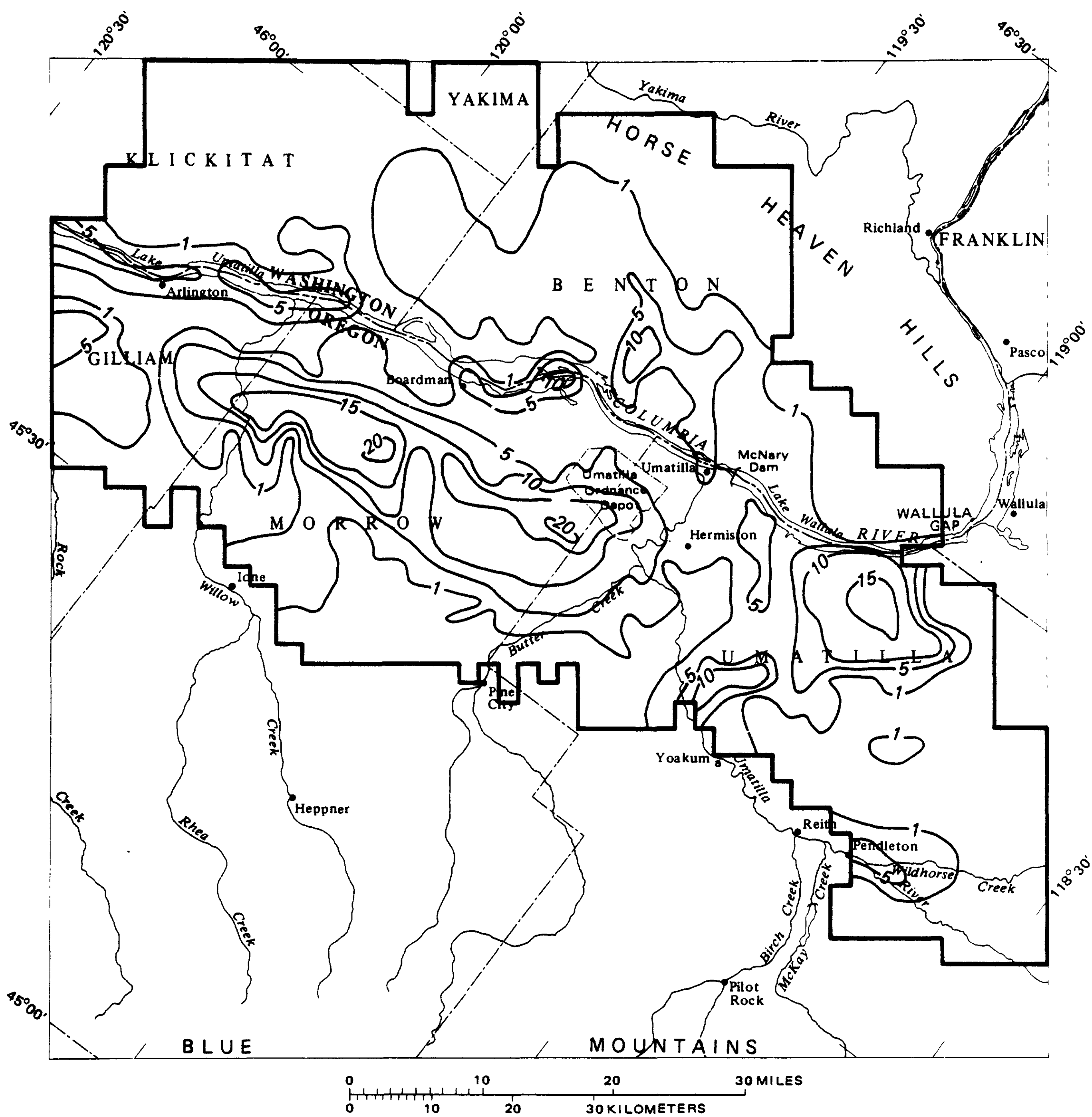


EXPLANATION

—— 10 —— LINE OF EQUAL TRANSMISSIVITY. Interval 4 and 5 X 10^{-2} ft^2/s.

▬▬▬ MODEL LAYER BOUNDARY

Figure 18b.--Transmissivity for layer 2.

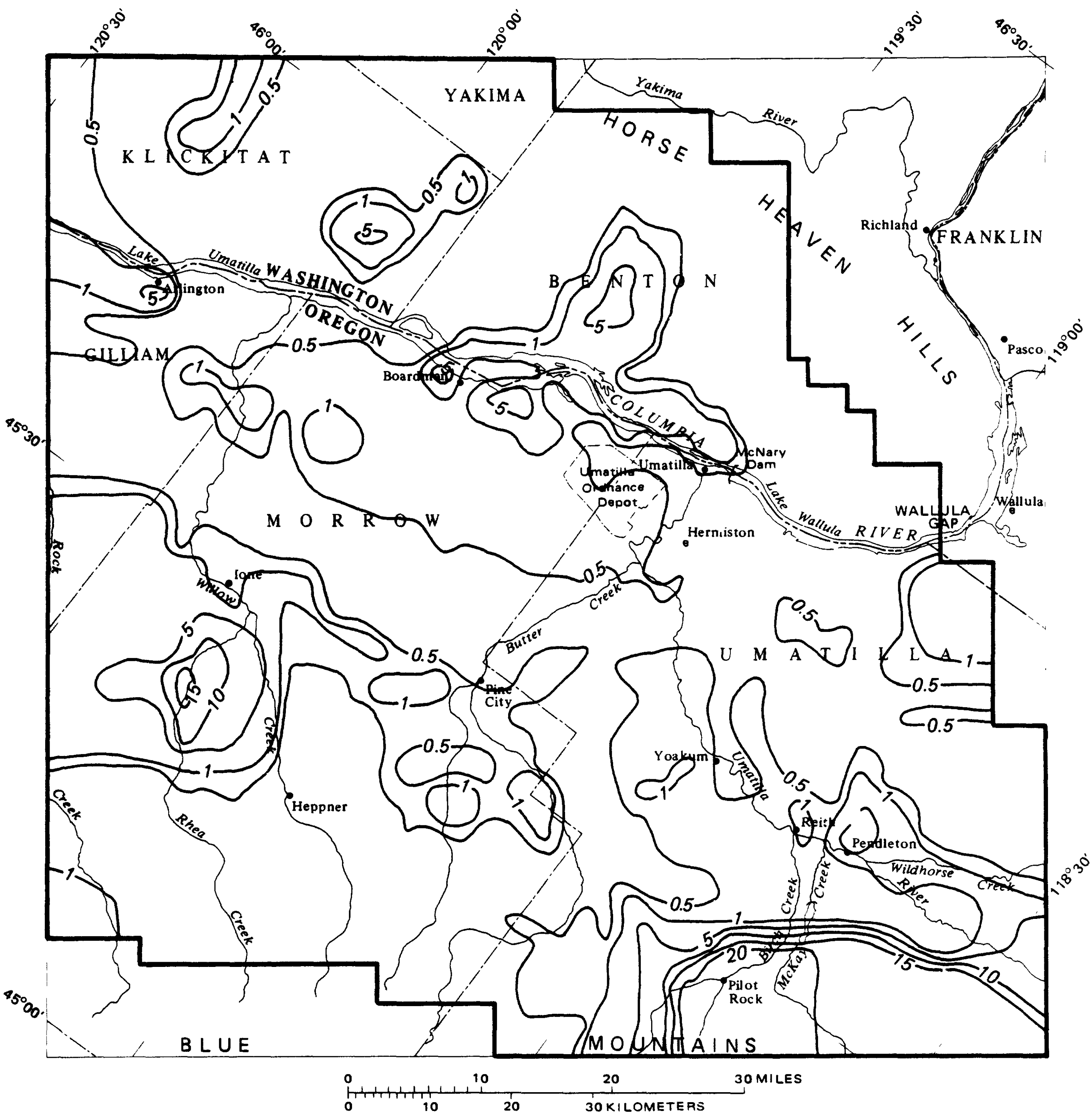


EXPLANATION

—— 10 —— LINE OF EQUAL TRANSMISSIVITY. Interval 4 and 5 X 10^{-2} ft^2/s.

▬▬▬ MODEL LAYER BOUNDARY

Figure 18c.--Transmissivity for layer 3.

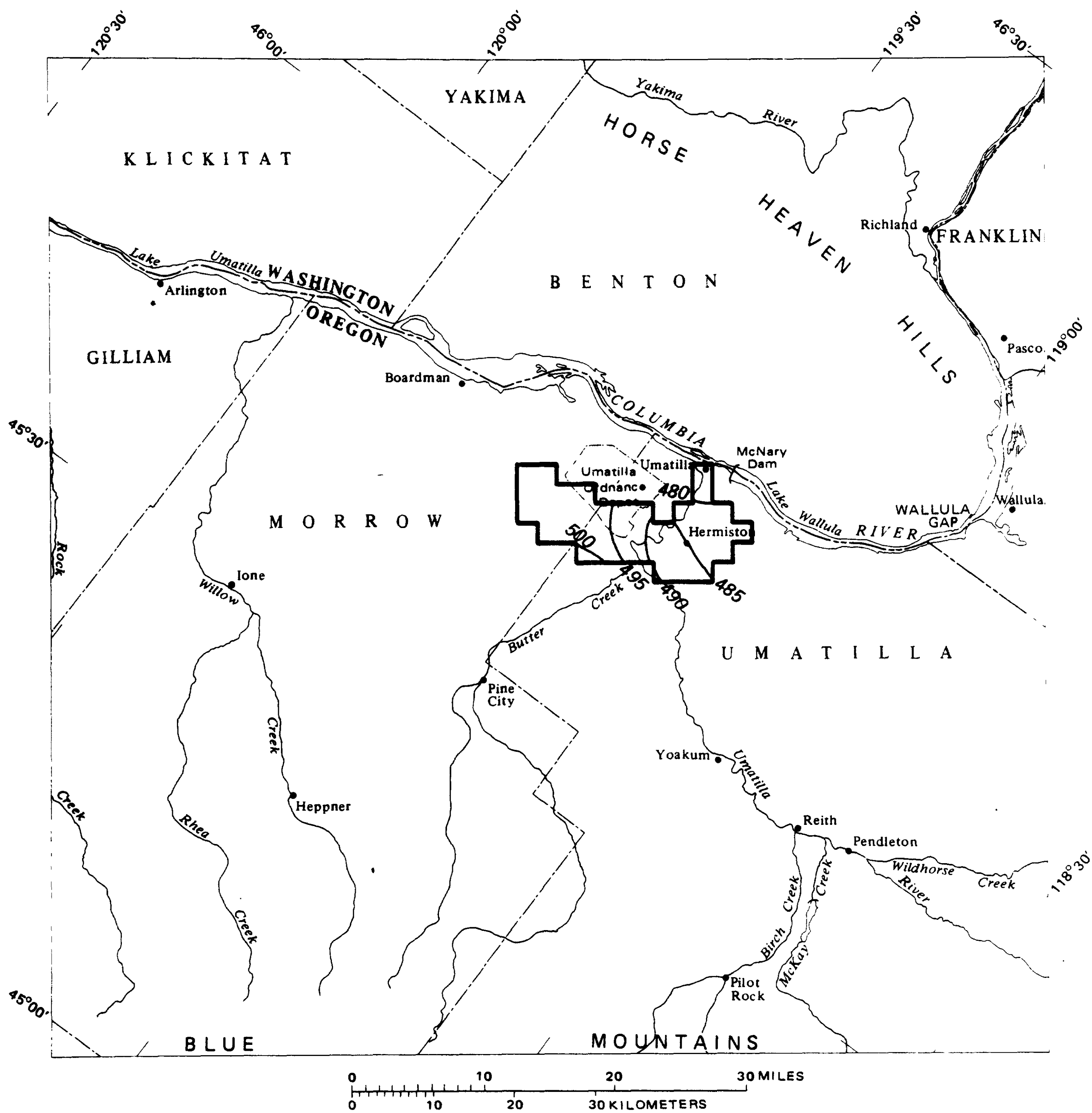


EXPLANATION

——— 1 ——— LINE OF EQUAL TRANSMISSIVITY. Interval 0.5, 4, and 5 X 10^{-2} ft^2/s.

▬▬▬ MODEL LAYER BOUNDARY

Figure 18d.--Transmissivity for layer 4.

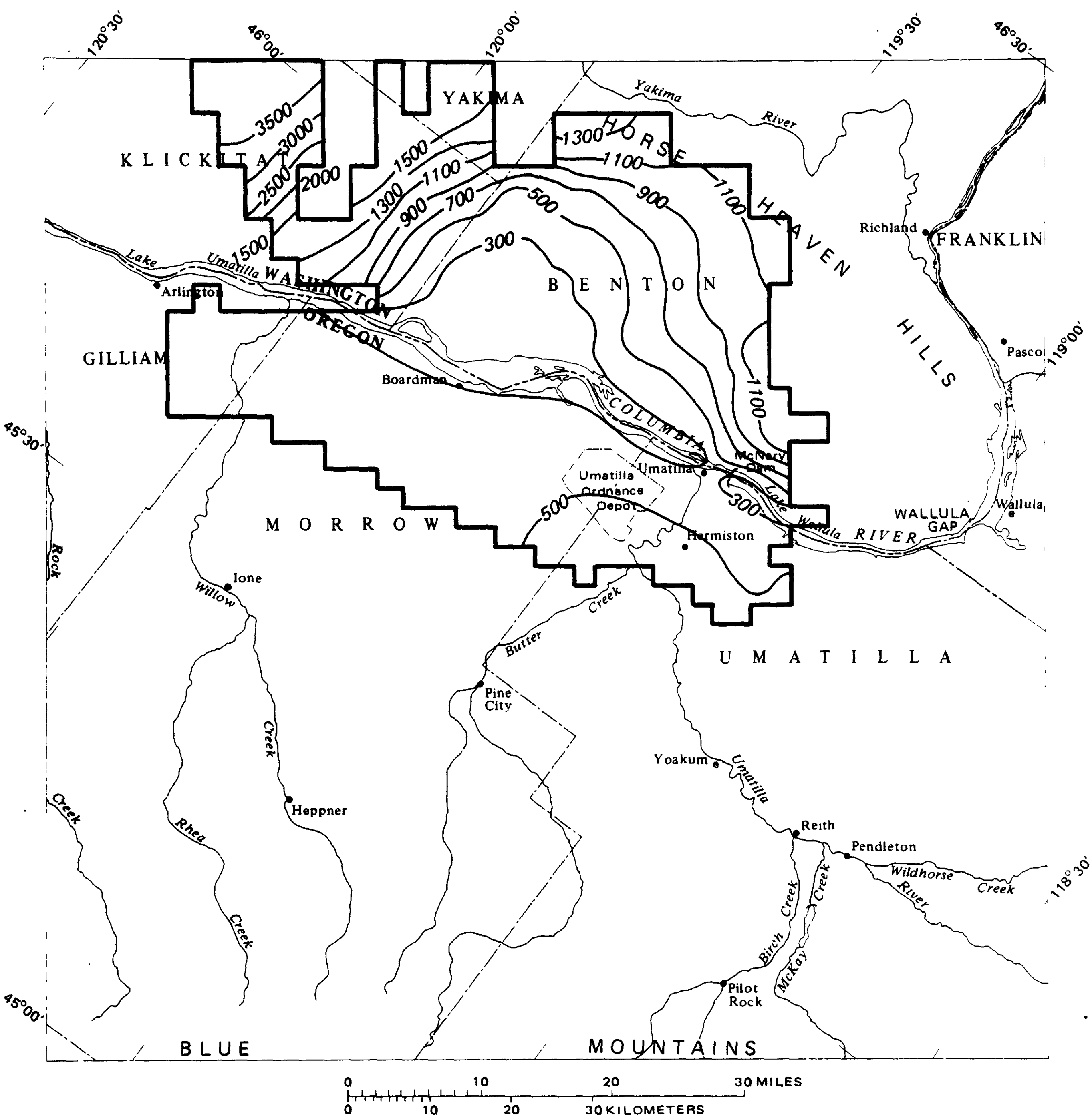


EXPLANATION

——500—— WATER LEVEL CONTOUR–Shows altitude of water level. Datum is sea level. Contour interval is 5 feet.

———— MODEL LAYER BOUNDARY

Figure 19a.--Head distribution as calculated during steady-state simulation for Layer 1.

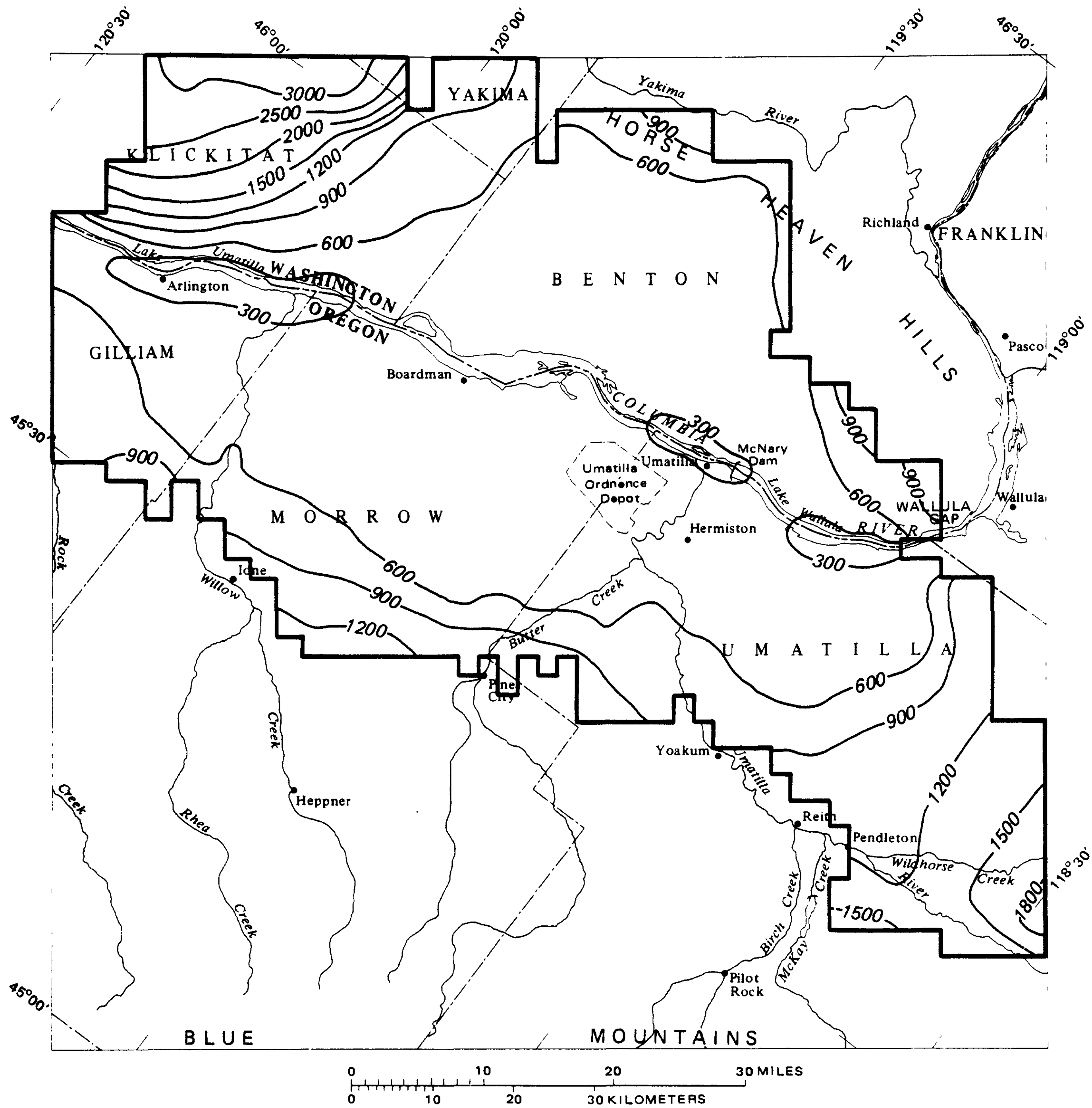


EXPLANATION

—— 500 —— WATER LEVEL CONTOUR–Shows altitude of water level. Datum is sea level. Contour interval is 200 and 500 feet.

———— MODEL LAYER BOUNDARY

Figure 19b.--Head distribution as calculated during steady-state simulation for Layer 2.

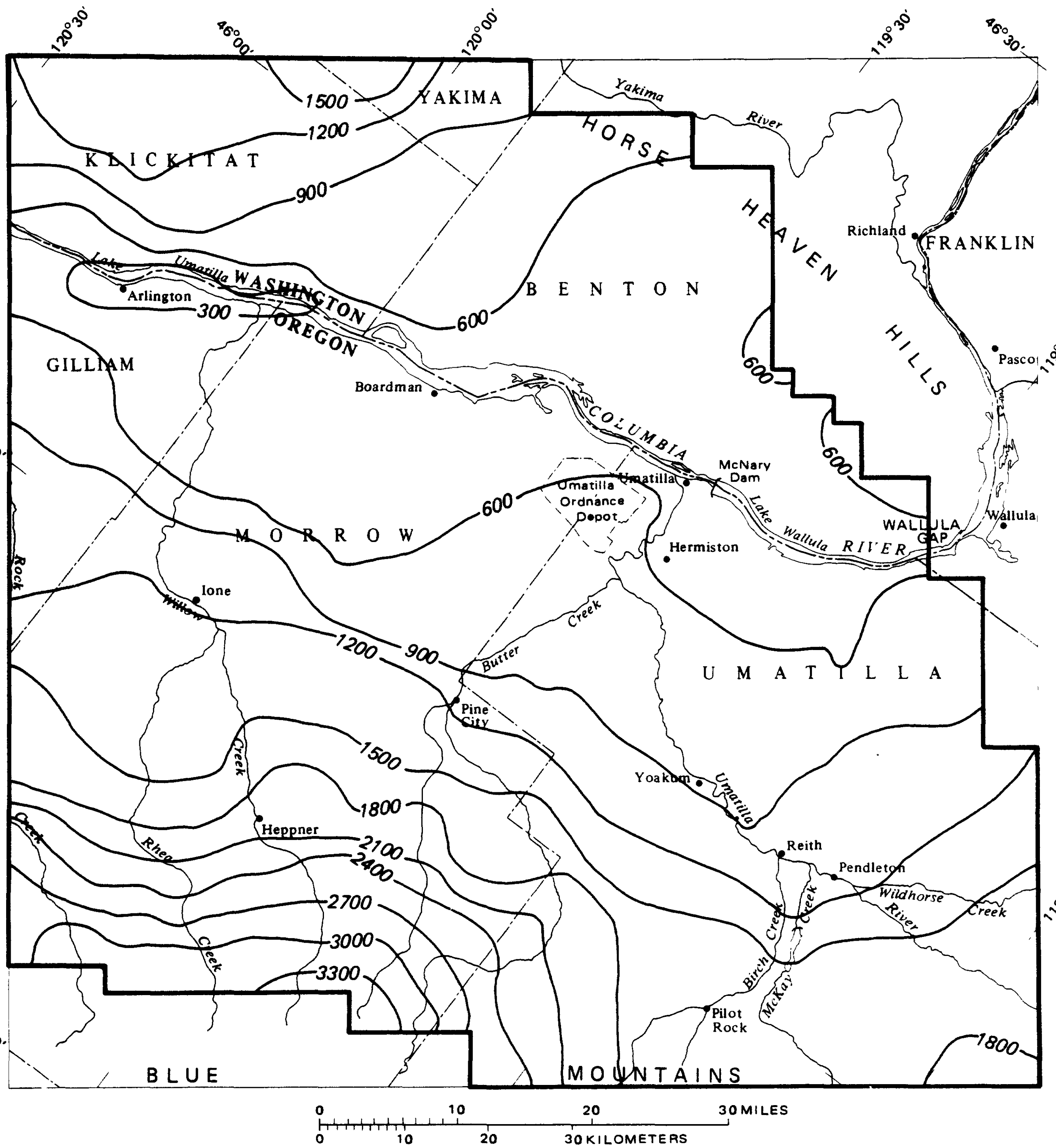


EXPLANATION

—— 900 —— WATER LEVEL CONTOUR–Shows altitude of water level. Datum is sea level. Contour interval is 300 and 500 feet.

———— MODEL LAYER BOUNDARY

Figure 19c.--Head distribution as calculated during steady-state simulation for Layer 3.



EXPLANATION

——300—— WATER LEVEL CONTOUR–Shows altitude of water level. Datum is sea level. Contour interval is 300 feet.

—————— MODEL LAYER BOUNDARY

Figure 19d.--Head distribution as calculated during steady-state simulation for Layer 4.

A streamflow analysis of the Umatilla Indian Reservation in the southeast part of the study area, made by Gonthier and Harris (1977, table 3), showed that the long-term average streamflow gain from the main reservation area was 60 ft^3/s. The model calculated about 55 ft^3/s from approximately the same area, even though the model river nodes do not conform exactly with the measuring sites on the boundaries of the reservation. Also, the calculation by Gonthier and Harris includes about 3 ft^3/s of flow to Wildhorse Creek. Wildhorse Creek was not simulated in the model. Thus, the model calculated values, at least in the reservation part of the modeled area, appear reasonable.

The steady-state, model-calculated gains and losses to and from major streams in the study area are as follows:

	Gain (+)/Loss (-) in ft^3/s
Columbia River	+72.55
Umatilla River	+34.22
McKay Creek	+32.27
Birch Creek	+8.59
Butter Creek	-9.30
Willow Creek	+13.55
Rhea Creek	+1.18
Rock Creek	+0.79
Total	+153.85

All of the streams with the exception of Butter Creek show a net gain in discharge as calculated by the model.

Model Budget

The flow rates as calculated by the steady-state model are as follows:

	Inflow (ft^3/s)		Outflow (ft^3/s)
Recharge	145.85	Drains	7.18
Leakage from streams	31.33	Leakage to streams	185.18
Boundaries	14.93	Boundaries	0.22
Totals	192.11		192.58

The difference between inflow and outflow calculated by the model (0.47 ft^3/s) is 0.24 percent, which is within the accuracy of the model parameters. Most of the flow enters the model area as recharge from precipitation, whereas most of the flow leaves the model area via streamflow.

According to the simulation, about 11 ft^3/s of flow enters the south and east boundary, about 3 ft^3/s enters the west boundary, and about 0.5 ft^3/s enters the north boundary of the model area as ground-water inflow to layer 4. About 0.1 ft^3/s enters the model area from layer 3 at the north boundary. About 0.1 ft^3/s leaves the north boundary in layer 3 and about 0.1 ft^3/s leaves the west boundary in layer 4.

Transient Analysis

Transient model analysis includes the period from 1950, when major pumpage for ground-water development began, through 1982. This period of 33 years was divided into 33 stress periods, with ground-water pumpage averaged for each individual period.

The method of analysis consisted of using the output from the steady-state simulation and adding the flow system stresses and storage properties for each layer, in order to allow the model to calculate water-level changes that result from the stresses through time (1950-82). The boundary conditions were the same as those used during steady-state analysis.

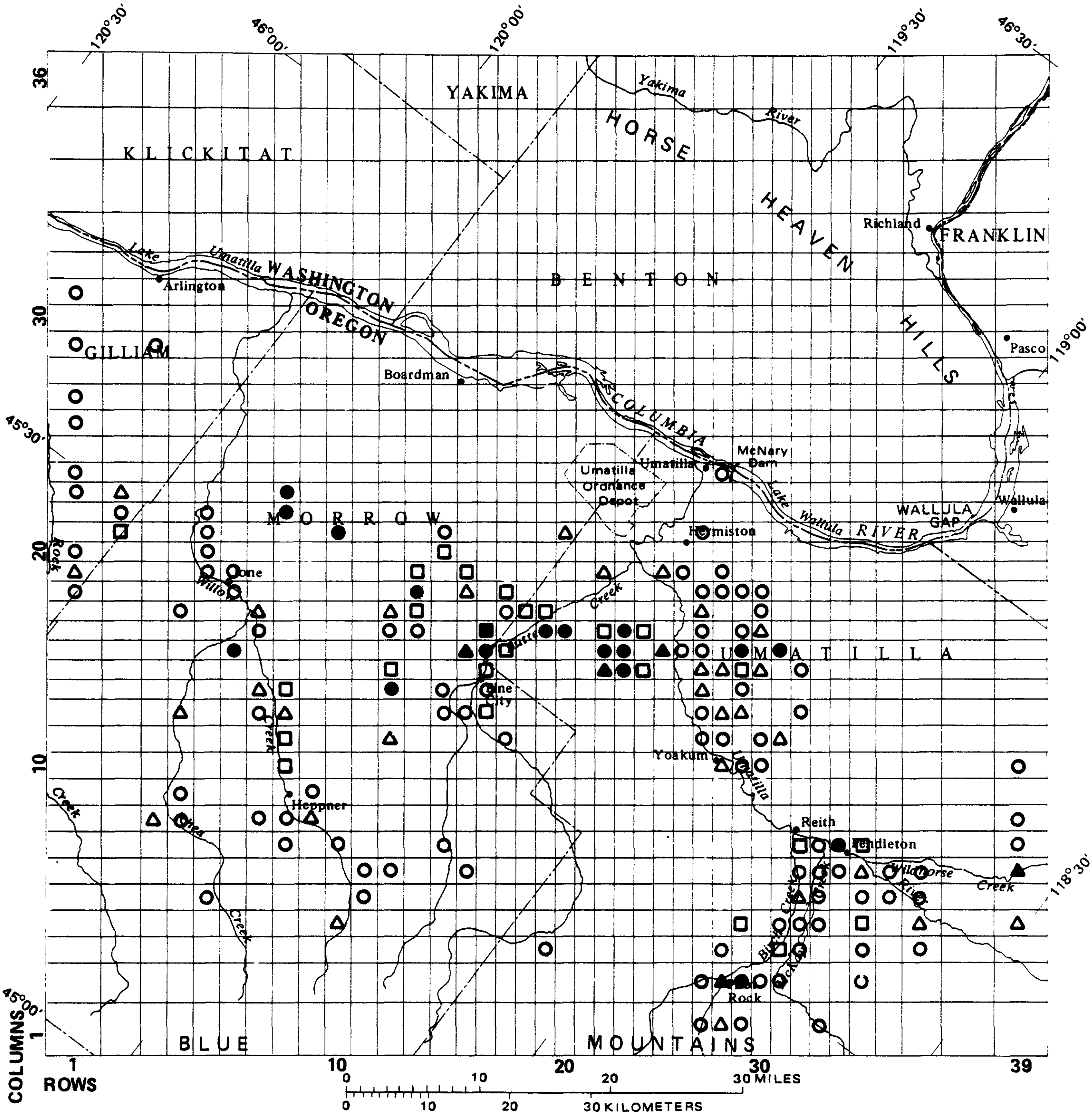
The outputs from the steady-state model are (1) calculated head distribution for each layer, (2) the adjusted lateral and vertical hydraulic conductivity values for each layer, (3) the average recharge for the model, and (4) the river and drain arrays.

The stresses for the transient model are (1) the quantity of ground-water pumped per year for each node in each layer (an example of pumpage distribution for 1 year for one layer is shown in figure 20), (2) the rise in stage for the reservoirs on the Columbia River behind the McNary Dam (completed in 1953) and the John Day Dam (completed in 1968), (3) the diversion of surface water for irrigation in the Horse Heaven Hills area in Washington, and (4) the diversion of surface water from the Umatilla River to a recharge canal in a small area west of Hermiston.

The storage properties of the model layers were initially estimated from ongoing work in the Horse Heaven Hills area and from other model studies in the Columbia Basin that have required values of storage coefficients as model parameters. Only a few individual aquifer tests are available in the area, and these are not adequate to areally define the storage coefficients. The initial specific yield or storage coefficients used for transient simulation were uniform values of 0.25, 0.01, 0.003, and 0.003 for layers 1 to 4 respectively. Additionally, because layers 2 and 3 are allowed to change from confined to unconfined during simulation, a secondary specific yield was estimated for use by the model when layers 2 and 3 were unconfined. The initial secondary specific yields for layers 2 and 3 were set at uniform values of 0.01.

Calibration

The temporal and spatial calculation of water levels by the model and their comparison with observed data is the essence of calibration for the transient model. The procedure consisted of adjusting values of the array multipliers for each storage array for each layer until reasonable water-level changes were obtained from the model. Subsequently, the storage coefficients were adjusted on a node-by-node basis until the differences between model-calculated water levels and measured water levels at control wells were minimized. Figure 21 shows hydrographs of selected wells in the study area and the model-calculated water levels. The calculated water levels, shown in 5-year increments, approximate the observed water levels in most wells. The differences are due partly to inaccuracies in the values of the hydrologic properties and partly to inaccuracies in pumpage estimates. The final storage coefficients as determined during model calibration, average 0.15, 0.01, 0.0045, and 0.0050 for layers 1 to 4 respectively.



EXPLANATION

GROUND WATER PUMPAGE IN CUBIC FEET PER SECOND

No pumpage
0.01–0.25
0.26–0.50
0.51–1.00
1.01–2.00
2.01–3.00
3.01–4.00

Figure 20.--Distribution of ground-water pumpage from public supply, industrial, and irrigation wells for layer 4 for 1980.

In the area west of Hermiston, in layer 1, large amounts of ground water are pumped for irrigation in addition to water spreading from surface diversions from the Umatilla River via a canal. Although data are not totally conclusive, it appears that the distributed canal water and return flow from surface-applied water recharges layer 1. This is indicated by the water-level rises in observation well G1 (fig. 21) from the mid-1970's to 1982. Some provisional data are available for the quantity of water distributed on the land surface from the canal, and ground-water pumpage data of reasonable accuracy are also available. However, data are not available to estimate the amount of return flow to layer 1. During model simulation the drawdown from 1962-74 was effectively simulated by adjusting values of the specific yield. Subsequently, the rises from 1975-82 were simulated by applying a recharge component to nodes in the model where canals are represented and then adjusting this component, by trial and error, until the model-calculated heads approximated the measured heads in layer 1. The maximum estimated pumpage for 1 year at any one node was about 4 feet per acre. About 70 percent of this water was simulated as return flow to layer 1.

Figures 22a, b, c, and d show model-simulated changes in water levels at the end of 1982. The calculated water-level declines for layer 1 for the period 1950-82 ranges from about 9 feet near Hermiston to about 24 feet in the area west of the Umatilla River. In layer 2, water-level declines of about 75 feet were calculated in an area near Hermiston. Water-level rises in layer 2 north of the Columbia River are due to imported surface water. Water-level rises of about 50 feet near the Columbia River are due to rising water levels behind the John Day and McNary Dams, as discussed earlier. Likewise, in layers 3 and 4 water-level rises of about 75 feet and 50 feet are due to leakage from these reservoirs. In layer 3 the largest water-level declines are about 200 feet in an area west of Butter Creek and about 100 feet in areas near Hermiston and north of the Columbia River. The largest water-level decline in layer 4 is near Butter Creek and is about 400 feet. Water-level declines of 100-400 feet in layer 4 are common in a large area from about 10 miles east of Hermiston to about 10 miles west and southwest of Butter Creek.

Model Budget

The flow rates as calculated for 1982 by the transient model are given below.

	Inflow (ft^3/s)		Outflow (ft^3/s)
Recharge	145.85	Drains	7.17
Leakage from streams	47.69	Leakage to streams	143.72
Boundaries	15.45	Boundaries	0.23
Infiltration from surface irrigation	36.43	Pumpage	181.77
Water taken from storage	102.29	Water added to storage	15.40
Totals	347.71		348.29

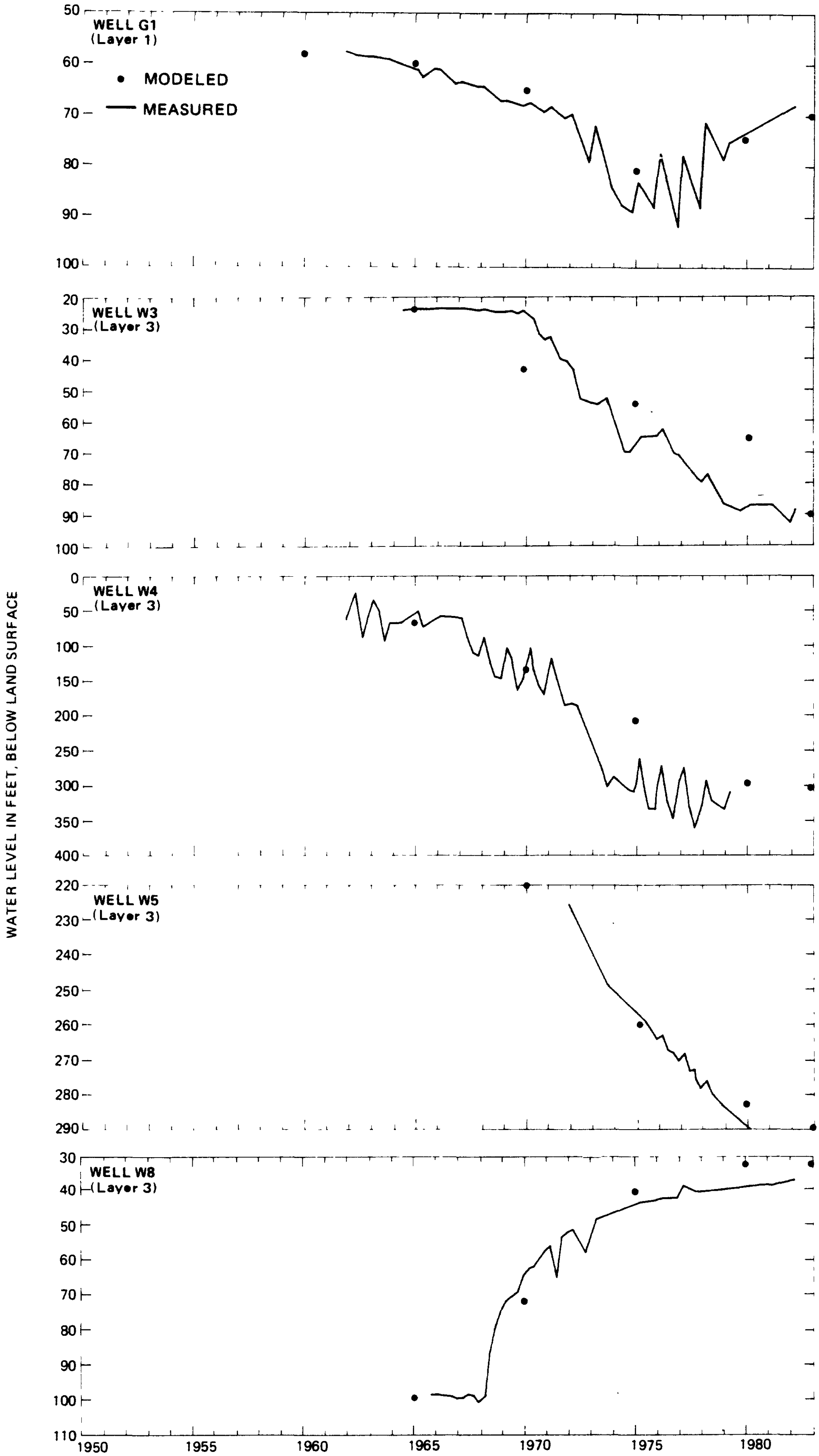


Figure 21.--Measured and model-calculated water levels, 1961-82.

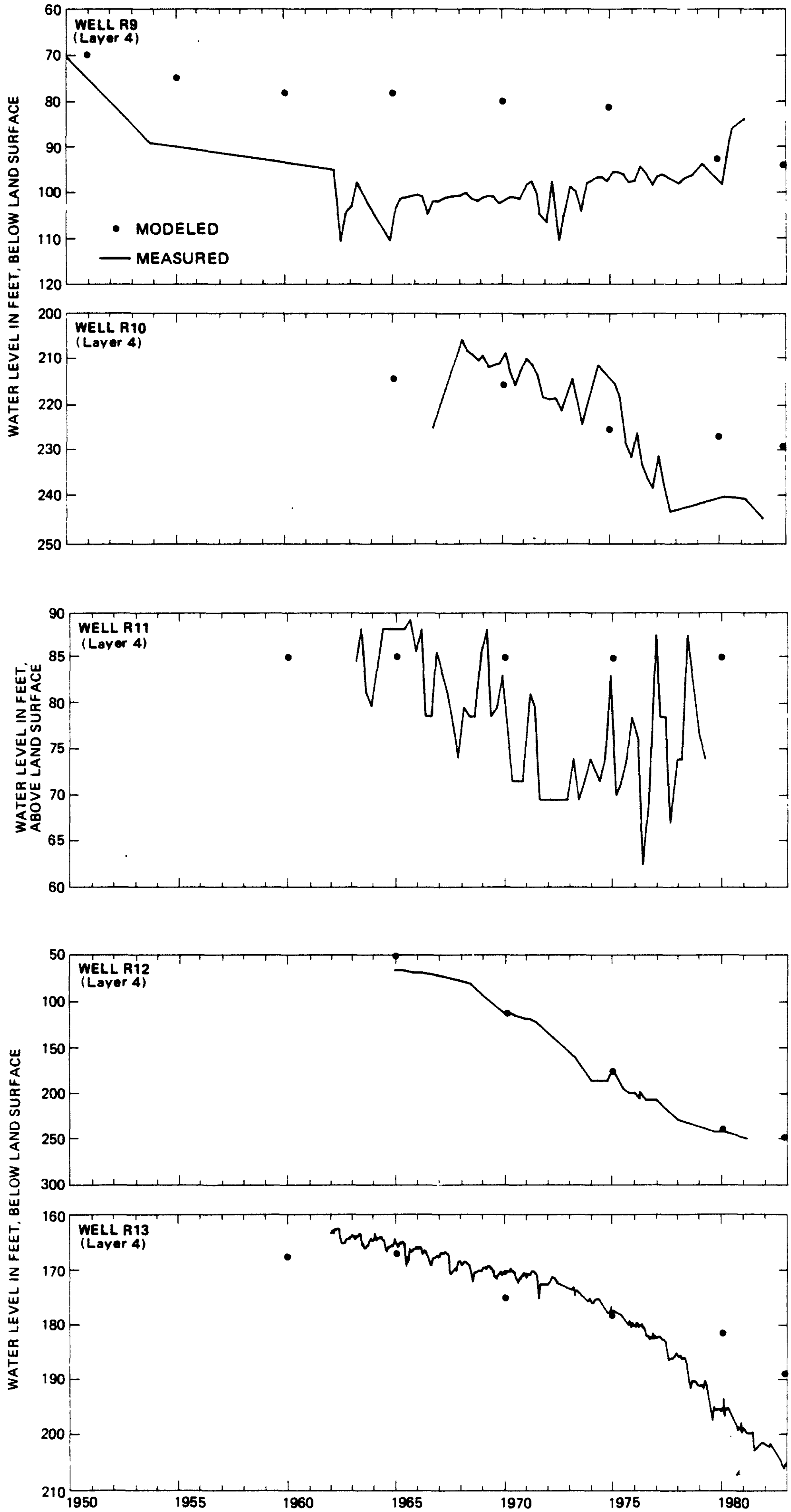
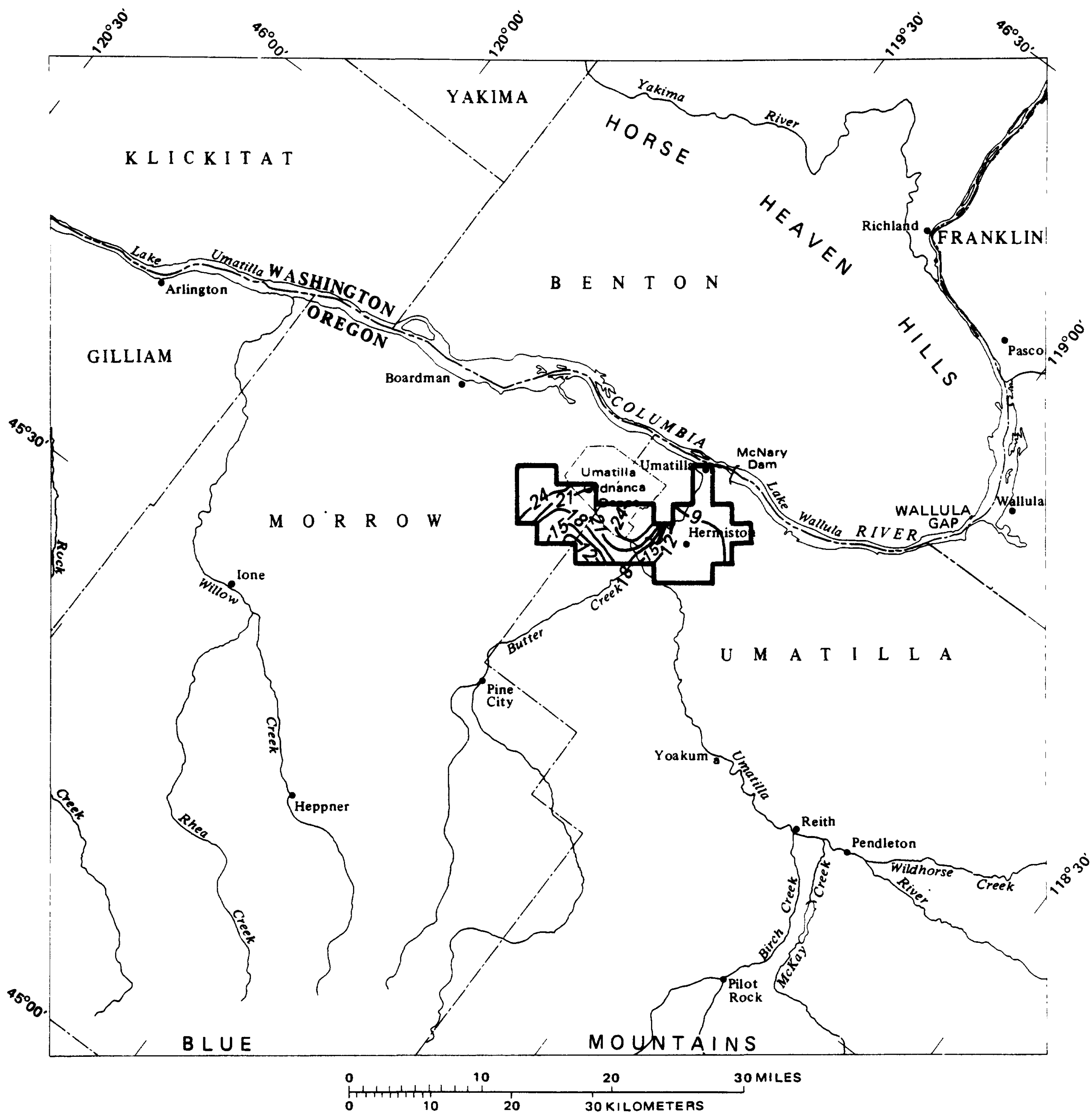


Figure 21.--Measured and model-calculated water levels, 1961-82--continued.

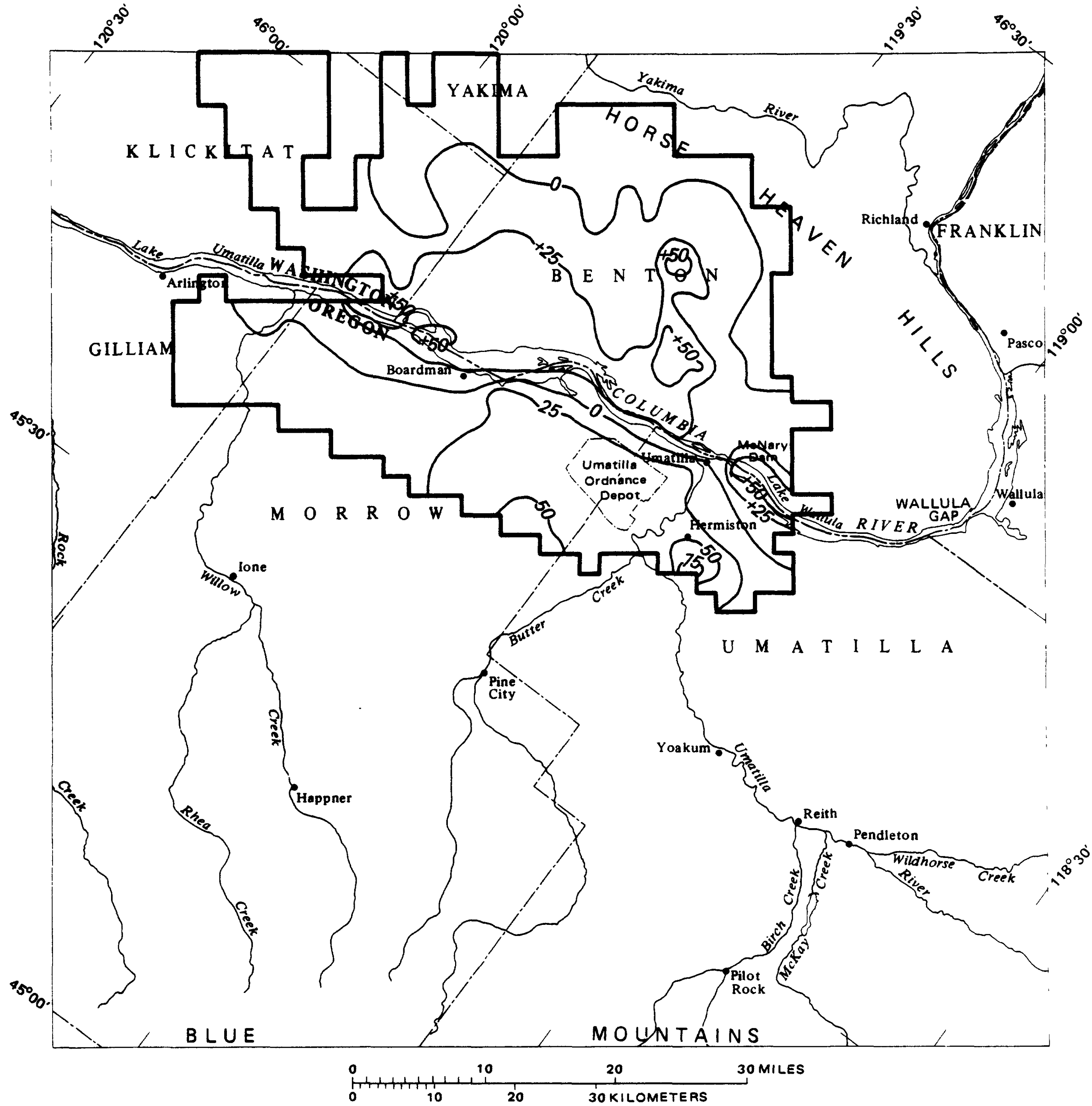


EXPLANATION

——15—— LINE OF EQUAL WATER-LEVEL CHANGE–Interval 3 feet.

━━━━ MODEL LAYER BOUNDARY

Figure 22a.--Distribution of water-level change as calculated during transient simulation 1950-82 for Layer 1.

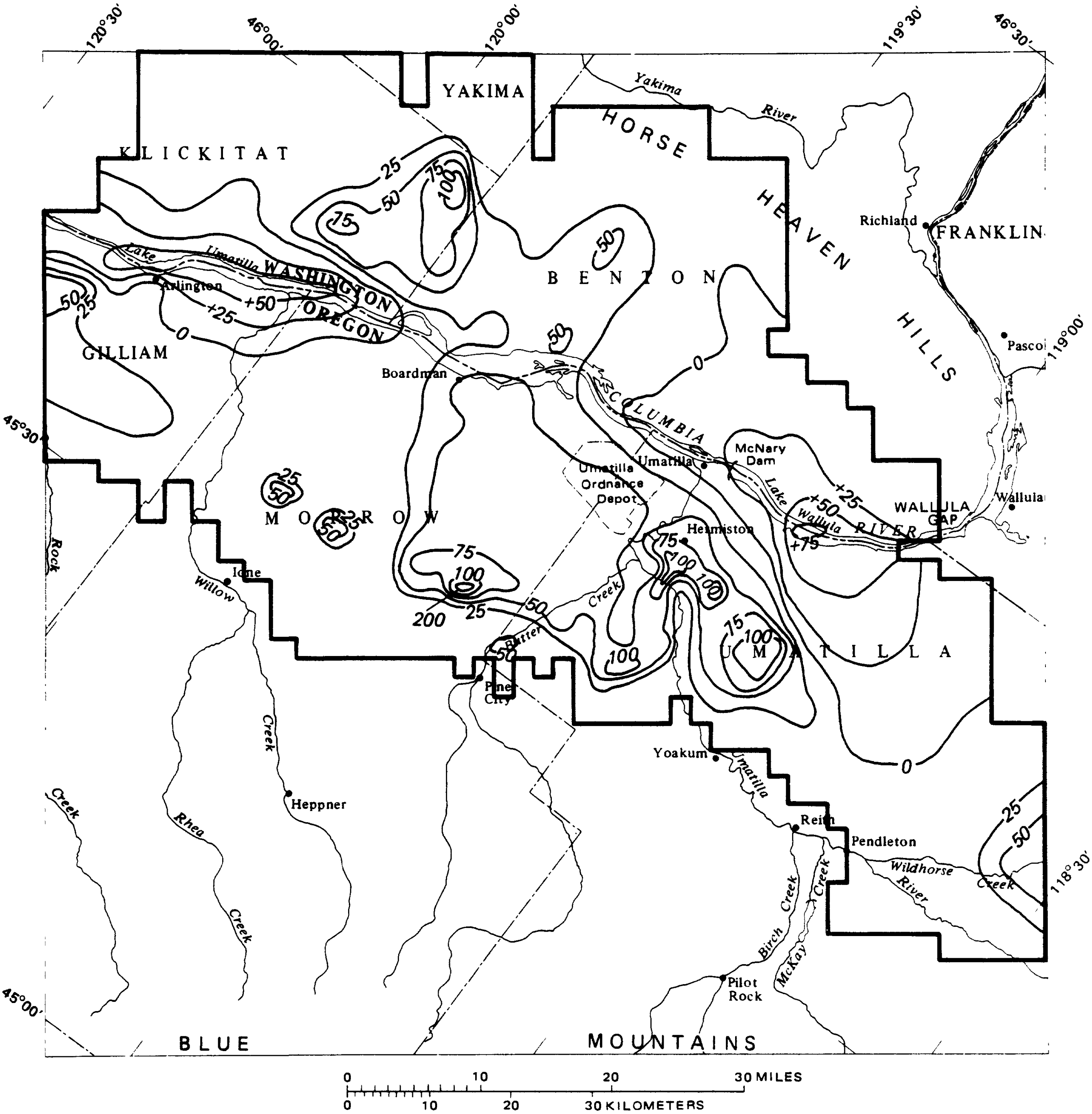


EXPLANATION

—— 25 —— LINE OF EQUAL WATER-LEVEL CHANGE–Interval 25 feet. (+) indicates water level rise.

———— MODEL LAYER BOUNDARY

Figure 22b.--Distribution of water-level change as calculated during transient simulation 1950-82 for Layer 2.

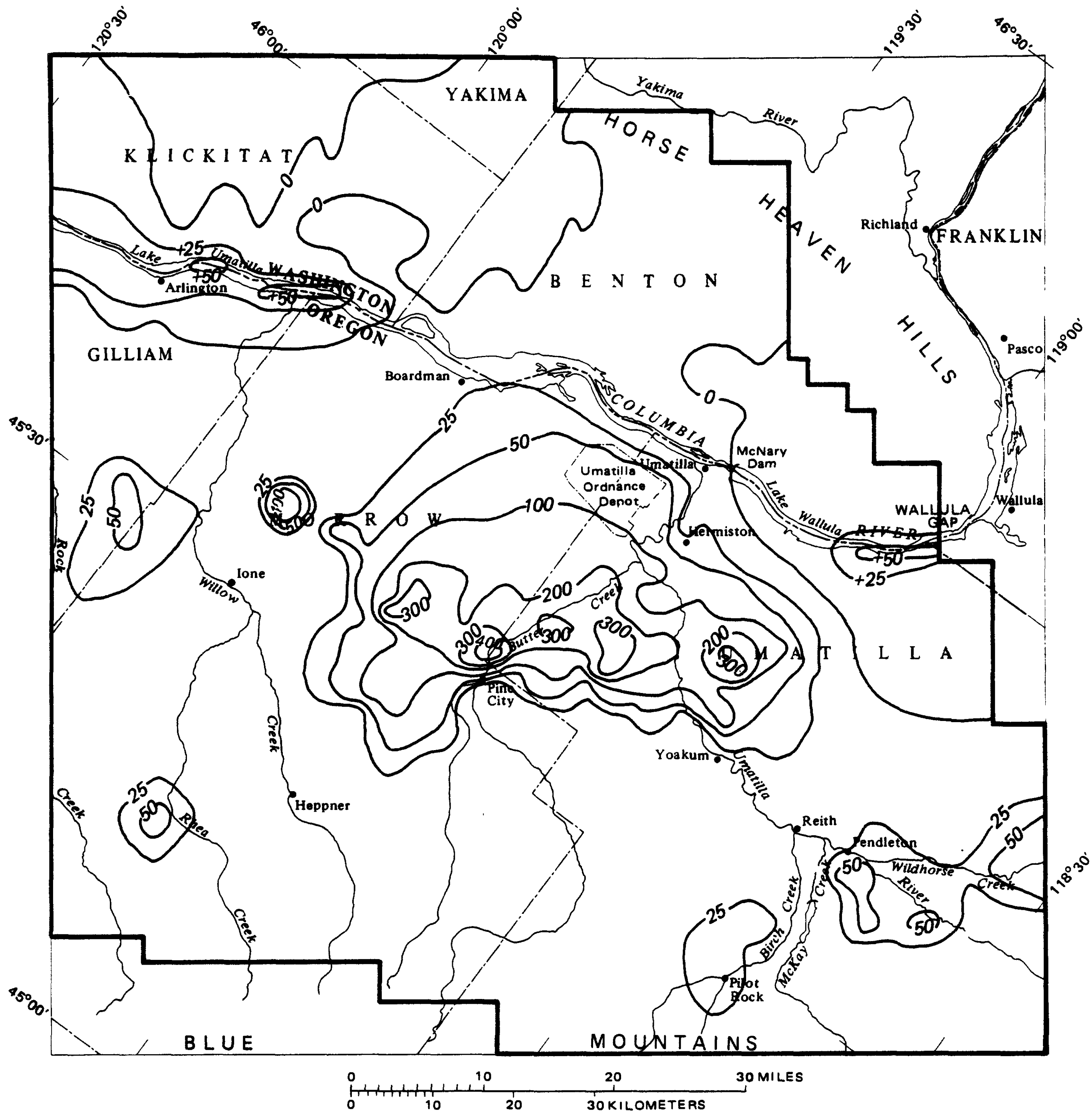


EXPLANATION

——100—— LINE OF EQUAL WATER-LEVEL CHANGE– Interval 25, 50, and 100 feet. (+) indicates water level rise.

MODEL LAYER BOUNDARY

Figure 22c.--Distribution of water-level change as calculated during transient simulation 1950-82 for Layer 3.



EXPLANATION

——300—— LINE OF EQUAL WATER-LEVEL CHANGE – Interval 25, 50, and 100 feet. (+) indicates water level rise.

——— MODEL LAYER BOUNDARY

Figure 22d.--Distribution of water-level change as calculated during transient simulation 1950-82 for Layer 4.

The difference between inflow and outflow is about 0.17 percent and is within the model limit (2 percent) for acceptable solutions. Acceptability is dependent on the assumptions that boundary conditions and adjustments to hydrologic parameters are correct. The transient simulation shows an increase at the end of 1982 of about 156 ft^3/s in inflow and outflow when compared with the steady-state simulation. The increased inflow is from ground-water storage (102 ft^3/s), infiltration from surface-water irrigation (36 ft^3/s), and leakage from streams (17 ft^3/s). The change in outflow is due to increased pumpage (about 182 ft^3/s), an increase in ground-water storage (about 15 ft^3/s), and a decrease in leakage to streams (about 41 ft^3/s). Flows to and from boundaries remained nearly unchanged.

The cumulative volumes for the water budget for the 1950-82 period are given below.

Inflow, in millions of acre-feet		Outflow, in millions of acre-feet	
Recharge	3.487	Drains	0.172
Leakage from streams	0.898	Leakage to streams	3.962
Boundaries	0.361	Boundaries	0.005
Infiltration from surface-water irrigation	0.188	Pumpage	1.632
Water taken from storage	1.098	Water added to storage	0.272
Totals	6.032		6.043

Sensitivity Analysis

The steady-state model was tested to determine how sensitive the model results are to changes in lateral and vertical hydraulic conductivity and to changes in recharge. The transient model was tested to determine its sensitivity to changes in storage coefficient. The results of the sensitivity analysis are shown in figure 23.

The results of the sensitivity analysis show that the steady-state model is most sensitive to changes in lateral hydraulic conductivity when the calibrated values were increased by as much as 20 percent. When the values were decreased by 20 percent, the model is most sensitive to recharge. For an increase of 20 percent in the lateral hydraulic conductivity values, the absolute value of the average water levels changed by 41 percent from calibrated values. For a decrease in recharge of 20 percent, the average water levels changed by 32 percent.

The results of the steady-state sensitivity analysis also show that the net stream leakage (outflow-inflow) is most sensitive to changes in recharge when the calibrated values are increased or decreased. The stream leakage is relatively insensitive to changes in lateral and vertical hydraulic conductivity. A change in recharge of ±20 percent increased or decreased the net stream leakage by ±17 percent.

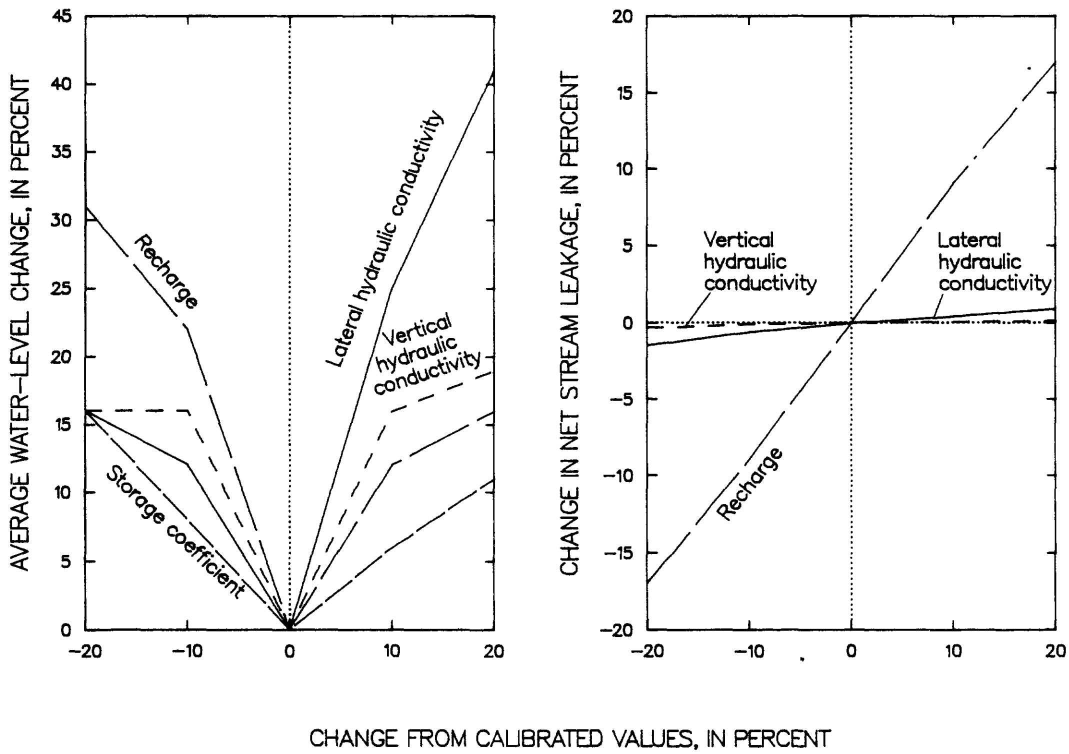
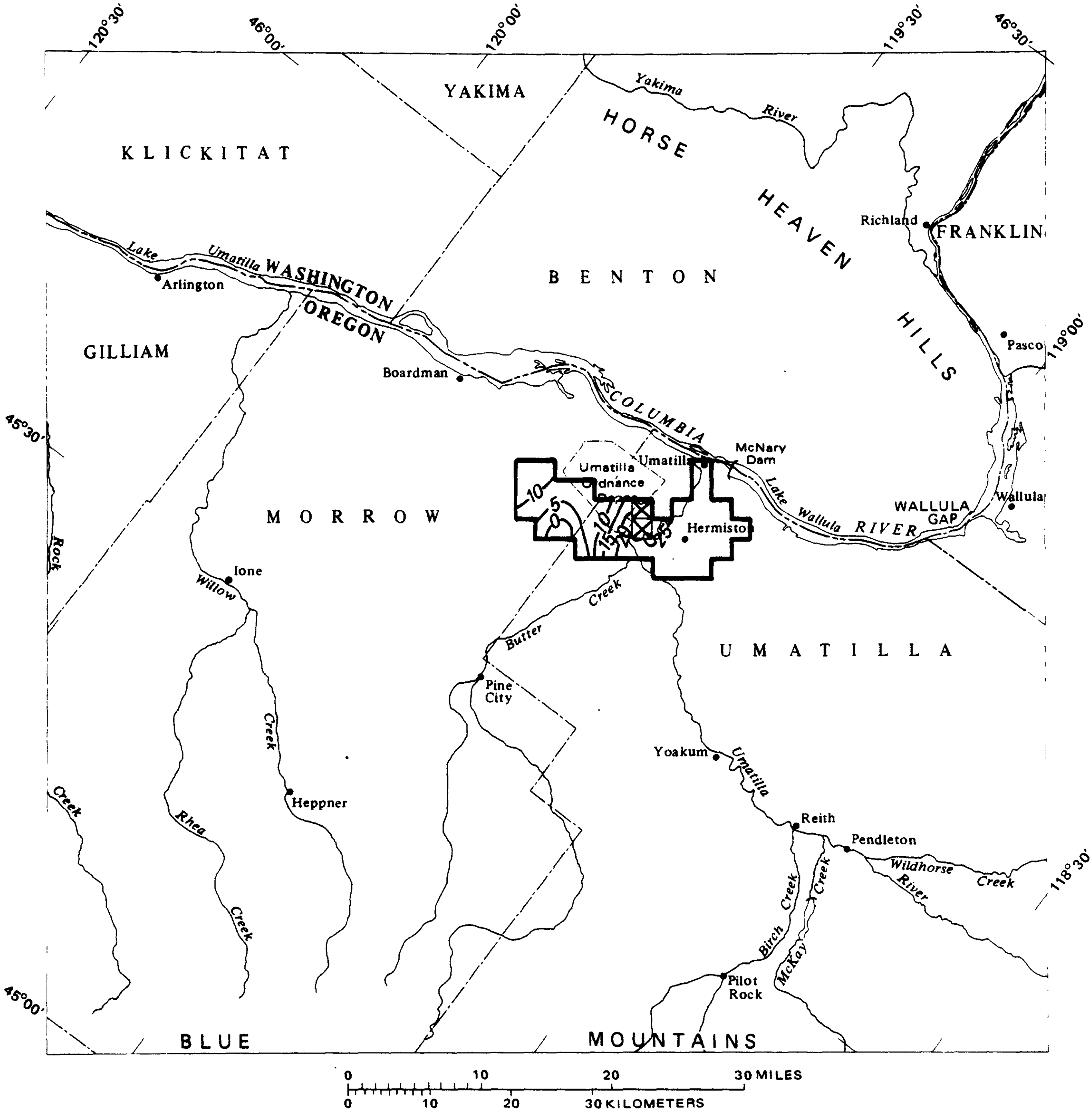


Figure 23.--Model sensitivity analysis.

For th.; transient model analysis it was found that increasing the storage coefficient values by 20 percent increases the absolute values of average water levels by 11 percent from the calibrated values. Decreasing the storage coefficient values by 20 percent increased the absolute values of average water levels by 16 percent from calibrated values.

Predicted Water-level Changes

The completed model was used to predict the water-level changes for the period 1983-2000 in each of the four layers, using the same hydrologic and boundary conditions as simulated for the 1950-82 period. The model was run for the period 1950-2000 with the actual calculated pumpage from 1950-1982 and with constant 1982 pumpage from 1983 to 2000. The resulting water-level changes for the period 1950-2000 are shown in figures 24a, b, c, and d. Water-level declines of about 1,000 feet were calculated for layer 4 north of Heppner between Willow and Butter Creek, and about 700 feet in areas north of Pine City near Butter Creek and east of Pine City between Butter Creek and the Umatilla River. Water-level declines of 100-700 feet were calculated for an area of about 450 square miles centered around lower Butter Creek. Near the east, northwest, and north part of the modeled area in layer 4, water levels remained unchanged or slightly rose. The largest water-level declines in layer 3, about 300 feet, were calculated for an area in Washington north of the Columbia River; and about 250 feet of decline was calculated for an area between the Umatilla River and Butter Creek in Oregon. Several nodes were dewatered as a result of pumping in layer 3 during the 1950-2000 period; these are shown in figure 24c.



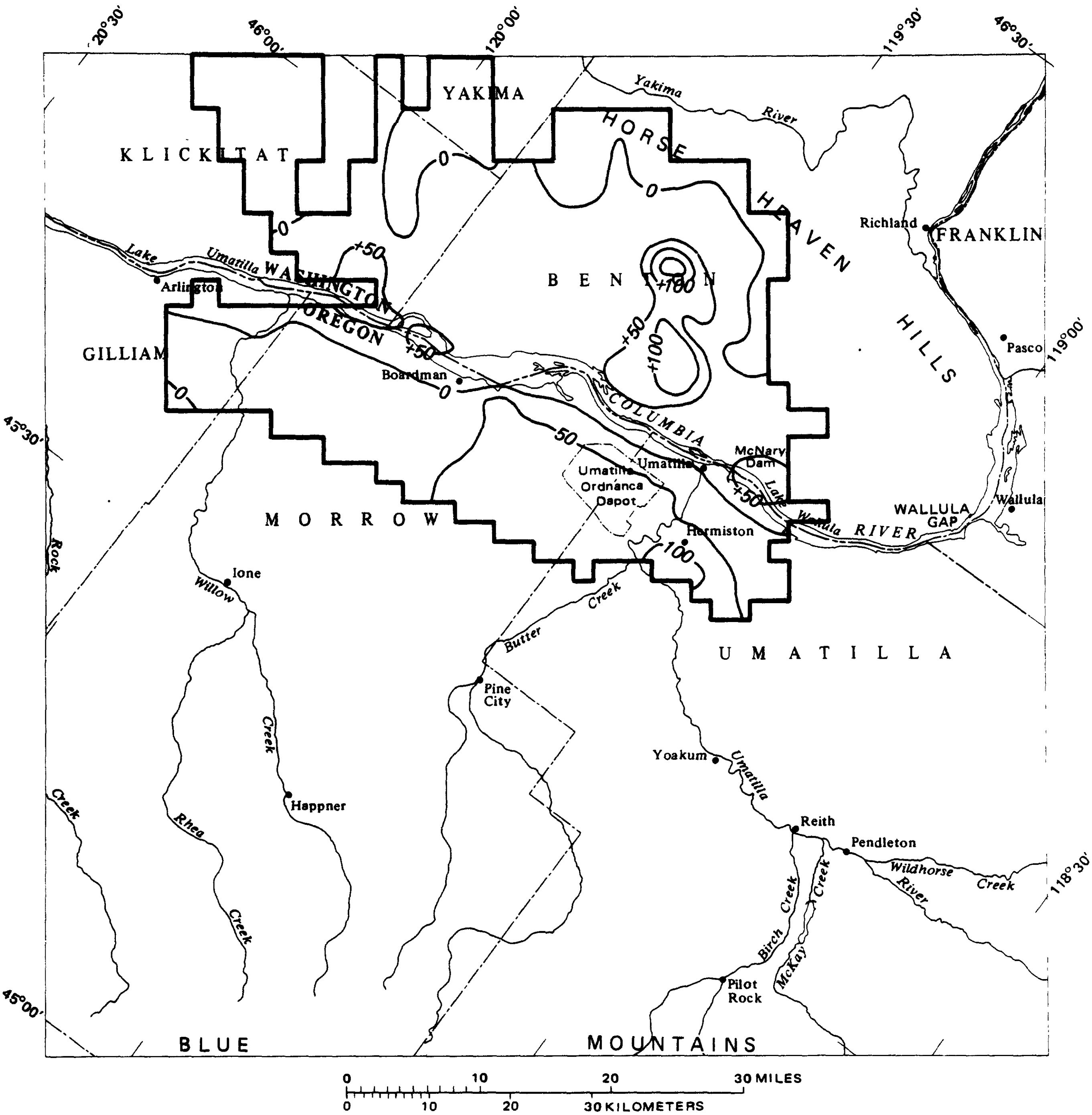
EXPLANATION

—— 10 —— LINE OF EQUAL WATER-LEVEL CHANGE–Interval 5 feet.

⊠ NODE DEWATERED DURING SIMULATION

▬▬ MODEL LAYER BOUNDARY

Figure 24a.--Distribution of water-level change as calculated during transient simulation 1950-2000 for Layer 1.

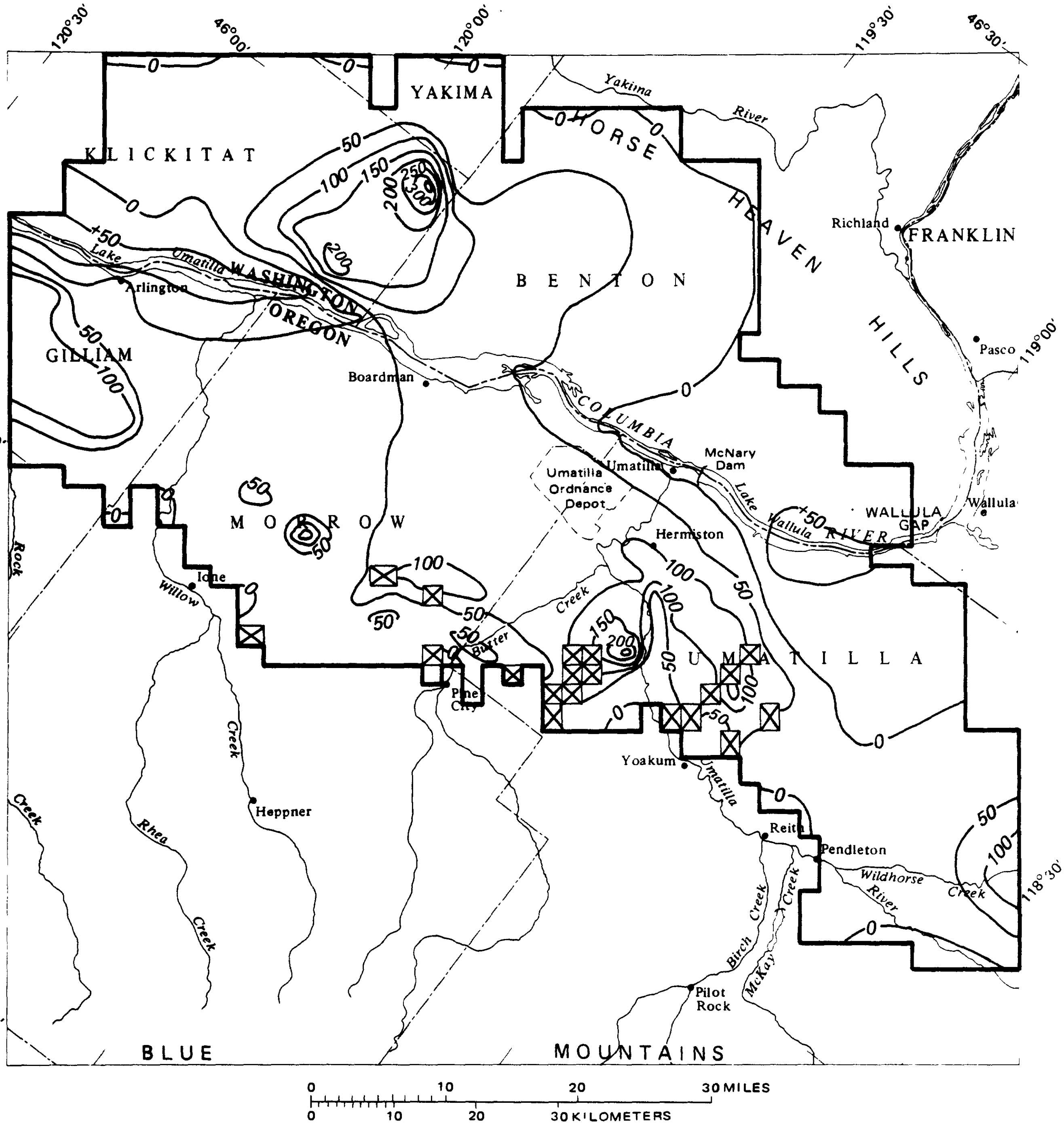


EXPLANATION

—— 50 —— LINE OF EQUAL WATER-LEVEL CHANGE–Interval 50 feet. (+) indicates water level rise.

▬▬▬▬ MODEL LAYER BOUNDARY

Figure 24b.--Distribution of water-level change as calculated during transient simulation 1950-2000 for Layer 2.



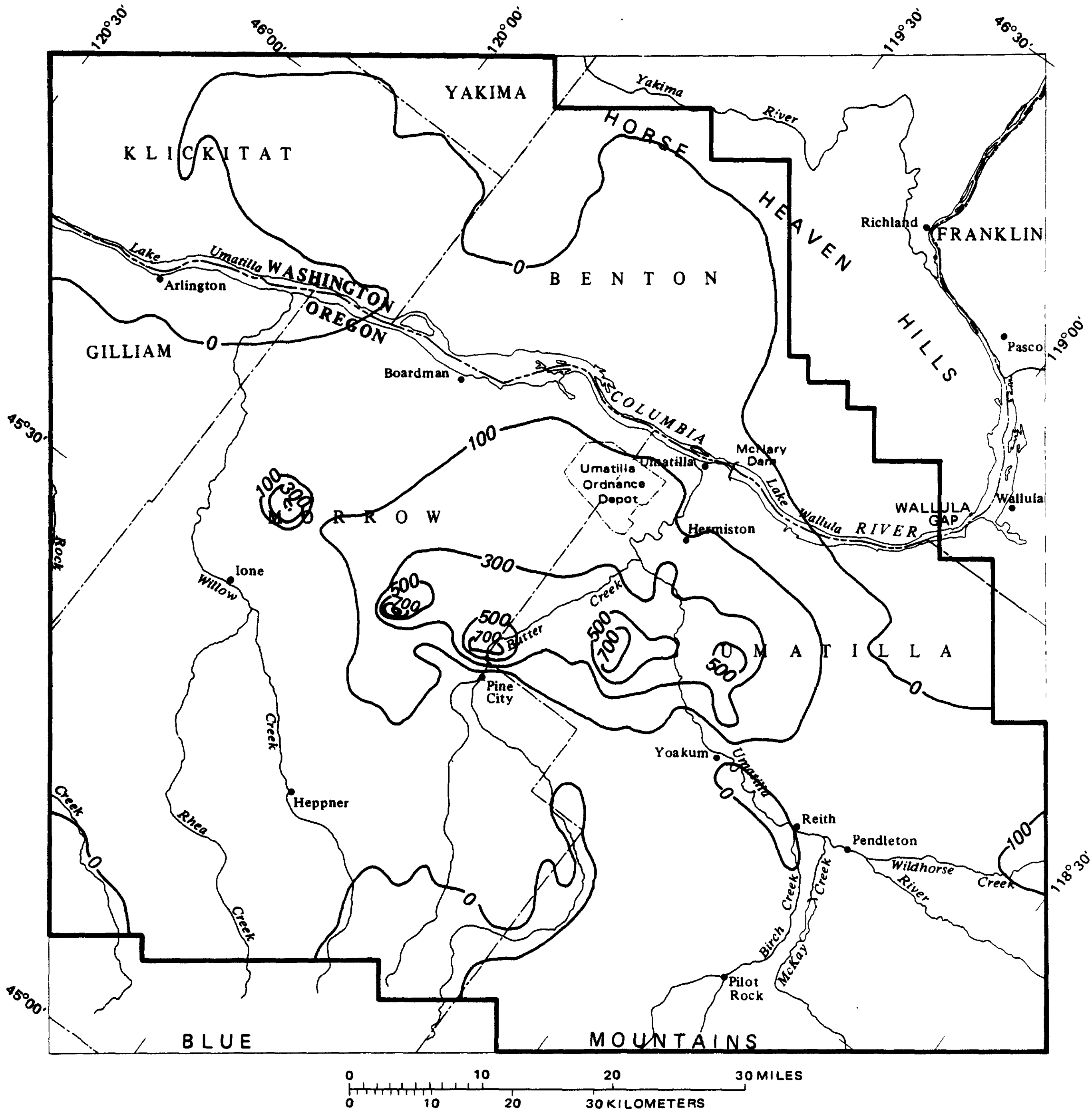
EXPLANATION

—50— LINE OF EQUAL WATER-LEVEL CHANGE–Interval 50 feet. (+) indicates water-level rise.

⊠ NODE DEWATERED DURING SIMULATION

▬ MODEL LAYER BOUNDARY

Figure 24c.--Distribution of water-level change as calculated during transient simulation 1950-2000 for Layer 3.



EXPLANATION

——300—— LINE OF EQUAL WATER-LEVEL CHANGE–Interval 100 and 200 feet.

▬▬▬ MODEL LAYER BOUNDARY

Figure 24d.--Distribution of water-level change as calculated during transient simulation 1950-2000 for Layer 4.

In layer 2 the largest water-level declines were calculated as about 125 feet near Hermiston. Layer 2 shows about 150 feet of water-level rise in areas north of the Columbia River, which is due to imported surface-water irrigation. The largest calculated decline in layer 1 was about 25 feet, west of the Umatilla River. Declines of about 20 feet were calculated in the area east of the Umatilla River near Hermiston. Two nodes went dry during the simulation; these are shown in figure 24a. No change or slight rises in water levels were calculated in the southwest part of layer 1 near the area where canals and return flow from ground-water pumpage provide recharge to layer 1 as explained earlier.

Calculated water-level changes do not represent changes at specific wells; rather, they are nodal changes and they indicate the effect of pumping throughout each cell as defined by the model grid. Water-level changes at specific wells would be greater than those shown for nodal changes. During the period 1983-2000, some of the pumping nodes became dewatered; the result was a decrease of as much as 15 percent in pumping rates for the period. When a node becomes dewatered, the model deletes that node from the analysis, along with any pumpage that may have been assigned the node.

The effect of long-term pumping at various pumping rates on water levels at one model node is shown in figure 25. This node was selected because it showed the largest model-calculated water-level declines for the calibration period (1950-82). Thus the predicted water-level declines shown in figure 25 are extreme values.

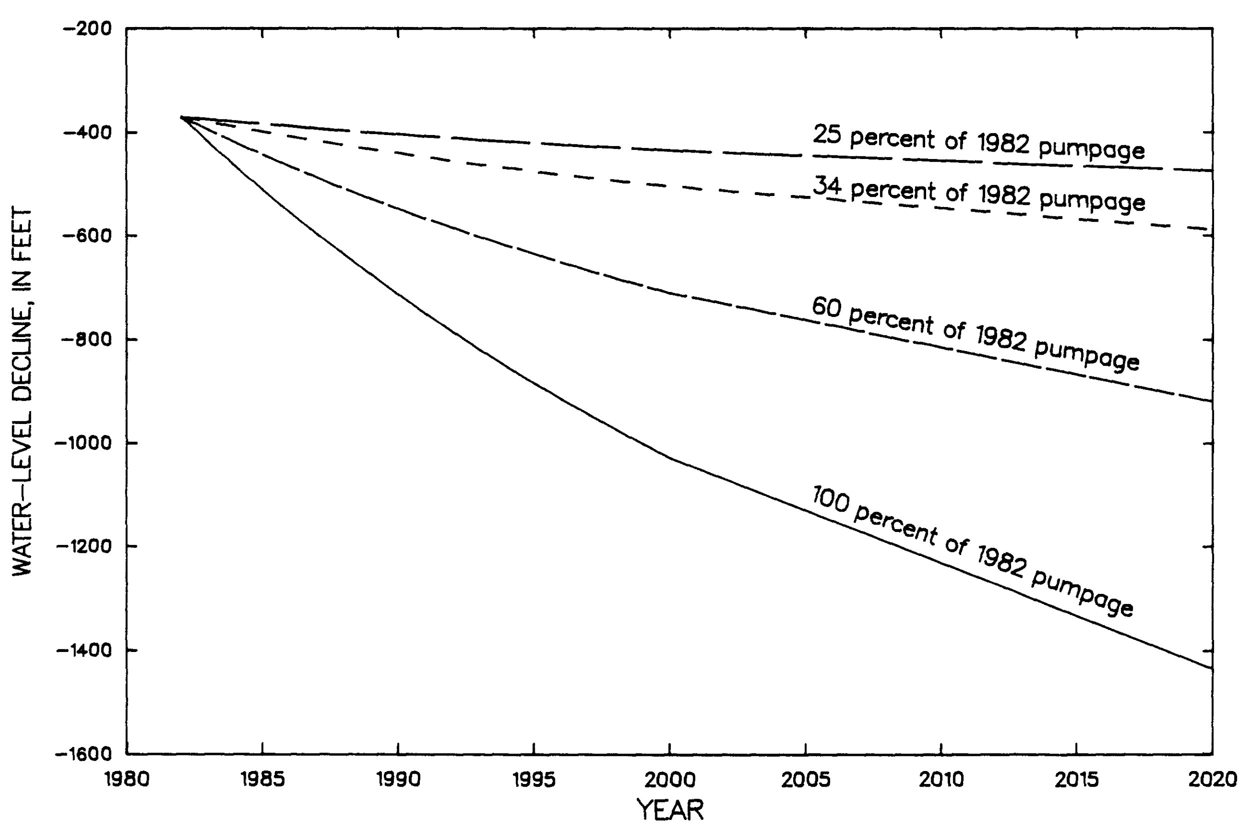


Figure 25.--Predicted water-level decline at model-node row 12, column 17, in layer 4, 1982-2020.

The difference between the model-calculated water-level decline at 25 percent of the 1982 pumping rate and the decline at 100 percent is about 600 feet at the end of 2000 and about 950 feet at the end of 2020. The rate of decline of water levels at the 25-percent pumping rate is about 3.6 feet per year for the period 1983-2000, as compared to about 2.0 feet per year for 2000-2020. At 100 percent of the 1982 pumping rate, the water-level decline for 1983-2000 is about 37 feet per year and is about 20 feet per year for 2000-2020. As shown in figure 25, at the 25-percent rate, near-equilibrium of water levels would be reached sooner than at the other rates of pumping.

The model-calculated water-level declines shown in figure 25 are based on the assumptions that long-term recharge from precipitation is constant, and that pumping would continue despite water-level declines greater than 1,000 feet.

NEED FOR FUTURE WORK

Future work in the study area needs to focus on several items related to the collection of additional data for better understanding of the mechanisms that control ground-water movement. Additional data would enhance model refinement and utilization.

A better definition of the stream-aquifer relation is needed for identifying the naturally occurring gaining and losing reaches of major streams (except the Columbia River) and the quantity of gains and losses. These data are needed for more accurate estimates of streambed conductances and stream stages as used in the model. Seepage from ground water that occurs from cliff faces above stream channels needs to be identified and quantified. This seepage represents an unknown quantity that was assumed negligible for this study, but it needs to be verified. All major stream diversions, return flow to streams, and flow to the ground-water system from surface-applied water also need to be measured.

Monitoring of ground-water pumpage and water levels needs to be continued. In particular, the installation of multiple piezometers in each of the four layers is needed for adequate definition of head distribution.

SUMMARY

Four geohydrologic units (layers) in the study area were delineated as to extent, thickness, hydrologic properties, recharge, and discharge. The maximum thickness of the Grande Ronde Basalt (layer 4) was estimated to be about 8,000 feet, the Wanapum Basalt (layer 3) about 1,000 feet, the Saddle Mountains Basalt (layer 2) about 800 feet, and the sedimentary deposits (layer 1) about 150 feet.

Steady-state and transient models were constructed and calibrated for the ground-water flow system. The steady-state model simulates conditions prior to major ground-water development, about 1950, and the transient model simulates conditions from 1950 to 1982.

Lateral hydraulic-conductivity values for each layer were estimated from specific-capacity data of wells. Transmissivity values were model calculated from hydraulic-conductivity and saturated-thickness values for each layer. Vertical hydraulic-conductivity values divided by thickness (k/b) values between layers were derived empirically from previous studies in the Columbia Plateau. Likewise, initial storage-coefficient values for each model layer were estimated from earlier studies. The initial values of the hydrologic properties were used to develop the model, but were subsequently adjusted during the calibration procedure. Final model-adjusted values of transmissivity range from 0.5 to 2 ft^2/s in layer 1, and from 0.005 to 0.25 ft^2/s in layers 2-4. Values of k/b range from 1×10^{-10} to 2×10^{-11} sec^{-1} for the interval between layers 1 and 2; 5×10^{-10} to 1×10^{-11} sec^{-1} for the interval between layers 2 and 3; and 3×10^{-13} to 1×10^{-14} sec^{-1} for the interval between layers 3 and 4. Values for storage coefficient average 0.15, 0.01, 0.0045, and 0.0050 for layers 1-4 respectively.

On the basis of model analysis, the major components of the water budget for inflow before ground-water development were recharge from precipitation, about 146 ft^3/s or 106,000 acre-feet per year (acre-ft/yr); ground-water leakage from streams, about 31 ft^3/s (22,400 acre-ft/yr); and subsurface flow entering the boundaries of the study area about 15 ft^3/s (10,900 acre-ft/yr). The major outflow component of the budget was leakage to streams, about 185 ft^3/s (134,000 acre-ft/yr).

During the period 1950-82, maximum water-level declines of about 300 feet occurred in parts of the Grande Ronde Basalt unit (layer 4). During the same period ground-water withdrawal increased to about 182 ft^3/s (132,000 acre-ft/yr). On the basis of model analysis, return flow from surface irrigation increased 36 ft^3/s (26,100 acre-ft/yr); ground-water storage decreased by about 87 ft^3/s (63,000 acre-ft/yr); ground-water leakage to streams decreased by about 42 ft^3/s (30,400 acre-ft/yr); and leakage from streams increased by about 17 ft^3/s (12,300 acre-ft/yr).

Predictions with the transient model indicate that maximum water-level declines for the year 2000 are about 1,000 feet in part of layer 4. This prediction was based on the premise that ground-water pumpage was the same as that for 1982 (182 ft^3/s) and did not increase or decrease during 1983-2000.

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