Specific Storage and Storage Coefficient

In a confined aquifer, the specific storage is related to the porosity and compressibility of the rock and water by the equation:

$$S_s = \gamma\ (\phi\ C_w + C_r\ ) \qquad (1)$$

where

S_s = specific storage; s
γ = specific weight of water;
ϕ = porosity;
C_w= compressibility of water; and
C_r = compressibility of rock.

Porosity of the sandstone strata in the eastern part of the area commonly ranges from 10 to 25 percent and averages about 20 percent. Compressibility of sandstone similar to that in the study area is about 1.5 $x10^{-6}$ in^2/lb (Fatt, 1958). These data, when used with the characteristics of water in the above equation, yield a specific storage of 9 $x10^{-7}$ ft^{-1}. This value is the volume of water the confined water-yielding sandstones release from or take into storage, per unit volume of rock, per unit change in head due to the compressive character of the water and rock.

In an unconfined aquifer, the volume of water released from or taken into storage by this process is insignificant when compared to the volume of water released by gravity drainage or filling of pore space in the rock. The storage coefficient of an unconfined aquifer is approximately equal to the specific yield of the water-yielding material and may be several orders of magnitude larger than the confined storage coefficient. No data are available to define the specific yield of the sandstones in the study area. However, sandstone that has a porosity of 20 percent could be expected to have a specific yield of about $1x10^{-1}$.

Storage coefficient in a confined aquifer is equal to the product of specific storage and aquifer thickness. Thus, a 100-ft-thick confined aquifer in the Twentymile Sandstone, or Trout Creek Sandstone Members, that has a specific storage of 9 $x10^{-7}$ per foot would have a storage coefficient of 9 $x10^{-5}$. Storage coefficient in an unconfined aquifer in either unit would be about $1x10^{-1}$.

Three storage-coefficient values obtained from pumping-well aquifer tests in the basal Williams Fork aquifer ranged from $2x10^{-4}$ to $lx10^{3-}$. The accuracy of such tests generally are poor, but results indicate confined conditions exist in this aquifer.

GROUND-WATER MOVEMENT

Ground-water movement occurs as a result of hydraulic-head differences in an aquifer. The head in an aquifer at a well is calculated from water-level-measurement data and normally is expressed in terms of the altitude of the standing water level in the well. Head determinations at many different sites define the altitude and areal distribution of head in the aquifer (a

potentiometric surface). Thus, the potentiometric surface indicates the altitude and distribution of the standing water level in wells. Ground water moves from points of higher head (near areas of ground-water recharge) to points of lower head (near areas of ground-water discharge) in a direction that generally is perpendicular to the equipotential lines on a potentiometric-surface map. Potentiometric-surface data and aquifer-characteristics data may be used to calculate the rates and distribution of recharge and discharge (water budget) in the aquifer. Detailed water-budget calculations normally are performed on a digital computer because of their number and complexity.

Ground-Water Recharge

Climate, vegetation, and geology have a direct effect on ground-water recharge. Because potential evaporation exceeds mean annual precipitation in the study area, most infiltration occurs only during snowmelt or intermittent periods of intense rainfall. Part of the water entering the soil is consumed by vegetation and lost to the atmosphere through transpiration. This process (evapotranspiration) is enhanced on south-facing slopes where greater insolation produces maximum evaporation and transpiration. The lower angle of incidence on north-facing slopes produces less evapotranspiration and increases the potential for ground-water recharge. Most recharge in the study area occurs in the spring at the higher altitude margins of the area when snowmelt eventually saturates the ground and enables deep percolation. Some recharge may result from thunderstorms in the summer; however, most of this water is lost to evaportanspiration and little can infiltrate to depth.

The ability of water to percolate to depth and recharge the bedrock aquifer also is controlled by the lithology of the soil and the underlying bedrock formations. Clayey soils or shaley bedrock commonly are of very low permeability and will retard water movement. By contrast, sandy soil or sandstone outcrops or subcrops are relatively permeable and may allow water movement to depth. Aquifer recharge zones in the study area are defined by the outcrop or subcrops of permeable bedrock units within the Williams Fork Formation and the underlying Trout Creek Sandstone Member. In the eastern part of the area, the middle member of the Williams Fork Formation is shale and is not considered to be a recharge zone. The extensive outcrops of the Lewis Shale and outcrops of the Iles Formation shale underlying the Trout Creek Sandstone Member likewise are not considered recharge areas.

The rate of recharge can be estimated from results of previous studies. Watershed modeling techniques were used by Weeks and others (1974) to define a relation between the rates of precipitation and ground-water recharge in the Piceance Basin (a mountainous area 60 mi southwest of Craig). The relation between precipitation and ground-water recharge for the Piceance Basin (fig. 32) initially was defined by a model that uses precipitation, solar insolation, and temperature data in calculating surface runoff and deep percolation (recharge) in a watershed of varied slope, aspect, vegetative cover, and soil type. Subsequent modeling of the ground-water flow system in the Piceance Basin indicated that the relation correctly defined the rate and distribution of recharge needed to properly simulate the geohydrology of the aquifers. Similar

surface-water modeling procedures were used by Parker and Norris (1989) in the Foidel Creek watershed. The precipitation-recharge results from Foidel Creek are less extensive than those from the Piceance Basin, but indicate that the relation defined in the Piceance Basin also applies to the Twentymile Park study area (fig. 32).

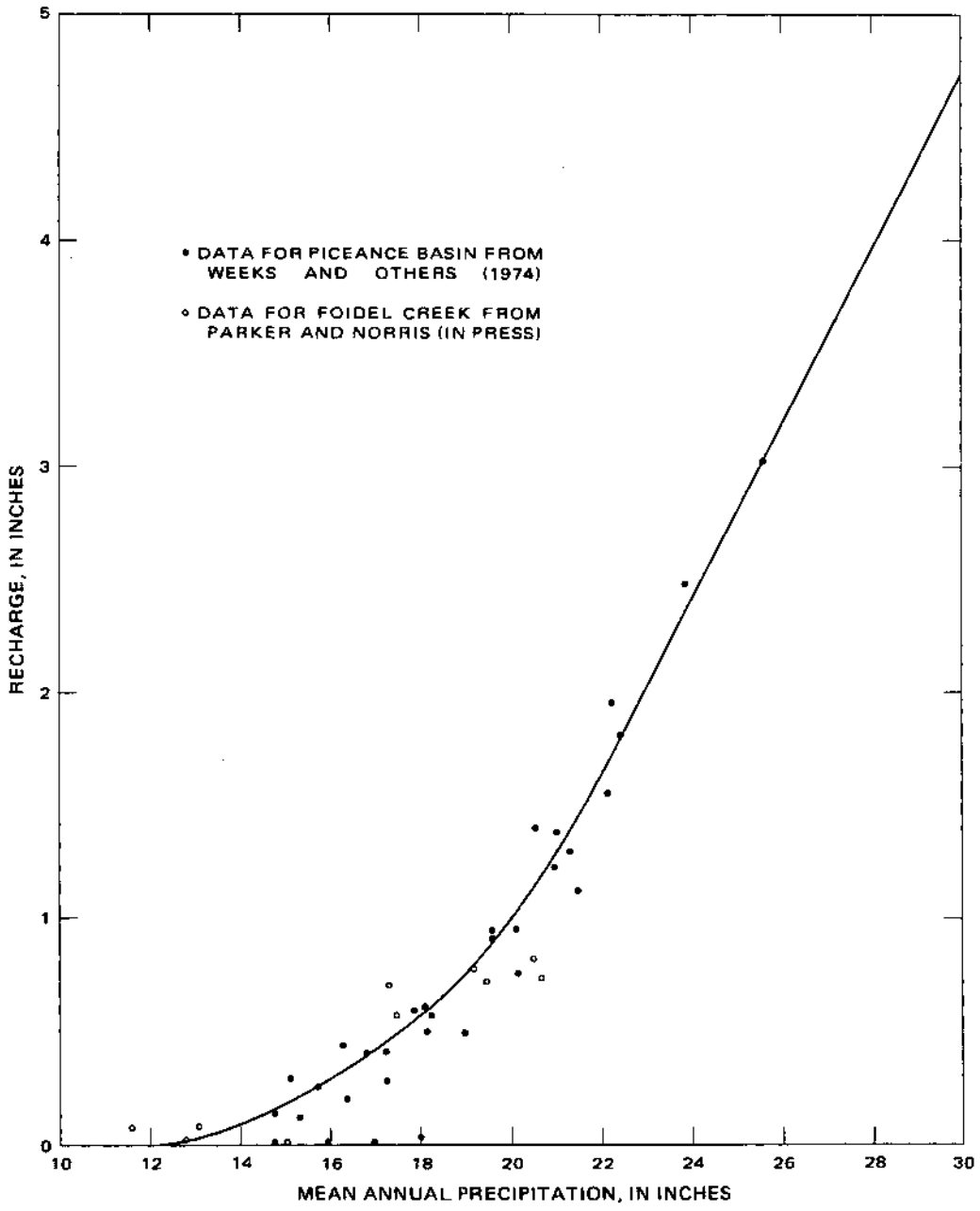

Figure 32.--Relation between mean annual precipitation and ground-water recharge.

Mean annual recharge to the bedrock aquifers in the eastern part of the study area was estimated by converting the precipitation rates shown on the isohyetal map (fig. 19) to potential recharge rates by using the relation shown in figure 32. The potential recharge rates then were multiplied by the size of the associated segment of the recharge zone to obtain distributed recharge. This technique produced larger estimates of recharge in areas that contain large recharge zones, and effects of altitude and aspect of the recharge zone were incorporated because these two factors are used in producing the isohyetal map. Mean annual recharge to the bedrock aquifers was 2.8 ft^3/s, or about 0.31 in/yr, in the 123 mi^2 eastern study area. Subsequent ground-water modeling in the area indicated that this rate of recharge is compatible with the known hydrology of the bedrock aquifers.

Potentiometric Surface

The potentiometric surfaces in the eastern part of the study are defined by water-level measurements in wells completed in the basal Williams Fork aquifer and the Twentymile aquifer. Too few data are available to define the potentiometric surface in the Trout Creek aquifer. The potentiometric-surface maps (figs. 33 and 34) are based on about 120 water-level measurements selected from an original group of more than 2,500 measurements. Most of the original measurements were made by mining company personnel and were released as part of public documents submitted with mine permit applications. Other measurements were made by U.S. Geological Survey and other Federal or State agency personnel. The maps are constructed to represent predevelopment (or steady state) water-level conditions in the two aquifers. The water-level measurements used on these maps were chosen to ensure that (1) they represent heads that are little affected by drawdown from mines or discharging wells, and (2) the measurements were obtained from wells completed only in one aquifer. These requirements eliminated many measurements from consideration even though heads in most of the area still are near predevelopment levels.

Seasonal changes in depth to water are most pronounced near the aquifer outcrops where recharge from spring snowmelt may cause 10 to 30 ft of water-level rise in the shallow aquifers. Seasonal changes in water level in the more deeply buried aquifers generally are less than 5 ft/yr but may exceed 50 ft/yr near a few pumping wells.

Heads in the basal Williams Fork aquifer are above land surface in much of Twentymile Park and along low-lying areas in most stream valleys (fig. 33). Wells completed in the basal Williams Fork aquifer in these areas can flow at land surface, although most wells are shut in to prevent loss of water from the aquifer. The potentiometric surface of the basal Williams Fork aquifer (fig. 33) ranges in altitude from more than 7,200 ft near recharge areas along parts of the aquifer margin to less than 6,500 ft near the two discharge areas near Hayden and the downstream reach of Fish Creek. The sinuous shapes of the potentiometric contours near Fish Creek, Foidel Creek, Grassy Creek, and Trout Creek are the result of ground-water recharge from, or discharge to, the streams. In some areas, the effects of recharge or discharge extend through the middle confining layer or through both the Twentymile aquifer and the middle confining layer. Heads in the basal Williams Fork aquifer generally were near steady-state conditions in 1986 except near mined areas or near uncapped

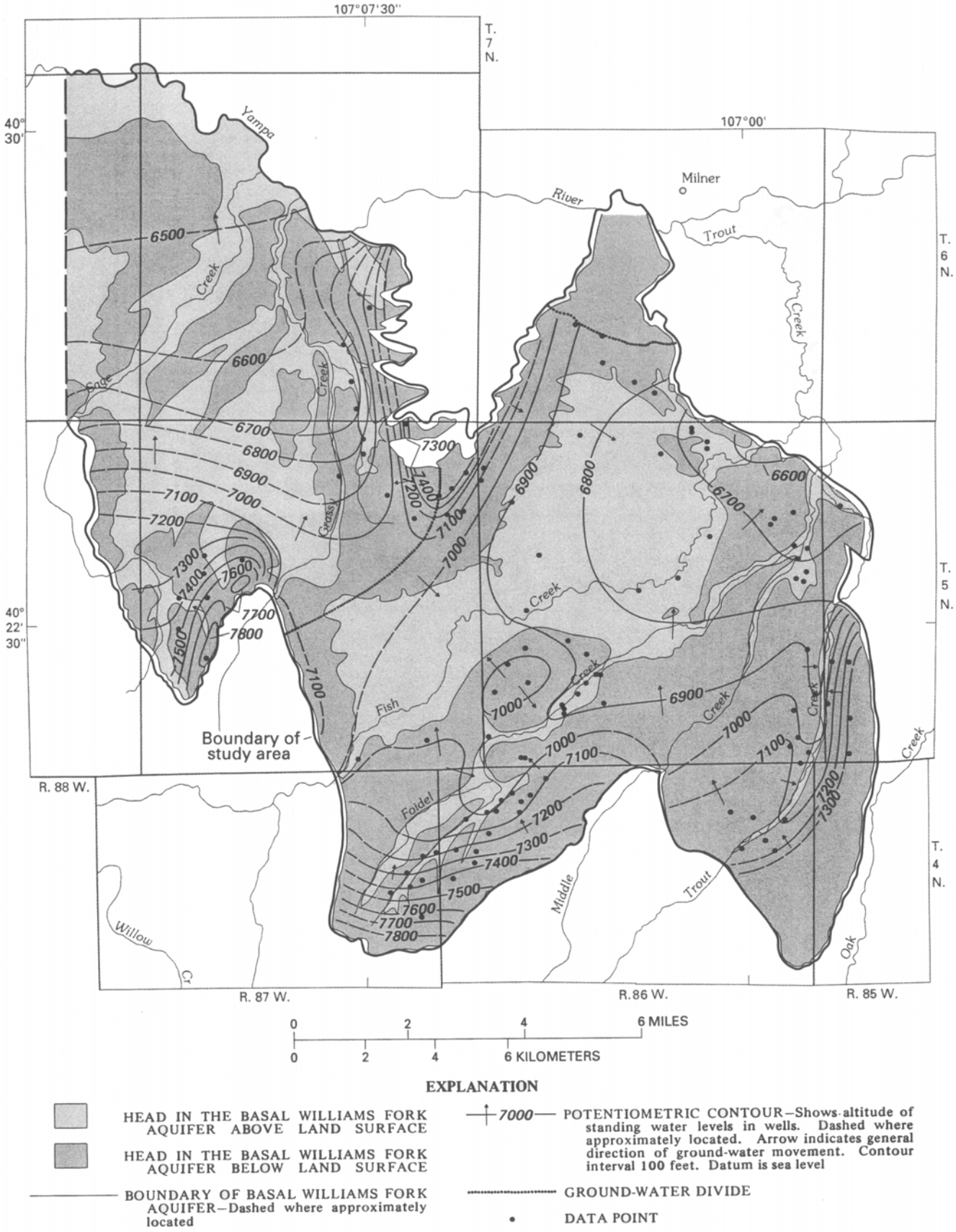

Figure 33.--Approximate steady-state potentiometric surface of the basal Williams Fork aquifer in the eastern part of the study area.

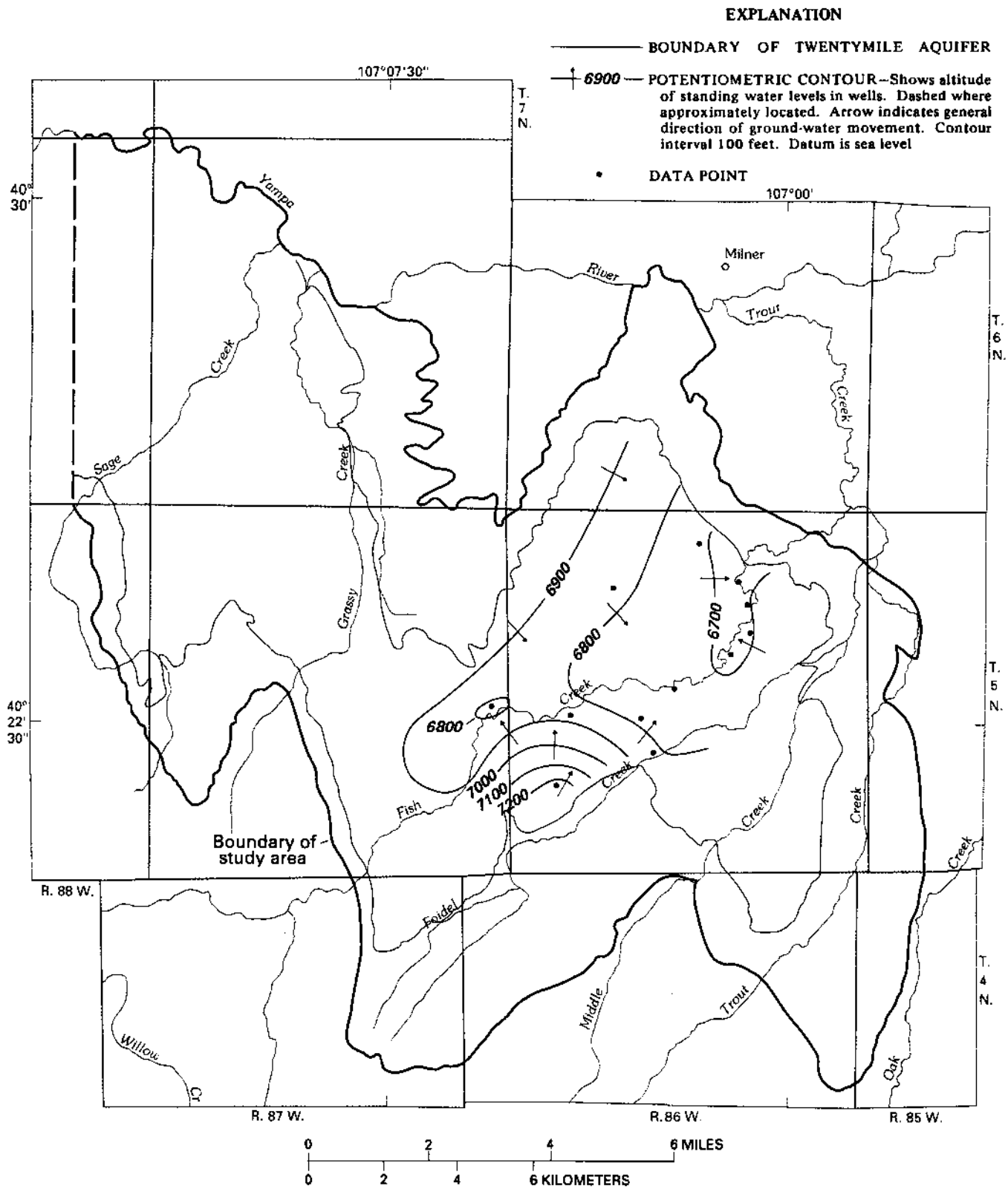

Figure 34.--Approximate steady-state potentiometric surface of the Twentymile aquifer in the eastern part of the study area.

flowing wells. In some instances, mining has removed part of the aquifer or replaced it with spoils material, either or both of which could disrupt steady-state conditions near the mine.

Ground-water divides in the potentiometric surface of the basal Williams Fork aquifer located near Grassy Gap and north of Twentymile Park form hydrologic boundaries in the flow system because ground water cannot move across the divides. As long as the divides remain in their present position, polluted ground water on one side of the divide can have no effect on ground-water quality on the opposite side of the divide.

The potentiometric surface of the Twentymile aquifer (fig. 34) is less well defined than that of the basal Williams Fork aquifer (fig. 33), for which more data are available. Heads in the Twentymile aquifer near the margin of the aquifer generally are higher than those of the underlying basal Williams Fork aquifer. Near Fish Creek, heads in the Twentymile aquifer generally are lower than those of the underlying basal Williams Fork aquifer. These head relations indicate that the potential exists for water to move from the Twentymile aquifer to the basal Williams Fork aquifer near the margins of the Twentymile aquifer and from the basal Williams Fork aquifer to the Twentymile aquifer near Fish Creek. It is likely that heads in the Twentymile aquifer also are higher near the margins of the aquifer and lower along the valleys of Fish, Foidel, Grassy, and Middle Creeks. Similar recharge-discharge conditions may exist in the Twentymile and basal Williams Fork aquifers. Recharge generally occurs in the outcrop areas of the aquifers and discharge occurs along the valleys of the principal streams draining the area.

Computer-simulation techniques were used to provide additional definition and corroboration of the hydrologic system as conceptualized for the eastern part of the study area. A multilayer model for simulation of quasi-three-dimensional flow (McDonald and Harbaugh, 1984) was constructed to simulate steady-state ground-water flow through the three aquifers. Model parameters, such as hydraulic conductivity and thickness of the aquifer and confining layers, were defined on the basis of the preceding data. Precipitation recharge through outcrops was defined by the previous estimates of distributed recharge. The effects of perennial streamflow and evapotranspiration also were simulated. A 2,000-ft interval grid consisting of 51 rows, 30 columns, and 3 layers was used to discretize the system; this grid also defines the scale at which hydrologic data were defined for use in the model. Model construction procedures are discussed in greater detail in the "Supplemental Information" section at the back of this report.

The model was calibrated to ensure its accuracy by comparing model-calculated heads and rates of discharge with measured values. An acceptable level of calibration was achieved after minor refinements were made to the model-input data. The resulting model-calculated potentiometric surface maps (figs. 35 and 36) are in good agreement with the maps based on measurements (figs. 33 and 34). The mean differences between calculated and measured values of head at 54 points in the model area was 9 ft in the Twentymile aquifer, 16 ft in the basal Williams Fork aquifer, and 20 ft in the Trout Creek aquifer. The differences were approximately randomly distributed over the model area. The model-calculated maps provide more complete definition of the potentiometric surfaces in the Twentymile and basal Williams Fork aquifers.

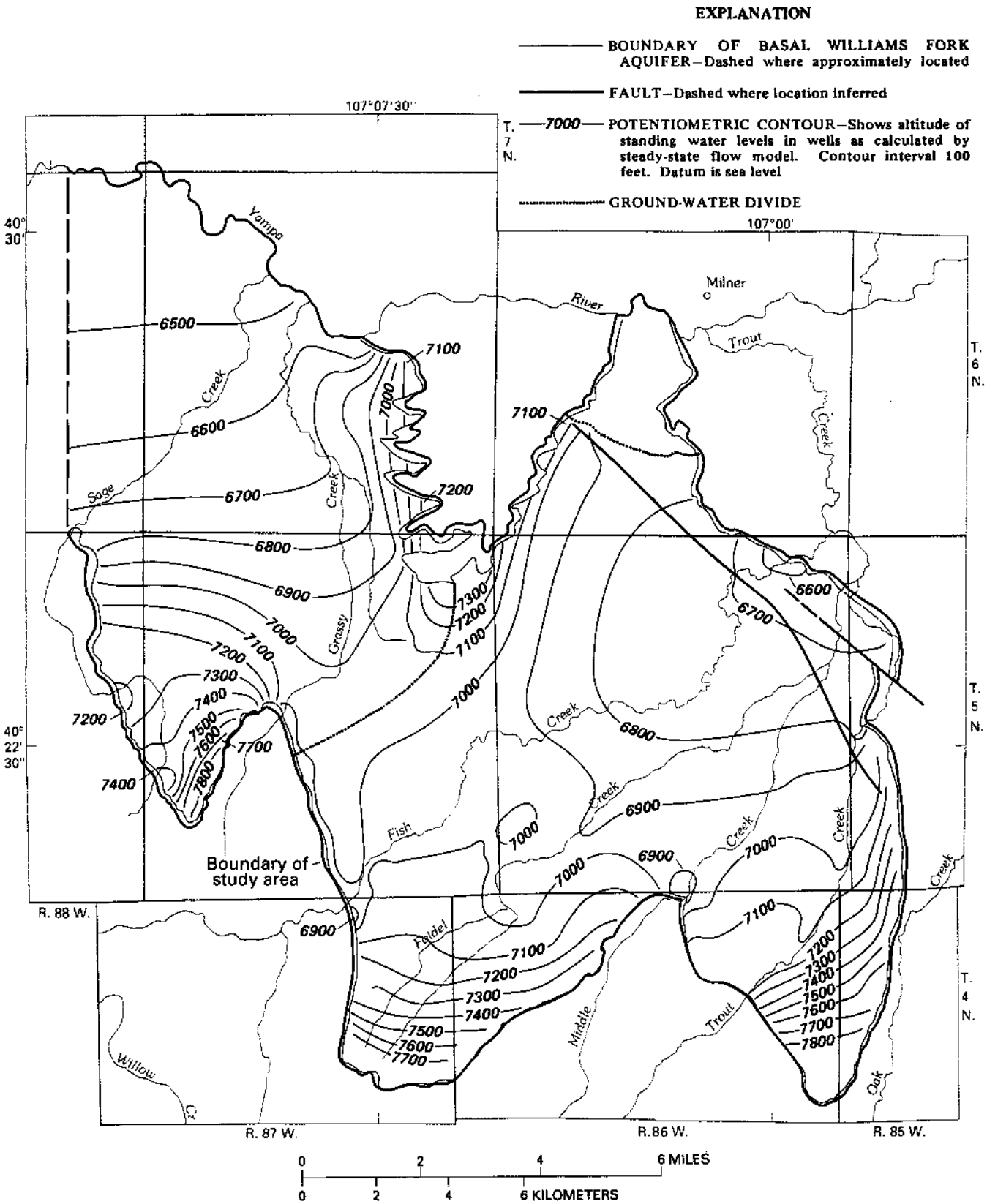

Figure 35.--Model-calculated steady-state potentiometric surface of the basal Williams Fork aquifer in the eastern part of the study area.

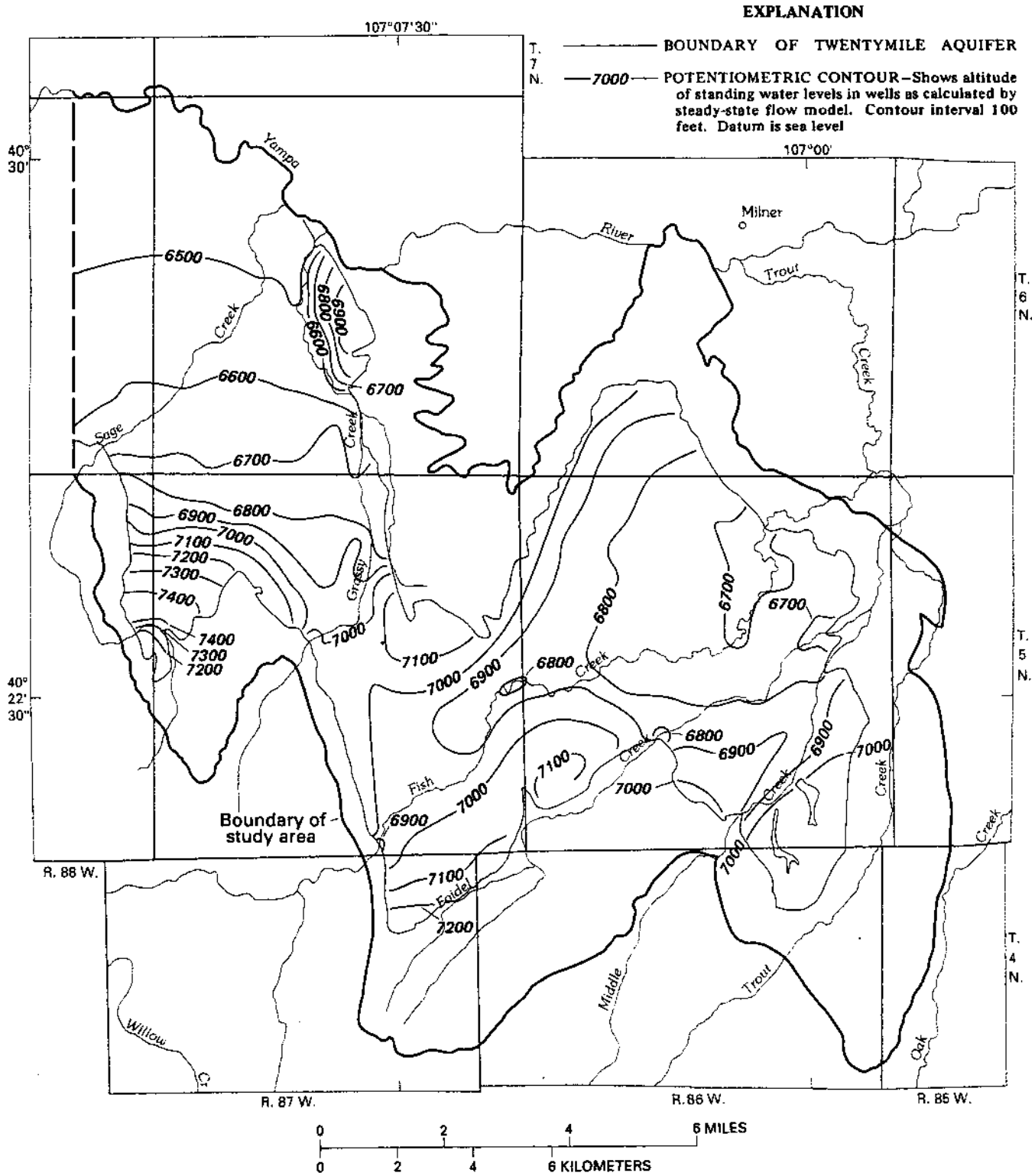

Figure 36.--Model-calculated steady-state potentiometric surface of the Twentymile aquifer in the eastern part of the study area.

Model results also indicate that the potentiometric surface in the Trout Creek aquifer essentially is identical to that of the basal Williams Fork aquifer. Model simulations that seem to more closely match the observed heads could be achieved by further changing model input in selected areas, but independent information is not available to justify such changes. As a result, the model was not forced to fit preconceived notions of the flow system.

Most of the water-level measurements made in the western part of the study area were made in uncased drill holes. As a result, the data do not define the potentiometric surface in a single aquifer. Instead, the water-level measurements represent a composite head that occurs in the water-yielding materials penetrated by the individual well. The approximate altitude of these composite heads is indicated in figure 37. The elevation of the heads generally conforms to topography; higher heads occur in the higher altitude areas to the south; lower heads occur in the lower altitude areas to the north. Heads are higher in the deeper aquifers because their recharge areas are at a higher altitude.

Flow Direction

The direction of ground-water flow in the eastern part of the study area is relatively well defined by data and is further corroborated by simulation results. Water moves from recharge areas along the elevated outcrops at the margin of the aquifers toward Twentymile Park and Hayden. A similar pattern of movement is indicated for the basal Williams Fork and Twentymile aquifers. In the higher outcrops of the Twentymile aquifer, the potential exists for interaquifer ground-water movement from the Twentymile aquifer to the basal Williams Fork aquifer. At lower altitudes, the potential is reversed and water may move from the basal Williams Fork aquifer upward to the Twentymile aquifer. In areas where heads are above land surface, water also may discharge to the surface, where it may be lost to evapotranspiration or flow into alluvial aquifers or streams.

The aquifers south and east of the ground-water divide at Grassy Gap form a closed basin because ground-water underflow into and out of the area is insignificant. Ground water in this area is derived from local recharge and moves through the area to discharge at the surface in Twentymile Park. To the northwest of the ground-water divide at Grassy Gap, ground water moves from the outcrop recharge areas to depth along the Hayden syncline. Water may discharge either by vertical leakage into the Yampa River and its alluvium or as underflow into the larger flow system associated with the Sand Wash basin to the north of the Yampa River.

Faulted areas occur near Eckman Park, on the west flank of the Tow Creek anticline, and near the eastern margin of Twentymile Park. Faults near Eckman Park and the Tow Creek anticline are parallel or subparallel to topographic gradients and the general direction of ground-water movement. If sufficient displacement has occurred on these faults, they may form barriers to ground-water movement across the fault plane. Such faults could have little effect on water movement parallel to fault planes. Thus, it is difficult to determine the effects of faulting on ground-water movement in these two faulted areas.

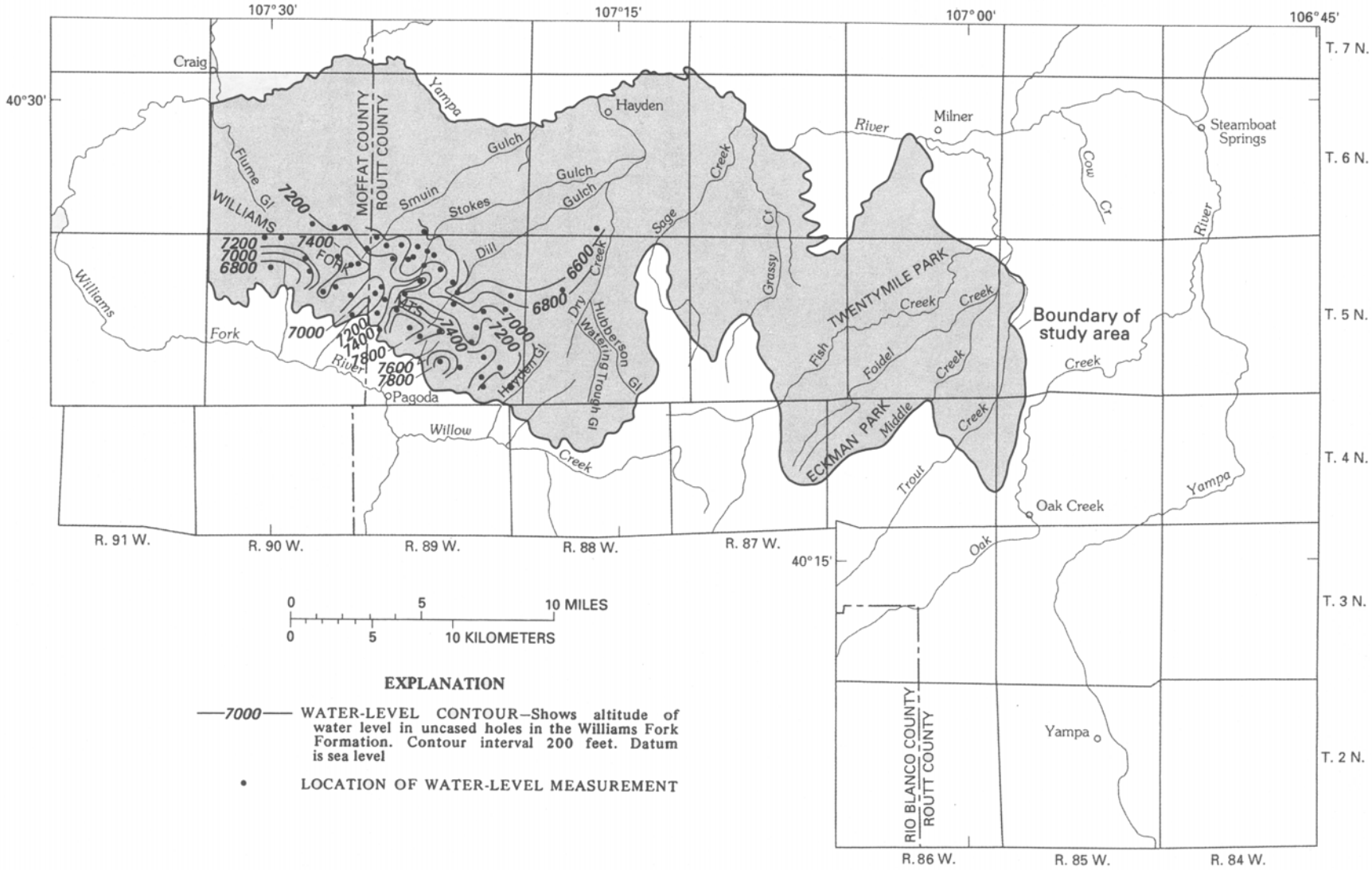

Figure 37.– Water-level altitude in the Williams Fork Formation in the western part of the study area.

Faulting is more extensive along the eastern margin of Twentymile Park. Subparallel fracturing associated with this fault zone has increased hydraulic conductivity parallel to the fault; water moves along the fault more readily than it moves across the fault. This flow pattern causes part of the elongated shape of the 6,700-ft potentiometric contour shown in figure 33.

Differences in hydraulic head in the three aquifers cause water to move vertically across the two confining layers that separate the aquifers. Heads that are above land surface in the uppermost aquifer also may cause upward movement of water across the upper confining layer (Lewis Shale). Although the volume of this interaquifer leakage is very small, it is an important component in the water budget of the area. The rate of vertical ground-water movement also is very small. The time required for a particle of water to move vertically across a confining layer was estimated from Darcy's law using the equation:

$$t = \frac{\phi L^2}{K \Delta h}, \qquad (2)$$

where

t = travel time;
ϕ = porosity of the confining layer;
L = thickness of the confining layer;
K = hydraulic conductivity of the confining layer; and
Δh = difference in hydraulic head between the aquifers immediately overlying and underlying the confining layer.

The parameters L, K, and Ah were defined by the corresponding model input data or model-calculated head. Porosity was assumed to equal 13 percent. As indicated in table 8, the mean traveltime for a particle of water to move vertically across the confining layers under steady-state conditions is on the order of thousands to tens of thousands of years. Thus, under steady-state head conditions, the interaquifer movement of ground water requires very long traveltimes and thus is not an important consideration in predicting shorter term movement of contaminants. Traveltimes across the lower confining layer are long, primarily because only small steady-state head differences are present across the confining unit. Under transient conditions, as in response to mine dewatering, larger head differences could develop and the resulting traveltimes could be shortened.

The direction of ground-water flow in the western part of the study area is more difficult to determine because of the lack of potentiometric-surface data for individual aquifers. It is probable that ground water moves in generally northeasterly and southwesterly directions from a ground-water divide that approximately coincides with the topographic divide of the Williams Fork Mountains. Southwestward-flowing ground water moves down the cuesta backslopes toward the Williams Fork. This flow is against structural dip and oblique to the strike of the aquifers. Flow paths likely are short because most water discharges at local springs and seeps. Most of the ground water in the western part of the study area flows to the north or northeast along paths subparallel to the dip of the regional structure. Recharge occurs in the highland outcrops along the Williams Fork Mountains. Down-dip movement

carries the water to greater depth in the Williams Fork Formation and under the Lewis Shale. Discharge occurs by upward leakage or by underflow. Water discharged by upward leakage is ultimately lost to evapotranspiration at the land surface or is tributary to streams or alluvial aquifers. Water discharged by underflow moves out of the study area and contributes to the larger regional ground-water flow system in the Sand Wash basin.

Table 8.--*Mean traveltime for steady-state flow of ground water across confining layers in the eastern part of the study area*

[NA, not applicable]

	Traveltime , in years, within surface drainage areas					
	Sage Creek	Grassy Creek	Fish Creek	Foidel Creek	Middle Creek	Trout Creek
Mean upward traveltime for water movement through confining layer separating Twentymile and basal Williams Fork aquifers .	30,300	4,900	11,000	6,300	2,900	NA
Mean downward traveltime for water movement through confining layer separating Twentymile and basal Williams Fork aquifers .	1,400	2,700	16,500	6,600	2,800	2,900
Mean upward traveltime for water movement through confining layer separating basal Williams Fork and Trout Creek aquifers.	98,400	32,700	63,000	28,700	44,300	69,900
Mean downward traveltime for water movement through confining layer separating basal Williams Fork and Trout Creek aquifers.	95,500	36,400	76,200	88,400	78,000	45,100

Lateral ground-water velocities in the eastern part of the area, computed during model simulations, range from less than 1.0 ft/yr to more than 30 ft/yr in the basal Williams Fork aquifer and from less than 0.1 ft/yr to more than 3.0 ft/yr in the Twentymile aquifer. The distribution of lateral velocities shown in figures 38 and 39 indicate that larger velocities generally are located near the margins of the aquifer and smaller velocities are prevalent in the central parts of the aquifers. The combined effects of relatively large hydraulic conductivity and potentiometric gradient result in a larger ground-water velocity along the fault area near the eastern margin of the aquifers. Other faults or interconnected fracture systems could have similar but less pronounced effects on local ground-water velocities. Lack of data prevents individual simulation of any such local features. However, the larger scale effects of local faults and fractures are incorporated in the model through use of spatially distributed hydraulic conductivity, which is based in part on results of aquifer tests in potentially fractured aquifers. Although ground-water velocities cannot be precisely determined because of this lack of data and associated uncertainties in the model parameter values, the magnitude of velocities shown in figures 38 and 39 are significant in that very slow rates of ground-water flow are indicated. A contaminant that enters the bedrock aquifer will not move rapidly to other parts of the area and could remain virtually immobile at some locations.

WATER BUDGET

The ground-water flow model was used to estimate the steady-state water budget for the eastern part of the study area. The simulated water budget (table 9) indicates that total recharge for, or discharge from, the bedrock aquifers is only about 2.6 ft^3/s. This small rate if flow us consistent withe small hydraulic conductivity and small well yields observed formations that are classified as marginal aquifers in this study. Recharge and discharge for each aquifer by major surface drainage area is listed in table 9. For example, the model calculated that the Twentymile aquifer receives 0.0946 ft^3/s of precipitation recharge from that part of the Sage Creek Drainage area that overlies the aquifer; the aquifer loses 0.0777 ft^3/s of discharge to evapotranspiration in the same area. Recharge may come from percolation of water in streams and ponds or from deep infiltration of precipitation. Discharge may be evapotranspiration or to streamflow and alluvial aquifer. Estimated total recharge to the basal Williams Fork aquifer is about 1.4 ft^3/s to the Twentymile aquifer is similar, but recharge is only about 0.02 ft^3/s in the Trout Creek aquifer. Recharge and discharge to the Trout Creek aquifer is limited by the small transmissivity and very limited outcrop area of the aquifer. Vertical leakage (the difference between total inflow and total outflow through the lateral boundaries of the aquifer) is the rate of flow through the confining layers that separate each aquifer. The Trout Creek aquifer receives about 5 percent of its recharge as leakage for the overlying basal Williams Fork aquifer and discharges about 90 percent of inflow into the basal Williams Fork aquifer in other areas.

The accuracy of the simulated water budget is affected by the size of the grid interval used in modeling, by the accuracy of the model parameters, and by the extent of the model calibration. The 2,000-ft grid interval used in this model provides a resolution of 930 nodes in the Trout creek aquifer, 920 nodes in the basal Williams fork aquifer, 530 nodes in the Twentymile