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Hydrology and Geomorphology of Green Pond – a High-elevation Depressional Wetland in the Blue Ridge of Virginia

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INTRODUCTION

Green Pond is a sphagnum bog situated on a broad summit in the Blue Ridge Mountains of central Virginia. In this research project, we analyzed the surficial aquifers and groundwater flow system that support this high elevation ecosystem in order to evaluate how water level fluctuations respond to changes in precipitation and evapotranspiration. Such studies provide the hydrogeological data needed for sound ecological management of sensitive environments.

Hydrogeologic Setting

Green Pond (Fig. 1) lies on Big Levels (elevation 976 m), the crest of a northwestward spur of the Blue Ridge Mountains. This upland, named the Greenville Salient by Moore (1952), is deeply incised by steep stream valleys that form the headwaters of the St. Marys River and South River. Immediately north of Flint Mountain, Big Levels plateau is a 600 m wide bench with less than 6 m of topographic relief in the Green Pond area. Green Pond sits within a depressional wetland that possesses no visible inlet or outlet streams; however, during periods of high rainfall, the pond is drained by overland flow to the north along foot trails. Downhill, numerous springs emerge from quartzite talus and bedrock fractures along valley toeslopes.

Groundwater within the study area appears to move through intergranular pores in residuum and sandstone as well as through fractures in quartzite and other bedrock. The Antietam Formation underlies the ridge summit and most of the Greenville Salient. Overall, this unit consists of resistant, thickly and very thickly bedded, white to gray quartzite interbedded with equal amounts of thickly stratified, less resistant sandstone. Subordinate amounts of laminated shale and siltstone (5-10%) and conglomerate (1%) occur throughout the thickness of the unit (Schwab, 1970; Gathright, 1977). Some authors separated the Antietam into an upper member and a lower member based on lithology and outcrop pattern (Moore, 1952; Caskie, 1957; Sweet, 1981). The lower member consists of up to 120 m of white, vitreous quartzite that crops out as ledge or cliff-making beds up to 30 m thick. The upper member of the Antietam consists of buff to brown friable sandstone in beds several cm to meters thick. The upper member generally creates an outcrop pattern of well rounded slopes covered by blocky sandstone and quartzite float.

Porous regolith transmits the shallow groundwater flows on the Big Levels plateau. The geomorphic history of the surficial materials and the characteristics of the rocks that produced them govern the permeability and distribution of shallow aquifers. Thus, we must understand how the landscape on Big Levels developed in order to evaluate the stratigraphic framework of this hydrogeologic setting.

The vast majority of wetlands within the Blue Ridge and adjacent provinces are not depressional but form in slope and riverine settings associated with stream valleys (Brinson, 1993; Smith et al., 1995; Cole et al., 1997). The enclosed basins where depressional wetlands can form in the Blue Ridge might be influenced by several geomorphic processes including periglacial sedimentation (Eargle, 1977; Delcourt & Delcourt, 1985), rotational slump blocks (Schultz & Southworth, 1989), structural controls (Diehl & Behling, 1982), or bedrock solution (Reed et al., 1963). Knechtel (1943) and Werner (1966) suggest that Green Pond is a solutional depression that lies on a small lens of dolomite on the axis of a very gentle syncline. Several problems exist with this interpretation. Stratigraphically, the Tomstown (Shady)

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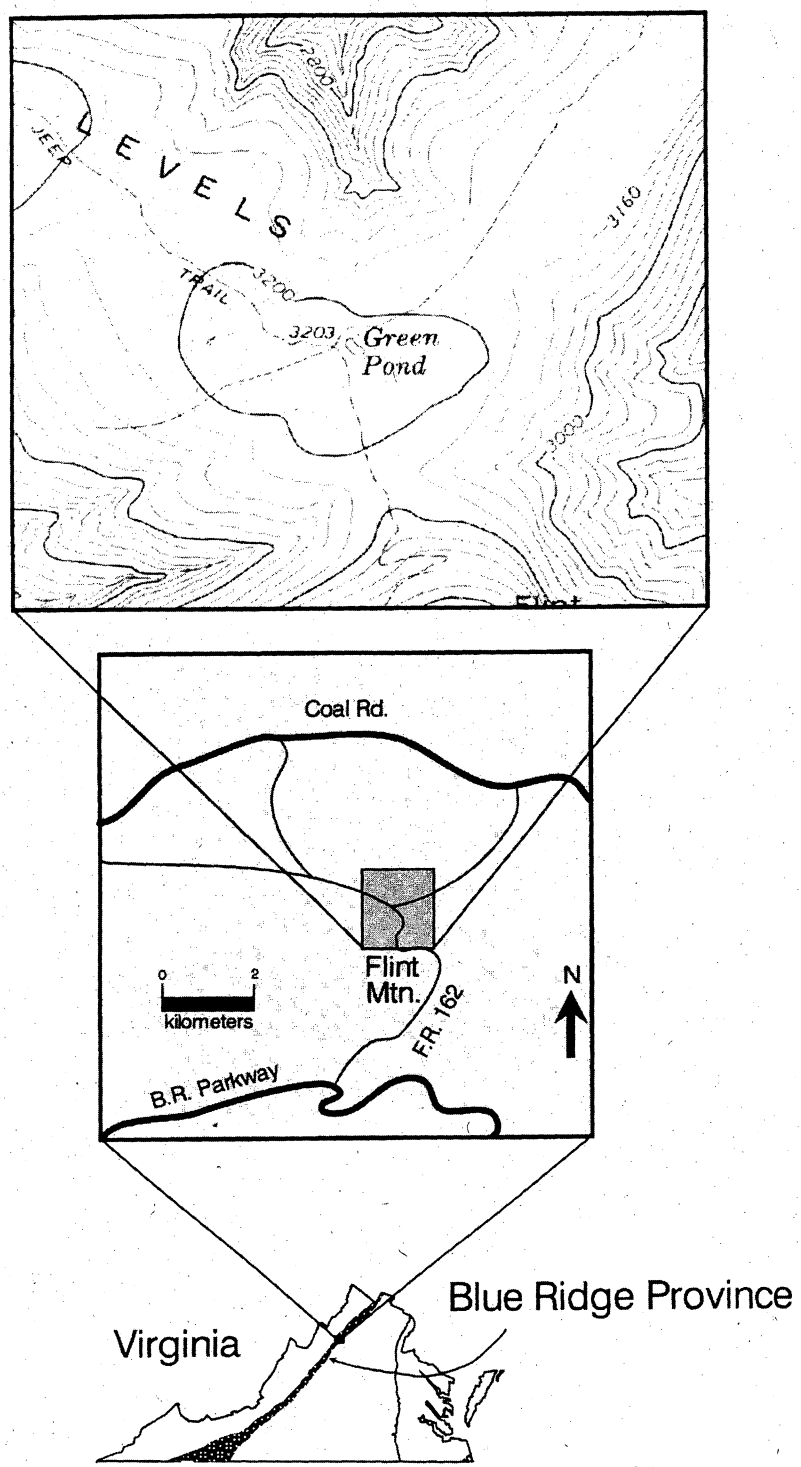

Fig. 1. Location of Green Pond and topographic map of study area. Topography taken from USGS Big Levels quadrangle; contour interval 40 feet. US Forest Service Road 162 intersects the Blue Ridge Parkway at Bald Mountain overlook. Valleys leading away from Big Levels and Kennedy Ridge are headwaters for Kennedy Creek (to the east), Cole's Run (to the north), and St. Marys River (to the west).

dolomite does overlie the Antietam quartzite; thus a dolomite-solutional origin for Green Pond is structurally possible. However, soluble exposures of dolomite are very rare near the surface throughout the region because the unit is usually strongly weathered to saprolitic residuum to depths of 30-60 m (Knetchel, 1943). The shallow dips on the limbs of the syncline mapped by Knetchel (1943) suggest that no soluble dolomite should be expected at depth below Green Pond. Also, based upon the results of three previous geologic studies, the precise geologic pattern of the Green Pond area is open to question - maps made by Knechtel (1943), Moore (1952), and Werner (1966) contain different numbers and locations of geologic structures there. According to Moore (1952), for example, Green Pond formed on quartzite on the limb between two folds.

Reed et al. (1963) suggested that shallow depressions on quartzite ridges in North Carolina, in settings very similar to those found at Green Pond, could have formed by silica solution if the surface was stable for a million years or longer. Noting that the pond waters contained appreciable quantities of dissolved silica, they concluded that acids generated in organic-rich sediments could aid in the long term solution of quartzose rocks and regolith. Bennet & Siegel (1987) and Bennet et al. (1991) demonstrated that in a peat bog the solubility of quartz increases sufficiently to enhance quartz dissolution due to complexing of organic acids.

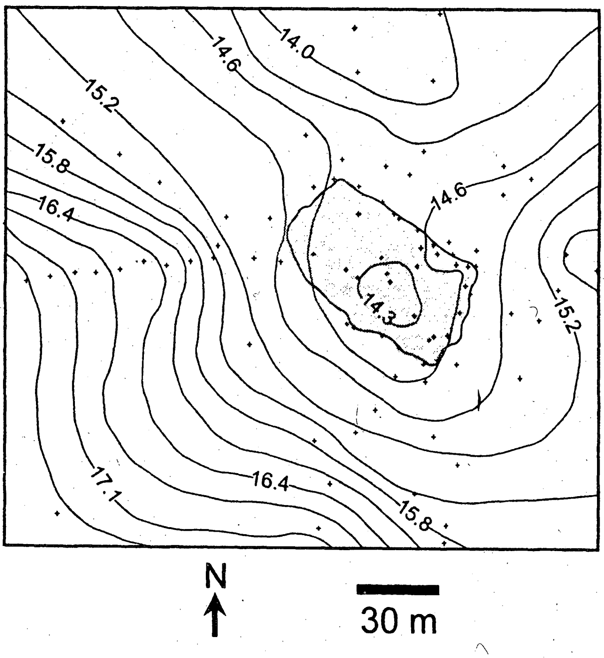

Fig. 2. Topographic map of Green Pond and surrounding area. Contours are in meters above an arbitrary local datum. Crosses indicated locations of surveyed points. Grey area is extent of open area around pond.

Hockman et al. (1979) mapped the soils in the study area as the Leetonia Series. This series occurs on broad ridges and side slopes, formed in siliceous materials weathered mainly from the Antietam quartzite. The Leetonia Series has been classified as a sandy, skeletal, siliceous podzol (mesic Entic Haplorthod). The predominant texture of these soils is very stony loamy sand. Wilson (1958) noted that the distinguishing feature of the Leetonia Series in Virginia is the presence of a bleached spodic (E) horizon due to the low base status of the siliceous parent material and the abundant acid vegetation (conifers).

Ecologic Setting

Non-alluvial wetlands of the central and southern Blue Ridge exhibit a considerable range of vegetation and wildlife, a diversity driven by the geology, soil, water chemistry and flow, elevation, age, and biogeographic history of the wetland and surrounding setting (Hanenkrat, 1980; Carter, 1986; Weakley, 1991). The Green Pond area is especially unusual, however. The forest surrounding Green Pond that developed in the 40-60 years since the latest clearcutting event is somewhat stunted, probably because of adverse climatic conditions and low soil fertility. It is dominated by pitch pine (*Pinus rigida*), white pine (*Pinus strobus*), black oak (*Quercus velutina*), scarlet oak (*Quercus coccinea*), red maple (*Acer rubrum*) and blighted chestnut (*Castanea dentata*). The understory shrub layer contains abundant mountain laurel (*Kalmia latifolia*) and Catawba rhododendron (*Rhododendron catawbiense*), huckleberry (*Gaylussacia* sp.), and Viburnum. The herb layer consists predominantly of cinnamon fern (*Osmunda cinnamomea*) and greenbrier (*Smilax* sp.) in wetter areas. Vegetation growing on the substrate of the pond consists of a sphagnum mat with numerous grasses and sedges such as *Carex* and the three-way sedge (A. Plocher, personal communication, 1993); several of the obligate wetland species in Green Pond are listed on the Virginia Rare Wetland Species List (C. Ludwig, personal communication, 1991). In the 1930s, a naturalist planted cranberries in the bog (Hanrenkrat, 1980).

True sphagnum bogs, such as Green Pond, are rare in Virginia, as are bogs and fens throughout the central and southern Appalachians (Schafale & Weakly, 1990). Of all the factors that form and sustain these rare ecologic settings, a critical, and maybe predominant, influence is the dynamic balance provided by the local wetland hydrology.

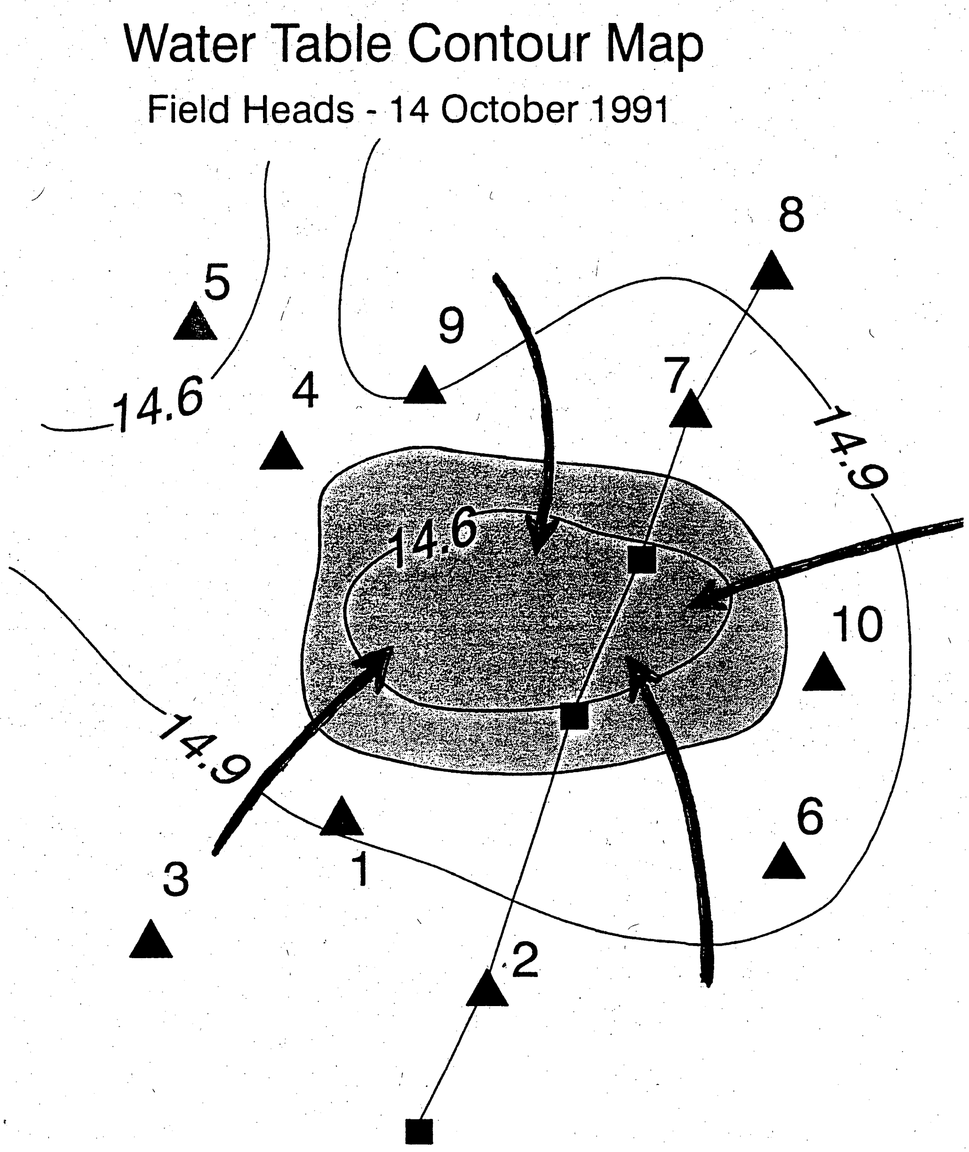

Fig. 3. Water table contour map for the Green Pond area on 14 October 1990. Contoured elevations are in meters above local datum. Triangles mark locations of monitoring wells. Arrows represent directions of groundwater flow. Line marks transect through well sites and boreholes (squares) used in Fig. 4.

FIELD METHODS AND RESULTS

Detailed elevation measurements taken along trails and open areas using a laser-based theodolite allowed us to analyze the surface morphology in and around Green Pond. An arbitrary local datum was set 50 feet (15 m, approximately) below a temporary benchmark. The large-scale (1"=50') topographic map constructed from the survey data (Fig. 2) became the base map for hydrologic analyses. Locations of bedrock outcrops, springs, drainage lines, and other features were plotted at 1:24000 using aerial photographs and U.S.G.S. topographic maps.

Ten groundwater monitoring wells installed around the perimeter of Green Pond permitted measurement of water table fluctuations in the shallow subsurface. We augered the boreholes to 1.5-2.0 m, close to the bedrock-residuum contact, and slotted the well risers along their entire lengths to within 15 cm of the surface. We placed well tops above the pond's seasonal high water mark and surveyed their elevations. Both chalk-and-tape and electronic probe techniques produced water elevation measurements that were repeatable to +0.5 cm. The water

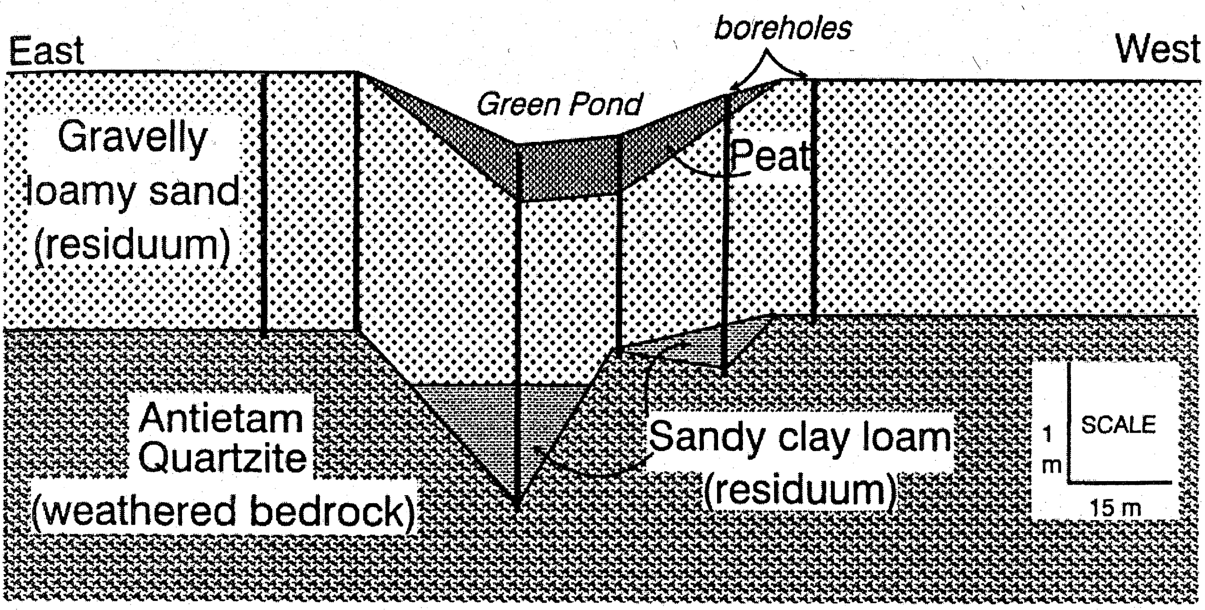

Fig. 4. Diagrammatic cross-section through Green Pond. Borehole locations noted in Fig. 3. In the gravelly sand loam residuum formed from quartzite, the coarsest sediments occur in the bottom 10-20 cm.

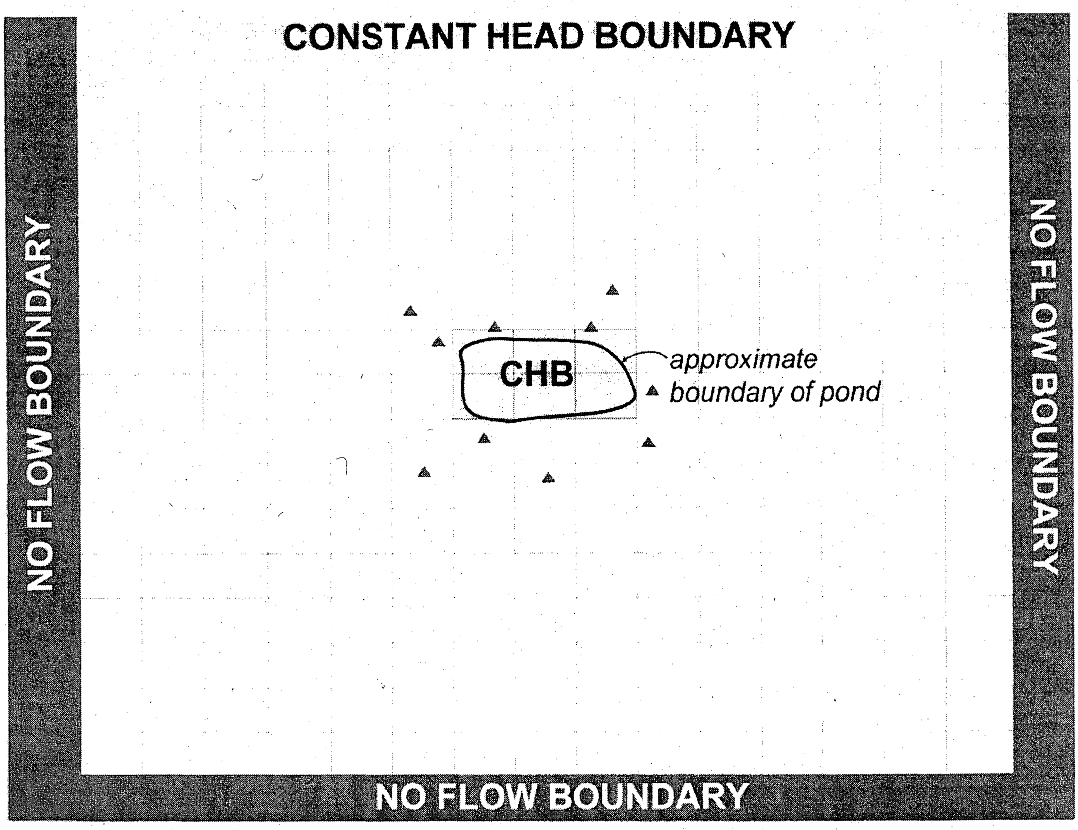

Fig. 5. Cells used for finite-difference groundwater flow model. Constant head boundaries (CHB) and no-flow boundaries noted by shading. Locations of monitoring wells used for calibration marked with triangles.

table maps constructed from these data allowed determination of groundwater flow directions (Fig. 3).

We described sediments taken from eight exploratory hand auger holes drilled along three transects within the clearing of Green Pond (Fig. 3). Data on the depth, Munsell color, texture, mottles, consistency, structure, cementation, and coarse fraction content proved most useful (see Jacobson, 1987 for techniques). The holes were drilled to bedrock, if possible, or to a maximum of 2.25 m; at this depth the overlying weight and pressure of the groundwater on the hand auger severely impeded removal of the auger. Stratigraphic cross-sections show the vertical and horizontal distribution of materials underlying Green Pond (Fig. 4).

We sampled pore water within the peat and underlying mineral material during one day and analyzed it for dissolved silica. Piezometers inserted various depths into the peat collected water that was extracted with a peristalic pump and analyzed in the field using a silica test kit accurate to within 2 ppm. Serial dilutions with deionized water permitted estimations of up to 300 ppm in silica rich areas.

These field data indicate that Green Pond lies in an elongated drainage basin that sits on a plateau underlain by nearly level Antietam quartzite. The land surface generally slopes to the north and overland flow exits the basin during and after some rainstorms through a low portion of the northern drainage divide. Groundwater flows into the pond, draining from a low ridge south of the pond and from the surrounding summit plain (Fig. 3).

Soils on the plateau surrounding the pond are usually less than 2 m thick and mostly consist of gravelly loamy sand to very gravelly sand. Loamy sediments usually occur near the surface, whereas the very gravelly sand is normally near the soil-bedrock interface. These geomorphic and textural characteristics suggest that the soils are residual materials derived from the underlying quartzite and have experienced little if any lateral movement. Peaty deposits up to a meter thick underlie the pond area. In some locations beneath the peat, fine grained mineral soil exists and can be quite clayey, especially in the southwestern portion of the pond area. These zones might be attributed to variations in parent material but their existence only beneath the peaty wetland soils suggests they are residual soil material with a history of more intense weathering, perhaps due to higher concentrations of organic acids. Our reconnaissance grade observations of silica concentrations in pore waters from piezometers 1-2 m away from the pond ranged between 20-300 ppm. Water collected from farther away was much lower. At Observation Well #10, silica concentrations were usually 2.5-3 ppm; springs at the headwaters of Cole's Run and Kennedy Creek averaged 3-4 ppm.

The lack of widespread clays or residual saprolitic dolomite suggests that Green Pond did not form by solution of Tomstown (Shady) dolomite. No direct evidence exists, either, to support a formative hypothesis related to periglacial processes. However, the levels of dissolved silica in groundwater near Green Pond that are an order of magnitude higher than in "background areas" indicate that Green Pond may have formed by dissolution of quartz with the aid of organic acid (e.g., Reed et al., 1963).

HYDROLOGIC ANALYSES

Analytical models or steady-state finite difference models can effectively simulate groundwater flows in some hydrogeological settings. The advantage of these models is that they are relatively simple. Where the geologic setting is relatively uniform and the only source of water is rainfall, the model may consist of only a single equation that relates the profile of the water table to simple estimates of the hydraulic conductivity (K) and geometry of the aquifer and the rate of recharge (W) (e.g. Fetter, 1980; p. 136). If the goal of the model is to describe heads in wells over large areas, a finite difference model can generate a contour map of the water table, but the initial data requirements are equally simple. The largest problem arises in estimating a single "steady state" value to represent recharge, a very "transient" variable. It is easy to obtain recharge estimates based on water budget calculations that use monthly temperature and rainfall data. However, our previous studies have shown that a monthly recharge value (W_{mo}) is insufficient when trying to simulate fluctuating water levels in ponds (Whittecar & Emry, 1992). Antecedent rainfall and evapotranspiration during several months must be incorporated in the calculation of the recharge value used in these simple models. As we will explain, we use a time-averaged recharge value, the "effective monthly recharge" (W_{em}), to overcome this problem.

We used a two-dimensional finite difference aquifer simulation model (PLASM) to analyze the groundwater flow in the Green Pond area (Prickett & Lonnquist, 1971). PLASM operates on a grid of nodes and calculates the discharge of water entering and exiting cells around each node. Each cell carries a value of hydraulic conductivity (K), aquifer thickness (b), and recharge rate (W). PLASM will predict the hydraulic heads that will occur across the area under given conditions. The results of this prediction should closely replicate field-measured heads; if a poor match results, one or more of the models parameters are changed and the model is rerun. This calibration process continues until a match falls within the

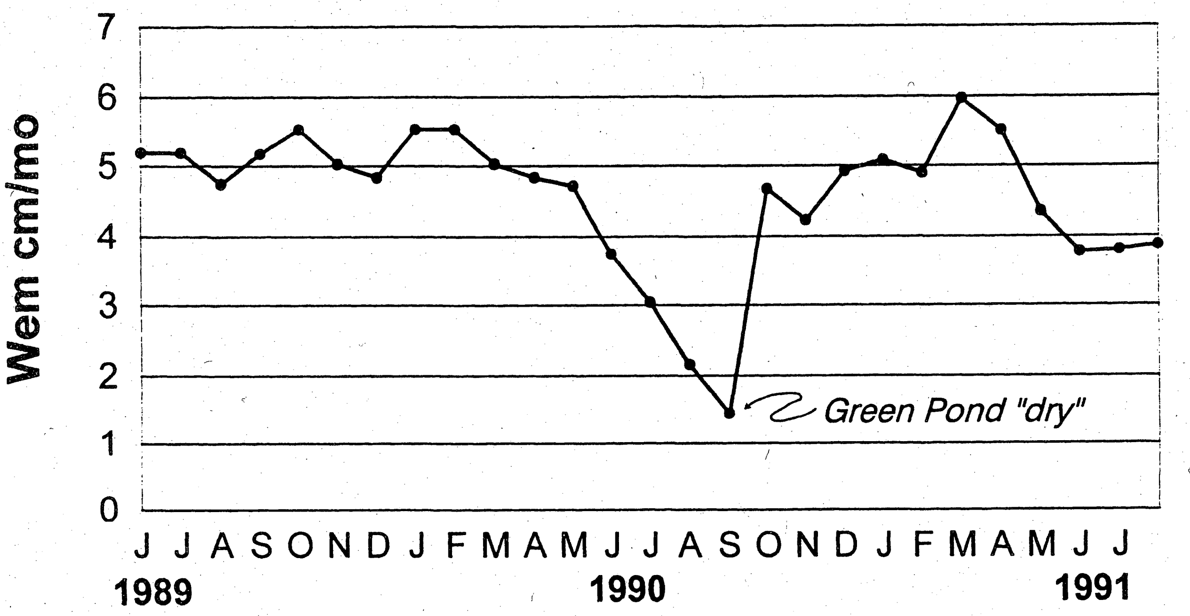

Fig. 6. Plot of Effective Monthly Recharge (W_{em}) for each month of the study period. Note that Green Pond contained no visible surface water in September 1990. W_{em} calculated using equations described in text. Recharge values for 12 months (n = 12) and a response-decay constant (D = 0.90) were used.

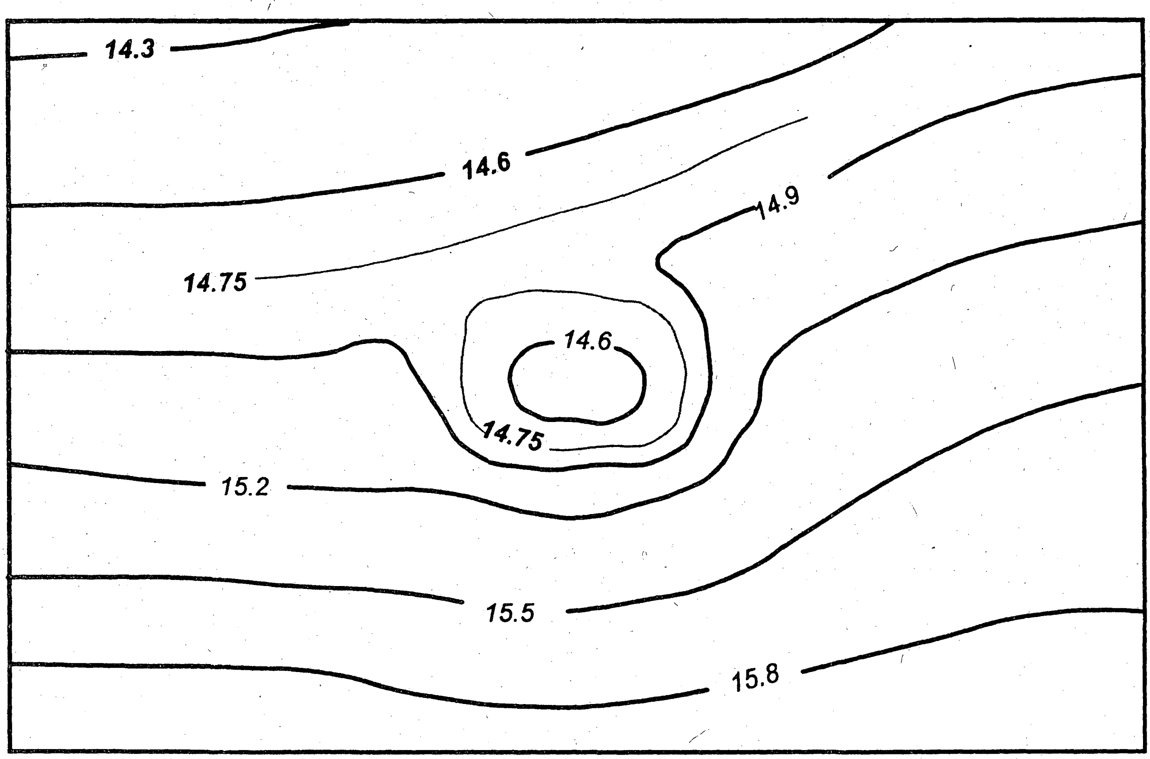

Fig. 7. Map showing distribution of heads predicted by the groundwater flow model for the Green Pond area for October 1990. Compare with Fig. 3.

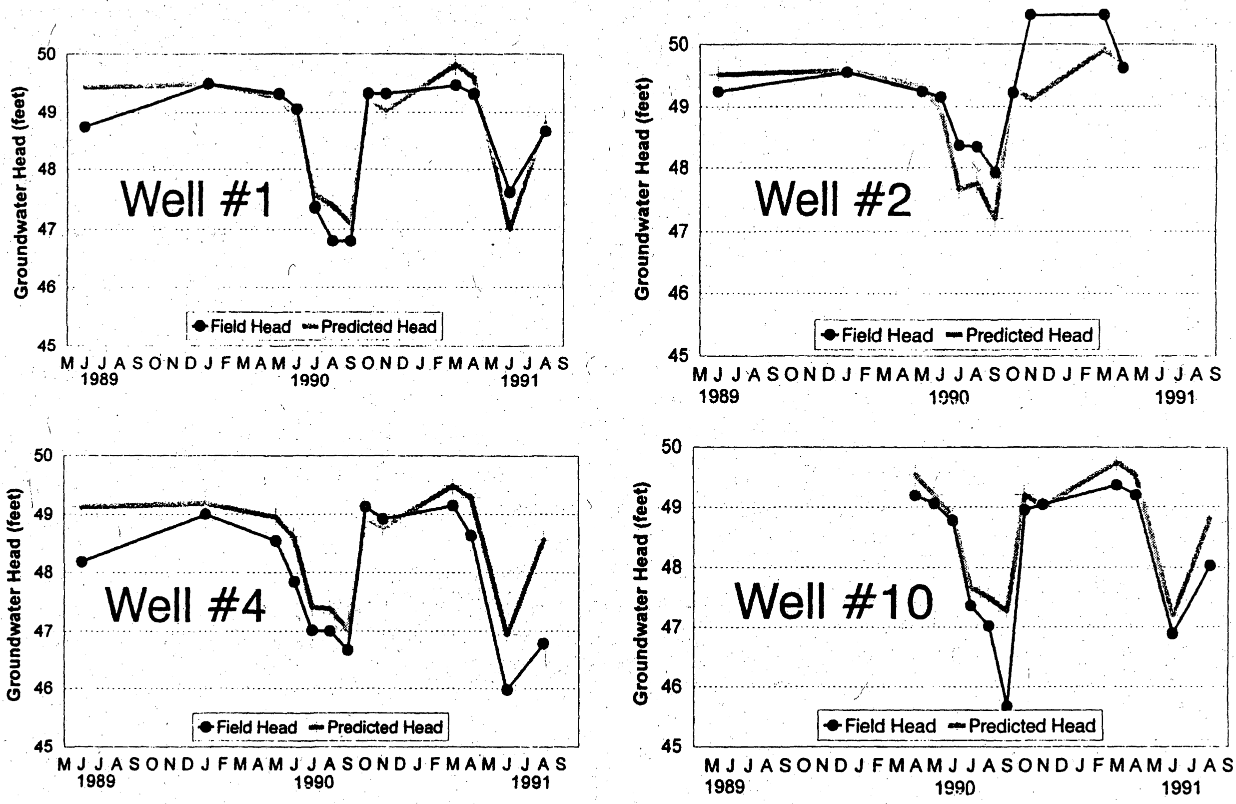

Fig. 8. Final calibration charts for results of finite difference groundwater flow model. Heads predicted by model are compared with measurements in observation wells 1, 2, 4, and 10.

error margin pre-established by the modeler (Wang & Anderson, 1982; Anderson & Woessner, 1992).

We discretized the study area into a 17x18 grid with 90-foot spacing between nodes (Fig. 5). The six cells that represent Green Pond lie at the center of the grid. The constant head boundary set along the northern boundary represents an inexhaustible supply of water located far enough from the pond to prevent interference with the pond fluctuations. No-flow boundaries along the other sides of the grid establish flow lines parallel to those margins in the areas away from the lake. These boundary condition replicate the field hydrologic setting caused by a drainage divide positioned on a low ridge more than 100 m south of the pond and the broad expanse of the plateau summit.

For hydraulic conductivity (K), our initial estimates (e.g. 100 gpd/ft^2) came from particle size analyses of soil samples at six sites around Green Pond and equations that relate mean grain size and sorting to hydraulic conductivity (Masch & Denny, 1966). The estimated value used for aquifer thickness (b = 1.5 m) in all cells came from our observations and soil survey data; we subtracted this value from the surface elevation taken from our detailed contour map (Fig. 2) to obtain the elevation of the aquifer bottom in each cell. The specific yield of unconsolidated sands, an estimate of storage (S), can range from 0.1 to 0.3 (Fetter, 1980); the best value of this parameter (S = 0.3) was obtained via trial and error.

We assigned a recharge value to each cell not in the pond by using precipitation records from Montebello, Virginia and temperature records from Big Meadows, Virginia, the closest sources of data from similar elevations. The equation used for recharge:

$$W_{mo} = P_{mo} - (P_{mo} \text{ x } I_{mo}) - Et_{mo}$$

where P_{mo} is the monthly precipitation rate, ${}_{mo}I$ is the interception factor based upon the vegetation present, and ET_{mo} is the evapotranspiration calculated using the Thornthwaite (1957) method based on air temperatures. Interception values for each month (0% for May-September; 20% for October-April) represent the reduction of rainfall due to evaporation off of the mix of hardwoods and conifers present in the area (Helvey, 1971).

Whittecar & Emry (1992) demonstrated the usefulness of the "effective monthly recharge" (W_{em}), calculated by this formula:

$$W_{em} = W_s / d$$

where

$$W_s = \sum_{a=1}^{n} (W_{mo} \times D^{a-1}) \text{ , and}$$

$$d = \sum_{b=1}^{n} D^{b-1}$$

where a and b are equal to the number of months prior to the month for which W_{em} is being calculated; and n is the number of months used in the calculation. D is a response-constant decay factor, usually between 0.90 and 0.99, that reflects the rate of reduction of water levels in the system; d is a normalizing factor (also see Linsley & Kohler, 1951). Fig. 6 shows W_{em} values for the months of this study.

With this capability to time-average recharge values, we ran the model in a steady-state mode for eight steps that covered 250 days, an arbitrarily large time chosen to assure convergence of solutions. Typically differences in head values calculated between successive iterations reduced to nearly zero within the first several time steps. The first step of calibration attempted to adjust the input parameters into the model so that the spatial distribution of predicted heads resembled the pattern of heads measured in the field. Heads gathered on 14 October 1990 reflect a relatively high water table and thus provided target values. The best comparison of spatial distribution came after raising the hydraulic conductivity value to 1,500 gpd/ft2. The success of this adjustment suggests

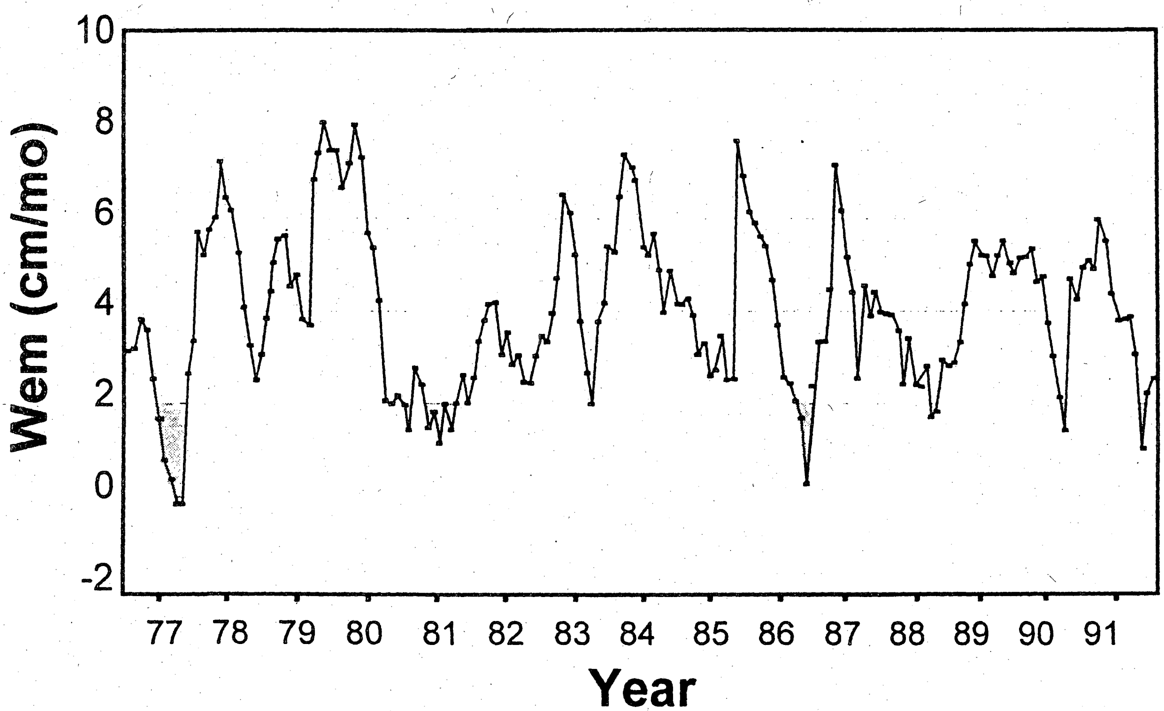

Fig. 9. Plot of Effective Monthly Recharge (W_{em}) for all months between 1977 and 1991. W_{em} calculated using equations described in text. A response-decay factor (D = 0.90) and 12 months of temperature and precipitation data from neighboring high elevation recording stations were used to calculate each W_{em} value. The shaded areas indicate the time periods that Green Pond would have been expected to go "dry," those with W_{em} less than 2 cm/mo.

that most groundwater moves through the higher permeability zones close to the base of the sandy soil profiles (Fig. 4).

The second step of calibration involved running the model for every month when we measured field heads (May 1989 through September 1991), using the effective monthly recharge values calculated with the preceding months of recharge. Measured heads in four wells near Green Pond (OW1, OW2, OW4, OW10) provided target values for each month's model prediction (Fig. 8). Use of different sets of effective monthly recharge values, calculated using different numbers of preceding months, suggest that 12 months of data (n = 12) and a response-decay constant (D) of 0.90 provided a reasonable match of predicted and measured heads. The largest consistent discrepancies of prediction occurred within the pond clearing and the surrounding forest during summer months, apparently indicating that actual evaporation exceeded the evapotranspiration calculated with the Thornthwaite equations. In their earlier study, Whittecar & Emry (1992) also noted an underestimation of evaporation during the summers. We compensated for this difference by increasing the losses from cells in the pond area by 0.2 gpd/ft2 during the summer months (June, July, August).

CONCLUSIONS

The groundwater head data gathered for this study (e.g., Fig. 3) indicate that Green Pond water levels reflect moderately rapid groundwater discharge from the surrounding wetland and surficial aquifer throughout most of the year. The results of the water budget analyses suggest that the shallow loamy sand soils and the small drainage area around Green Pond would be expected to pass water from a recharge event over a scale of one year or less. The gravelly zone at the base of the sandy regolith soils on Big Levels plateau may have higher conductivity than the overlying loamy sand that would contribute to the response of the Green Pond basin. For comparison, an analysis of pond levels in a basin on a thick, sandy barrier island aquifer needed 1.5 to 2 years of recharge data to calculate a usable effective monthly recharge value (Whittecar & Emry, 1992).

The usefulness of the W_{em} calculations lies in using long term records of precipitation and temperature to estimate the frequency of pond drying droughts, those that might place the greatest stresses upon aquatic species. During our study, Green Pond virtually disappeared during the summer of 1991, a time when the calculated W_{em} dropped below 2 cm/mo. Fig. 9 shows the W_{em} for all months from 1977 to 1991, as calculated from weather records kept at Montebello, Virginia. During that period, W_{em} values dropped below 2 cm/mo in six summers, usually for only a month or two but once for a period of 5 months. These data suggest that species thriving in Green Pond must have adapted to an episodic fluctuation of water levels that can dessicate the pond every few years.

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