STEADY-STATE SIMULATION AND CALIBRATION OF PRESTRESSED CONDITIONS

A steady-state simulation of prestressed (nonpumping) conditions is necessary to determine initial conditions for transient simulation. Initially, transmissivity values for siltstone and sandstone were obtained from aquifer test results reported in previous studies (Johnston and Larson, 1979, and by Leggette, Brashears, and Graham, 1980). These values, integrated with geologic information from Leavy, Froelich, and Abram (1983) are used as the basis for estimating relative transmisssivity for other bedrock types in the basin. The bedrock map was used as a guide to delineate zones of inferred equal transmissivity (fig. 3).

Initial streambed leakance values were calculated using a general ground-water flow equation (Darcy's law) to relate mean annual baseflow (ground-water discharge) to streambed permeability and the hydraulic gradient across the streambed. If no water is artificially moved (piped or pumped) out of a drainage basin, then all water entering the basin should equal the amount leaving the basin. Thus, stream baseflow will aproximately equal effective aquifer recharge (ground-water recharge less ground-water evapotranspiration).

Mean annual baseflows were estimated at 15 sites in the basin (table 2) by a method that employs streamflow-duration curves (Trainer and Watkins, 1979). Rating curves relating stream discharge to water levels in a representative well within the basin were developed at 5 long-term stream-gaging stations in the basin. The equations defining these curves were used to separate the baseflow component from the streamflow hydrograph (Schicht and Walton, 1961). These separations indicate that the 68-percentile discharge on the flow-duration curve is representative of mean annual baseflow for streams in the Culpeper basin. This value was used to estimate mean annual baseflow from streamflow-duration curves at the remaining 10 sites.

Differences in baseflow among the various subbasins is related primarily to differences in geology. The higher values are generally associated with streams that drain pre-Triassic rocks in their headwater areas west of the Culpeper basin. Those rocks are more deeply weathered than the fractured sedimentary rocks within the basin and can store relatively more water to be released during extended dry periods and consequently maintain higher baseflows.

Initially, a uniform, basin-wide, aquifer recharge rate was assumed to equal the mean annual baseflow (2.70 in/yr) computed for the basin (table 2). Trial simulations indicated that the above assumption would be incorrect because of the large transmissivity contrasts in the basin, especially between the diabase-thermally metamorphosed bedrock complex (units J and T) and the unaltered sedimentary (units S1, S2, S3, S4, and S5) rock units. Thus, the recharge value assigned to any particular node was based on the previously assigned transmissivity for that node. These different recharge values were then distributed uniformly over each transmissivity zone to produce an average basinwide recharge rate approximately equivalent to the computed mean annual baseflow.

The model was calibrated by trial and error adjustment of input parameters until simulated water levels and ground-water discharge values were in general agreement with values derived from field measurements (fig. 4). Figure 5 shows steady-state water levels computed by the calibrated model under assumed prepumping conditions. Adjusted values of transmissivity and recharge determined by model calibration are given in table 3. An anisotropy factor related to the directional transmissivity (T_{xx}:T_{yy}) of 4:3 was used to produce an acceptable match of simulated and observed water levels.

Differences between simulated and observed water levels range from 0 to 30 percent. Differences greater than 10 percent occur near streams that were too small to be properly incorporated into the model, and near no-flow boundaries. In most cases, differences between simulated and estimated ground-water discharge (baseflow) are less than 40 percent. Larger discrepancies occur in areas affected by upstream impoundments and other controls on streamflow. An overall value of 2.60 in/yr of ground-water discharge to stream nodes was calculated by the model. This value compares favorably with the estimated value of 2.7 in/yr shown in table 2.

TRANSIENT SIMULATIONS OF HYPOTHETICAL WELL FIELDS

Transient simulations of pumping using hypothetical well fields in different bedrock types were run to study the effects of ground-water withdrawals and to demonstrate the potential use of a model in planning future ground-water development. However, because historical pumping and water-level data were inadequate to calibrate the model properly under transient conditions, head and/or drawdown values computed by the model should be used with caution.

Model runs for all transient simulations treated the ground-water system as being confined (artesian). Values of transmissivity and storage coefficient are held constant when operating the model in a confined mode. In actuality, the ground-water systems of the basin is only partially confined and elsewhere has a free surface (water table). In a water-table system, values of transmissivity and storage coefficient are directly proportional to the saturated thickness. As water levels decline, saturated thickness decreases which results in a decrease in the above mentioned aquifer properties. However, if the change in saturated thickness is small relative to the initial saturated thickness, confined simulation of water-table conditions is expected to produce reliable results. During transient simulations, model results were considered unreliable if this change was excessive.

In the transient simulations, a well field was represented by a single node. The hydraulic properties assigned to the well field were determined by the bedrock type in which the node was located. Grid and boundary conditions were identical to those shown in figure 3, and transmissivity, recharge, and initial head values were obtained from the steady-state model. Storage coefficients used in the model were assigned according to bedrock type and ranged from 0.0007 to 0.008. These values were based on results of aquifer tests reported by Johnston and Larson (1979) and by Leggette, Brashears, and Graham (1980).

The transient model was used to simulate one year of pumping at rates ranging from 0.1 to 0.5 ft^3/s. Simulations were run under both normal and extreme drought conditions. Normal conditions imply aquifer recharge is continuous over the period of simulation and stream recharge or discharge (equivalent to ground-water discharge or recharge) occurs at all stream nodes. Extreme drought conditions imply no aquifer recharge, and stream recharge or discharge occurs only in nodes representing the Potomac and Rappahannock Rivers. The remaining streams are assumed to have dried up.

A representative series of computer-generated potentiometric maps show the model simulated water-level declines after pumping individual fields in sandstone, siltstone, and limestone-conglomerate-dominated subbasins (figs. 6*a–d*). Well fields were pumped independently for a year and then superimposed onto one map for each indicated pumping rate and condition.

In general, the configuration of the drawdown cone is influenced by the bedrock type in and adjacent to the well field (particularly the diabase-thermally metamorphosed zone), the distance to and the distribution of stream nodes in the model, and the pumpage rate. The maps indicate little change, however, in the size and shape of the drawdown cone resulting from pumping the limestone conglomerate (unit S4) at different pumping rates under both normal and drought conditions. This result is attributed to the high transmissivity and storage coefficient of the limestone conglomerate and the effect of the nearby Potomac River, which functions as a ground-water sink during normal conditions and as a ground-water source during drought conditions. This same result is quantitatively shown by the model in the mass-balance calculations. During both normal and drought conditions, the Potomac River contributed 65 percent of the water induced by pumpage. The remaining 35 percent was accounted for through storage contribution. Simulations of pumping from nodes representing the diabase and thermally metamorphosed bedrock type (units J and T) indicate that the rocks are extremely poor aquifers compared to the unaltered sedimentary rocks (units S1 and S2).

SUMMARY AND CONCLUSIONS

In the Culpeper basin, the use of ground water is technically feasible for public supply in small towns, satellite systems (subdivisions and trailer courts), and fringe areas of existing distribution systems. Ground water is perhaps the only available source for individual, widely scattered homes in rural areas of the basin.

These supplies might be provided by individual large-yield wells or by a field of several smaller-yield wells. Systematic exploration for the most favorable drilling area should be able to locate adequate supplies close enough to points of intended use to eliminate the expense of supply lines to some distant surface-water source.

Even in larger communities that use surface water as their primary source of supply, ground water can be either a supplemental or a reliable emergency supply during periods of drought. Standby wells could be pumped at maximum possible rates for periods as long as 90 to 120 days, even though severe local declines in water levels would probably occur. Ground-water levels would recover after pumping stopped and precipitation returns to normal.

Water levels, streamflow data, and a bedrock map were used to develop a ground-water flow model of the Culpeper basin. The model was used to improve understanding of the ground-water flow system. Aquifer properties assigned to each node describe the combined fracture and rock properties of the rock. Water levels and ground-water discharge values computed by steady-state simulation are in close agreement with values derived from field measurements. Transient runs were made to simulate drawdown distributions that would result from various rates of pumping in different bedrock types. The transient model was not calibrated, however, because historical pumping data were inadequate.

Additional hydrologic and geologic information are needed to develop a more reliable ground-water model for predicting water level declines due to pumping. Geologic sections within the basin are needed to define the thickness of the "active" ground-water system. Data on fracture size and density would permit evaluation of spatial and directional variations in transmissivity. Finally, historical data adequately describing the time dependency of such hydrologic factors as precipitation, evapotranspiration, baseflow, pumpage, and water levels are needed as input data for reliable transient calibration and prediction.

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FIGURE 4.—Distribution of water-level and stream-flow data sites used in model calibration.

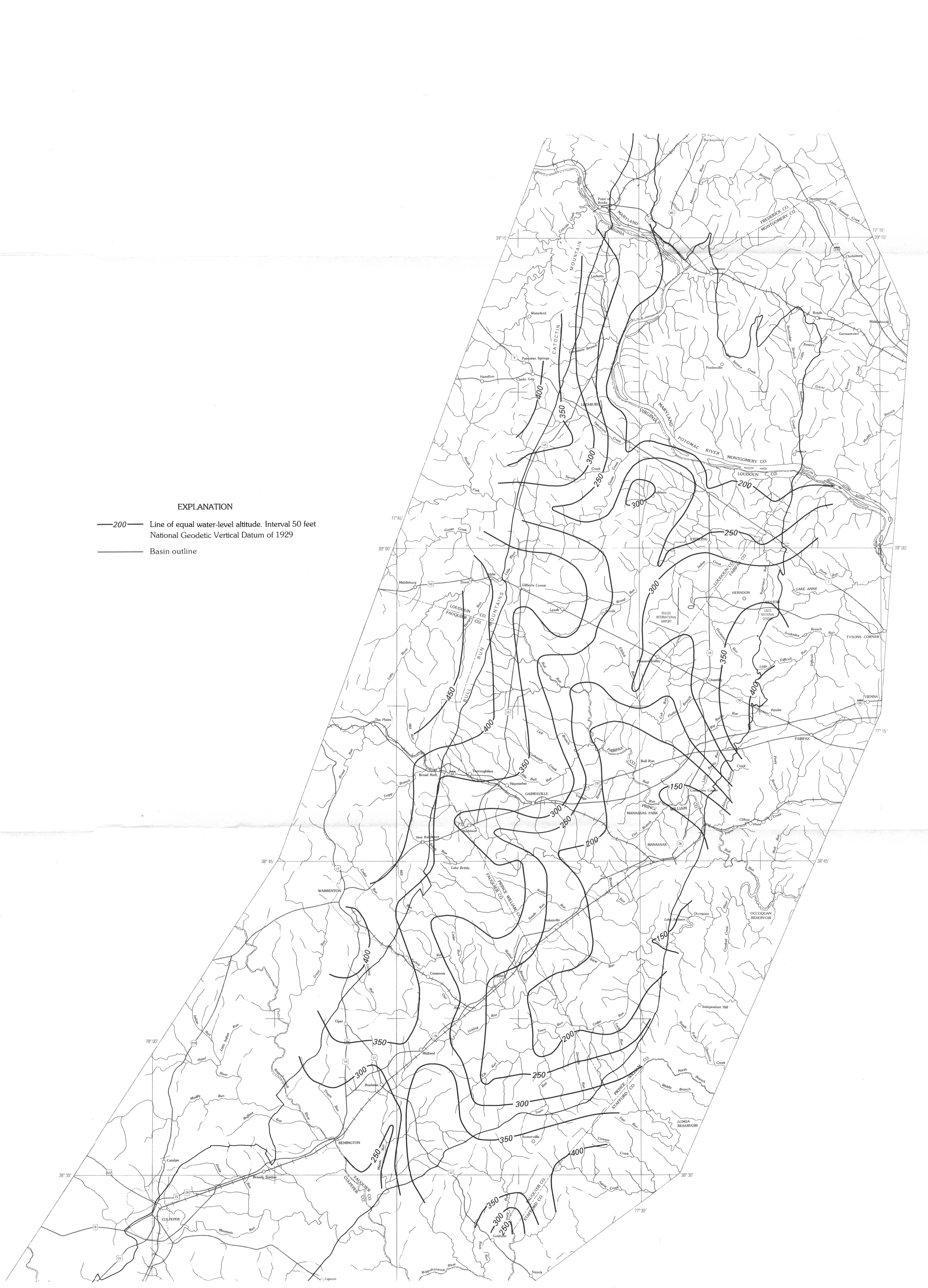

FIGURE 5.—Simulated steady-state water-level altitudes map of prestressed conditions.

A.—Pumped at 2 ft^3/s under normal conditions.

B.—Pumped at 4 ft^3/s under normal conditions.

C.—Pumped at 2 ft^3/s under drought conditions.

D.—Pumped at 4 ft^3/s under drought conditions.

EXPLANATION

—10— Line of equal water-level drawdown. Interval 10 feet, except where dashed and explicitly stated. National Geodetic Vertical Datum of 1929

Basin outline

FIGURE 6.—Simulated drawdown distributions resulting from 1 year of continuous pumping from well fields in indicated rock type during normal and drought conditions. Location of node pumped is shown in figure 3.

TABLE 2.— Summary of estimated mean annual baseflows of principal streams in Culpeper basin. Baseflows estimated using 68-percentile discharge on streamflow.

Name of stream	Location	68-percentile discharge on streamflow-duration curve (cubic feet/second)	Area of watershed (square miles)	Estimated mean annual baseflow rate (inch/year)
Rappahannock River	Remington	248.00	620.0	5.4267
Cedar Run	Aden	16.50*	62.6*	3.6310
do.	Catlett	7.21*	81.1*	1.2038
do.	Warrenton	2.79	12.3	3.0780
Broad Run	Bristow	10.00*	39.1*	3.4713
do.	Buckland	12.50	50.5	3.3597
Bull Run	Clifton	41.20*	159.1*	3.4922
do.	Catharpin	4.80	25.8	2.5241
Goose Creek	Ashburn	1.42*	50.0*	0.3852
do.	Leesburg	95.58	332.0	3.9054
Broad Run	Leesburg	3.71*	70.8*	0.7111
do.	Arcola	0.59	5.3	1.5111
Cub Run	Broad Run	8.50	49.9	2.3122
Sugarland Run	Herndon	0.26	3.36	1.0501
Potomac River	Point of Rocks	3,150.00	9,561.00	4.4692
			Total	40.5016
			Average	2.7001

*Value associated with upstream gaging station subtracted out in computation.

TABLE 3.— Model calibrated values of transmissivity and recharge.

Principal lithology at node	Model transmissivity (square feet/day) T_{xx}	T_{yy}	Model Recharge (inches/year)
Limestone conglomerate	5,040	3,700	5.25
Siltstone	3,600	2,700	4.84
Sandstone	3,000	2,250	4.18
Basalt	2,520	1,890	3.93
Conglomerate	2,500	1,875	3.91
Pre-Triassic rocks	800–1,300	600–975	2.00–2.57
Diabase and thermally metamorphosed rocks	80	60	.17

SCALE 1:250 000

5 0 5 10 MILES

5 0 5 10 15 KILOMETERS

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