

Hydrogeologic Controls on Lake Level at Mountain Lake, Virginia

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ABSTRACT

Mountain Lake in Giles County, Virginia, has a documented history of severe natural lake-level changes involving groundwater seepage [*Jansons*, 2004] that extend over the past 4200 years [*Cawley*, 1999], and as of December 2010 the lake was about 2% full by volume. Situated in the Valley and Ridge physiographic province on the axis of a plunging anticline and straddling contacts between three upper Ordovician and lower Silurian formations, the lake is one of two natural lakes in Virginia.

A daily water balance, geophysical surveying with dipole-dipole electrical resistivity, and chemical sampling have shed light on the nature of flow to and from the lake, including: 1) the steady nature of net groundwater outflow, 2) the seasonal response to precipitation of a forested first-order drainage system in fractured rock, 3) the influence of a fault not previously discussed in literature regarding the lake, and 4) the possibility of flow pathways through karst features.

Results from a water balance indicate steady lake drainage and significant recharge when vegetation is dormant, particularly during rain-on-snow melt events. The resistivity profiles display a highly heterogeneous subsurface and reveal low-resistivity areas that suggest flow pathways to and from the lake. Well logs, satellite images, and outcrop observations appear to confirm the presence of a fault to the east of the lake. Chemical evidence suggests that karst features may be present in the upper Reedsville-Trenton formation underlying the lakebed.

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List of Abbreviations

AET	actual evapotranspiration
AGI	Advanced Geosciences, Inc.
amsl	above mean sea level
ArcGIS®	Geographic information system software by Esri
CFC	chlorofluorocarbon
DEQ	Department of Environmental Quality
ERT	electrical resistivity tomography
ET	evapotranspiration
GIS	geographic information system
GNIS	Geographic Names Information System
GPS	global positioning system
IDW	inverse distance weighting
ML	Mountain Lake
MLBS	Mountain Lake Biological Station
MLC	Mountain Lake Conservancy
MLH	Mountain Lake Hotel
NOAA	National Oceanic and Atmospheric Administration
NCDC	National Climactic Data Center
NOHRSC	National Operational Hydrologic Remote Sensing Center
NWS	National Weather Service
PAR	photosynthetically available radiation
PET	potential evapotranspiration
SWE	snow-water equivalent
TDS	total dissolved solids
USGS	United States Geological Survey

1. Background and History of Mountain Lake

Mountain Lake is located in the Valley and Ridge physiographic province [Fenneman, 1938] in southwestern Virginia. The lake is situated near the top of Salt Pond Mountain in eastern Giles County (Figure 1), six miles east-northeast of Pembroke, Virginia, on private property owned by the Mountain Lake Hotel (MLH) and Mountain Lake Conservancy (MLC). The location was operated as a resort as early as 1857, and still contains structures from this period; the current hotel building dates to 1936. The hotel has three public supply wells, which includes one new well not yet in operation, and serves a transient population of up to approximately 4200 people per month [DEQ, 2010]. The location of the lake and historic hotel within a privately-owned nature preserve is not to be confused with the adjacent Mountain Lake Biological Station [MLBS] belonging to the University of Virginia or the adjoining 16,500 acre Mountain Lake Wilderness, Virginia's second largest federally-designated wilderness area.

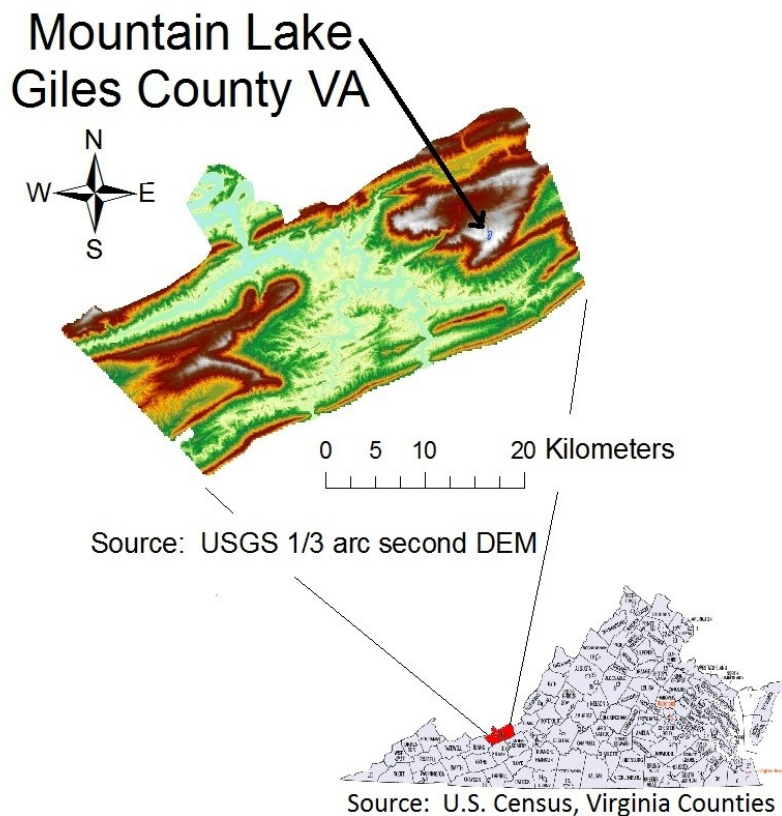


Figure 1: Location of study area

[Map sources: USGS 1/3 second digital elevation model, 2009 and
U.S. Census Virginia Counties Map, 2009]

Mountain Lake is one of only two natural lakes in Virginia, represents the only natural lake in the unglaciated Appalachians, and is the highest-altitude named U.S. lake east of western Kansas and south of New England [GNIS, 2010]. When full, the lake covers an area of $1.94 \times 10^5 \text{ m}^2$ and has been measured to have a maximum depth of 33.5 m [Williams, 1930].

The lake is located 1.2 km from the eastern continental divide. Its position with respect to USGS-defined hydrologic units is shown in Figure 2, and three first-order streams in the vicinity of the watershed are depicted in Figure 10. To the east of the lake and across the eastern continental divide, Sartain Branch drains into the Mid-Atlantic Hydrologic Unit (02) [Seaber, 1987]. The lake itself is located in the Ohio Hydrologic Unit (05), where Pond Drain drains to the north and west and receives the lake's surface outflow when full, eventually draining to the New River. To the southwest, the first-order stream, Doe Creek, also drains toward the New River.

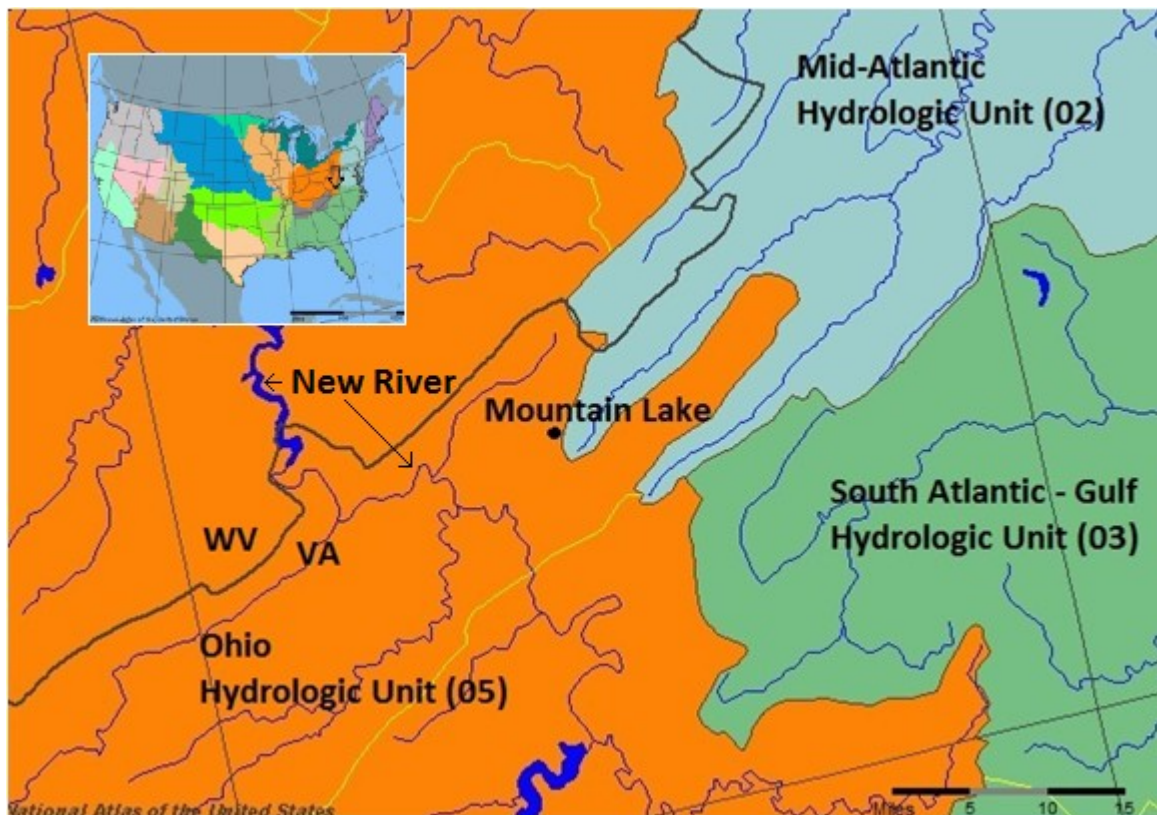


Figure 2: Location of Mountain Lake within USGS hydrologic units

[Map source: National Atlas of the United States, 2010]

Mountain Lake overlies contacts between three different geologic formations (Figure 3). At points near the southwestern end of the lake the Upper Ordovician Reedsville-Trenton

formation is exposed, consisting of interbedded siltstone, shale, and limestone. The Ordovician Juniata formation is exposed in several points along the lakebed stream that feeds the lake when the level is low, and consists of sandstone, siltstone, and shale. The Silurian Tuscarora formation is an orthoquartzite that is locally conglomeritic present in place along Mountain Lake Road to the west of the lake, collapsed in the lakebed, and in the form of 5 m high boulders to the north of the lake. The Silurian Rose Hill formation, a hematite-cemented sandstone and shale, overlies the Tuscarora and is also present as float within the lakebed [Eckroade, 1962]. Structurally, the lake lies on the crest of the doubly-plunging Bane anticline that extends across the length of Giles County (Figure 34) [McDowell and Schultz, 1990].

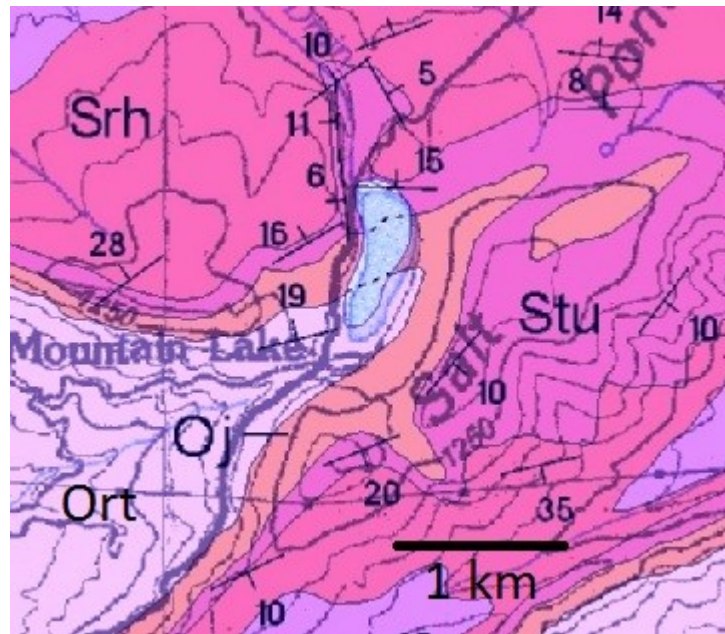


Figure 3: Mountain Lake and geologic formations

Mountain Lake in blue; Ort, Ordovician Reedsville-Trenton; Oj, Ordovician Juniata; Stu, Silurian Tuscarora; Srh, Silurian Rose Hill [Bartholomew et al., 2000]

Past Lake Levels and Hydrologic Studies

Mountain Lake has had a history of lake-level fluctuations. Historical evidence for past changes in lake level begins with a description of a small lake fed by springs by the first European explorer in 1751 [Johnston, 1898]. A drawing of the lake in a 1789 deed for the property shows a low lake elevation, and the mountain to the south of the lake, in a possible reference to the state of the lake at the time the mountain acquired an English name, is Salt Pond Mountain. Biological evidence of lake level changes includes both sediment core and tree-ring analyses. Marland [1967] analyzed sediment cores and found zones of littoral rather than

planktonic microfossil assemblages indicating past low lake levels. Parker [1975] analyzed a tree stump rooted in the lake bed, documenting the existence of a yellow pine rooted 10 m below the level of full lake with 22 annual growth rings with a ^{14}C date of 1655 ± 80 , showing the lake must have been low during this period for over two decades. A second 20-year old tree was dated to 1892 ± 50 years [B C Parker, 2003]. Cawley [1999] used sediment core analysis and radiocarbon dating of pollen, spore, and microfossil assemblages to both refine and extend the period of analysis back in time and concluded that six extended periods of low or empty lake levels had occurred around 1900, 1600, 1100, 800, and 200 AD, and 2200 BC. The earliest radiocarbon date indicated that the lake was full in 4000 BC. The reader is referred to Parker [2003] for a comprehensive review of this historical evidence.

While it may be thought that this hydrologic system is a simple one and that lake levels are simply sensitive to recent precipitation, there are at least two examples from the historical record that suggest otherwise. First, the lake was measured to have a depth of 33.5 m in the summer of 1930 [Williams, 1930], indicating the lake was full, despite a drought in 1930-1931, which the Palmer Drought Severity Index for southwestern Virginia's division 6 shows to have been "extreme" and by this metric is the most severe drought over the period of record from 1895 to the present [NCDC, 2011]. During the summer of 1931, a paper resulting from two visits by Hutchinson and Pickford [1932] does not make any mention of lower lake levels except to say that the then-director of the Mountain Lake Biological Station [MLBS] thought that "the alleged going dry [of ML] in the past was probably false or exaggerated" (Parker 2010, pers. comm.). In addition, two publications from the years before and after the drought do not indicate that levels were changing [Williams, 1930], [Sharp, 1933]. However, in 1935, a few years after this drought had ended, Lewis [1957] reported a moderate drop in lake level, suggesting a longer lag time for lake response than would be expected if immediate precipitation alone were responsible for declines in lake level. The second example is from 1969-1970 when the lake is reported to have been full [B C Parker, H.E. Wolfe, R.V. Howard, 1975] despite coinciding with a moderate drought representing one of the three longest lasting droughts (two and a half years) over the period of record [NCDC, 2011]. Thus, the lag time response of the lake to precipitation is not well understood.

After a period of time in which the lake was continuously full (1969-1997) [B C Parker, 2003], the lake more recently experienced fluctuations in level that correlate with periods lower rainfall [Jansons, 2004], including a complete drying up of the lake for three days in September 2008. During this period, parts of the deep lake experienced a fish die-off, but dead fish were not present in and around the deepest hole when it dried up, and the fish population returned after the lake began filling [Parker, pers. comm. 2009], indicating that some type of reservoir likely extends below the lake bed and/or underneath the lake shore. Figure 4 displays a photo of a dry crevice measuring 0.8 m by 0.5 m.



Figure 4: Photograph of a lakebed crevice

Crevice was exposed in September 2008 when lake was near empty. Approximate dimensions: 0.8 m x 0.5 m [credit: Greg Zielske 2008, used with permission]

Other hydrologic features have also been documented. Surface streams and springs in and around the lake have also been mentioned in historical documents [Johnston, 1898]. An engineer reports that a spring discharging at the south end of the lake was said to discharge below the level of the lake “during normal times”, without specifying the nature of the change from abnormal times [Meredith, 1934]. Scuba divers documented springs along the lakebed [Cawley, 1999]. Deevy, Jones and Daly [Roth, 1964] produced a bathymetric map of the lake bottom using a plumb line, and Cawley [1999] produced a bathymetric map from seven sonar transects.

A monthly water balance was calculated using lake levels from February 2002 to September 2003. During this time the lake level dropped to a maximum of 6.6 m below capacity in November 2003 [Jansons, 2004]. Lake volume changes were calculated based on lake level measurements taken by hand and Cawley’s bathymetric map. The National Weather Service (NWS) Automated Flood Warning System weather station data derived from the MLBS rain gauge sensor was used for monthly precipitation totals. Evapotranspiration was calculated using monthly mean values for input parameters to the Penman-Monteith method. The water balance took into account basin precipitation, basin evapotranspiration, well abstraction, and lake volume change to calculate the net groundwater outflow. Calculated on a monthly basis, net outflow ranged between 2.6 and 107 liters per second, with an average over the 18-month period of 43 liters per second.

Lake Formation and Geologic Studies

The presence and location of Mountain Lake is anomalous for this part of the Appalachians. Past geologic work that has focused largely on mechanisms of lake formation

may also be relevant to understanding possible pathways of drainage from the lake. Rogers [1884], Hutchinson and Pickford [1932], Sharp [1933], Eckroade [1962], Mills [1988], and Parker [2003] discussed evidence to support the theory that Mountain Lake is a product of the damming of a stream by some type of rock slide or creep. Parker [1975] states that a preexisting stream along the axis of the breached anticline would have caused preferential erosion underneath the Tuscarora that could have caused a slide. He also states that documentation of seismic activity [Hopper, 1971] centered in Giles County points to possible influence of earthquakes on the formation of the lake, and reports anecdotal evidence of a rise in lake level following an April 1959 earthquake. Mills [1988] characterized types of boulder streams in the area that formed by collapse of resistant sandstone formations including the Tuscarora, and addressed possible mechanisms of movement, including sliding, creep, periglacial processes, and debris flows, all of which could be relevant to the structure of drainage pathways from the lake.

Several authors have discussed the possibility of karst dissolution in relation to Mountain Lake. Holden [1938] first suggested dissolution of a carbonate layer in the upper Reedsville-Trenton as the reason for collapse and subsequent formation of the lake. Butts [1940] describes a 2-m thick layer of carbonate at the top of the Martinsburg along U.S. 460 (then Rt. 8) just north of Narrows, Virginia. Marland [1967] describes the Juniata and Martinsburg [Reedsville] formations together as being primarily shale, “some of which are calcareous, with thin beds of limestone.” Parker [1975] described an outcrop of Reedsville-Trenton 60 m below the bottom of the lake outside of the lake’s watershed that was 50% carbonate. Cooper [1964] and Fiedler [1967], however, argued that no sinkholes have been found to exist elsewhere in the Martinsburg [Reedsville], and based on evidence in a tunnel that cuts through a complete section of the formation, argues that even if dissolution occurred the formation would retain its structural integrity.

Another factor that has been discussed as both a mechanism of lake formation and a geologic control on lake levels is a posited northwest-trending fault or joint through the lake at the location of the deepest crevice. Cawley [1999] mapped a lineament through the crevice and conducted a geophysical survey using electrical resistivity along four transects in an attempt to confirm the presence of a fault or feature corresponding to such a lineament. The resistivity survey, which had a maximum apparent depth of 15 m, was interpreted to confirm this feature.

2. Purpose and Scope of this Investigation

While lake level fluctuations have been and are ongoing, the precise hydrologic and geologic mechanisms for these fluctuations have not been demonstrated. An understanding of the hydrogeologic system at Mountain Lake could indicate whether changes in groundwater outflow are necessary to explain past lake level changes, and will provide a much needed baseline so that potential future changes in outflow caused either by natural events or manmade interventions can be readily evaluated.

Four Hypotheses

In an attempt to clarify the hydrologic and geologic mechanisms for lake level fluctuations, this study responds to and builds on previous investigations, addressing four hypotheses. The first two hypotheses address the hydrology of the lake: 1) net groundwater outflow is constant over time; 2) groundwater outflow is dependent on lake level. The third and fourth hypotheses address how geologic structure and stratigraphy may be impacting the water balance of the lake: 3) a northwest-trending fault or joint extends through the lake and is a structure of hydrologic significance; 4) karst dissolution is a viable possibility for groundwater flows into and out of the lake. The individual hypotheses are discussed below.

First Hypothesis: Net groundwater outflow is constant over time.

The current study first aims to reevaluate the results of the water balance conducted by Jansons et al. [2004]. In that study, average net groundwater outflow over the 18-month period was calculated to be 43 liters per second, but on a monthly basis, net groundwater outflows were calculated to have varied by a factor of as much as 40. According to this study, net outflow appears to have local lows in the months of January, March, May, and August, and local highs in February, July, and October and thus does not seem to correspond to any obvious seasonal pattern. As the author acknowledged, it is not clear if the variation in monthly net outflow reflects true hydrologic conditions or is an artifact of the monthly time step and the lag between precipitation and lake recharge. Also, errors appear to have been introduced in evapotranspiration calculations and may also have been introduced by not taking into account snowmelt or storage changes in the soil and groundwater. In the current study, a more complete, higher-resolution water balance is constructed in order to understand the variation in net groundwater outflows over time.

Second Hypothesis: Groundwater outflow is dependent on lake level.

The second research problem considers the effect of lake level on groundwater outflow. Questions of how the change in storage of the lake is related to groundwater outflows, whether groundwater outflows may be head-dependent, and how lake leakage may be hydraulically and geologically controlled have not been evaluated. Lake level variations during this study provided an opportunity to attempt to address these questions.

Third Hypothesis: A northwest-trending fault extends through the lake and is a structure of hydrologic significance.

A posited northwest-trending fault running through the lake at the location of the deepest crevices was investigated with a geophysical survey using electrical resistivity [Cawley, 1999]. The four published resistivity lines were limited to a length of 250 m, which provided an apparent imaging depth of only 15 meters, which is not sufficient to establish the presence of a potential hydrologically active fault in what is presumably a fractured and very heterogeneous geologic medium. The Wenner array used in Cawley's work is a preferred array for investigating horizontal stratigraphy whereas the dipole-dipole array has been found to be more effective at imaging vertical structures [W] Seaton and Burbey, 2002]. In this study a more

extensive geophysical survey, combined with lineament analysis and joint sampling at outcrops, is used to investigate geologic structures around the lake.

Fourth Hypothesis: Karst dissolution is a viable hydrogeologic pathway for groundwater flows to and from the lake.

In the summer of 2008 as the lake's level was lowering prior to completely drying up in September, Parker [2008] described an unusually high pH in the month of July and an abnormally large presence of a community of algae that use bicarbonate as their photosynthetically available carbon source, suggesting flowpaths to the lake through carbonate rocks. In addition, the form of the lake bottom as it was revealed that fall was compared to sinkholes and may have similarities with a collapse structure in quartzite on Roanoke Mountain which is not itself in carbonate rock but lies 213 m above a dolomite formation and appears to have formed via piping [Henika 2009, pers. comm.]. The possibility of karst dissolution at the lake will be addressed again in light of chemical sampling and geophysical surveying.

3. Possible Hydrologic Controls Influencing the Lake

In an effort to quantify net groundwater outflow from the lake to address the first two hypotheses, a water budget was developed for the lake watershed for the period from May 2009 to November 2010 with the following components:

$$In - Out = \Delta Storage \quad (1)$$

$$R + S - Abs - E_L - AET - GW_{out} = \Delta S_L + \Delta S_{SGW} \quad (2)$$

where

R is rainfall

S is snow water equivalent

Abs is well abstraction

E_L is lake evaporation

AET is actual evapotranspiration through soil and vegetation

GW_{out} is net groundwater outflow

ΔS_L is the change in lake storage

ΔS_{SGW} is change in soil and groundwater storage.

This water balance differs from that used by Jansons [2004] in two ways: First, snowmelt and change in soil and groundwater storage are included in the water balance. Second, whereas Jansons used the water balance to calculate the unknown net groundwater outflow, in the

current study this element is estimated independently, and the mass balance equation is used to solve for the soil and groundwater storage term ΔS_{SGW} .

Change in Lake Storage

The lake level (or *stage* or *head*) at Mountain Lake changes over time, responding to changing hydrologic and perhaps geologic conditions. In order to evaluate factors that may be affecting lake stage, the change in lake volume must be determined. Quantifying that change in volume, or *change in lake storage*, is central to the water balance. To calculate the change in lake storage, the volume of the lake was calculated as a function of increasing lake stage. Because of the irregular three-dimensional shape of the lake, an accurate bathymetric map of the lakebed was required. Then, lake stage measurements were made over an 18-month period, from which hourly and daily changes in lake storage could be derived. Lake level changes at Mountain Lake are then associated with changes in the volume of water in the lake and used in a mass water balance.

To create the bathymetric map, elevation data points were collected from three sources. A Hummingbird sonar was used in conjunction with Trimble® differential GPS to register depths of 1142 points on the portion of the submerged portion of the lakebed in July 2010, with concentrated sampling near the deep holes. A further 744 points with altitude readings were obtained using the differential GPS on the dry portion of the lakebed, with concentrated sampling along the thalweg of the southern stream. Vegetative canopy prevented sufficient communication with satellites for accurate altitude measurements along the rim of the lake, so 270 points from the USGS 1:24,000 Newport quadrangle were digitized around the edge of the lakebed in following the 3880 and 3900-ft. contours. The complete set of points used is depicted in Figure 5.

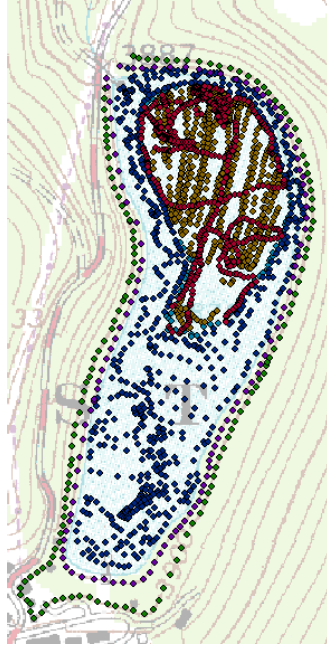


Figure 5: Data points used in creation of bathymetric map

**[Background map source: USGS 7 ½ minute Digital Raster Graphic (DRG),
Eggleston]**

For purposes of comparison, four methods were used to create raster interpolations of the bathymetry of the lakebed in ArcGIS®: natural neighbor, inverse distance weighting (IDW), and universal and ordinary kriging. As shown in Figure 6, all four methods give a similar map of the lakebed, and results were confirmed visually with aerial photos of low lake levels from 2008 and as lake levels declined again in 2010. The aerial photos suggest that the lake bottom map previously developed by Cawley [1999] placed the holes further southeast than subsequent emptying of the lake revealed them to be, and that the holes are somewhat less elongate than previously depicted. These differences are of interest geologically but would not have significantly influenced lake volume calculations. However, it is noted that the maximum lake depth is only 29.33 m, and this depth occurs only in one pixel of the interpolated (kriging) raster. It is not known if the sonar did not register the deepest part of the lake, or if previous measurements (for example, 33.5 m as measured with a plumb line [Williams, 1930]) were incorrect, or if the shape of the lake bottom has changed due to sedimentation. Further

discussion of issues related to lake volume measurements can be found in Appendices C and D.

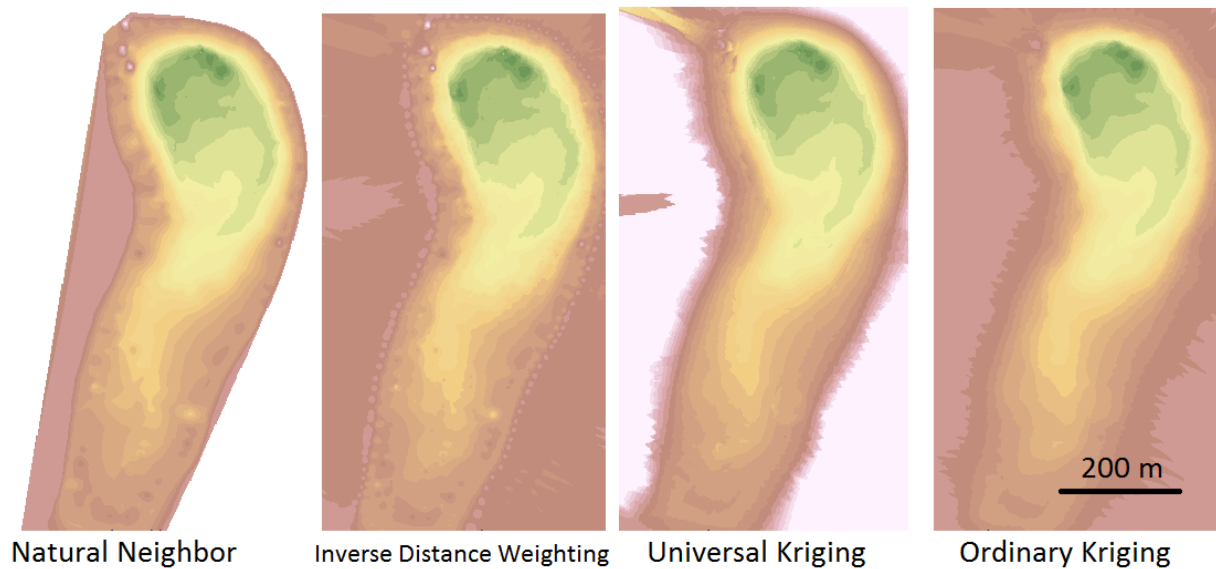


Figure 6: Raster interpolations of lakebed bathymetry

Because lake level is directly measurable, lake volumes were calculated as a function of lake level. Volumes were calculated in ArcGIS® with a cut-and-fill function that takes as its inputs the raster interpolation of the lakebed and a second raster of the lake surface at a given water level. A workflow was created in ArcGIS® to automate this process and iterate through lake levels at one-meter intervals. The volumes for full lake for each interpolation method are given in Table 1. The maximum difference between the four calculated volumes corresponds to 1.8% of the smallest total volume. Much of the volume discrepancies are likely due to differences in interpolation near the edges of the lake, and would only introduce error into a water balance when the balance is being calculated for a near-full lake, which has not been the case during this period of study. The standard IDW method for estimating topography was chosen for subsequent calculations in this study, though the universal kriging method is used in figures because it provides a smoother perimeter. The IDW lake volume calculations were subsequently interpolated linearly to correlate lake level with lake volume (Figure 7). Linear interpolation was chosen because other fits did not accurately reproduce the volume curve at the low lake levels that were encountered in the fall of 2010. Millimeter precision accommodates the hourly sampling rate and resolution of the transducers and ensures that the level-to-volume transformation is not the limiting factor in the precision of lake volume calculations.

Method	Natural Neighbor	IDW	Universal Kriging	Ordinary Kriging
Volume of full lake, m ³	1,859,495	1,883,626	1,855,985	1,889,280

Table 1: Full lake volumes according to different interpolation methods

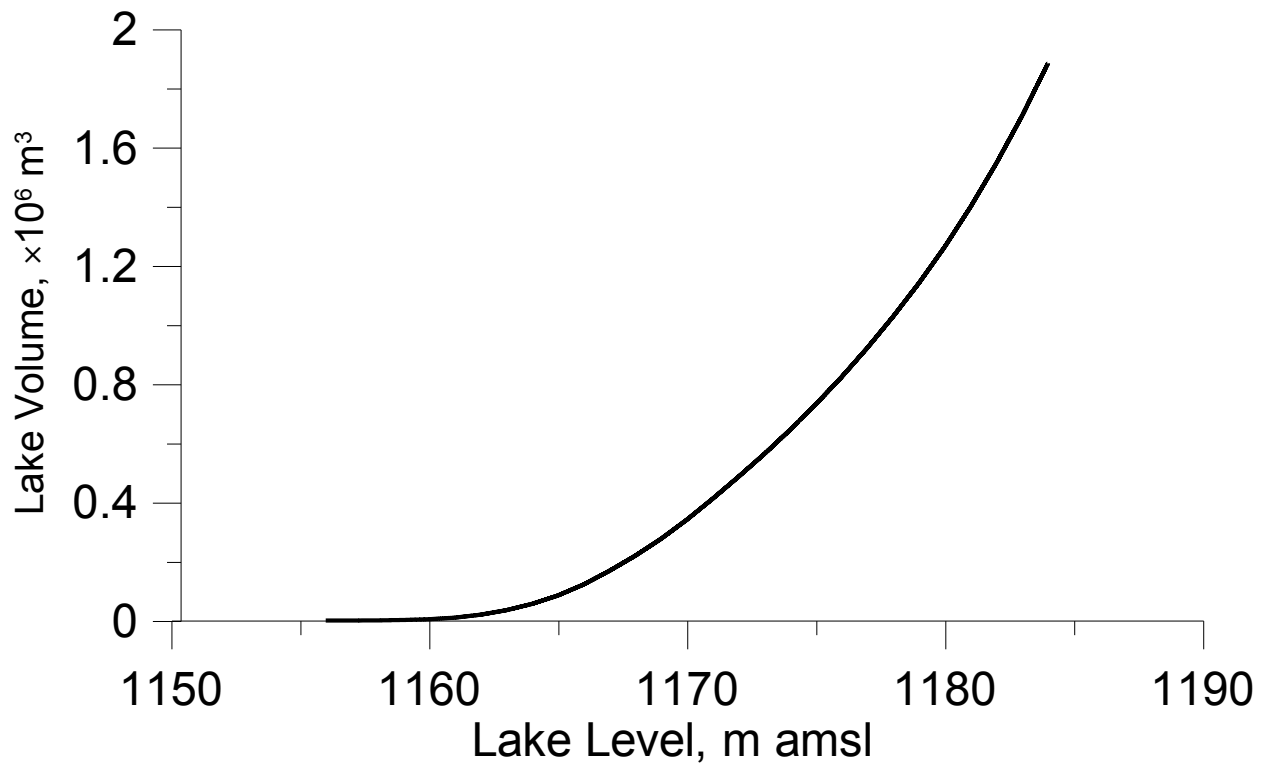


Figure 7: Lake volume as a function of lake level

Lake level in meters above mean sea level (amsl)

The stage height of the lake is the primary focus of this study and was measured directly during the course of this investigation and lake levels were then correlated with lake volume. Daily lake levels were recorded from May 2009 to May 2010 using an In-Situ pressure transducer, and hourly levels were recorded from May 2010 to November 2010 using a Solinst pressure transducer. Both records were processed to remove barometric pressure loading (Appendix F) and calibrated to points on the lakebed surveyed with a Leica® Flexline TS06

Total Station. The lake levels encountered over the period of study are shown in Figure 8. Daily (and hourly, when available) changes in lake volume were obtained, resulting in lake hydrographs that show the change in volume of the lake over time. A lake hydrograph showing the daily change in storage of the lake over the period of study is shown in Figure 9, where positive values indicate that the lake volume increased during the day, and negative values indicate a decrease in volume.



Figure 8: Lake level changes at Mountain Lake
May 2009 to November 2010

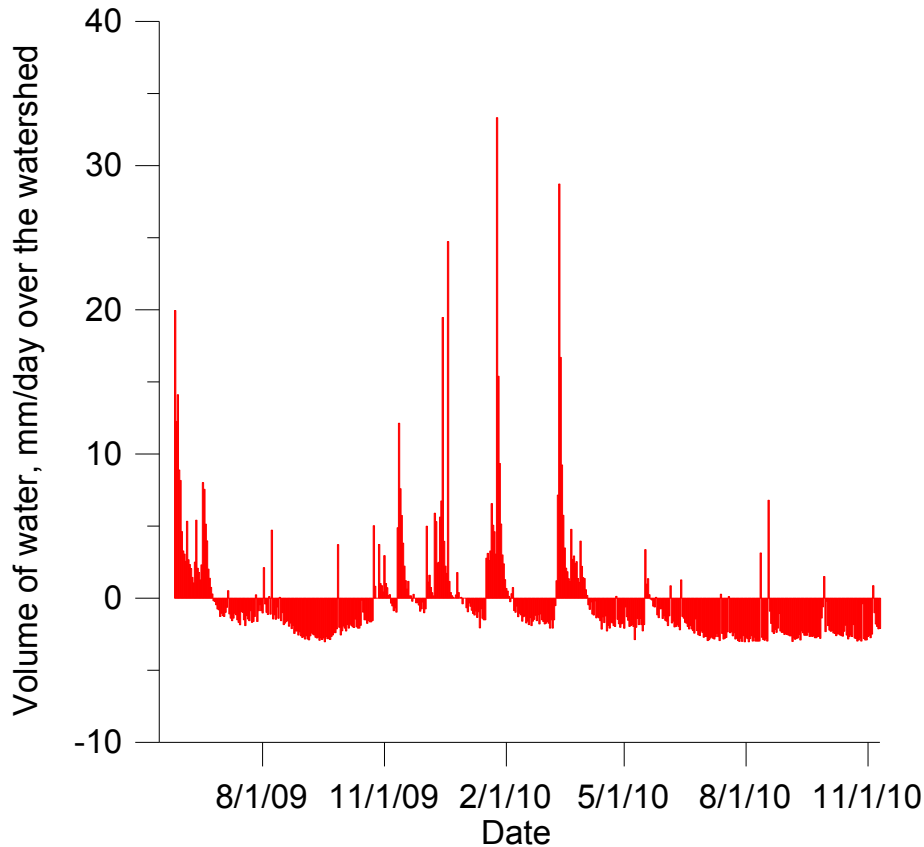


**Figure 9: Daily change in lake storage
May 2009 to November 2010**

Past accounts of the lake having been ‘full’ may not be entirely accurate. For example, a May 2000 satellite image (Figure 47) shows that the lake does not reach the spillway, though it was reported as being full for that year [B C Parker, 2003]. Without instrumentation, lake level changes are much more easily observable when lake levels are low. Due to the skewed funnel shape of the lake, thirty-three percent of the volume of the full lake is found in the top five meters of the lake. By contrast, the bottom 5 meters of the lake account for only 0.5% of the lake volume. The lake with 80% of its lakebed area covered with water is only 2/3 full by volume. Therefore the morphology of the lake may contribute to uncertainty regarding historical lake levels.

The changes in lake volume need to be expressed in units compatible with units used to measure precipitation over the watershed. To convert between a volumetric rate of change in the lake and a depth rate of precipitation, the lake’s watershed was delineated and its area measured in ArcGIS® as shown in Figure 10. Some of the sharp curves in the delineation to the west and south of the lake are due to roads or paths observed to have manmade drainage

structures or erosional gullies which channel surface water into the lake's watershed. The calculated area of the watershed is 1,334,900 m², or about 1.3 km². The surface area of the full lake as measured in the ArcGIS® is 194,200 m², leading to a ratio of the surface area of the full lake to the area of the watershed of 1 : 6.87.

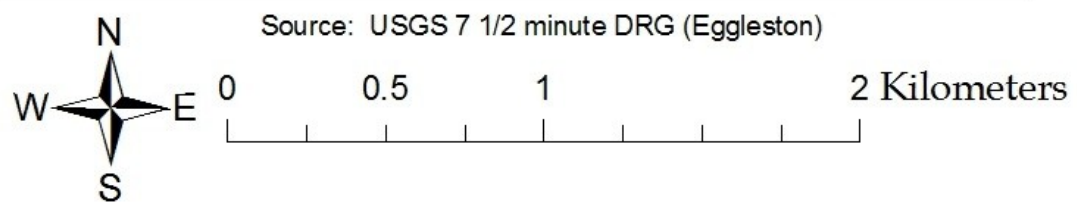
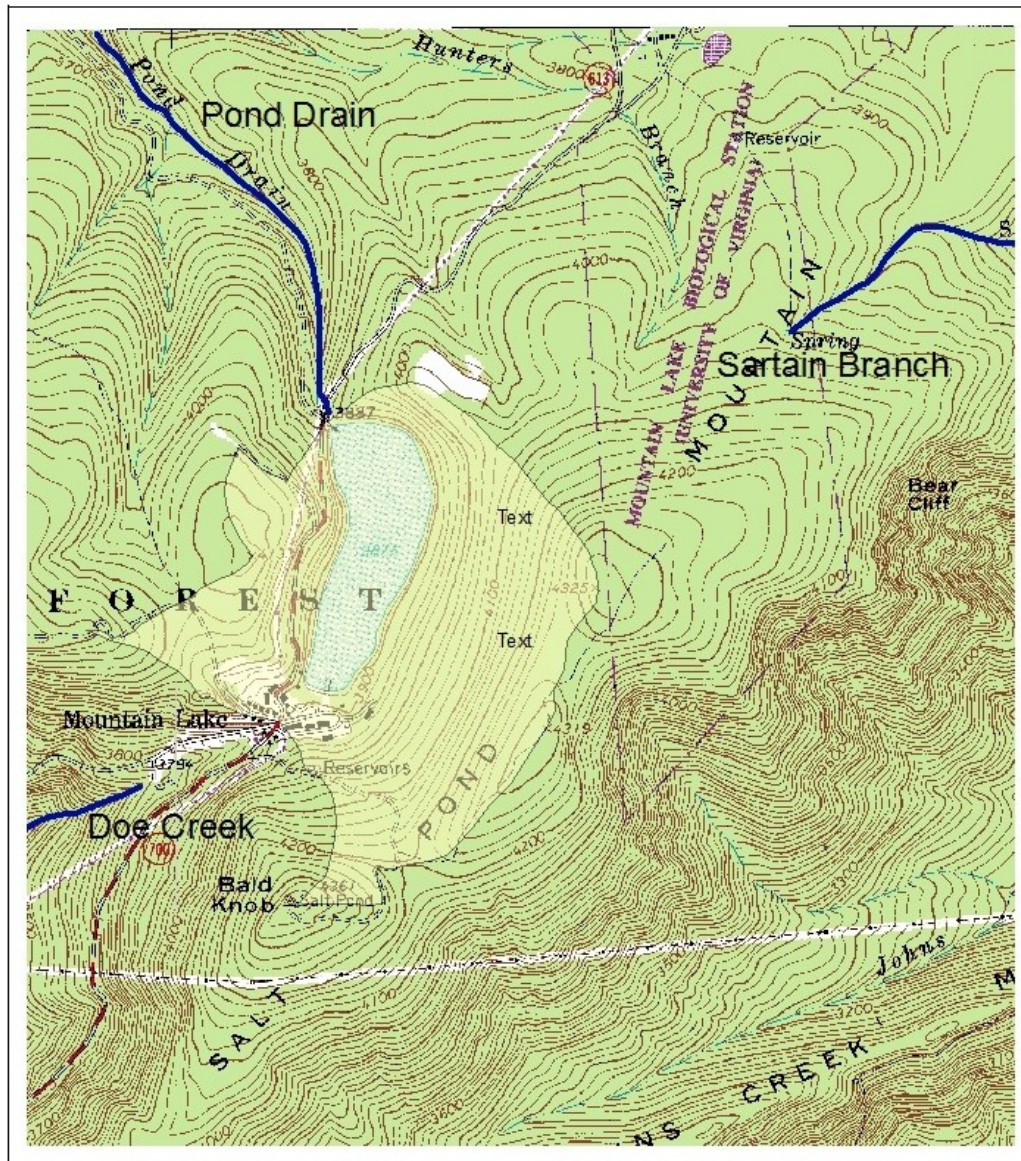


Figure 10: Mountain Lake and its watershed, with three first-order streams
 [Background map: USGS Eggleston 7 1/2 min. digital raster graphic (DRG),
 Eggleston]

Precipitation

Precipitation, in the forms of both rain and snow, is a key component in developing a water budget of Mountain Lake. At the outset of this investigation existing data sources for these components were not thought to be adequate. The closest rain gauge to the lake until May 2010 was located outside the watershed 2.4 km northeast of the Mountain Lake Hotel (MLH) at the Mountain Lake Biological Station [MLBS]. Though snowfall was a significant form of winter precipitation in 2009-2010, weather data at MLBS did not include snowfall measurements over the period of the study. Furthermore, snow depth data available from other sources in and of itself is not sufficient for use in a water balance. The previous water balance [Jansons, 2004] had used rain data from MLBS but had not considered snowfall. To improve upon the previous water balance, more complete sources of data were needed for both rain and snow inputs.

To improve upon and evaluate the existing data set, new rain data was obtained within the lake's watershed and compared to data from MLBS. A Campbell Scientific® TE525WS tipping bucket rain gauge with a CR1000 data logger recording at hourly intervals was installed at the Mountain Lake Hotel in May 2010 near the southern end of the lake. After installation, overlap of precipitation observations at the two stations allowed for direct comparison to evaluate possible spatial and temporal variations due to microclimatological or orographic effects. The rain measured at the two sites from late May to early November 2010 is shown in Figure 11. During this period of overlap, MLBS registered 2.0% less total rain than MLH, but the timing of the rain events is clearly similar. On a monthly basis, variations were more significant, with MLBS registering more rain in June and October by 21% and 16%, respectively, while in July, August, and September, MLH rain exceeded MLBS rain by 11%, 14%, and 11%, respectively. These results suggest that, at least for summer and fall rain events, there is significant spatial variation in rainfall for individual events, but that the timing and average quantities are similar. Therefore, using the MLBS rain data may be an acceptable substitute for the period before which MLH data is available. The benefits of analyzing the data over the full 18-month period for which daily lake volume measurements exist far outweigh the risks of introducing some error due to moderate spatial variations of rainfall.

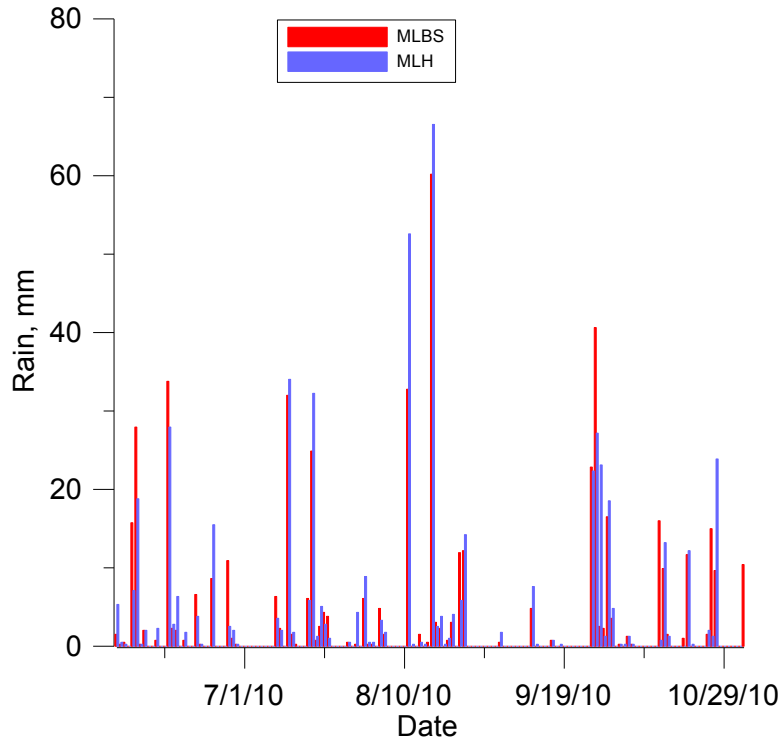


**Figure 11: Comparison of MLBS and MLH rain gauge data
May 2010 to November 2010**

Snowfall was well above average during the winter of 2009-2010 in southwestern Virginia, which heightened the need for a methodology for quantifying effective precipitation from snowfall events. The Mountain Lake area received snow totals of 1092 mm from 6 events between December 2009 and February 2010, according to National Oceanic and Atmospheric Administration winter storm summary maps [NOAA, 2010]. Whereas rain generally evaporates, runs off, or infiltrates immediately after falling, frozen precipitation does not enter into surface or groundwater systems until it melts. Thus the quantity of hydrologic interest for frozen precipitation in a water balance is the melt rate. Snow depth measurements such as those available through NOAA are not sufficient inputs for a water balance because they do not measure snow density or indicate timing of snowmelt. The desired quantity, melt rate, is not normally measured directly in the field, and is a function of weather parameters such as air temperature and solar radiation, as well as properties of the snowpack such as thickness, density, and albedo [ASCE, 1996]. These snowpack characteristics change with time due to partial melting and refreezing, compaction, and additional snowfall. To quantify the melt rate, this study makes use of a remote sensing technique to quantify the depth of potential meltwater and an existing mathematical model to calculate the timing of the melt events.

Snow-water equivalent (SWE), the total depth of potential meltwater contained in the snowpack¹, is measured in this study with a gamma absorption technique. Since radioactive decay of small amounts of uranium in the soil profile releases measurable gamma radiation and water absorbs some frequencies of that radiation, a comparison of gamma radiation readings before and after a snow event can be used to quantify the total volume of water deposited over the course of the event [Carroll, 1988]. In the fall and winter of 2009 and early spring of 2010, the National Operational Hydrologic Remote Sensing Center [NOHRSC] flew aircraft equipped with gamma ray sensors over parts of the eastern U.S., including the area in the southern Appalachians where Mountain Lake is located. Data from a fall flyover before the season's first snow provided a background gamma signature for the area, and subsequent flyovers were used to calculate SWE deposited by winter storm events. Using gamma radiation for snow-water equivalents is thought to be a superior method to field measurements, which typically suffer from sampling problems and do not measure ice lenses, perched water, or ice and water in the upper soil profile, all of which contribute significantly to snowmelt totals [NOHRSC, 2010]. The spatial resolution of the gamma absorption method produces an average SWE over an area that is approximately four times larger than the Mountain Lake watershed. This resolution on the one hand avoids large-scale sampling problems due to drifting but may also ignore true microclimatological and orographic differences between adjacent watersheds. The principal stated source of error for the method is the unknown background soil moisture in the top eight inches of the soil profile in the initial fall flyover. However, for the reasons given above it is thought to be a superior method than a traditional field sampling technique.

Daily snowmelt rates were obtained from a mathematical model [NWS, 2010] that took as inputs SWE derived from gamma absorption and weather data from a regional network. The energy and mass balance is computed using multiple layers, a 1 km² cell size, and hourly time steps. Snow depths, as would be measured by a ruler, are also back-calculated by the same model, and measured snow depths at selected locations and times were compared to modeled snow depths and then used for calibration. Although MLBS was not a calibration site for snow depths, it was included in the network of weather data and as such, snowmelt rates were calculated by the model for that site. Modeled daily snowmelt is displayed together with measured daily rain in Figure 12. The data are stacked, showing total daily inputs to the water balance from the two sources. Note that days of major snowmelt coincide with days of heavy rain, which is typical for snowmelt in the Appalachians [ASCE, 1996]

¹ Note that snow-water equivalent (SWE) has two related but distinct definitions. The definition used in

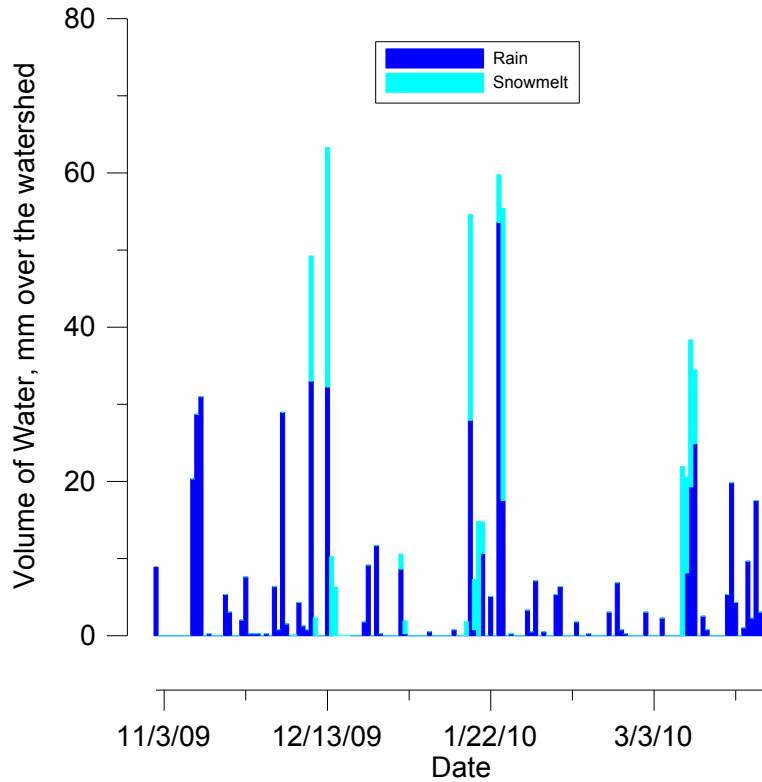


Figure 12: Winter snowmelt and rain
November 2009 to March 2010

In order to assess the accuracy of this method of calculating snowmelt rates, new snowfall totals that were back-calculated from the National Weather Service (NWS) model were compared to published winter storm summary maps [NOAA, 2010], and sources of uncertainty were considered. The comparison showed that for smaller storm events, new snowfall totals were similar. For three larger storms for which data are available, however, NWS totals are 1.2 to 2.5 times larger than NOAA totals. NOAA totals do not, however, include ice accretion, and this may be a significant source of water from larger storms. For example, NOAA documents note that in a storm on February 5-6 2010, isolated higher elevation areas may have accumulated up to 13 mm of ice, but that these values are not included in the storm summary. Since the MLBS weather station is incorporated as a calibration site for NWS and is no longer included in the NOAA network, the NWS values are therefore considered to be the most reliable snowfall values available. It is unknown whether precipitation gauges are double-counting some snowmelt or possibly miss some rainfall at the beginning of a rain-on-snow event resulting from a layer of snow and ice on the top of the collecting bucket. It is clear from Figure 24 that most snowmelt events are temporally well-placed by the NWS model and are coincident with changes in lake storage. However, on two occasions in February 2010 the lake

response indicates that the actual snowmelt occurred two to three days later than the modeled snowmelt. While the calculation of snowmelt rates may introduce some errors into the water balance, the inclusion of a term that accounts for frozen precipitation is likely a significant improvement over not including one at all.

Well Abstraction

Abstraction from groundwater pumping by hotel wells is one possible mode of outflow within the Mountain Lake watershed, and such pumping could theoretically have some impact on lake levels. To evaluate this possibility, pumping records were examined and waterworks outflows were located. Of the wells on the property, only two wells are both within the watershed and actively used for abstraction. The outflow for hotel water is a water treatment system that transports wastewater outside the watershed, although a small percent of extracted water may periodically be recharged to the watershed from the flushing of storage tanks. Daily abstraction ranged from approximately 7,000 to 14,000 gallons per day [DEQ, 2010], or 0.02 to 0.04 mm per day extrapolated over the area of the watershed, depending on the season, with yearly totals of approximately 11 mm. This total yearly abstraction represents less than 0.8% of the yearly precipitation and is well within the uncertainty of rainfall inputs given the variability in the two rain gauge readings used in the study. By way of comparison, rainfall was measured in increments of 0.254 mm, and the maximum magnitude of unexplained daily changes in lake level was similar at 0.23 mm over the watershed, so the amount of groundwater pumped for use at the hotel on a daily basis is an order of magnitude smaller than the daily measurement uncertainty of the precipitation and storage terms. Pumping is a known cumulative loss, however, and values for abstraction of 0.04 mm/day for May through November when the hotel is open and 0.02 mm/day for the off season is included in the water balance.

Direct Lake Evaporation

Lake levels may be affected by direct evaporation of water from the surface of the lake. Since the surface area of the lake changes significantly with lake stage, this value may change over the course of a season as lake levels fluctuate. Direct evaporation may also be affected by seasonal, daily, or hourly changes in weather parameters such as temperature, solar radiation, and wind speed. Two techniques are used to assess the influence of direct evaporation on lake levels: pan evaporation calculations and a weather parameter correlation analysis.

The annual Class A pan evaporation rate of 1092 mm (43 inches) was used for an estimate of lake surface evaporation [ASCE, 1996]. A pan coefficient of 0.7 was multiplied by this rate in an attempt to account for differences in heat transfer mechanisms between an evaporation pan and a deeper water body. This corresponds to a daily average direct evaporation of 2.1 mm per day averaged over the year and over the lake surface. However, if little evaporation takes place in the winter and the total evaporation is evenly distributed through the months of March through October, this would correspond to an average daily evaporation rate of 4.5 mm/day over the surface area of the lake. The pan evaporation rate would only apply to the area of the lake itself. As the lake stage increases, so would the area

from which direct evaporation can occur. Figure 13 shows the approximate daily direct lake evaporation as a function of lake stage. It should be noted that while direct evaporation decreases from the lake as the level decreases, evapotranspiration may continue from the lakebed itself, where significant grass and vegetation still grow, and the total change in evapotranspiration due to lower lake levels may not be significant.

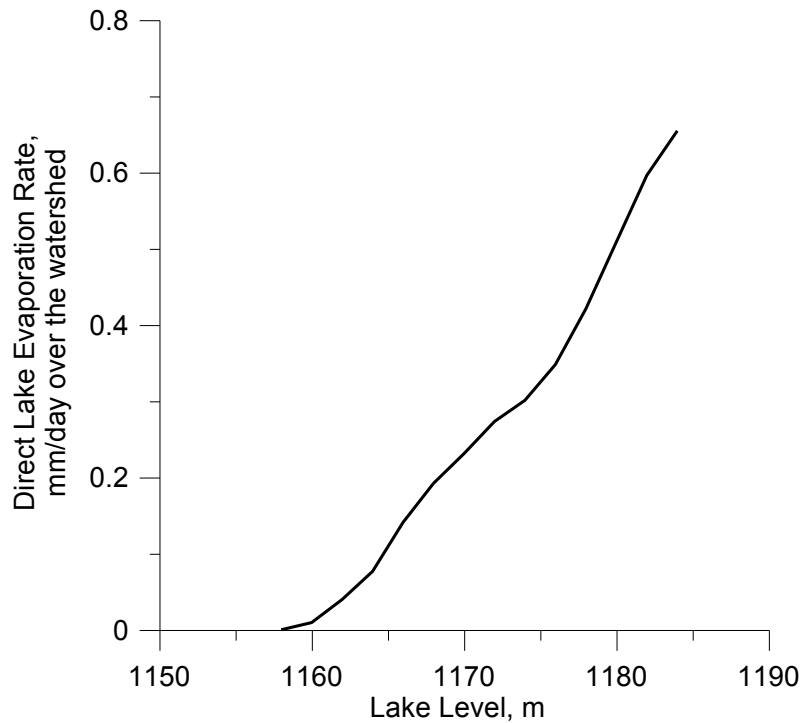


Figure 13: Direct lake evaporation as a function of lake level

A second analysis of direct lake evaporation was performed using observed data to assess possible correlations between evaporation and weather parameters. While the previous evaporation analysis presumes that direct evaporation is constant during the warmer part of the year and evaporation rates do not depend on weather parameters, observed lake levels may indicate otherwise. For example, it is possible that evaporation increases with increasing air temperature, decreasing humidity, and increasing wind speed, and that lake evaporation may accordingly be observed to be higher under these conditions.

To assess this possibility, a correlation analysis was performed. First, the lake level data had to be filtered to remove days where observed changes in lake level may have resulted from recent precipitation. Precipitation may either cause lake levels to rise, or cause the rate at which lake levels fall to decrease. Thus, the first and second derivatives of lake volume were used to filter out days when the changes in storage were influenced by any previous days' precipitation. Using the filtered dataset, any observed changes in lake level that are correlated with certain

changes in weather parameters may indicate a causal relationship. The same procedure was used with an hourly time step to assess hourly evapotranspiration during the day. Data were filtered from the In-Situ dataset from May 2009 to June 2010 retaining only days with both a negative change in storage and a negative second derivative of storage. Days with average temperatures below 10°C were also excluded to avoid days of the year when surface water was likely to be partly frozen and evaporation could be expected to be minimal.

The analysis failed to show any correlations between weather parameters and direct lake evaporation. A correlation matrix of the daily filtered data set is shown in Figure 14. The correlation matrix confirms that decreased solar radiation is correlated with increased humidity, and that warmer days tend to be sunnier, qualitatively confirming the validity of the analysis, but there is no visual evidence for daily temperature, humidity, pressure, wind speed, or solar radiation changes being correlated with accelerated decline of lake level. Likewise, the hourly analysis showed no correlation between mid-day temperature highs and increased hourly declines in lake volume. Therefore, the presumption that direct evaporation from the lake surface is constant over the warm part of the year is probably an adequate one.

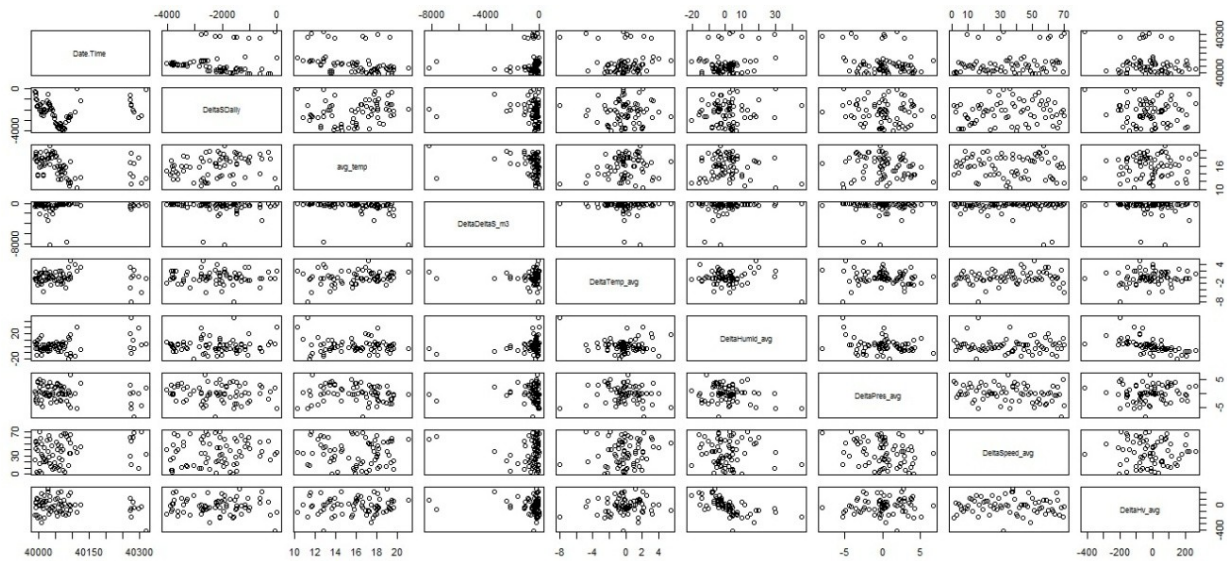


Figure 14: Correlation matrix for change in storage and weather parameters

Net Groundwater Outflow

Lake levels at Mountain Lake are presumed to be heavily influenced by water flowing from the lake and out of the watershed, possibly through the crevice features at the deep end of the lake. The data from the previous water balance [Jansons, 2004] suggested that the net rate of this groundwater outflow may vary by to a factor of up to 40 when calculated on a monthly basis. To assess this possibility, in this study the net groundwater outflow (defined as groundwater outflows from the lake minus any groundwater inflows, neither of which can be

directly measured) is determined by separation of precipitation-related quickflow from baseflow in the daily lake hydrograph. This baseflow is interpreted as the net groundwater outflow.

To evaluate baseflow, a hydrograph separation was performed on data from the period between July and November 2010. Figure 15 shows that the lake volume appears to decline steadily throughout the recording period whenever rainfall is minimal. Figure 16 presents the same information in the form of the lake hydrograph, and shows that when rainfall and subsequent runoff are minimal, the daily change in lake volume is nearly constant from July through November 2010. This constant negative change in volume is interpreted to be the baseflow. The baseflow reaches a minimum value of 46.7 L/s or 3.02 mm/day in depth over the watershed.

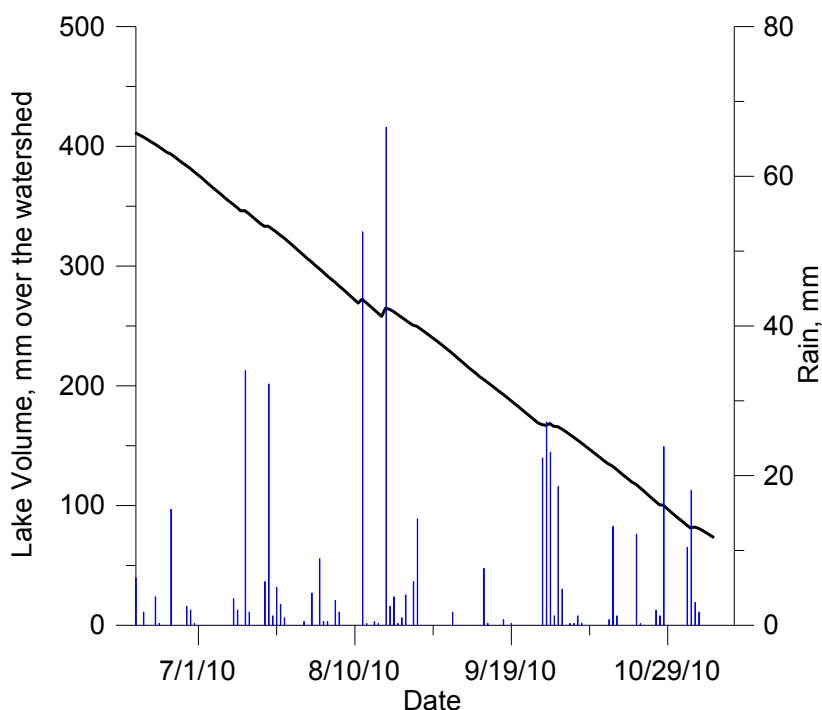


Figure 15: Lake volume and rainfall
July to November 2010

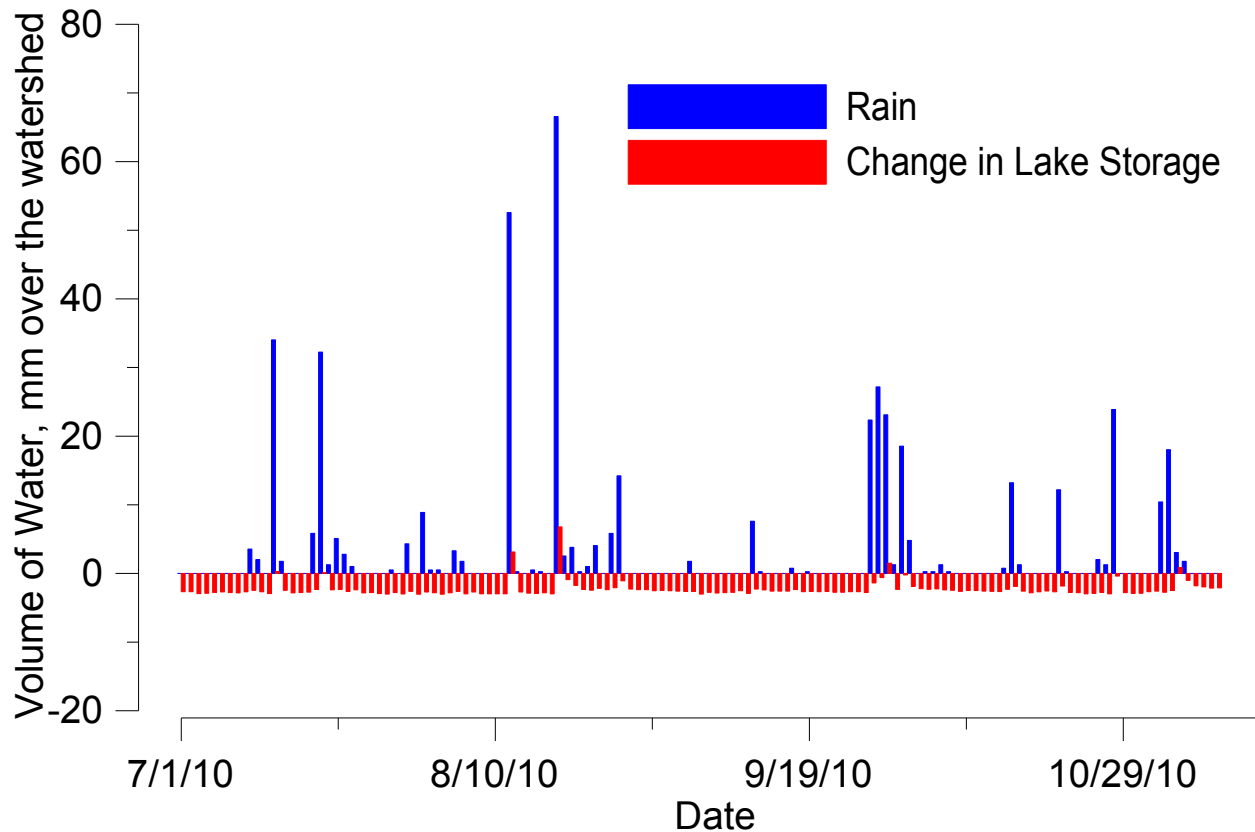


**Figure 16: Daily change in lake storage and daily rainfall
July to November 2010**

Small variations in this baseflow that are not attributable to antecedent rainfall do occur and limit the precision of the baseflow measurement. Hourly lake level measurements show variations of up to 5 mm that are not correlated with the overall trend of the daily water levels. This volume of water corresponds to an uncertainty of 0.23 mm over the watershed for a lake covering 30% of the lakebed area. If the lake were full, this would correspond to a maximum uncertainty of 0.72 mm over the watershed. Therefore, if the same amplitude of background noise in the pressure transducer measurements were seen when the lake is full this may limit the precision of a water balance to a larger extent than it does when the lake is low. At lake levels encountered during this study, the uncertainty in the baseflow also happens to be similar to the detection limits of the rain gauges (0.254 mm). The 5 mm variation in lake level measurements cannot be attributed to the 0.6 mm precision of the pressure transducers, but may be the result of small waves on the surface of the lake and possibly barometric pressure differences between the lake and barometric pressure transducers.

Before equating this loss of lake water seen in the near-constant negative baseflow to net groundwater outflow, other system outflows need to be considered. These include direct lake

evaporation and abstraction from pumping; there was no surface outflow from the lake during the period of study. It was previously estimated that the evaporation from the lake surface for the lake levels encountered over the period of study could be expected to account for 0.1 to 0.3 mm/day over the watershed, or approximately 3-10% of the daily change in storage. It is not known if productive wells on the property are hydraulically connected to the lake and thus if abstraction from wells indirectly affects lake levels. If a connection exists, the highest abstraction rate would occur during the summer months when hotel occupancy, and subsequent water usage, is highest. Daily pumping was calculated to account for a maximum of only 0.04 mm over the watershed and thus would not be considered significant on a daily basis. However, if this pumping is not constant but performed over a typical period of 4 hours, during this time 0.24 mm over the watershed may be removed from the system. This could be a possible explanation for some of the noise seen in the hourly baseflow of the lake, but could only account for declines in lake level and not subsequent increases.

Based on these considerations, a value of 3.0 mm/day over the watershed was chosen as representative of the maximum rate of baseflow. During this period of study, 93-97% of this baseflow is accounted for by net groundwater outflow from the lake; the remaining 3-7% is accounted for by direct lake evaporation. The uncertainty associated with this outflow rate is approximately 0.23 mm/day or about 8% of the maximum baseflow rate. Thus, responding to the first hypothesis, it is shown that over a period of four months, daily net lake outflow did not vary by more than a factor of 1.08 (or 8%). This corresponds to approximately 3804 m³/day or 44 L/s. This value is in close agreement with that calculated by the prior water balance of 43 L/s [Jansons, 2004] as an annual average, but shows that outflow is constant on a daily basis.

Actual Evapotranspiration

Evapotranspiration may influence lake levels at Mountain Lake by controlling the amount of runoff received by the lake from precipitation in the watershed. Numerous methods for evaluating evapotranspiration (ET) on different scales in a variety of environments have been developed. Most of these methods evaluate potential evapotranspiration, which expresses the atmospheric demand for water vapor, regardless of the presence of sufficient water to fill that demand. The quantity required for the water balance, however, is the actual evapotranspiration (AET), which is the total amount of water evaporated from soil and vegetative surfaces and transpired through vegetation. AET is always lower than or equal to PET, and depends on the availability of water in the system. In this study, the necessity of improving ET calculations was established. PET was calculated using seven different methods and the results were compared and evaluated. However, only one method was chosen for use in the water balance. Next, a subset of the study period was used to calibrate the calculated PET to a representative AET, and this transformation was applied to the entire study period. Finally, results were compared to a regional statistical model to assess the accuracy of the AET values.

In the prior water balance [Jansons, 2004], evapotranspiration calculations appear to have contained three errors that may have caused some of the calculated variability in the net groundwater outflow rate. First, the evapotranspiration component was an estimate of potential (PET) rather than actual (AET) evapotranspiration. Second, calculations appeared to use monthly average values for inputs, but the Penman-Monteith method that was used is most appropriate for daily or hourly calculations [ASCE, 1996]. Third, the current study shows that the radiation sensor used in the prior study was not appropriate for evapotranspiration calculations, and conversion techniques were not found to produce adequate substitutes. This study aimed to improve upon evapotranspiration calculations by addressing each of these problems.

In this study, six methods were initially used to calculate PET. The Penman-Monteith method was chosen because it is the agricultural industry standard. The other five methods were chosen because they had been previously evaluated in a study of PET in forested watersheds in the southeastern U.S. [Lu *et al.*, 2005]. The reader is referred to Crago and Brutsaert [1992] for background on the combined mass and energy balance equation that forms the theoretical basis for several of these methods, to [ASCE, 1996] for details of the combined mass and energy balance equations, and to Lu [2005] for a summary of other equations. Results from each method calculated with data from May to November 2010 are shown in Figure 17. A summary of each applied method is presented as follows:

- The Penman-Monteith method [Allen, 1998], the current standard method of the agricultural industry, represents a mass and energy balance method that attempts to account for all the physical mechanisms of PET. It calculates PET for a reference crop and requires estimation of several parameters by the user. In this and other methods, alfalfa was chosen as the reference crop because it has a leaf area index sufficiently high for direct penetration of radiation to soil to be minimal and therefore mimics forest under full canopy. The method was designed for inputs at hourly time increments to reduce problems from phase differences in diurnal cycles of vapor pressure deficit, wind speed, radiation, and air temperature. In this study, the method was applied twice using the software Ref-ET [Allen, 1998] once using hourly time increments, which were then averaged over the day, and once using daily time increments.
- The Makkink method [Makkink, 1957] represents a version of the Penman-Monteith equation that is simplified on the basis of field calibrations in agricultural settings. Its only inputs are solar radiation and mean daily temperatures.
- The Priestley-Taylor method [Priestley, 1972] represents a simplification of the combined mass and energy balance method that assumes losses due to advection are minimal. It requires only radiation data as input, and was calibrated with 10-

day intervals. It is designed for estimating ET in large land areas following region-wide rains and was calibrated in locations in Australia and the U.S.

- The Hargreaves method [*Hargreaves, 1985*] represents a temperature-only method designed to use the daily temperature range as a proxy for radiation input. It was calibrated to a grass reference surface in California with 10-day time intervals, although for purposes of comparison here the daily results are displayed.
- The Hamon method [*Hamon, 1963*] represents a radiation and temperature-dependent model that requires as input only daily duration of direct solar radiation, temperature and latitude. Hamon's method was developed specifically for an experimental forested watershed at Hubbard Brook, NH, and later calibrated at a Coweeta, NC and resulted in an empirical coefficient of 1.0 at Hubbard Brook and 1.2 at Coweeta [*Federer and Lash, 1978*]. Daily solar duration was calculated in ArcGIS® on a weekly basis taking into account calculated topographic shadowing in the basin.

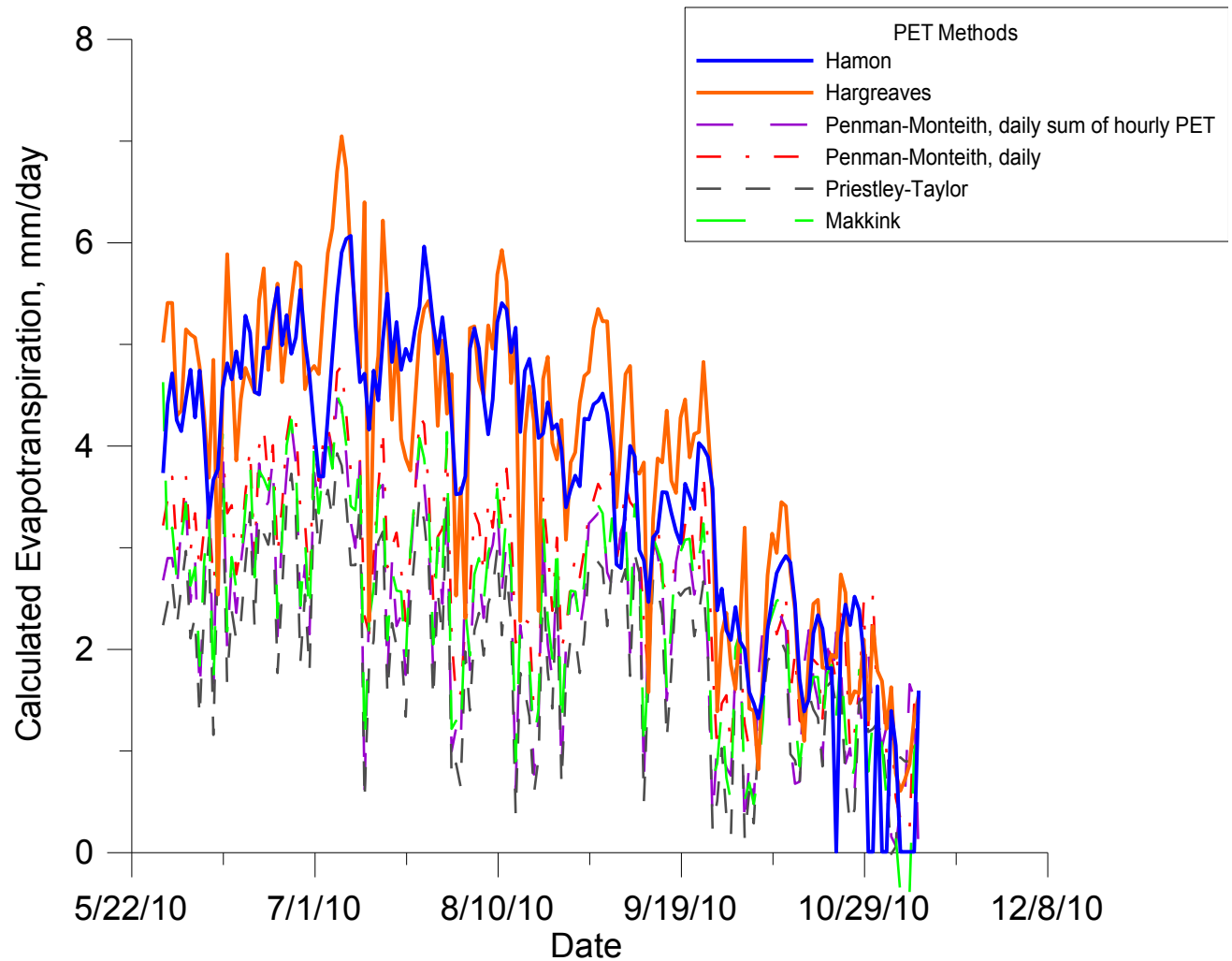


Figure 17: Comparison of potential evapotranspiration methods

May to November 2010; dotted lines represent radiation-based methods

The results in Figure 17 show that most of the methods that take solar radiation as an input resulted in distinctly lower calculated PET values than the non-radiation methods. Therefore, the radiation data required further evaluation. The radiation sensor at MLBS is a photosynthetically active radiation (PAR) sensor designed to measure the amount of light from 400 to 700 nm that is useful to plants specifically for photosynthesis. The preferred type of radiation sensor for evapotranspiration calculations, however, is a pyranometer which measures total energy from radiation across a wider part of the electromagnetic spectrum from 300 to 1100 nm. Since a day may be cloudy and register very low PAR but transmit significant heat energy from the microwave portion of the spectrum, PAR sensors and pyranometers do not measure equivalent units, and no formal conversion between the two can be defined. A PAR sensor returns a photon count per unit area per unit time, whereas a pyranometer returns

units of power per unit area. If the incoming spectral distribution of the photons were known, the energy carried by each photon can be converted to units of energy by Plank's law, $E=hc/\lambda$ where h is Planck's constant, c is the speed of light, and λ is the wavelength. An estimated average wavelength of 552 nm has been suggested for making this conversion [McCree, 1972]. This leads to an approximate value for the ratio, K , of the PAR-derived irradiance to the integrated irradiance across the wider spectrum [Ross, 2000]. The commonly used K for global radiation is 0.434, but because the true ratio changes for direct, diffuse, and reflected components as a function of cloud cover, the use of any single K represents an oversimplification. Though no reason exists to question the functioning of the MLBS PAR sensor itself, the use of the MLBS radiation data may then be inappropriate for PET calculations because the true transformation function between the PAR and wide-spectrum radiation of interest is unknown.

The radiation data used in a PET calculation can be validated by comparing the measured PAR radiation data to theoretical clear-sky radiation values. If the converted PAR radiation data are correct, they should periodically reach (within a few percent), but not exceed, the theoretical clear-sky irradiance envelope. Figure 18 shows measured and calculated irradiance over a representative two-week period. Nights and periods of cloudiness were indeed registered and did fit to some similar envelope-like maximum that itself varied seasonally. However, the measured radiation consistently fell far short of the clear-sky irradiance envelope calculated based on date, latitude, and the most conservative values for atmospheric scattering effects. Therefore, available PAR radiation data underestimates actual wide-spectrum radiation and therefore underestimates PET.

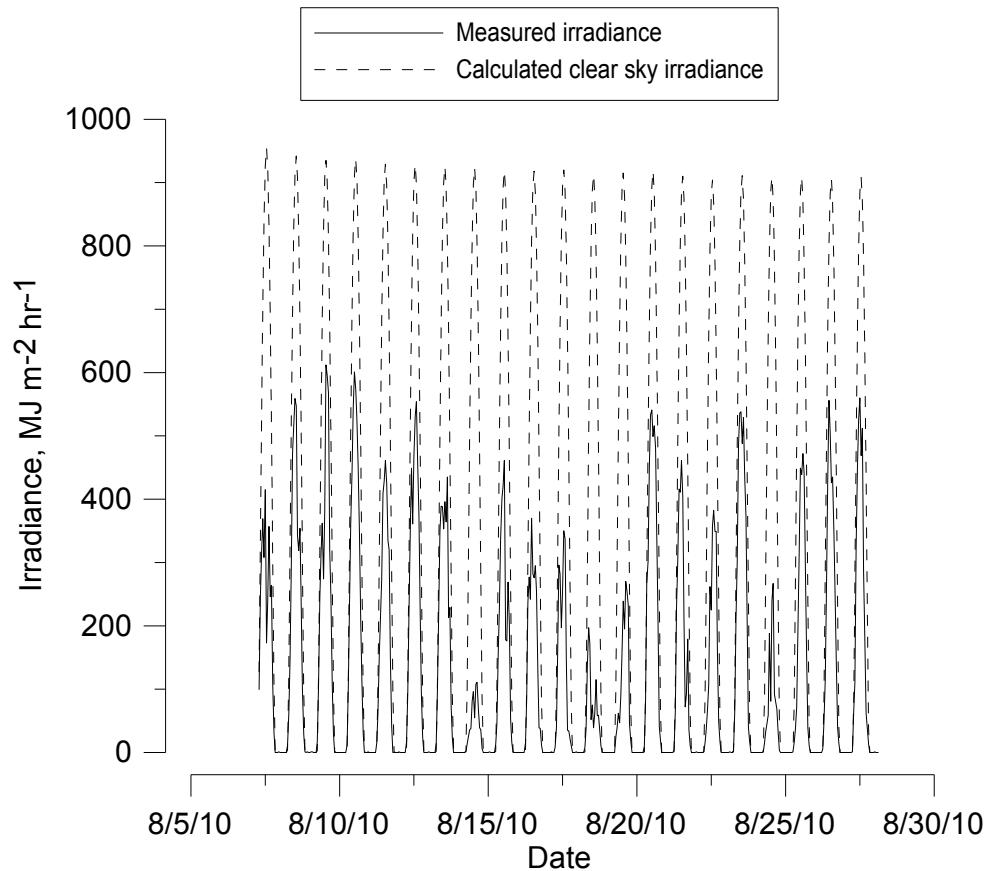


**Figure 18: Integrity assessment of radiation data
using representative period in August 2010**

Next, a method was chosen for calculating PET. The method should not require radiation data for reasons stated above, leaving only the Hargreaves and the Hamon methods for consideration. The Hargreaves method was developed for a dry climate in California and is found to be valid in northern Spain [Martínez-Cob and Tejero-Juste, 2004], but was found to overestimate PET in the more humid southeastern U.S. [Lu *et al.*, 2005]. The Hamon method was developed specifically for forested catchments in the eastern U.S., and therefore was selected for use in this study.

As a final step for obtaining an estimate for evapotranspiration, actual evapotranspiration (AET) was calculated to be a fraction of potential evapotranspiration (PET) based on a period of time over which the total change in watershed storage is known. Two days, one in the fall of 2009 and one in the fall of 2010, were chosen to frame a period during which the change in soil and groundwater storage could be approximated as zero. On October 19, 2009 the lake level was 1169.32 m. The day with that same level during the fall of 2010 occurred on September 5. The lakes were in periods of similar seasonal decline, little or no rain had fallen in the previous 10 days, and the lake levels were equal at some point during the day. Given these similar seasonal and antecedent moisture conditions it may be reasonable to assume that the shapes of

the local water tables and/or piezometric surfaces were also the same on these days, and that the total change in soil and groundwater storage could then be approximated as zero over this time. The total water balance for that 321-day period was then solved for AET. The fraction AET/PET was then multiplied by the daily PET calculated from Hamon's method to obtain estimates for daily AET values for the entire period of study. Results of both AET and direct lake evaporation are displayed in Figure 19. This method of estimating AET does not directly take into account availability of water for evapotranspiration, and in times of drought it may be expected that AET would decrease as water availability decreases. However, since forests under drought conditions are typically warmer, drier, and experience more radiation than under non-drought conditions, a higher atmospheric demand can also be expected and some researchers have suggested that forest evapotranspiration tends to be conservative under a variety of conditions [Roberts, 1983] and that AET decreases only slightly even during periods of drought [Oishi et al., 2010].

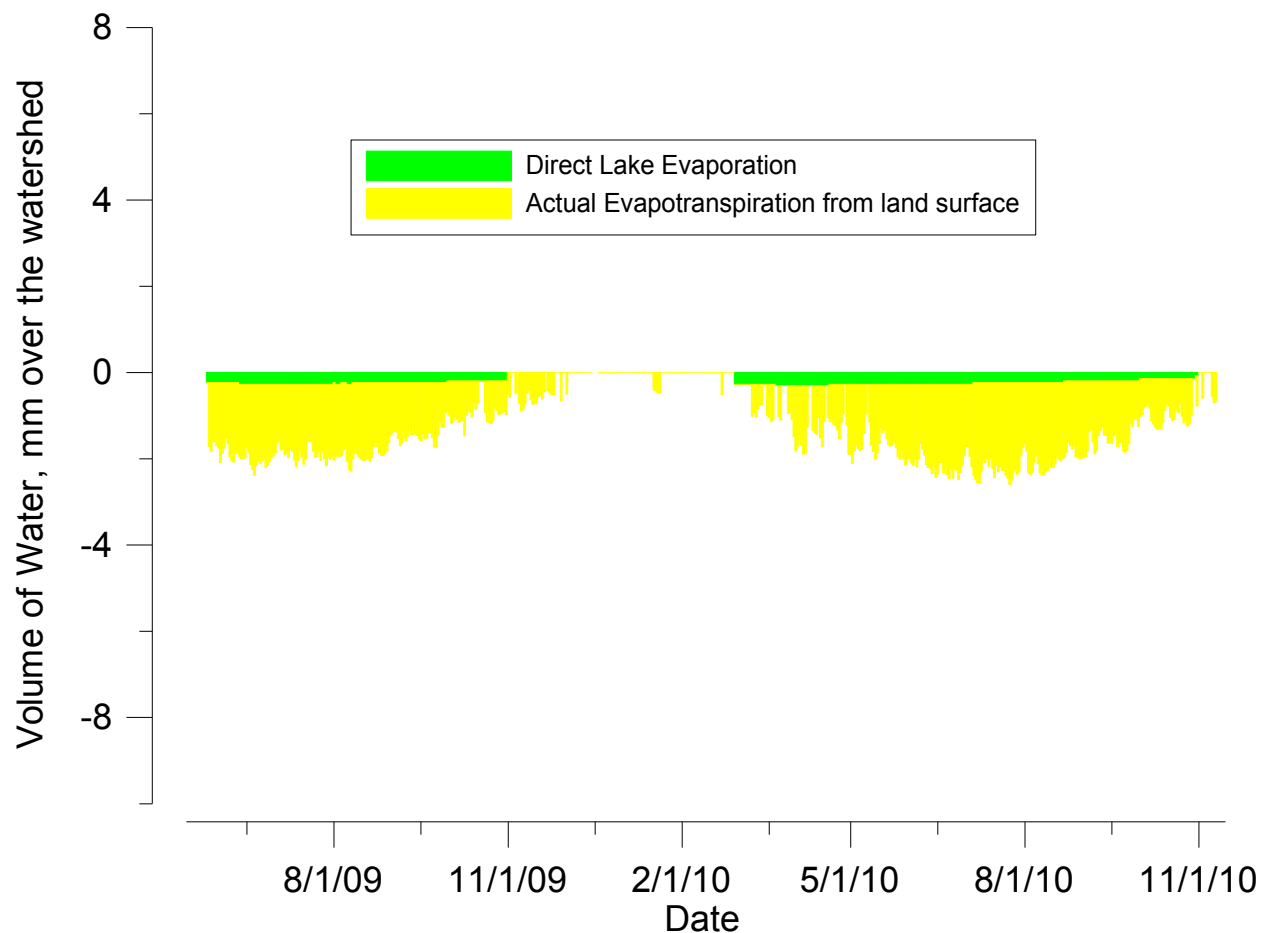


**Figure 19: Actual evapotranspiration and direct lake evaporation
May 2009 to November 2010**

A proxy for AET is also calculated according to a regional statistical regression model [Lu *et al.*, 2003] that is developed with data from 39 forested watersheds in the southeastern U.S. In lieu of AET, Lu calculated the difference between annual precipitation and annual runoff. The difference was averaged over multiple years to minimize effects of changes in storage over each year. The regression model associates higher total precipitation, lower elevations, and lower latitudes with an increased AET, with an additional small correction for the percent of forest cover in the watershed. The developed regression model is expressed as

$$AET = 1098.786 + (0.309)P_t - 0.289(Elev) - 21.840(Lat) + 1.96(F_o) \quad (3)$$

where *AET* is the actual evapotranspiration as represented by precipitation minus runoff in mm/yr, P_t is annual mean watershed precipitation in mm/yr, *Elev* is the land elevation in meters, *Lat* is the latitude in degrees, and F_o is the percent of the watershed that is forested. The AET value calculated by this statistical model differs from the results of all the other methods used in this study in that it represents an annual estimate rather than a daily estimate, and is based on the assumption that AET is equal to precipitation minus runoff.

Evapotranspiration values produced from the Lu model were compared to values obtained from the water balance used here. Results of the Lu statistical model indicate that values of AET are significantly higher than those calculated from the Mountain Lake water balance, even if direct lake evaporation and known changes in storage are included in evaporation totals. Results are summarized in Table 2 using an elevation of 1190 m amsl, a latitude of 37.36°, and a forested area of 80.4% (not including the lakebed). It is worth noting that Mountain Lake lies near the northwestern geographic boundaries of the dataset used to develop the regression model. Its mid-range watershed elevation is more than 200 m above the highest elevation site included in the regression model, and 950 m above the average elevation in the study. A shortened growing season at higher elevations is one possible explanation for the smaller yearly evaporation totals calculated in the water balance. Extrapolation of the Lu regression model to sites such as Mountain Lake which are above the elevations used in model development may be inappropriate.

Method	P _t (mm/yr)	ET (or proxy) (mm/yr)	Comments all data from Water Year 2009, Oct 2009 – Sept 2010
Lu model	1427	536	Snowmelt data included
	1197	465	Snowmelt data excluded
	1197	432	Snowmelt data excluded, adjusted for changes in storage calculated by water balance
Water balance		337	AET from land only
		397	land AET and direct lake evaporation

Table 2: Comparison of Lu [2003] statistical AET model and lake water balance

Change in Soil and Groundwater Storage

The storage capacity of the soil and groundwater system surrounding the lake may influence lake levels because storage of water in the geologic media influences the partitioning of precipitation into evapotranspiration and runoff to the lake. As a final step in the water balance, the mass conservation equation was solved on a daily basis to obtain the change in soil and groundwater storage.

A distinct seasonal pattern is observed in the storage term, which is influenced largely by evapotranspiration. In Figure 20, days of positive storage changes correspond to the transfer of rain and snowmelt into the ground. Days showing negative storage changes indicate days when the soil and groundwater system is losing water, either to the atmosphere via evapotranspiration (largely a summertime phenomenon), or to the lake via direct or indirect runoff (largely a wintertime phenomenon). Figure 21 shows the typical winter and summer responses of lake levels to these different soil and groundwater storage behaviors.

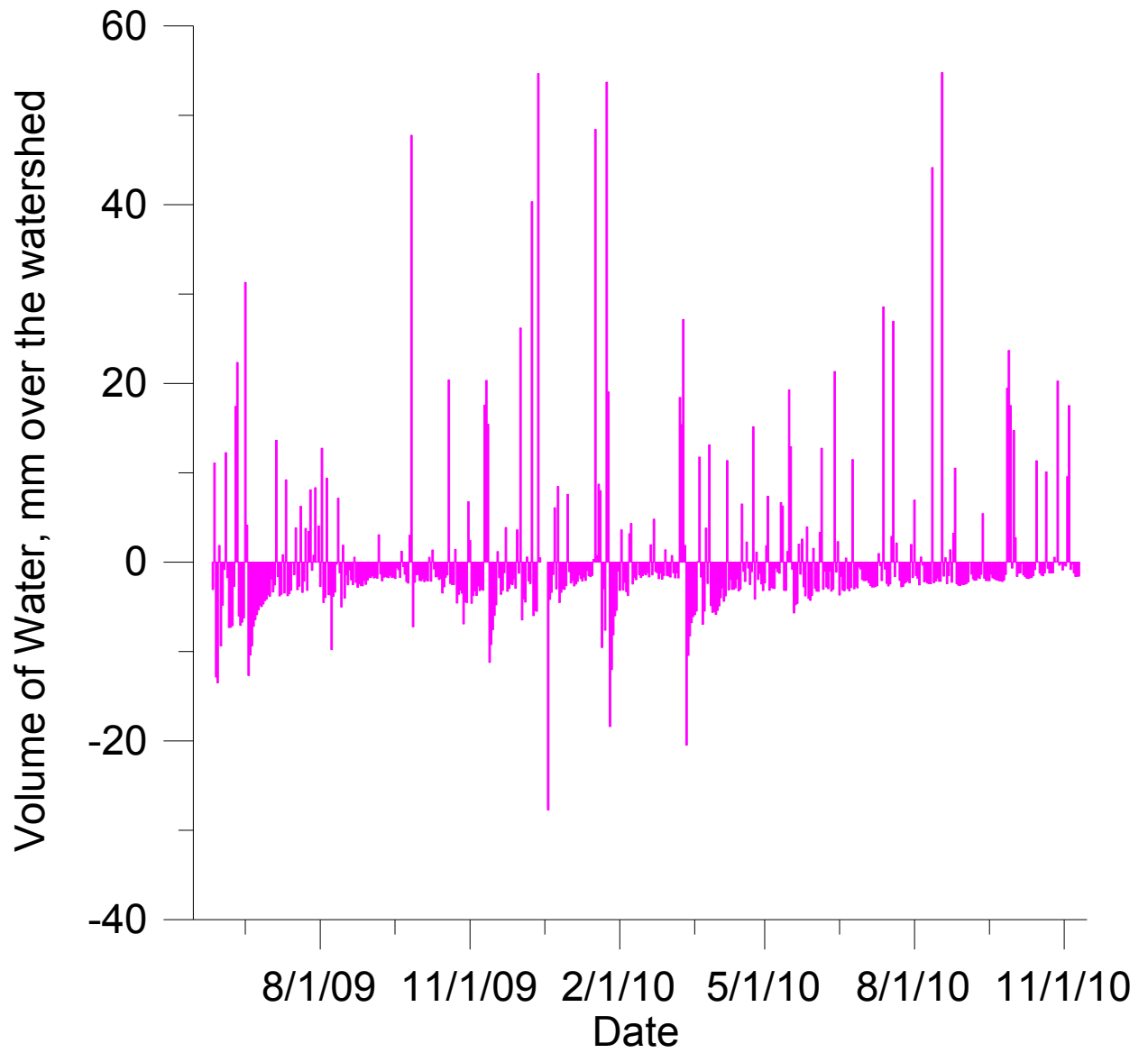


Figure 20: Daily change in soil and groundwater storage
May 2009 to November 2010

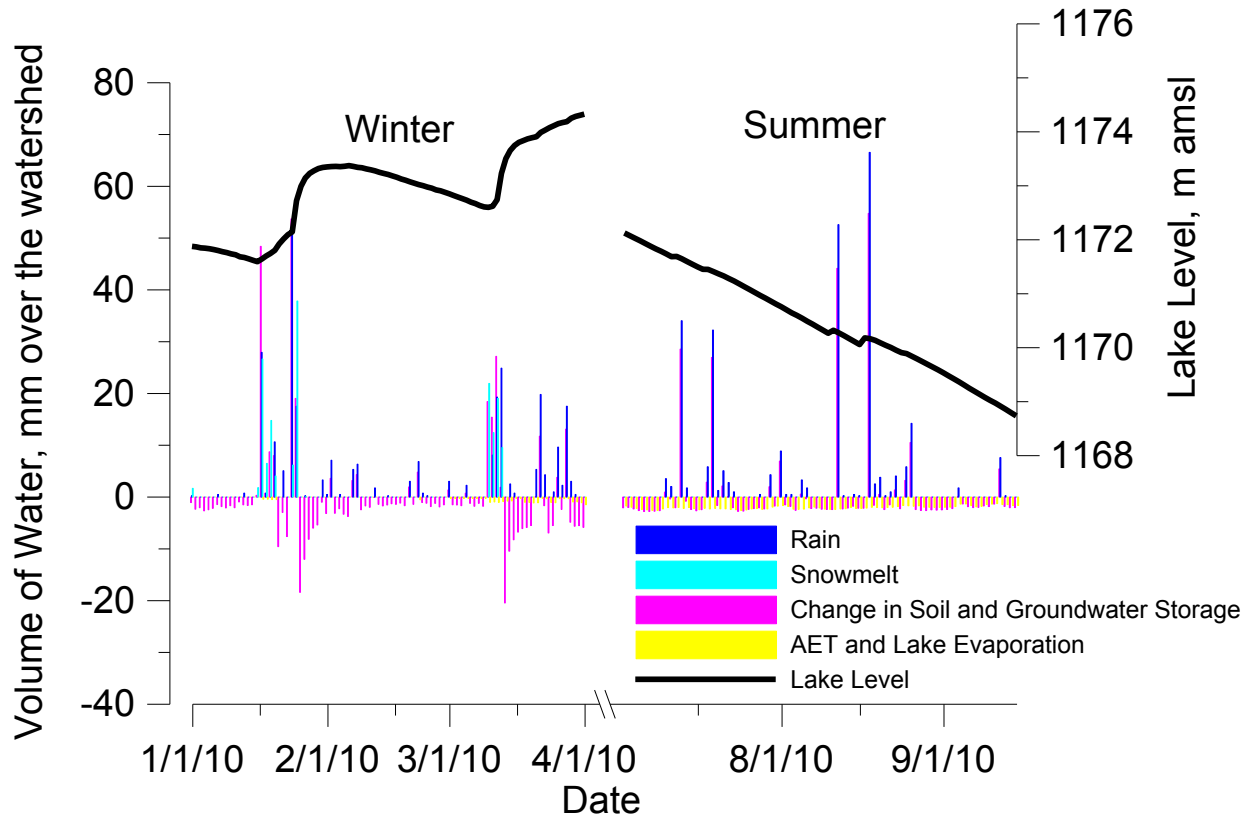


Figure 21: Seasonal differences in lake level response

The soil and groundwater storage term is solved in this study as the unknown term in the water balance equation (equation 2), and therefore includes all errors. However, the daily storage term for which the current water balance is solved does seem to correspond to expected behavior, suggesting that serious or systematic errors contained in the storage term are likely minimal. For example, the Lily-Bailegap soil complex that is found in most of the watershed has an available water capacity of about 110 mm (4.3 inches) [NRCS, 2010], which appears to be sufficient to store the quantities of water arriving in large precipitation events before evapotranspiration occurs in the summer.

Although water levels were monitored in all accessible wells, estimations of groundwater storage were not calculated from these measurements. The well hydrograph (Figure 22) from the Horse Barn 1 well (HB in Figure 47) indicates rapid response to precipitation in a perched water table approximately 30 m higher than the level of full lake. A second well hydrograph from White Pine Well 1 km north of the lake along Pond Drain (Figure 23) indicates a deeper aquifer with response to precipitation within hours of an event but also displays a distinct tidal signature, suggesting an unconfined aquifer with very low storage. This finding is consistent with regional studies that found very low primary porosity in these

formations [Bird, 1981] Most wells on the property were either outside of the watershed, influenced by pumping, or not accessible. Accessible wells were self-selected to be low producing wells, accessible only because they had been abandoned. Even if all wells on the property had been accessible, they could not be used to estimate aquifer storage because of lack of the necessary data on secondary porosity and on the areal extent of any identified response in a given part of the system.

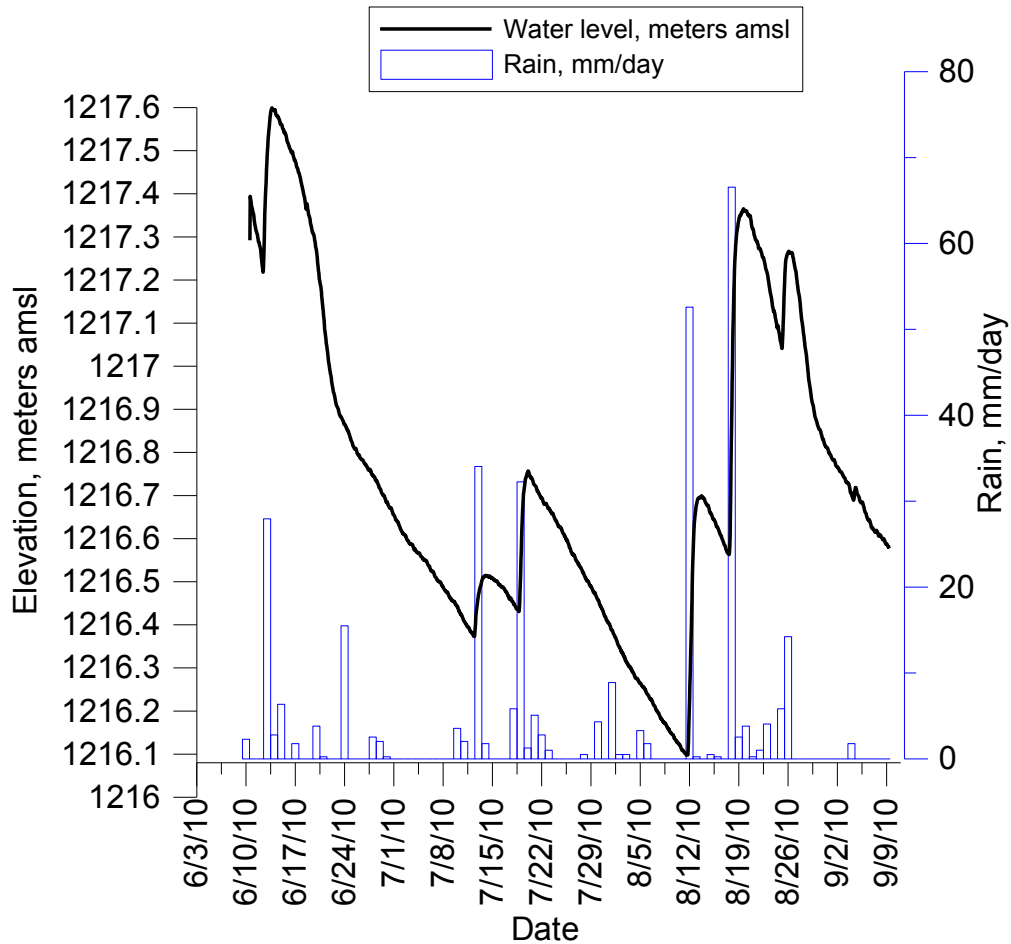


Figure 22: Well hydrograph with precipitation, Horse Barn 1 Well

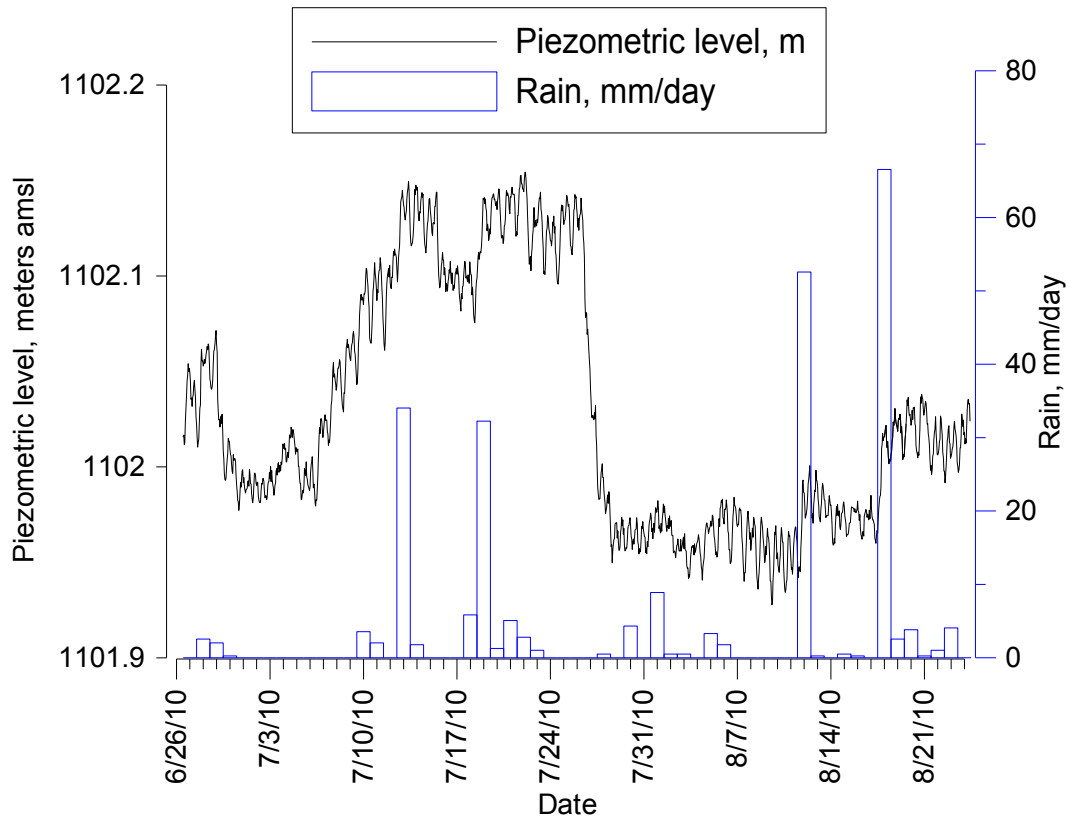


Figure 23: Well hydrograph with precipitation, White Pine Well

Figure 24 summarizes the results of the water balance and shows clear recharge periods during the season from November to April when vegetation is dormant, with particularly significant recharge episodes following rain-on-snow events. A rainy midsummer in 2009 was able to stabilize lake levels for a short period of time, but near-constant declines in lake volume occurred during late summer and early fall of 2009 and all of summer of 2010 despite a small number of very heavy rain events. The water balance suggests that lake levels respond quickly to precipitation but the magnitude and duration of the precipitation response varies seasonally. In order for the lake to fill, precipitation must occur in sufficient quantity outside of the growing season to overcome the steady leakage rate and make up for what may have been lost at the height of the growing season.

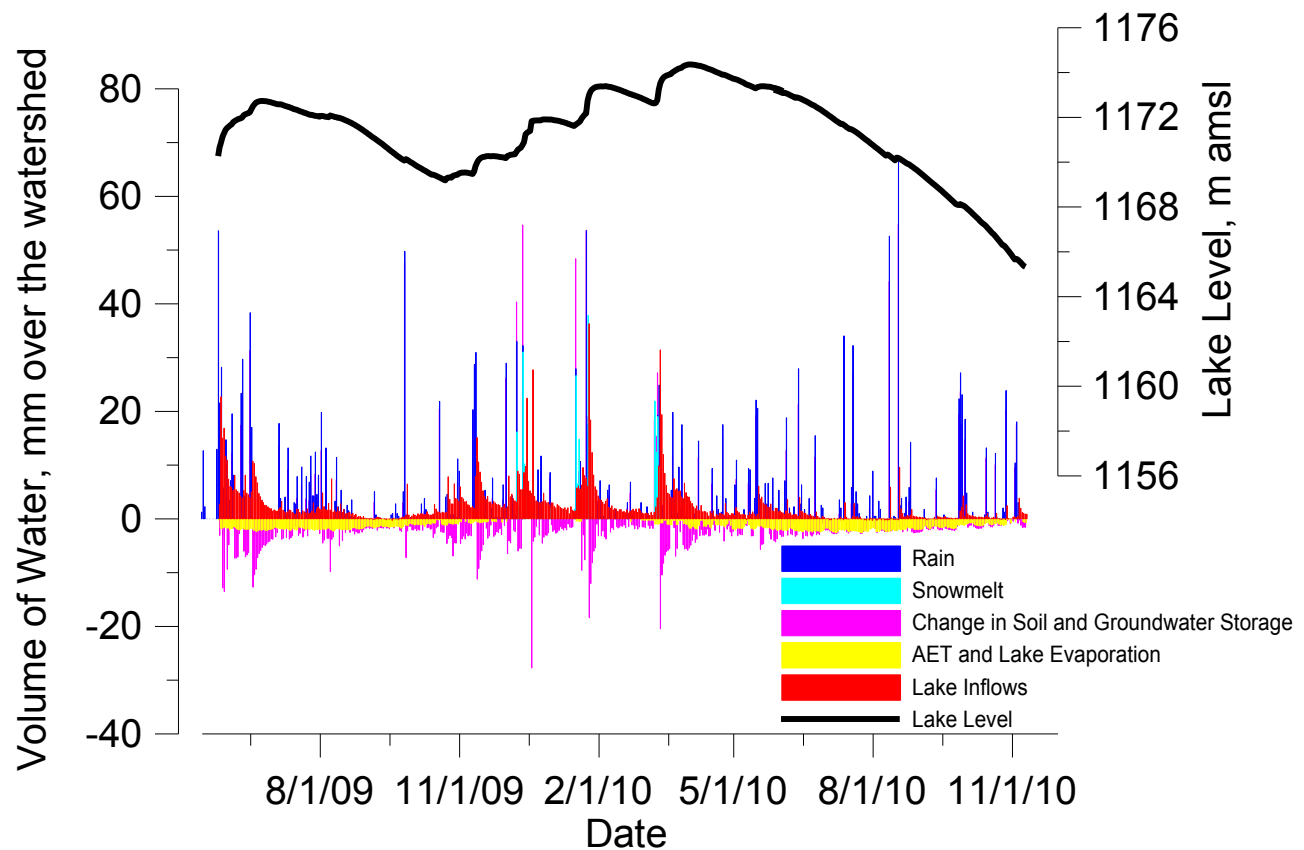


Figure 24: Mountain Lake water balance

May 2009 to November 2010

Models for Lake Level Response

In an attempt to use lake response observed during this study to both explain past lake levels and predict what might be required for the lake to fill, two models are defined. The first is a simple annual summary of the water balance, and the second is based on precipitation-driven recession curves.

Annual Summary Model

Rough approximations of the time it might take to drain or fill the lake can be made based on results from this study together with estimates of annual average precipitation rates at the lake. However, these approximations are based on regional long-term average precipitation values and presume that the assumptions made about the behavior of the lake from May 2009 to November 2010 apply equally well to other periods of study. This calculation uses only values of rainfall, snowmelt, actual evapotranspiration, and groundwater outflows from the lake as measured over water year 2009 (October 1 2009 to September 30 2010). It presumes that direct lake evaporation was included in the term for change in lake storage. It also presumes

that over a long period of time, the change in soil and groundwater storage is minimal, which may be reasonable as this term varies widely on the scale of days and weeks. Most importantly, it assumes that the net groundwater outflows from the lake do not vary over time. If the distribution of precipitation in water year 2009 were typical, the rainfall required to maintain the lake at a steady level with no change in soil and groundwater storage based on these assumptions is 1459 mm/year.

Average values for rainfall at Mountain Lake had to be estimated in order to make projections of what type of departure from average rainfall would be required to empty or fill the lake. Although precipitation data from MLBS were available for 1971 to the present, changes in the methods of measuring snowfall as well as gaps in the published data prevent this record from being used as a reliable source of long term averages. Data from 1971 to 1989 which did not have significant gaps and included separate snow depth measurements was selected for calculation of an average value at Mountain Lake. Snow depth was converted to snow water equivalent with a 10:1 ratio. A second source of climate data was used [NCDC, 2011] to extend the record of annual averages over a period ranging from 1900 to 2010, based on Mountain Lake's location with respect to Virginia's six climate divisions. Mountain Lake lies within division 06 but near the border with division 05, so values of annual precipitation from both divisions were averaged. The average annual precipitation from 1971 to 1989 was 356 mm less than the reported values from Mountain Lake, which include snowfall. A long-term proxy for annual values of precipitation at Mountain Lake was obtained by simply adding 356 mm to the averaged NCDC value. Over the period of record from 1900 to 2010, the average annual precipitation was then calculated to be 1409 mm/year.

An attempt was made to draw some conclusions about the behavior of the lake based on these NCDC precipitation values, but these calculations predict lake levels to decrease more than observations confirm, and it appears that either these precipitation values or some presumptions used in the formulation of the water balance cannot be applied over longer times without some modification. If the lake requires 1459 mm/year of precipitation to remain at a constant lake stage, then for the period of record it has only met this requirement 32% of the time. Hence, the lake should not have been as full during this century as it has been. Table 3 shows some implications of these assumptions. By comparison, the lake drained to empty within four years leading up to September 2008, but the annual precipitation, by the same modified NCDC measurements, was no more than 10% below average. It is therefore possible both that this average precipitation value derived from the NCDC dataset is an underestimation of true annual precipitation, or that some behavior of the system has changed over time. Possible conditions that may have changed over time are groundwater inflows via springs in the lakebed and groundwater outflows via blockage and/or changes in sedimentation. Past unmeasured snowmelt could also explain this behavior, as could overestimation of historical lake levels. In addition, small errors in the USGS topographic map around the lake would have

been incorporated into lake volume calculations for the top several meters of the lake, which could overstate the full volume of the lake.

The lake would drain from full to empty if precipitation were:		for a period of:	
	average		28 years
	10% below average		7 years
	20% below average		4 years
	30% below average		3 years
	absent		1.3 years
Similarly, the lake would fill from empty if precipitation were:		for a period of :	
	10% above average		15 years
	20% above average		6 years
	30% above average		4 years

Table 3: Projections of lake behavior

based on NCDC-derived average annual precipitation values and lake behavior as observed May 2009 to November 2010

Recession Curve Model

Unit hydrographs based on lake response to precipitation were developed to model lake behavior as a function of precipitation and month of the year. A constant net lake outflow of 3.0 mm/day over the watershed is subtracted out of the daily change in lake storage, so that the values analyzed are positive and represent lake inflows only. These inflows are taken to include all overland flow, interflow, stream flow, spring and groundwater flow, and direct recharge to the lake.

As hydrographs typically recede logarithmically after a rain event, a recession curve may be used to model hydrograph behavior. A recession equation [ASCE, 1996] may be given by

$$Q(t) = Q_0 K^t$$

where

$Q(t)$ is discharge through time [m³/day]

Q_0 is the measured discharge at the beginning of logarithmic behavior [m³/day]

t is elapsed time in days [dimensionless], and

K is a recession constant characteristic of the watershed [dimensionless]

In a typical hydrograph this occurs in the time between the inflection point on the recession limb of the hydrograph and the return of base flow when the obvious effects of the precipitation event have subsided. The equation presumes an ideal unit hydrograph where one rain event, uniform temporally and spatially over the watershed, is isolated from the effects of previous and subsequent precipitation events.

To obtain recession constants, a nonlinear least-squares estimate was applied using the open-source code R for different rain events. The events were chosen based on having a multiple-day recession with logarithmic behavior. Recession constants K (Figure 25) were found to vary seasonally. The rain-on-snow events have K values of 0.59 and 0.65, respectively. Isolated late summer and early fall events have low K values between 0.05 and 0.21.

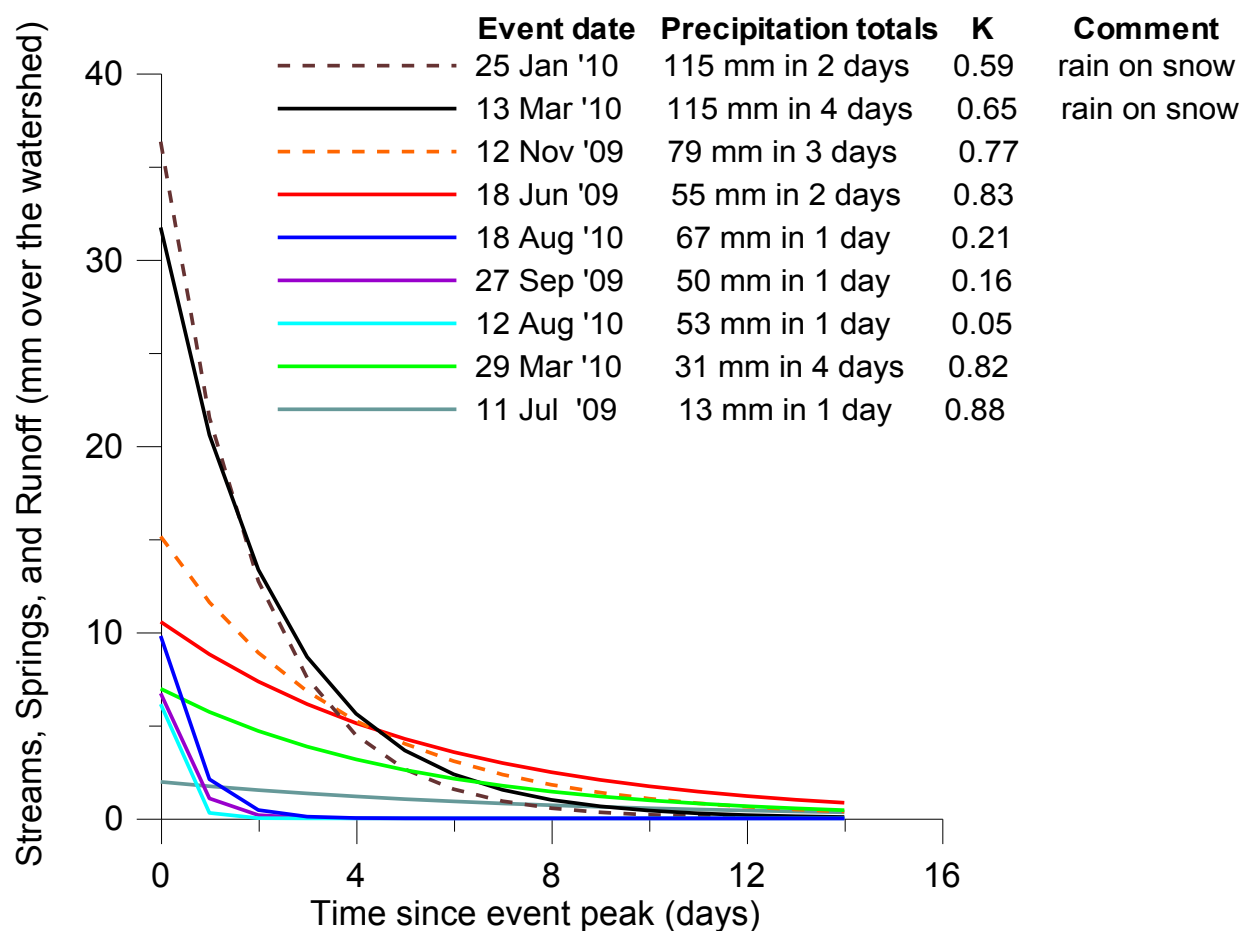


Figure 25: Recession curves for individual precipitation events

These recession curves developed for isolated events were then used to reconstruct the cumulative effects of precipitation events on lake level. The effects of past days' rainfall were summed for each day, and the resulting modeled inflows were compared to the measured inflows from the lake hydrograph (Figure 26). The previous analysis provided initial values for K , and initial Q_0 values were calculated as a fraction f of event precipitation. These K and f values were allowed to vary on a monthly basis and manually calibrated in order to match observed lake inflows (Table 4).

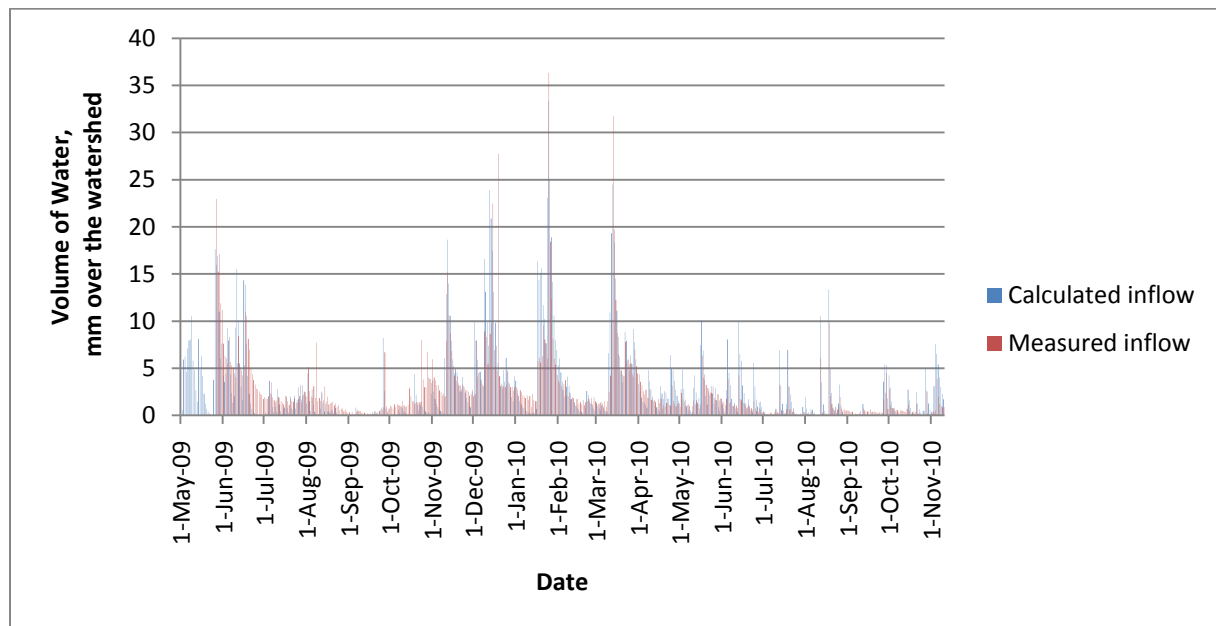


Figure 26: Recession curve model

Month	f	K
January	0.29	0.75
February	0.29	0.75
March	0.29	0.75
April	0.29	0.75
May	0.29	0.55
June	0.35	0.55
July	0.2	0.4
August	0.2	0.33
September	0.16	0.3
October	0.2	0.5
November	0.29	0.75
December	0.29	0.75

Table 4: Parameter values used in recession curve model

The recession curve model can also be evaluated by calculating predicted lake levels and comparing them with measured lake levels. Figure 27 shows that these calculated values for lake level underpredict lake levels for July through November in the summer of 2009 and overpredict them during the same period in 2010. There appears, therefore, to be no combination of monthly K and f values that allows for a closer match of both years' data. The recession curve model was also applied to the lake from 2002 through 2003 using Jansons [2004] original lake level and precipitation data (Figure 28). A similar underprediction of lake levels occurs in September through November.

The reasons for this underprediction in lake level appear clearer in 2009 than in the 2002-2003 dataset. Rainfall at MLBS in May 2009 was extremely high, with a monthly total of 283 mm, compared to an average monthly rate of only 100 mm for the period from May 2009 to November 2010. It appears that the soil and groundwater storage response of the watershed in late spring of 2009 (Figure 24) is more similar to a typical winter response than a summer response. Correlation of parameters K and f to evapotranspiration and antecedent precipitation, respectively, rather than fixed monthly values, might yield better results. While initial attempts in this direction did improve lake level predictions for 2009-2010, the methods were not confirmed by similar improvements in the 2002-2003 prediction of lake levels. Further work would be required to establish optimal physically-based parameters for K and f .

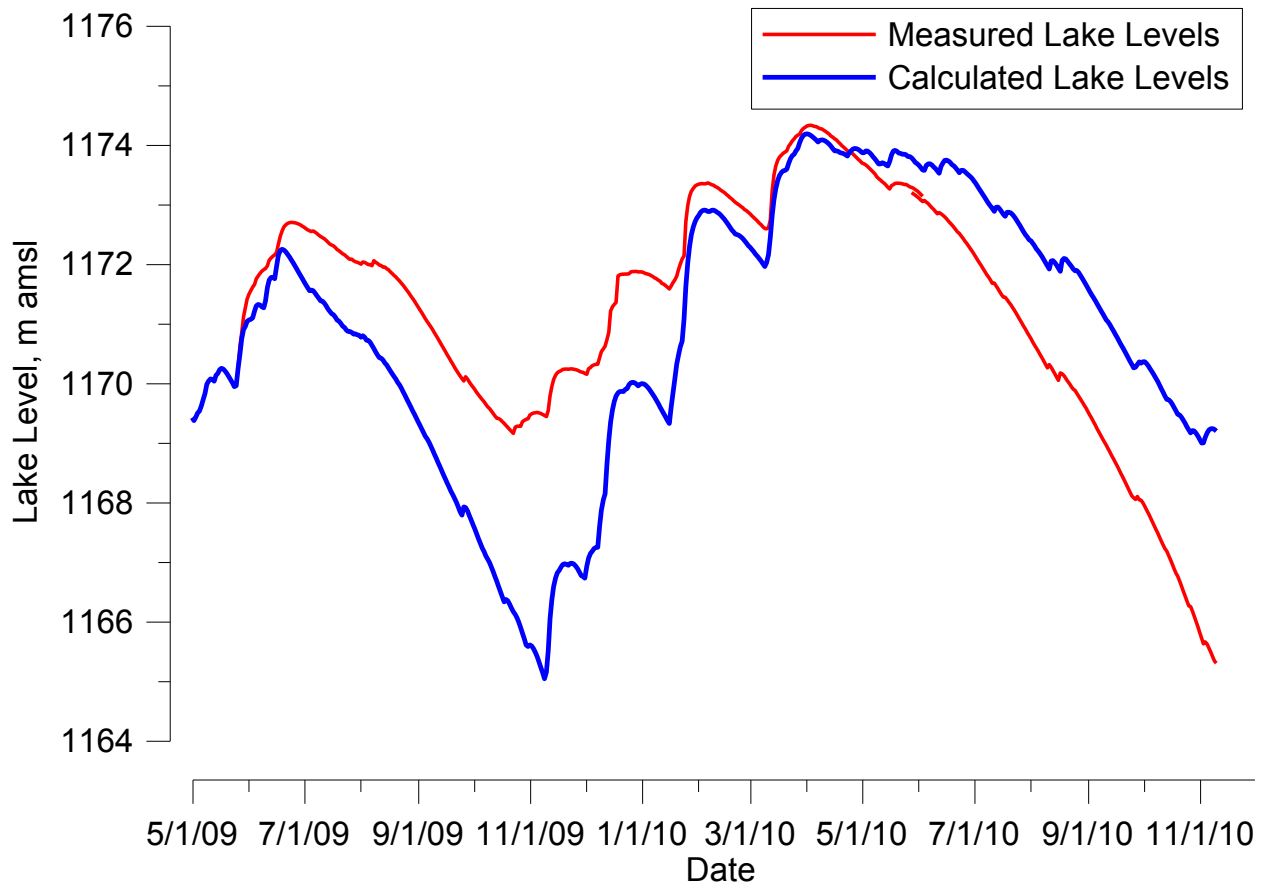


Figure 27: Predicted and measured lake levels
May 2009 to November 2010

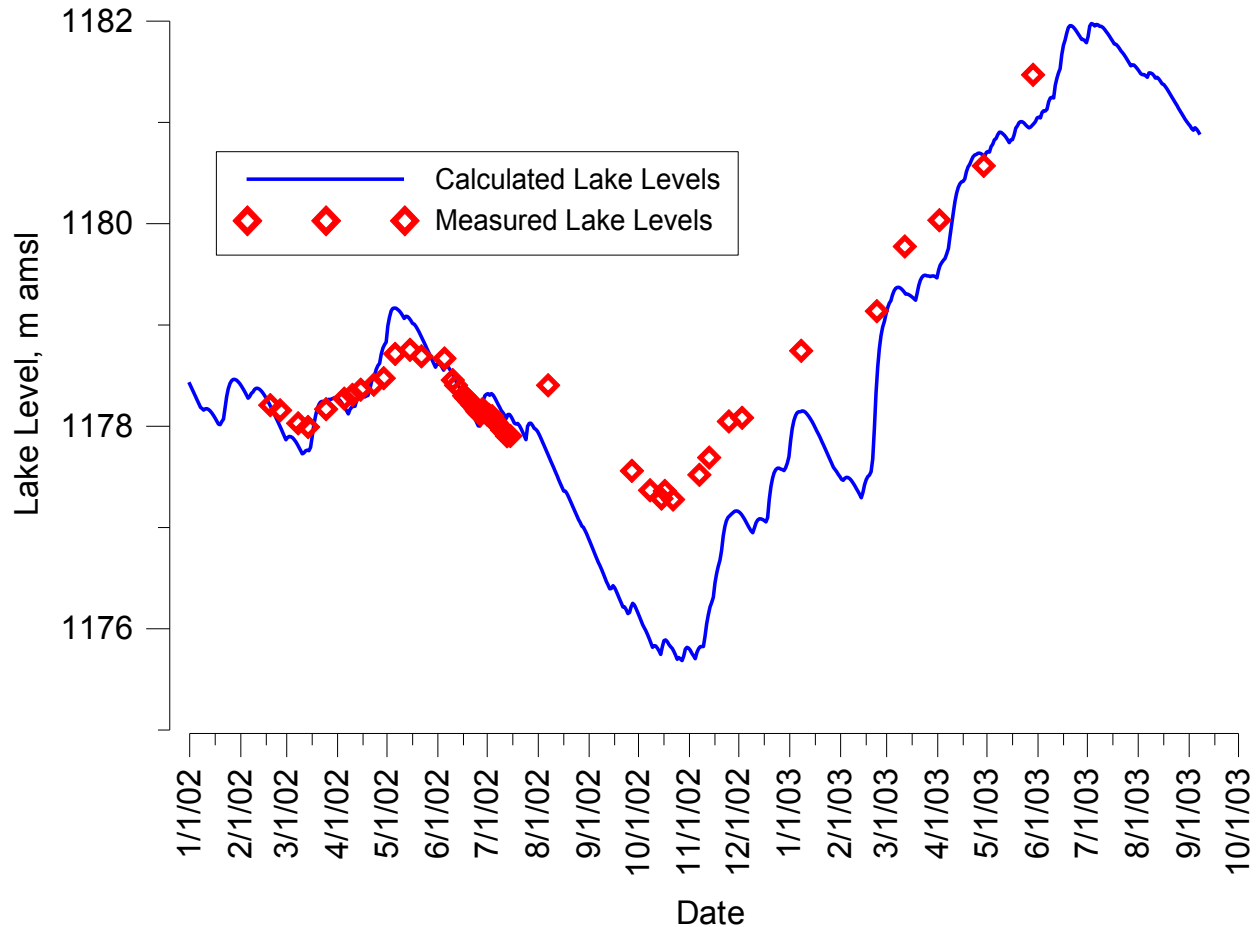


Figure 28: Predicted and measured lake levels
January 2002 to September 2003

Implications of Constant Net Groundwater Outflow

Conservation of energy in a hydraulic system dictates that higher head differences lead to higher flows through the system, and it had been anticipated that the effects of higher groundwater outflows from the lake might be visible in the lake hydrograph when lake levels were higher. Namely, the magnitude of the baseflow of the hydrograph (the daily change in storage, minus effects of recent precipitation) might increase at higher heads. However, no such effect was measured. Three possible reasons for this are that 1) groundwater inflows to the lake, which are not directly related to recent precipitation and also increase with higher heads in the lake, obscure changes in actual groundwater outflows, 2) measurement uncertainty of the change in storage term is larger than any change in groundwater outflows, or 3) the hydraulic geometry of the drain could theoretically be sensitive to head differences. While a combination of all three is also possible, the first explanation might best account for the observed constant baseflow, based on measurements of spring flow during the summer of 2010. These

considerations must be explored when addressing the second hypothesis, which considers whether lake levels have an effect on groundwater outflows from the lake.

Monitoring of flow in streams, seeps, and springs in and around the lake bed was attempted in the summer and early fall of 2010. However, the continuous measurements from the summer of 2010 are not considered reliable for technical reasons discussed below, and isolated measurements made at other times are not sufficient to determine how baseflow might have changed over the period of study. Accounting for these inflows is further complicated because as the lake level fluctuates, springs that are unmeasured groundwater flows when the lake is high become surface flows when the lake is low. Two of the major springs (SS, ES in Figure 47) were monitored with weirs (at a man-made point of flow restriction and with a v-notch weir, respectively) and pressure transducers for a period of two months during the summer of 2010. These springs, however, tend to be located over areas that make weir installation and maintenance problematic, as the geologic material is composed of either boulders and cobbles, or fine-grained unconsolidated sediment, and weirs appear to have been subject to occasional piping failures as the unconsolidated sediment between cobbles became eroded beneath the weirs. In addition, it is difficult to ascertain whether flow observed above ground constitutes the majority of flow, or if more flow is present at depths that the weir did not intercept. Calibration of weirs using saline tracers was also attempted in order to account for water that might pass underneath or around a weir at a given spring site, but calibration is recommended spanning the range of flows, and due to the flashy flows in streams following rain events, the highest flows were never calibrated. Qualitatively, however, both of those springs showed trends of decreasing flow and both had gone nearly dry by mid-September 2010, beginning with the higher-elevation ES location. Some minor flows were present during a rain event on September 27, 2010, when hourly monitoring of these same two locations (ES, SS in Figure 47) as well as three others showed a total peak inflow to the lakebed that did not exceed 0.30 L/s. The area of reported springs [Cawley *et al.*, 2001] near the Juniata contact on the lakebed became exposed in the fall of 2010 as the lake continued to recede, as did other small springs on the east side of the lakebed. Meanwhile, the lake continued to lose water at a steady rate. The only observed inflow was from the small stream that appeared to initiate incrementally from diffuse seeps in the middle of the lakebed and flowed at a rate of approximately 3-4 L/s, by estimation with an uncalibrated v-notch weir plate in early November 2010. Continuous monitoring of this flow to establish values for baseflow was not feasible in the easily disturbed fine sand sediments over which the lakebed stream flowed, and even if this baseflow could have been ascertained, other springs that may feed the lake from below the waterline could not be monitored.

Qualitatively, springflow data suggest that combined surface and groundwater inflows to the lake decreased as the lake levels declined from May to November 2010, despite the problems with springflow measurement described in the preceding paragraph. If the inflow rate of 3-4 L/s were to represent a value of baseflow from springs and streams in November

2010, it would mean that the actual groundwater outflow would have to be 3-4 L/s, 0.25 mm/day over the watershed, or about 8.3% above the measured constant baseflow of 44 L/s. It is not known exactly how the total inflow rate value may have changed over the period of declining head in the lake, but the magnitude of this flow is similar to the uncertainty in net outflow measurements (0.23 mm over the watershed).

Even if groundwater inflows to the lake did not change significantly with lake level changes, there is a second possible explanation for why the daily baseflow of the lake hydrograph might remain constant. The second explanation is that the measured baseflow of the lake hydrograph, though it appears constant, is not, but any long-term variation is obscured by the limits of precision of the measured volume changes in the lake. The limits of detection of flow for the system were calculated to be approximately 0.23 mm/day over the watershed in the summer and fall of 2010. Using a simple application of Darcy's law, an exit head from the system can be calculated such that the difference in flow between when the lake level is high and when it is low is below the limits of detection of the equipment.

$$Q_{high} = -KA \frac{(h_{high} - h_{exit})}{L} \quad (4)$$

and

$$Q_{low} = -KA \frac{(h_{low} - h_{exit})}{L} \quad (5)$$

Assuming K, A, and L to be constant, the equations are combined and solved for h_{exit} .

$$h_{exit} = \frac{Q_{high}h_{low} - Q_{low}h_{high}}{Q_{high} - Q_{low}} \quad (6)$$

An exit head could then be calculated based on known heads and known and projected flow values. Using values from the June and November heads of 1172 m and 1165 m, and considering volumetric flows of 3 mm and 2.77 mm over the watershed as our high and low Q's which are indistinguishable by our method of instrumentation, an exit head is calculated to be approximately 1081 m. Interestingly, this elevation corresponds to one of the steepest parts of the Pond Drain profile between Mountain Lake and the New River, second only to the Cascades. This same equation, if the observed constant outflows were to continue over lake levels reaching the spillway, would predict a lower limit for an exit head at 936 m, at the elevation of Little Stony Creek in Little Meadows, which is located where the stream profile flattens significantly prior to reaching the Cascades (Figure 29). This explanation provides a second possible reason that the net lake outflow may have appeared to be constant despite head changes in the lake.

The report of a spring one kilometer north of the Pond Drain spillway, posited to be a source of groundwater outflow deriving from the lake [Cawley *et al.*, 2001], was also

reevaluated. The posited spring had been found to have lake-bottom diatoms present in sediment. This author suggests that the location tested may have been the juncture of another small stream from an adjacent watershed which was slightly misrepresented on the USGS topographic map. No significant differences in major ions were found between this location and locations along Pond Drain upstream and downstream of the juncture (PDS, PDBch, PDBdg in Figure 47), and reexamination of the diatom data indicated that the location (PDBch) did not have significantly higher lake-bottom diatom counts than other sites closer to Pond Drain (Parker, 2010, pers. comm.) No springs along Pond Drain were found during the course of this study. However, the presence of shallow lake diatoms had been reported in all locations. If the stream in question is only a surface stream, this would suggest that transport of lake diatoms would have occurred during high water events in the stream, which may be possible as the stream profile gradient is low in the area.

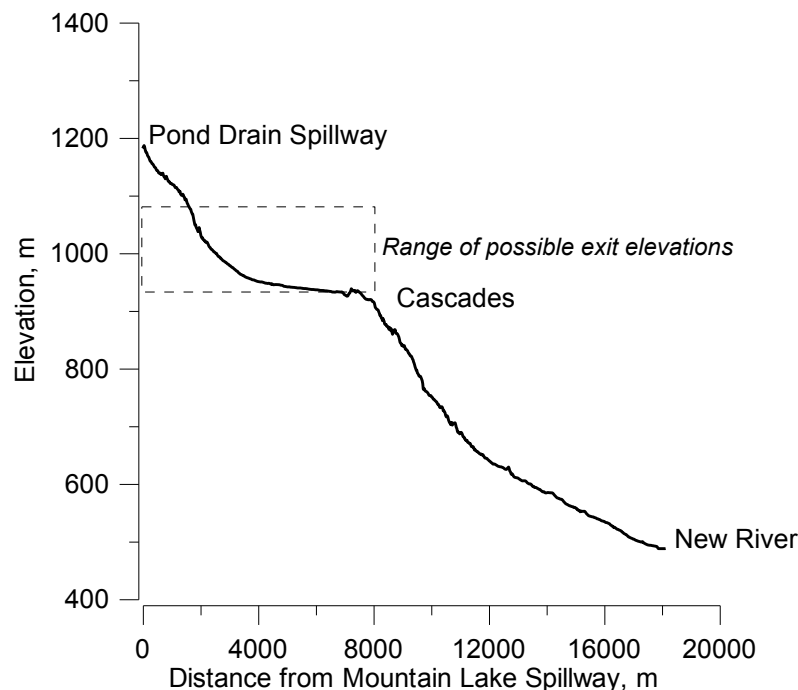


Figure 29: Pond Drain stream profile

A third theoretical explanation for observed constant net outflow is that the geologic medium in the vicinity of the outlet may change in some way in response to head changes. This change could be lumped together with either the area or the hydraulic conductivity term in Darcy's Law, depending on nature of the closure with respect to the representative volume that describes K. For example, higher pressure could cause a tighter seal, at some restriction in the flow path, thus reducing the effective area of the flow tube. Though collapse or sedimentation may play a role in the nature of drainage from the lake, there is currently no direct evidence that supports this third explanation.

4. Possible Geologic Controls Influencing Lake Levels

Lake levels are influenced significantly by groundwater flows to and from the lake; therefore, geologic controls on pathways for this water entering and leaving the lake are worthy of investigation. This study focuses primarily on the possibilities of fault- and karst-related flowpaths, based on current and previous investigations that suggest support for these mechanisms.

The third hypothesis addresses the existence of a specific fault at the lake, which may have hydrologic significance. A fault could affect lake levels because it could be associated with high conductivity flowpaths from the lake leading to lake drainage, as well as enhancing flowpaths into the lake, allowing for the lake to fill, or both. Alternatively, a fault could also be associated with a region of lower hydraulic conductivity that prohibits rather than enhances flow. Cawley [1999] posited a northwesterly fault through the lakebed intersecting the area of the deepest crevice in the lake, as shown in Figure 30. Preferential flow paths in consolidated rocks are determined by fractures and joints, which may have local values of hydraulic conductivity orders of magnitude larger than the values of the surrounding host rock [Fetter, 2001]. Because stresses involved in joint and fracture formation have effects across a wide variety of scales, fractures observed at the surface may be linked to fractures at depth [NRC, 1996]. Therefore, to address the third hypothesis confirming the existence and evaluating the hydrologic significance of this posited fault, lineaments were analyzed using satellite imagery, joints were sampled at exposed outcrops, electrical resistivity sections were profiled in order to characterize subsurface resistivity as a proxy for hydraulic conductivity, and well logs were used to aid our geologic understanding of the site.

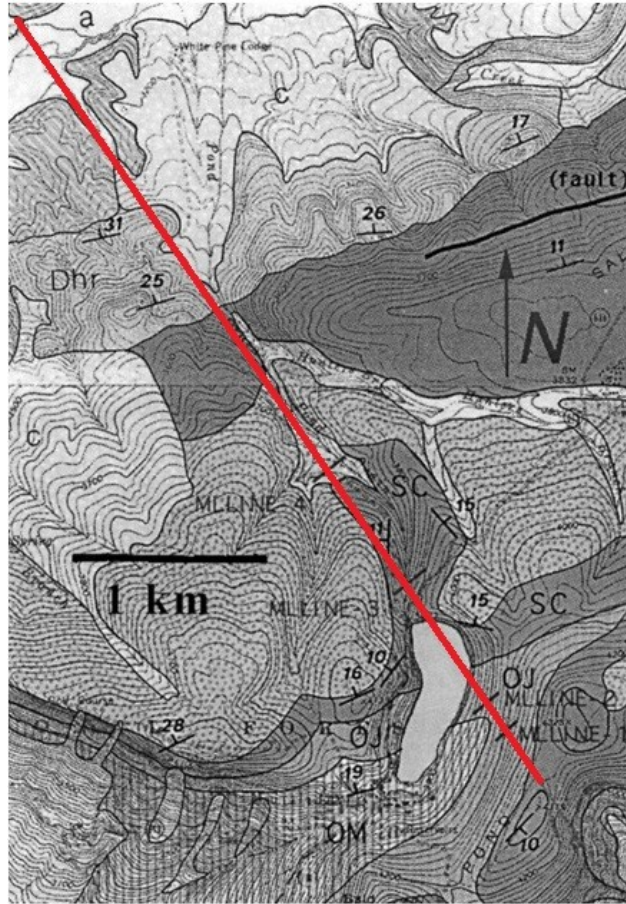


Figure 30: Posited location of lineament according to Cawley [1999]

[Cawley, 2001, used with permission]

The fourth hypothesis addresses the possibility of karst dissolution as providing a dominant pathway for flow. To address this, water chemistry was analyzed to assess possible types of flowpaths to and from the lake. In addition, the formation suspected to have sufficient limestone for dissolution to occur must be located, which requires detailed knowledge of the stratigraphy and structure at the site. Engineering-scale geologic maps of this site do not exist. Well logs were used in conjunction with identified outcrops and existing geologic maps to determine the physical location of a possible carbonate unit influencing lake discharge. Literature on regional stratigraphy was reviewed to aid in delineation of observed formations in well logs.

Lineament Analysis

A lineament analysis presents a statistical summary of linear features seen in remote sensing images and highlights areas of interest for subsequent field investigation. Lineaments are straight or nearly straight topographic features that are observable from aerial and satellite images [Marshak and Mitra, 1988]. Analysis of lineaments in a region may help characterize a region's drainage pattern and may be related to joint sets seen in the field. While a given

interpretation of lineament locations is highly subjective [Wise, 1982], the technique is routinely used for siting water wells in areas of fractured rock with some statistical success [Mabee, 1992], [Lie, 2002]. In the case of Mountain Lake, lineament analysis may indicate the location of a fault or preferential orientation of fractures associated with flow that is related to changing lake levels at the lake.

The lineament analysis was conducted using standard methods [Marshak and Mitra, 1988]. A color-infrared 3.5-minute digital orthoquad (Figure 31) was used to identify lineaments that extend from Salt Pond Mountain northward to the stream junction of Pond Drain and Hunter's Branch (Figure 10). Each lineament analysis should be repeated with similarly-processed images taken at different times of day to mitigate sun angle bias. All available images with significantly different sun angles were low quality aerial images that were inferior to the digital orthoquad. Since mixing image types in a single analysis is not recommended, the aerial images were not included. Azimuths and lengths are summarized in rose diagrams. Figure 32 shows the results of the lineament analysis with 5° bins and color-coded lineament lengths, and Figure 33 shows the same data using 20° bins and no color-coding.

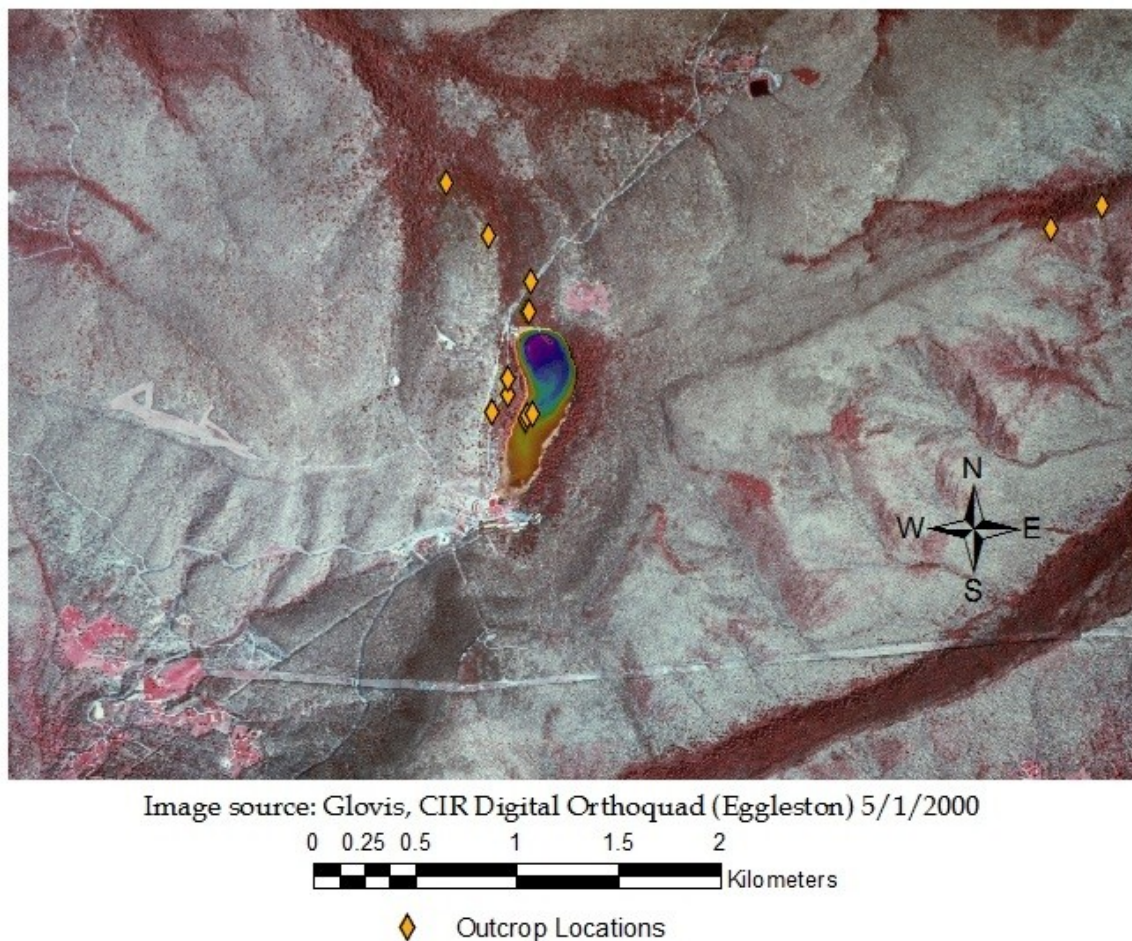


Figure 31: Satellite imagery used in lineament analysis

and locations of outcrops included in joint sampling
[Source: Glovis, USGS CIR Digital Orthoquad, 2010]

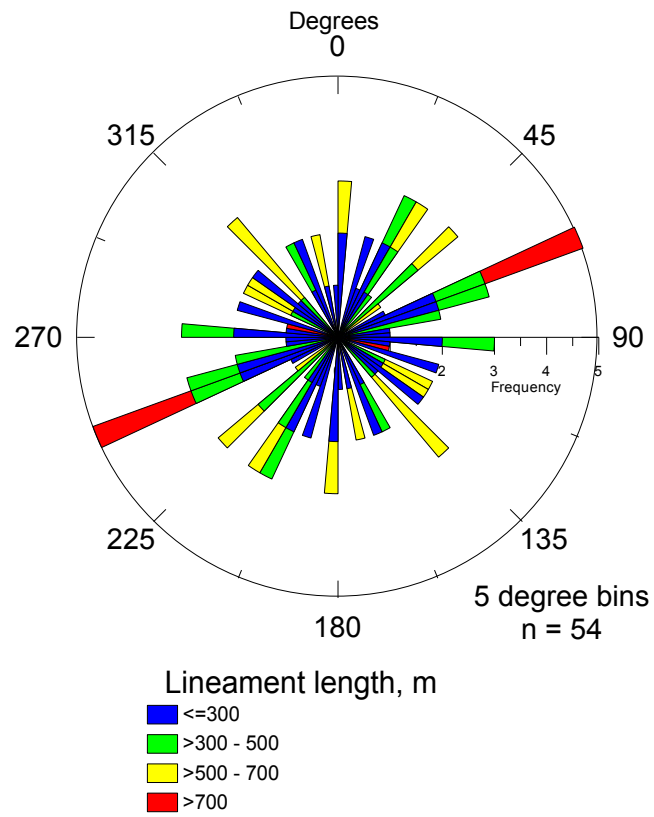


Figure 32: Lineament directions and lengths, using 5 degree bins

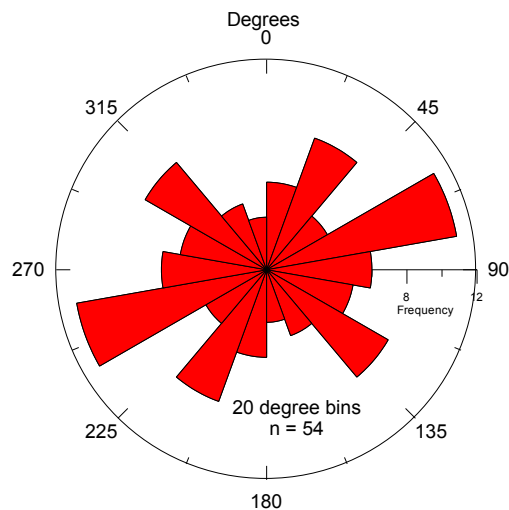


Figure 33: Lineament orientations, using 20 degree bins

Predominant lineament orientations were not found in the northwesterly direction corresponding to a postulated fault by Cawley [1999]. Instead, the prominent lineament direction is found to be in an east-northeast direction. Three predominant orientations emerge when using relatively wide 20 degree bins: 1) near 70° E of N, 2) near 45° W of N, and 3) near 30° E of N. However, using the same data with 5° bins and color-coded lineament lengths, the bin from 65-70° predominates. The longest lineament, over 1.2 km long, is oriented at 66° E of N. This long lineament was determined in the field to be a buried cliff-like break in the Tuscarora formation, which is marked by a steep rise or bench structure with Tuscarora boulders in the vicinity of the base. Other more subtle benches subparallel to Sartain Creek can also be seen in the orthophoto and were confirmed in the field. Similar but shorter double benches are also found to the south and southeast of the lake. A geologic map [Bartholomew, 2000] indicate that some outcrops had been noted and logged, but these bench features were not apparent on 20-ft contour topographic maps or existing geologic maps and are located where no roads and few trails pass. McDowell and Schultz [1990], however, mapped a minor thrust fault at this location as well as a 14° northerly change in the direction of the Bane Anticline to the north and east of the lake (Figure 34). This mapped fault corresponds to the preferred lineament direction found in the orthophoto, but no map overlay had been published that identified the fault in relation to the location of Mountain Lake. Therefore, the lineament analysis corroborates the previously mapped fault to the east-northeast of the lake rather than the predominant fault to the northwest.

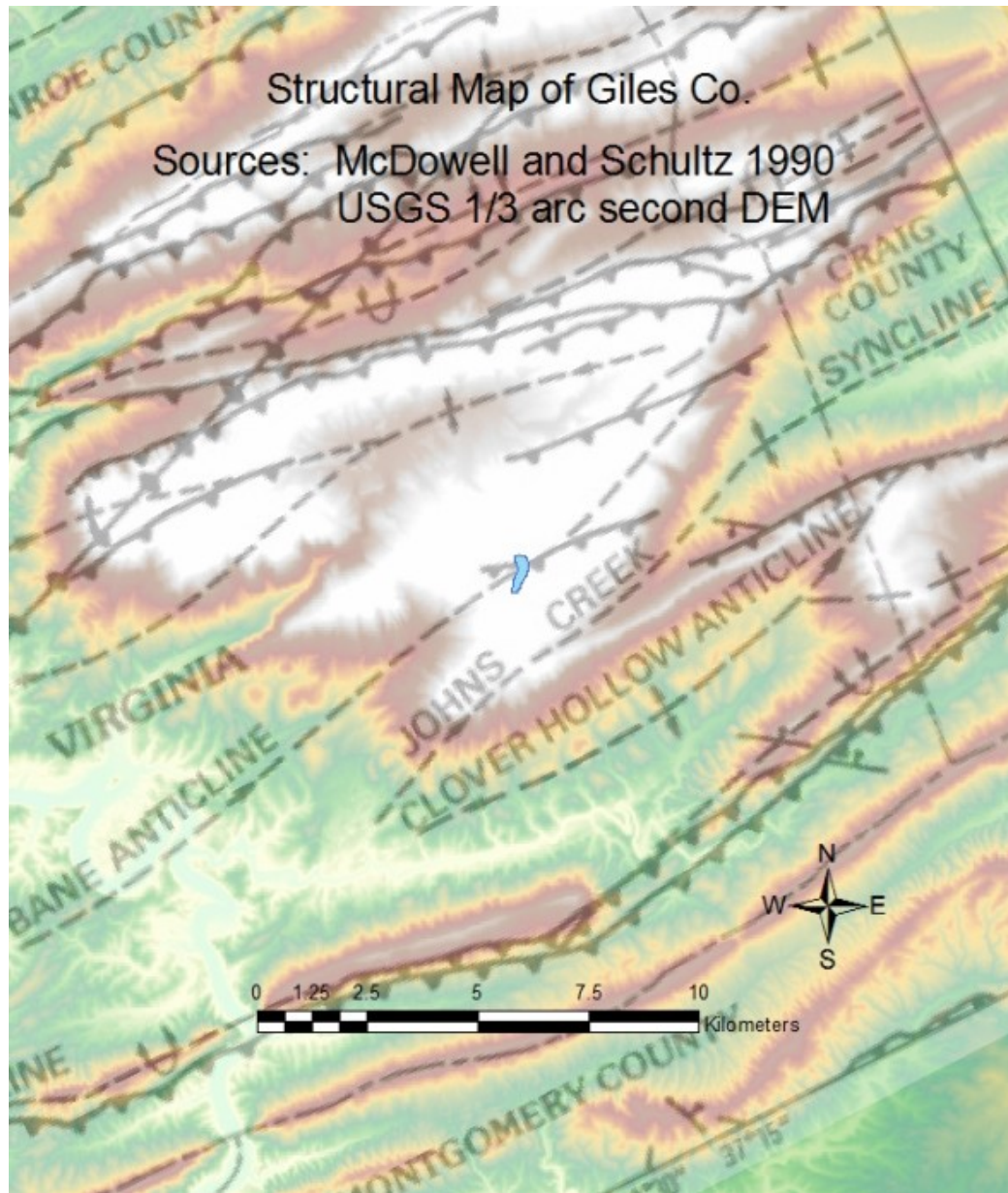


Figure 34: Structural map of Mountain Lake area

Mountain Lake at center; digital elevation model as background

[Sources: McDowell and Schultz, 1990 and USGS 1/3 arc second DEM, 2009]

Joint Sampling

Joint sampling was conducted to provide a statistical summary of joint orientations that are visible in outcrops. Fractures observed at outcrops may be representative of regional stresses associated with joint formation. As such, fractures may indicate directions of

preferential pathways for flow to or from Mountain Lake because stresses involved in joint formation are likely to have been applied across a variety of scales. In this study, strike and dip of bedding were measured and representative sampling of open joint sets was performed along a traverse [Marshak and Mitra, 1988] at 13 outcrops (Figure 31) in the study area and include eight outcrops from the Tuscarora formation, four from the Juniata, and one from the Rose Hill. Figure 35 presents a summary of the strike of vertical and sub-vertical joint planes in the sampled outcrops, and results are discussed below.

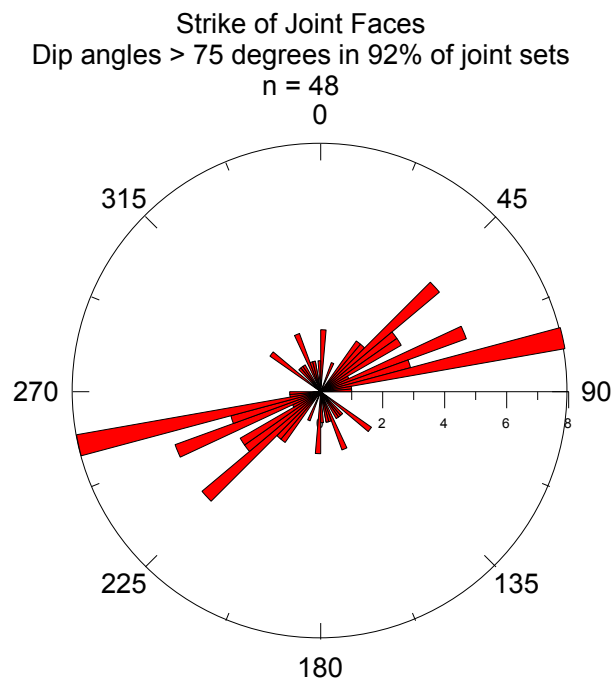


Figure 35: Joint directions from outcrops

Results of the joint sampling method used show that a joint trend exists in a north-northeasterly direction between 50° and 80° . These joints may be associated with a breached Bane Anticline or the identified fault to the east of the lake. Two outcrops were included in the inventory that did not present any noticeable joint sets; both of which were located on White Pine Road farther from the crest of the Bane anticline. Although joint sets in the northwesterly direction may exist and be of structural significance for a groundwater outlet for the lake as proposed by Cawley [1999], such a joint orientation was not evident with the sampling method used in this study.

Several sources of possible bias in this joint sampling are noted, however, that might have misrepresented the true joint population. First, to obtain true rather than apparent strike and dip of joint faces, sampling was performed on open joints rather than fracture traces, and

only if both joint faces were present. Following this method, and because outcrops exposed by roadcuts were thought not be representative of jointing that would be observable under natural conditions, the outcrop faces themselves were not sampled. Second, no correction factor was applied in this study to counter undersampling of joints at small angles with respect to the traverse. Since five of the outcrops were largely exposed along north-south lying outcrop faces, ignoring these exposed faces may have undercounted joints close to this direction [Terzaghi, 1965], [Wang and Mauldon, 2006]. Thirdly, it is difficult to ascertain with certainty that a block, even a large one, is in place. At one location in the study area, for example, a massive Tuscarora block ten meters high and twenty meters long had broken off the side of a cliff. If this block had been partly buried in soil rather than exposed on all sides, it would not be apparent that it is not in place, and measured joints would not represent properties of in-place rock. The fourth source of sampling bias lies in the relation of individual joints in a joint set to bedding planes. Since traverses were generally parallel to bedding planes, a joint set repeated across multiple bedding planes would only be counted once, but a joint set consisting of multiple joints through the same bedding plane would be counted multiple times. This sampling bias may minimize the importance of some joint directions. Consequently, while evidence for east-northeasterly joint sets is clearly present, the lack of evidence for northwesterly joint sets should not be taken as proof of their absence.

Electrical Resistivity

Electrical resistivity tomography (ERT) uses direct electrical currents to probe the electrical resistivity of the subsurface, which, in hydrology, may be associated with the degree of fluid saturation [Parasnis, 1997]. The technique has been used to locate the water table, water- or air-filled fracture zones, faults, and karst conduits [W J Seaton, Burbey, T. J., 2000]. ERT does not distinguish moving water from ponded water in the subsurface, but increased hydraulic conductivity may be inferred if differences in apparent elevation of low-resistivity areas are associated with head differences. In this study, ERT was applied with the objective of confirming the presence of a northwesterly fault through the lake.

A number of techniques are available for electrical resistivity surveying, each of which has properties that make it more or less suitable for a given geologic environment and study objective. At Mountain Lake, the study site is fractured rock and the principle objective of using ERT is location of a hydrologically active fault. In a series of numerical as well as field tests in a fractured crystalline study area in the Blue Ridge Province, Seaton and Burbey [2002] found the dipole-dipole array to yield the deepest profile and most detailed resolution of several common arrays shown in Figure 36. While the study area at Mountain Lake consists of sedimentary rather than metamorphic rock and thus primary porosity may have a relatively larger influence on flow, fracture flow is still expected to dominate and so the dipole-dipole method was considered to be the best configuration for identifying fluid-filled fracture pathways.

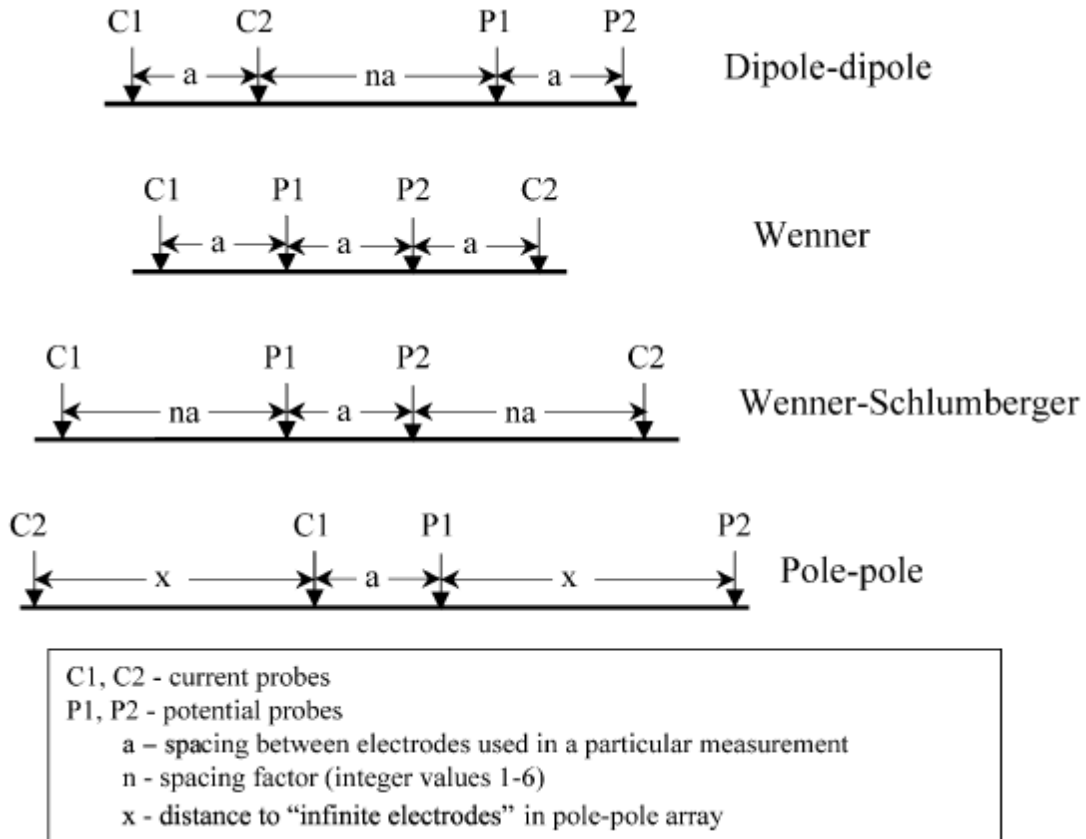


Figure 36: Commonly used arrays in electrical resistivity tomography
[Seaton and Burbey 2002, used with permission]

In this study, data for thirteen subsurface electrical resistivity profiles surrounding the lake were obtained using an Advanced Geosciences, Inc. (AGI) Supersting® 8-channel resistivity meter. Measurements were made with 64 electrodes and a maximum line length of 630 m. Finding adequate contact between the electrodes and the ground and maintaining straight lines in dense forest and over rubble fields also influenced location and curvature of profiles. Latitude and longitude of electrodes were marked with a handheld GPS and topography data was input using a combination of field measurements, the USGS Newport Quadrangle, and the lakebed bathymetric map. Local variations in elevation were not entered into model topography where not significant on the scale of the line. Locations of profiles are shown in Figure 37.

Attaining a deep subsurface profile with reliable results was important to the Mountain Lake study. Therefore, measures were taken to ensure that the array spacing used did not create undue noise. One weakness of the dipole-dipole array is a decreasing signal-to-noise ratio with increasing depth of study that occurs with increasing distance between the point of current

injection and the point at which potentials are measured. The signal is subject to both interference from telluric currents and detection limits of the receiver [Ward, 1990] and a maximum value of $n=6$ is suggested [Loke, 2010]. However, this suggestion is based on one-dimensional depth soundings where the a -spacing is held constant. Automated resistivity systems that acquire data for two-dimensional profiles, however, result in an increase in both a - and n -spacings in various combinations. In addition, the AGI instrument increases signal strength automatically when no signal is detected. Whereas validation of the n -spacing in Seaton and Burbey's field tests was based on repeated measurements and exclusion of data points that varied by more than 3% between measurements, validation of appropriate n -spacing in this study was based on visual comparison of inverted results from separate repeated measurements. In the present study, two sets of measurements were taken for three of the profiles, then the apparent resistivities are inverted, and the results are then visually compared. No qualitative difference was found between the versions in any of three cases for which repeated measurements were taken. Though this did not eliminate the possibility of influence of currents that change diurnally or on a time scale larger than the time between the two measurements, it did provide a check on random telluric noise and problems with readings below detection limits. Therefore, in this study the dipole separation factor in the dipole-dipole array was set to a maximum² of $n=8$.

The purpose of collecting resistivity data is to create a model of the subsurface geology as it is reflected in regions of differing electrical (and by inference, hydraulic) conductivity. In this study individual model profiles were created by inversion of the measured resistivities with the software Res2DInv® using a non-linear least squares optimization technique [Loke and Barker, 1996]. Since flow in fractured rock can be expected to lead to wide variations in resistivities on small spatial scales, a robust inversion using the absolute value rather than the square of the difference between calculated and measured values was employed to avoid artificial smoothing of true sharp boundaries between regions of differing resistivity. Because the dipole-dipole array is most sensitive in the short distance between each pair of current or potential electrodes, and because large differences in near-surface resistivity were often present, the surveys were inverted using a model cell size of half the distance between electrodes. True depths of subsurface features are not known, but if the parameters used to calculate apparent depths are uniformly applied, then apparent depths can be compared between profiles and interpretations involving head differences can be made. Apparent depths in this study were calculated based on an initial depth of 0.35 times the electrode spacing with an eleven percent increase in each subsequent model layer.

² The N2 line which curved around the north end of the lake bed was inverted with a subset of the collected data to remove the effects of the curvature, using only segments that were equivalent to straight lines, and subsequently reducing the apparent depth of the profile.

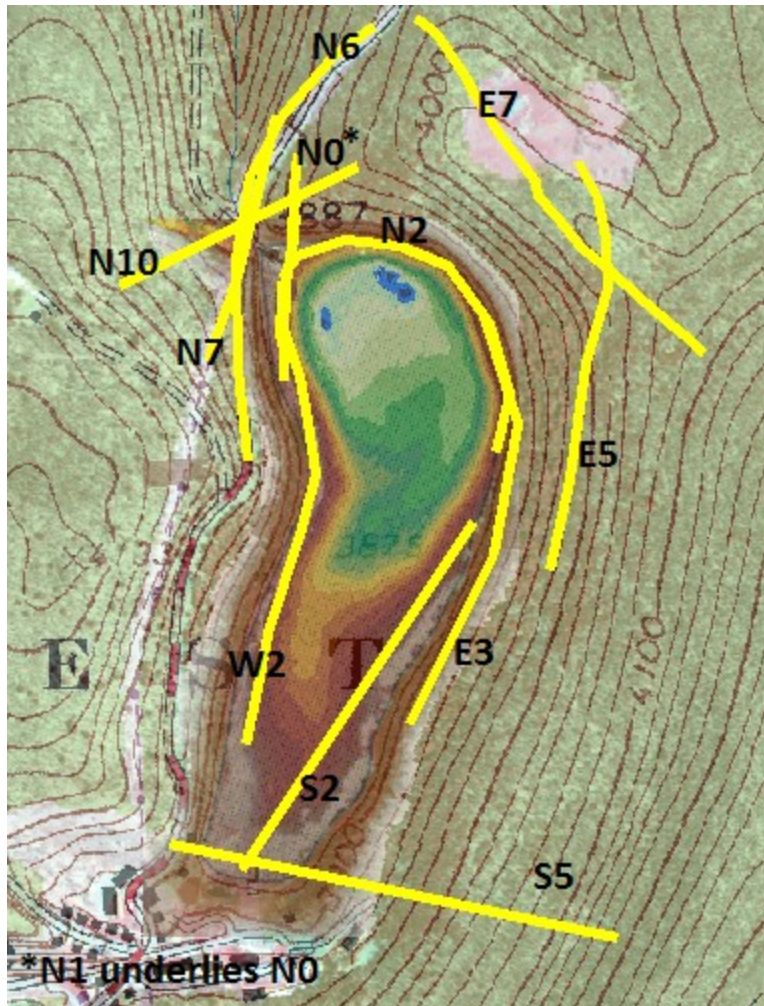


Figure 37: Locations of resistivity profiles

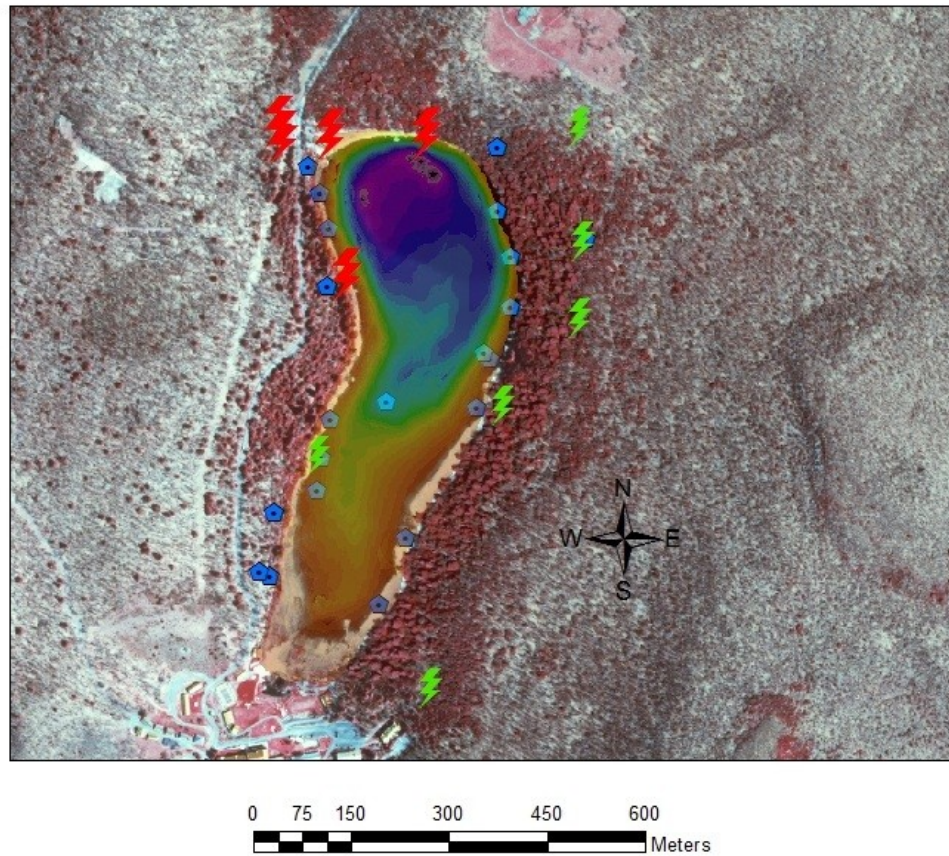
[Background map source: USGS 7 ½ minute DRG, Eggleston]

The most significant results of the resistivity survey are summarized in the discussion below, and additional profiles can be found in Appendix G. Model resistivity sections for eight of the profiles are displayed in the section that follows. Lines were labeled with one letter and one number, where N, S, E or W represent the north, south, east, or west side of the lake where the profile was measured, and a lower number indicates a line closer to the lake. Note that the results display resistivity variation from the average on a log scale, but the scale is not fixed from survey to survey. All lengths along the profiles are given in meters from the array's first electrode. The profile is measured from the most southern and/or western point of the line, unless noted otherwise. Apparent elevations are given in meters.

ERT profiles reveal approximately 12 low-resistivity zones around the lake suggesting potential subsurface flow pathways. Locations of these areas are shown in Figure 38 and have

been divided into two groups: those with apparent elevations deeper than the bottom of the lake and those with elevations above the bottom of the lake. Even though elevations of model sections are approximate, the elevations of these low-resistivity areas are generally higher to the south and east of the lake, ranging from 1165 to 1180 m in elevation and thus falling within the range of lake levels. The low resistivity areas to the north and west of the lake are generally lower in elevation, ranging from 1120 to 1175 m. Springs and streams with the highest flows are on the lakebed itself or to the east and south; many of the other locations indicated in Figure 38 have very minor and intermittent inflows. The results suggest that the low-resistivity areas to the south and east may crop out as springs that feed the lake and the general slope of the water table or piezometric surface decreases from the south and east, and that lake outflow appears to be to the north and west. This correlates with evidence of water levels in drillers' logs from the public supply wells along the southern end of the lake (W1 and W3 in Figure 47), which show that water levels at the time of drilling were higher than lake levels at the time. No clear evidence is present of a deeper drainage structure to the east of the lake that would drain across the eastern continental divide towards Sartain Creek.

Image source: Glovis: CIR Digital Orthoquad (Eggleston) 5/1/2000



Explanation

-  Deep Low-Resistivity Features
-  Shallow Low-Resistivity Features
-  Springs, Streams, or Seeps

Figure 38: Locations of subsurface low-resistivity areas
[Image source: Glovis, USGS CIR Digital Orthoquad, 2010]

Drainage Pathways from the Lake

Lines N0, N2, N6, N7, and N10 all show low-resistivity areas near Pond Drain below the level of the lake that may represent drainage pathways from Mountain Lake near Pond Drain. Lines N6, N7, and N10 all show that a low-resistivity area is not centered directly below the present-day stream channel³ that emerges from the spillway, but rather underlies an area south and west of the stream channel. Figure 39 displays profile N6, which clearly indicates the area of interest with respect to lake drainage.

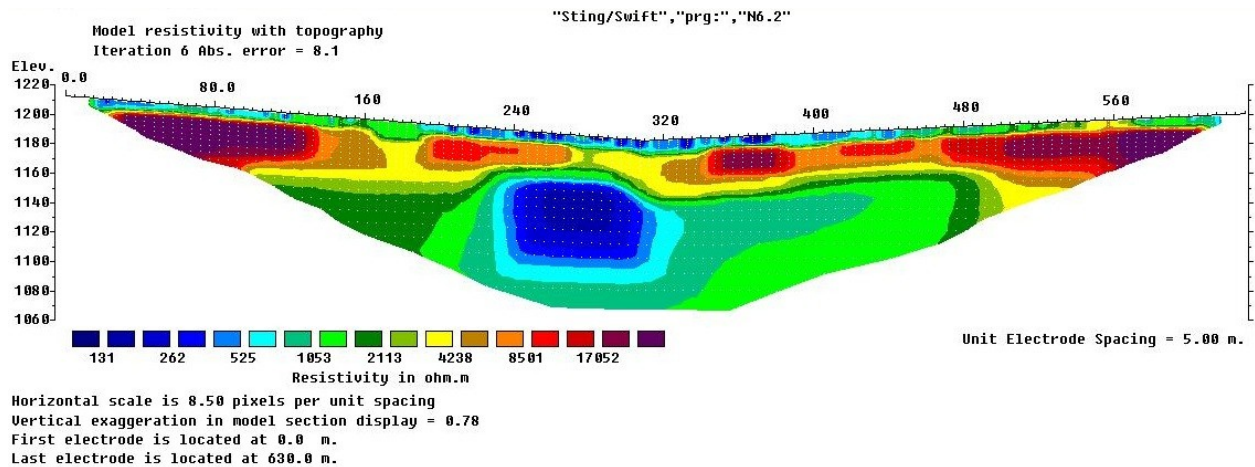


Figure 39: Resistivity profile N6

Figure 40 shows profile N7, which displays a more detailed view of the upper section of the same low-resistivity area shown in N6.

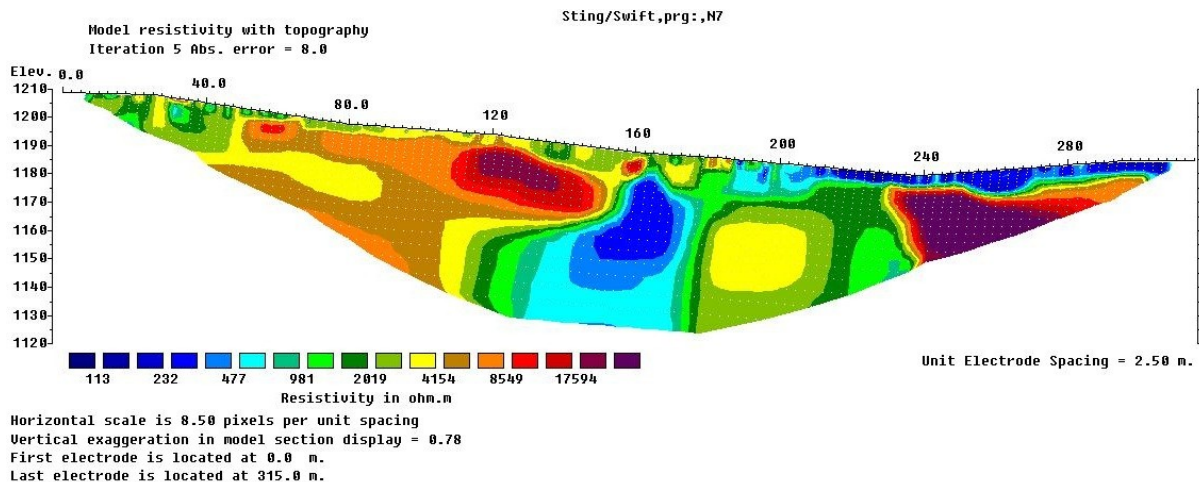


Figure 40: Resistivity profile N7

³ The stream channel emerging from the lake spillway at Pond Drain is incorrectly placed on the USGS topographic map. It crosses underneath Mountain Lake Road north of where it is drawn.

Profile N10 (Figure 41) shows the same low-resistivity area located beneath a point west of the present-day stream. The low-resistivity area beneath meter 60 on this line corresponds to the location of an abandoned spring box. The boulder field on the eastern part of this line prevented adequate contact being made at some electrodes in the upper part of the line, and so the high-resistivity portion of this line east of meter 192 should not necessarily be construed as an excessively high-resistivity area.

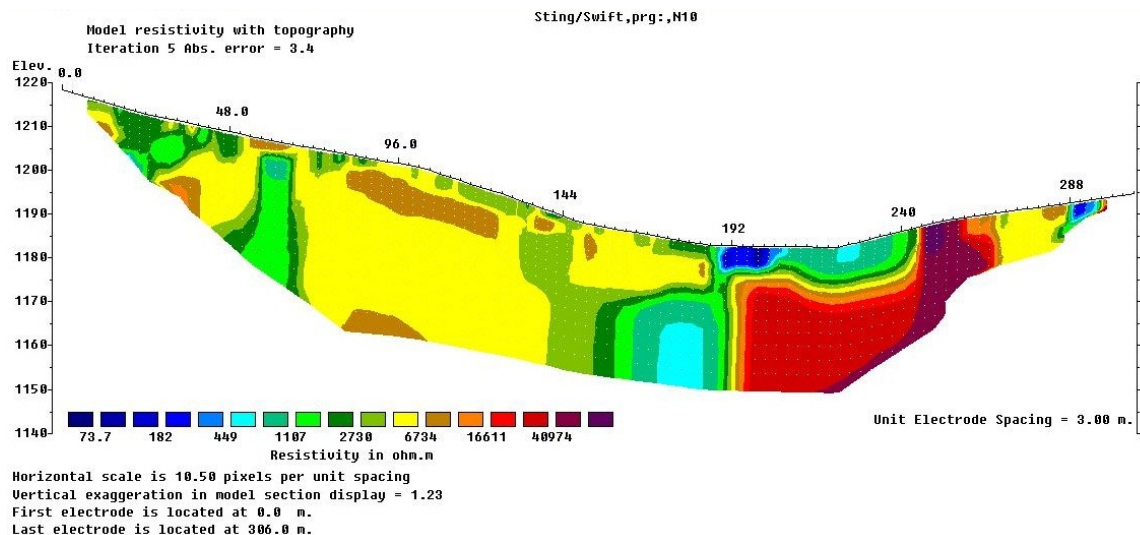


Figure 41: Resistivity profile N10

Profile N0 (Figure 42) reveals a low-resistivity area at an apparent elevation of 1120 m centered to the north of Newport House and well below the bottom of the lake. This is the deepest low-resistivity area of any of the profiles.

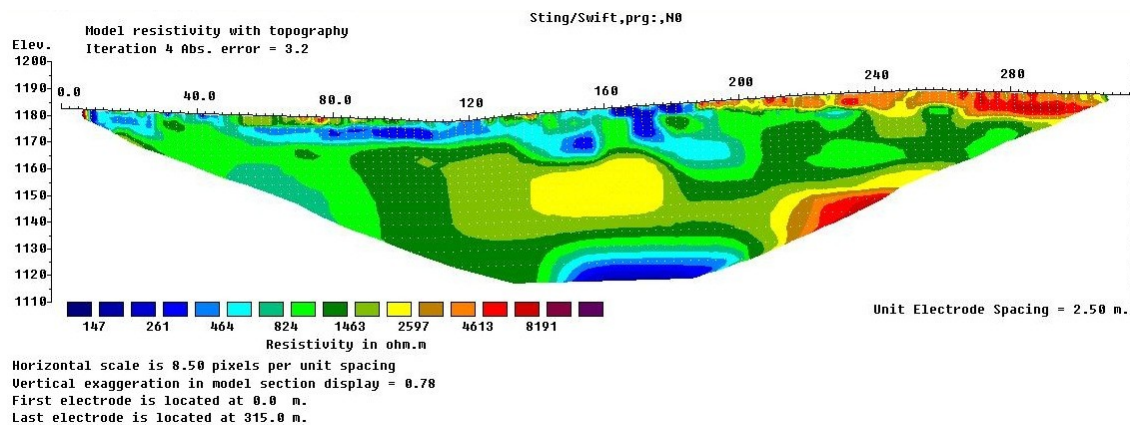


Figure 42: Resistivity profile N0

The N2 line profiled the north side of the lake shore (Figure 43). Although significant curvature occurs in this line, the data file was edited to only include data from shorter (<200 m) segments that approximated a straight line and the apparent depths are consequently much shallower than those of other profiles. The profile clearly shows a low-resistivity area due north of the deepest crevice and low-resistivity areas corresponding to incoming springs on the northeastern side of the lakebed.

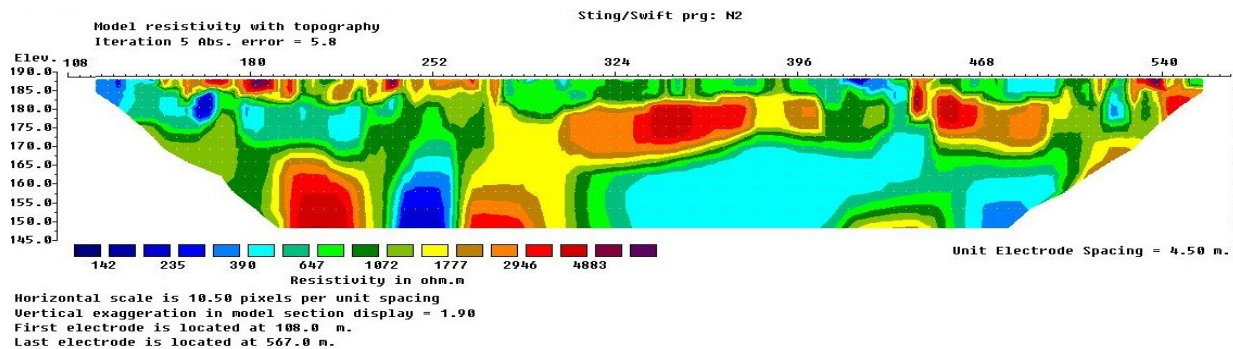


Figure 43: Resistivity profile N2

Figure 44 displays the W2 profile, which features two low-resistivity areas. The southern area corresponds to a wide and marshy area with a stream channel that joins the other major surface inflow on the lakebed. The larger and deeper low-resistivity area is not located near the other deep holes at the northwest corner of the lake but has a similar apparent depth.

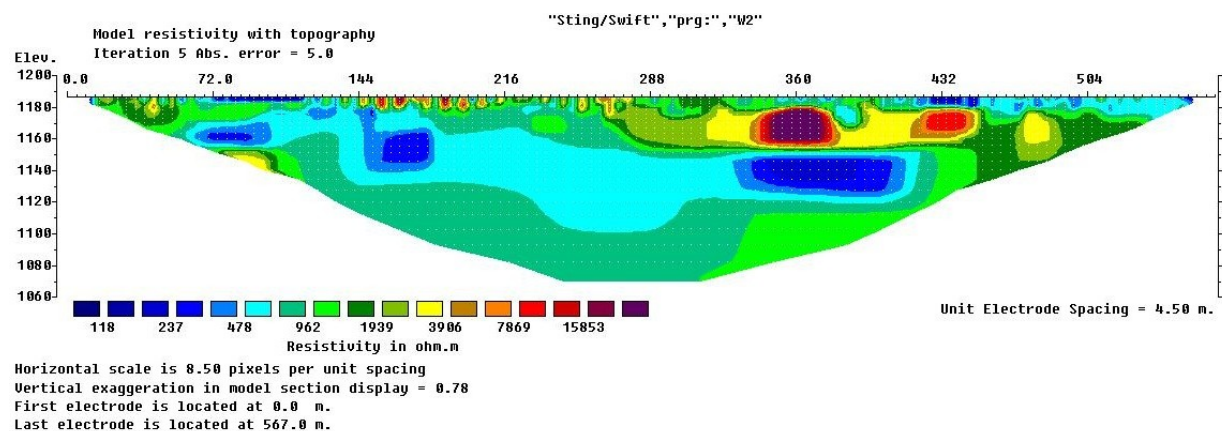


Figure 44: Resistivity profile W2

Recharge Areas

Profile E5 (Figure 45) is centered on the spring to the east of the lake (ES in Figure 47). The source area from the spring appears to extend northward past the topographic divide with

the next small watershed to the northeast, suggesting that recharge for this small region of the watershed is structurally rather than topographically controlled.

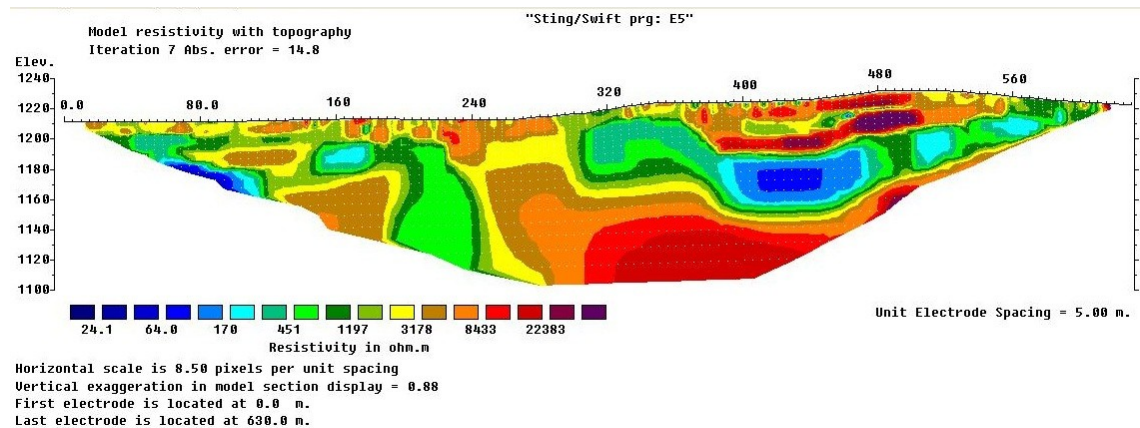


Figure 45: Resistivity profile E5

Profile E7⁴ (Figure 46) reveals a distinct resistive block with vertical sides and an apparent dip towards the southeast, with a zone of low resistivity that strongly suggests drainage over the top of the block and to an area that corresponds with one of the springs on the eastern side of the lake. A second recharge zone from 470 to 510 m appears to show the continuation of a similar feature from the E5 line.

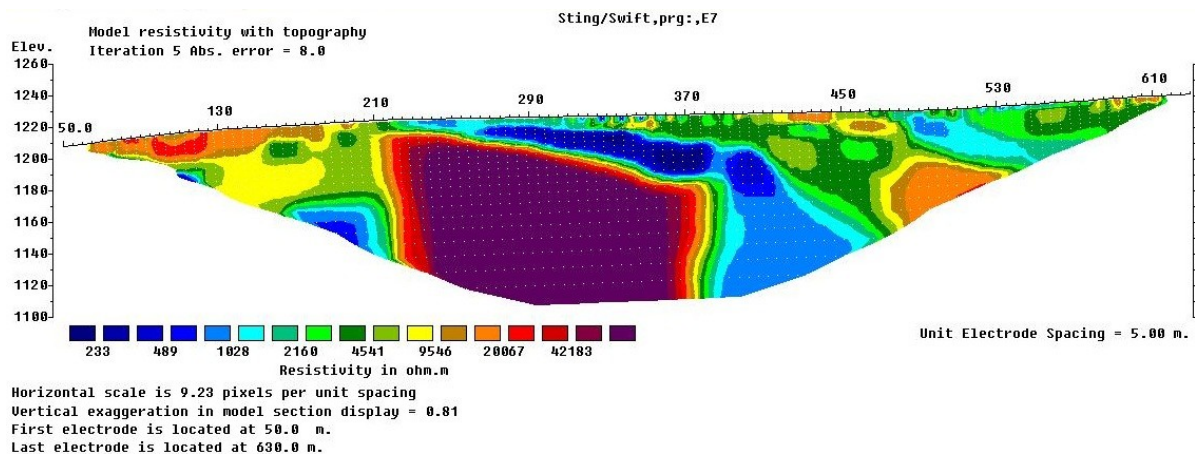


Figure 46: Resistivity profile E7

In summary, the resistivity results reveal a highly heterogeneous subsurface at Mountain Lake. Likely subsurface flowpaths from the northwest corner of the lake have been located. Whether or not these flowpaths are associated with a fault as posited by Cawley [Cawley, 1999], however, is unclear. No unique feature was found that intersected the lake at

⁴ Note that in this profile only, distance along the line increases to the south.

the deepest crevice and continued both to the northwest and southeast as depicted in Figure 30. Evidence exists for significant springflow on the eastern side of the lake, possibly associated with the eastern fault mapped by McDowell and Schultz [1990]. These springs, however, largely ceased flowing as the lake level declined in the fall of 2010.

Well Logs

Three wells on the hotel property were logged with a gamma tool, an optical televiewer, and calipers in an effort to understand how structure and stratigraphy might affect flowpaths in the vicinity of the lake. The complete log from the Horse Barn 1 well lies near the axis of the anticline at the northeast corner of the lake, and is well-placed for characterization of the geology present at the north end of the lake. This well log was used both to locate the elevation of contacts between formations and to identify the formations present at the elevations of the deep low-resistivity sections seen in the resistivity profiles. The contact elevations were used to confirm the presence of a structural feature as mapped by McDowell and Schultz [1990]. The other two wells logged had some blockages that allowed for only partial logging of the boreholes, and aside from demonstrating that soft and/or fractured rock is present in the subsurface, the logs from these wells are not discussed further here.

The Tuscarora formation was identified in the well log as the formation between the elevations of 1169.7 and 1188.7 m on the basis of its color, texture, gamma signature, and thickness. The Tuscarora is composed largely of quartzarenite which is an arenite with a mineralogical framework of at least 95% quartz grains, and includes conglomeritic sections [[Pettijohn, 1957; F J Pettijohn, 1975]. This quartzarenite is associated with a low gamma signature which was clearly visible in the well log between these elevations. The Tuscarora also may contain fine-grained red iron-bearing sandstone and shale in the upper part of the section in southwestern Virginia [USGS, 2010] which may be associated with some darker bands and small increases in the gamma signature towards the top of the formation and also at 1176 m where the caliper showed a widening of the well bore. The thickness of the Tuscarora in Giles County is close to 30 m and 8.6 m of upper Juniata were found to be transitional with Tuscarora at Gap Mountain [Diecchio, 1985]. These descriptions are compatible with the Horse Barn 1 well log, which showed the thickness of the section to be 19 m, with an abrupt transition to Rose Hill above and what may be intertonguing with the Juniata below where intermittent low gamma signatures and light colors extend for an additional 5 m. The Tuscarora is found from field mapping to outcrop above the boulder field to the north of the lake at approximately this elevation.

The position of the Tuscarora in the well aids in the interpretation of the structural feature mapped by McDowell and Schultz (1990). While the well log shows nearly horizontal strata compatible with its location near the crest of an anticline, the elevation of the top of the Tuscarora differs significantly from the elevation of the Tuscarora as it appears 850 m away (at 37° 21.5' N, 80° 31.6' W) across the fault as drawn by McDowell and Schultz [1990], where the top of the Tuscarora is at least 1318 m [USGS, 2009]. This corresponds to a difference in

elevation of the top of the Tuscarora of at least 129.6 m, which is significant given that the thickness of the section is only 19 m and the strata appear horizontal in the wellbore. The feature identified by McDowell and Schultz was mapped as a thrust fault. The confirmed evidence for thrust faulting in the area is from a possibly repeated 4 m section of Tuscarora higher in the well (1206 to 1210 m) together with the dip angle of strata suggested by resistivity lines E5 and E7. However, the well logging data do not, on their own, provide enough information to conclude whether the structural feature is a thrust fault or simply a breached anticline that has undergone some vertical displacement. Regardless, the presence of either type of structure to the east of the lake may provide a preferential pathways for groundwater flow through the Silurian and upper Ordovician siliciclastic formations found in the lake watershed.

The well log was also used to identify the formations present at the bottom of the lake and at elevations suggested by the deep holes interpreted from the electrical resistivity profiles at the north end of the lake. The Juniata exhibits cyclical bedding where a typical cycle has three sections that begins with arenite, followed by an arenite to wacke with interbedded mudstone, and ending with a mudstone. These cycles vary in thickness from 0.3 to over 4 m [Diecchio, 1985], and lead to a variable gamma signature with low-gamma sections linked to arenite followed by sawtooth-like sections linked to interbedding of arenite with mudstone. This gamma signature is present below the bottom of the Tuscarora to a depth of 1102 m and is particularly distinct in the lower half of this section. Near 1100 m, the gamma signature changes to one that is less variable and remains in the higher range, which may indicate that arenites are no longer present and may represent the contact with the Reedsville-Trenton formation. Measurements at Narrows, Virginia and at Gap Mountain show the Juniata formation to be approximately 75 m thick [Diecchio, 1985]; this interpretation would indicate that the formation at Mountain Lake is 70 m thick. However, the contact between the Reedsville-Trenton and the Juniata has been found to be gradational [Blue, 2011].

The stratigraphy present at the bottom of the lake and at the location of the deeper low-resistivity sections observed in the electrical resistivity profiles may put certain constraints on a conceptual model of flowpaths to and from the lake. The level of the lake, when full, approaches, but does not reach, the top of the Tuscarora formation. The bottom of the lake is at an elevation equivalent to the upper third of the Juniata formation, with approximately 50 m between the bottom of the lake and the top of the Reedsville-Trenton. Photos of the dry lake from 2008 (Figure 4) suggest that the crevices at the bottom occur in the Juniata, and the Tuscarora float lining the northern half of the lake is consistent with the lithology interpreted from well logs. The apparent depth of the deepest low-resistivity pathway at the northern end of the lake as indicated in resistivity profiles occurs at approximately 1120 m, which is about 20 m above the elevation of the Reedsville-Trenton. Though the apparent depth should be treated as an approximate number, it appears that the depth of penetration of the resistivity meter does not reach the Reedsville-Trenton, and therefore these flowpaths may be occurring in the Juniata formation, suggesting that flowpaths from the lake may be fault or collapse-related, rather than

directly karstic. The location of a fault with a displacement of almost 130 m intersecting the center of the lakebed, however, might mean that the Reedsville-Trenton at the southern end of the lake is significantly higher in elevation, as confirmed by the presence of Reedsville-Trenton outcrops above lake level at the southwestern end of the lake. Accordingly, water entering the lake through flowpaths from the southern end of the lake may be more likely be associated directly with karst features.

Water Chemistry

If subsurface pathways at Mountain Lake are formed by karst features in the upper Reedsville-Trenton formation, it is possible that the water chemistry in the lake may be rich in dissolved calcium bicarbonate. Therefore, water was sampled for major ions at one well, three springs, five streams, and deep and shallow areas of the lake. Because the three public supply wells on the property were not available for sampling during the present study, chemical data were used from sampling conducted by MLH in 2008. The use of older data is justified because the USGS calculated apparent ages of the water in two of the public supply wells to be 15 years old (Appendix B) so it is not likely that the well water chemistry had changed significantly in the intervening two years. Results of the chemical analyses were used to draw conclusions about flow paths in the watershed.

Water samples were collected on several occasions in the fall of 2010. Sampling locations are presented in Figure 47, where the orange Stiff diagrams are centered over the sampling locations, and Table 5 specifies the type of sampling location. Lake water was sampled both at the northwestern edge of the lake and from the deepest part of the lake using a bailer. All samples were filtered with a 0.45 μm filter, preserved appropriately (HNO_3 for cations and hydrochloric acid for dissolved organic carbon), and analyzed for major ions and Al, Fe, and P. The samples were titrated after a 36-hour holding time, then alkalinity was calculated using the inflection point method and presented as bicarbonate equivalent. Appendix H provides tables with sampling dates, analysis of charge balances, pH, conductivity, temperature, and chemical data.

The results of the chemical analysis (Figure 48) show that the water samples range along a spectrum, from calcium bicarbonate water in the wells to water with a more balanced ionic distribution in the stream to the east of the lake in siliciclastic rocks. Total dissolved solids decrease along this spectrum from 65 mg/L as found in well W1 to 5 mg/L as found in Sartain Branch stream. The lake water samples reflect an average composition between well water and surface water. From these data, it could be suggested that the springs and streams have shallower sources separate from those of the public supply wells, and that the lake is a mixture of both end members.

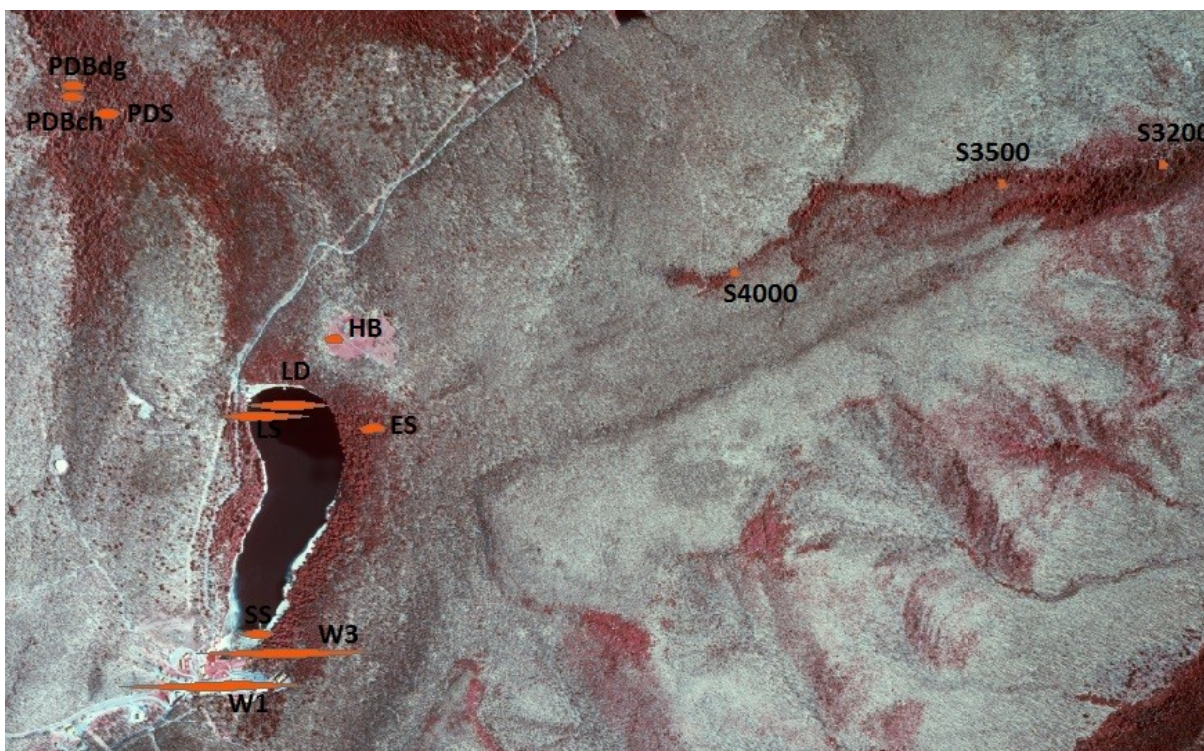


Figure 47: Locations of chemical sampling with Stiff diagrams

[Image source: Glovis, USGS CIR Digital Orthoquad, 2010]

Abbreviation	Type of site	Name
W1	well	MLH Well #1
W3	well	MLH Well #3
HB	well	MLH Horse Barn Well #1
SS	spring	South Spring
ES	spring	East Spring (Dodsworth Spring)
LS	lake	Lake, shallow
LD	lake	Lake, deep
S4000	spring/stream	Sartain Creek at elevation of 4000 ft.
S3500	stream	Sartain Creek at elevation of 3500 ft.
S3200	stream	Sartain Creek at elevation of 3200 ft.
PDBch	stream	Pond Drain, at branch
PDBdg	stream	Pond Drain, below bridge
PDS	stream	Pond Drain, south of branch and bridge

Table 5: Sampling locations

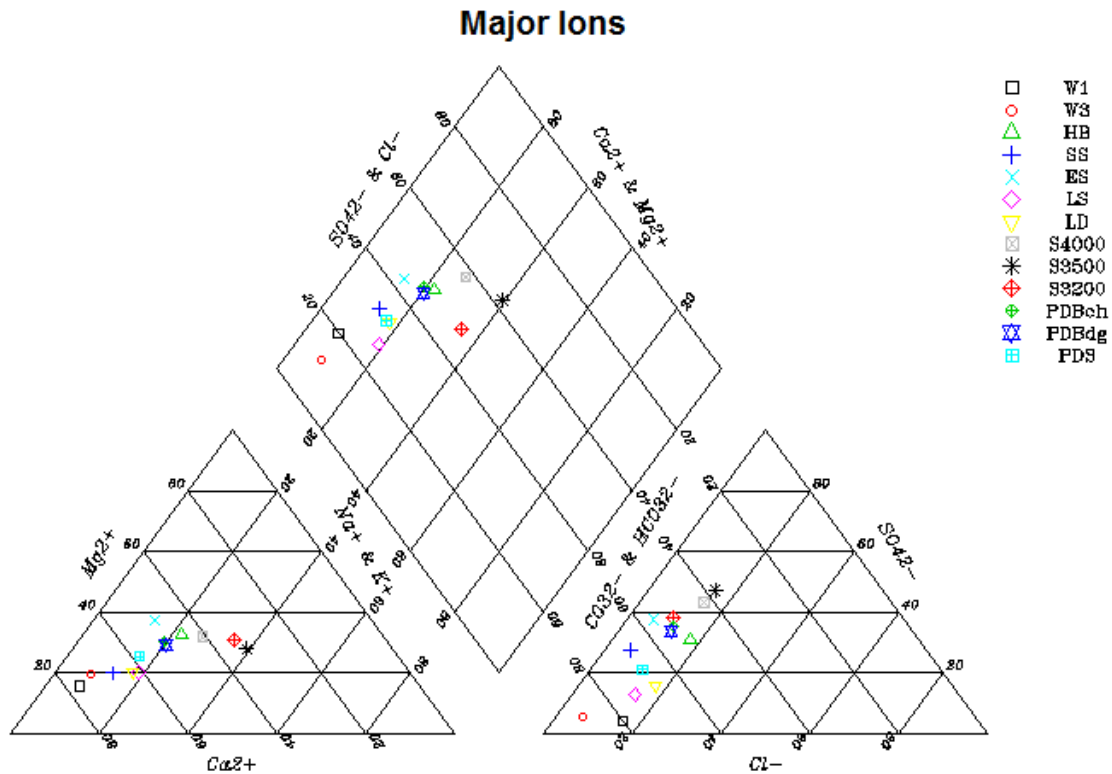


Figure 48: Water chemistry: Piper diagram

Chemical concentrations were then compared to water typical of karst dominated waters, which typically has dissolved calcium carbonate concentrations of 30 to 440 mg/L [Drew, 1985]. The water chemistry data shows that the wells in the Reedsville-Trenton have calcium bicarbonate concentrations of 55 and 57 mg/L and the lake itself in 2010 had concentrations of 24 and 28 mg/L. By comparison, measured streams and springs have concentrations in the range of 2-10 mg/L. In contrast, lake water samples from the summer of 1998 as measured by Cawley [1999] had alkalinities of only 4-8 mg/L (calcium was not measured). Thus it is likely that groundwater inflows high in calcium bicarbonate are present and that they become relatively more important when lake levels are low and recharge from precipitation is minimal, as has been shown to be the case during the summer of 2010. It appears that groundwater inflows to the lake make contact with formations high in calcium-carbonate.

If karst dissolution were to control flow pathways that affect lake levels at Mountain Lake, first and foremost the structure and stratigraphy would have to allow for those pathways to develop. The term karst describes landscapes affected by the dissolution of calcium carbonate by slightly acidic precipitation, and karst terrain is typically found in limestone and dolostone. Karst features are widespread in Giles County, although most of these are located in middle and not upper Ordovician formations [Holsinger, 1963]. However, confirming a previous report by Butts [1940], the upper section of the upper Ordovician Reedsville-Trenton at Rt. 460 north of Narrows, Va. was found to have significant highly reactive limestone beds throughout the top 10 m of the formation, with individual beds up to 25 cm thick. A section of Reedsville-Trenton one kilometer from the lake (37° 20.83 N, 80° 32.65 W), possibly the same one referred to by Parker [1975], was also found to have multiple 5 cm beds that were somewhat reactive to hydrochloric acid. The lake crevices, then, may lie approximately 50 m above this potentially reactive formation, based on the location of the Reedsville-Trenton interpreted from the Horse Barn well log. The fault to the east of the lake may be providing a pathway for fresh precipitation to reach any existing limestone beds that may exist at the location of the lake, and resistivity results in profiles N0, N6, and W2 suggest approximate apparent depths of low-resistivity areas at the elevation of the top of the Reedsville-Trenton formation. Therefore, structurally and stratigraphically it seems that karst dissolution pathways and related collapse structures may be a possibility at Mountain Lake.

Since karst dissolution could be expected to widen drainage pathways over time, the temporal scale required for enhanced dissolution needs to be addressed. Thin shale beds, like those seen at an outcrop in the Reedsville along Mountain Lake Road, are commonly the site of bedding plane partings where dissolution commences [Waltham *et al.*, 2005], and dissolution occurs more rapidly near the top of a limestone layer as incoming water has its greatest concentration of carbon dioxide. The total amount of water is also directly correlated to dissolution rates, and allogenic water as runoff from nearby clastic rocks can supplement infiltration from direct precipitation [Waltham *et al.*, 2005]. Dissolution along bedding planes and fractures in karst begins very slowly as laminar flow in fissures but can reach rates of 0.1 mm/year when fissures become wide enough for the occurrence of turbulent flow, which is approximately 5-10 mm wide under normal hydraulic gradients [White, 1988]. Applying these dissolution rates to the known history of lake level changes extending back 4200 years [Cawley, 1999], it implies that any given fissure could possibly have widened by as much as 42 cm during this period. Given multiple fissures, this could significantly increase the cross-sectional area of any karst drainage pathways from the lake.

Collapse debris and sediment accumulation would need to be invoked in order to explain the fact that Mountain Lake not only drains but also refills, if Mountain Lake's high rate of drainage is due in part to karst dissolution of limestone beds in the Reedsville-Trenton. However, collapse may be a reasonable proposition. Thinly-bedded limestone is prone to collapse before dissolution results in extensive caverns [Waltham *et al.*, 2005]. Breakdown and collapse debris that accumulates both in larger caverns and below narrower *breccia pipes* may

subsequently lead to side wall undercutting and further lateral collapse. Also, the lowering of the water table associated with a dry lake episode and subsequent increase in effective stress on the drained formation may induce incipient collapse. Once a cavity is present, breakdown can also occur due to seismic events. For example, Tawney's Cave in southeastern Giles County (37° 18' 45" N, 80° 30' 42" W) had an extensive lower level that could be reached by "a fairly roomy slide" but due to breakdown from a 1959 earthquake became a "very tight 25 to 30 foot crawl" [Holsinger, 1963]. Debris-filled cavities could act as natural filters, holding in place finer sediment drained from the lakebed. All of these processes of breakdown could contribute to a change of the hydraulic geometry that may affect draining and a refilling of the lake.

Thus, in response to the fourth point of investigation, some evidence is found to suggest that karst dissolution may play a role in the nature of groundwater flows at the lake. If this is the case, the geologic feature at the deep end of the lake could be classified as a kind of caprock sinkhole, which is seldom more than 100 m across and is typically characterized by initial dissolution of interstratal carbonate layers and progressive upward collapsing of overlying siliciclastic layers [Walsham *et al.*, 2005].

5. Summary of Conclusions

Hydrologic Controls on Lake Level

Precipitation and Evapotranspiration:

- Lake levels respond quickly to precipitation, and are particularly sensitive to winter precipitation. This was especially evident in winter 2009-2010 rain-on-snow melt events. Summer and early fall rain events, by contrast, do not lead to a marked increase in lake levels. During the summer and early fall, the lake level was observed to decline at a steady volumetric rate, absent recent precipitation.
- Evapotranspiration, mediated by temporary soil (and possibly groundwater) storage during the growing season accounts for the subdued response of the lake to rain events in the summer and early fall.
- When rainfall is unusually high and in excess of evapotranspirative demand, as occurred in May 2009, the behavior of the soil and groundwater term and subsequent response of the lake may resemble typical winter behavior irrespective of season, with significant recharge to the lake continuing in the month following the high precipitation.

- The amount of total winter precipitation including snow, sleet, and ice storms as well as rain may be a determining factor in the long-term maintenance of lake levels. Lack of long-term records of snow, ice, and snow water equivalence introduces uncertainty into the historical study of lake response to precipitation.
- Direct lake evaporation increases slightly with the increased surface area of the lake when levels are high, but this is minimally significant with respect to total watershed evapotranspiration.

Baseflow of the Lake Hydrograph:

- Responding to the first hypothesis, the baseflow of the lake hydrograph was observed to remain constant on a daily time step from July to November 2010 and also during part of the summer of 2009. The daily baseflow was 3.00 ± 0.23 mm/day over the watershed. After accounting for direct lake evaporation, 93 to 97 percent of this change in storage, or approximately 44 L/s, is attributed to net groundwater outflow (subterranean outflows minus subterranean inflows). This value was found to remain constant on both a daily and a monthly basis, a result which differs from the previous water balance [Jansons, 2004].
- The baseflow of the lake hydrograph is calculated to be 44 L/s and is in very good agreement with the average value calculated by Jansons (43 L/s) over the previous 18-month period of study. Given different formulations of the water balance used in the two studies, this suggests that 1) snowfall in 2002-2003 was not significant, 2) two of the deficiencies in the calculation of evapotranspiration (using PAR rather than pyranometer data, which would underestimate ET, and using PET rather than AET, which would overestimate ET) are of opposite signs, and may have partly cancelled each other out when averaged over the 18-month period, and 3) the total change in soil and groundwater storage over the watershed calculated on an annual basis is not significant.
- Minimum rates of groundwater inflow to the lake are estimated to be 3-4 L/s, based on measurements of exposed springs as the lake level declined in the fall of 2010. This would correspond to a minimum rate of groundwater outflow from the lake of 47-48 L/s, given that the net groundwater outflow was observed to be constant. Changes to the rate of actual groundwater outflow, however, were not measurable. The second hypothesis involving head-dependence of groundwater outflows, then, was only considered from a theoretical perspective.

A simple annual model based on a water balance summary for the water year 2009 suggests that some condition(s) must have been different earlier in the last

century for the lake to have been full or near full as often as the historical record indicates. A recession curve model based on 2009-2010 data also underpredicts lake levels when applied to 2002-2003 data. Possible conditions that may have changed over time are groundwater inflows via springs in the lakebed and groundwater outflows via blockage and/or changes in sedimentation. Past unmeasured snowmelt could also explain this behavior, as could overestimation of historical lake levels.

- The historical data record, both for on-site precipitation and for lake levels, is not complete enough to undertake a detailed year-by-year study of past lake levels. This period of study is the first time that daily lake levels have been measured continuously.
- There is evidence that long-term changes to groundwater inflows to the lake have occurred in the last century. A spring that was once a principal water source for the hotel had been considered perennial in 1934 despite a severe drought several years earlier, but was observed to be only intermittent in 2010. Effects of precipitation on spring or lake inflows observed during this study, however, did not appear to extend longer than approximately one month, and the nature of the transition from perennial to intermittent was not observable on the time scale of this study.

Geologic Controls on Lake Level

Lake morphology:

- Lake level changes are much more easily observable when lake levels are low. Due to the shape of the lake, past accounts of the lake being 'full' may not have been uniformly accurate.

Faulting:

- A fault to the east of the lake which had not previously been discussed in the literature on the lake was confirmed in the field. It had been mapped as a thrust fault [McDowell and Schultz, 1990]. Evidence of a repeated section of Tuscarora was present in the Horse Barn well log.
- A well log, satellite imagery, and the DEM suggest that the Tuscarora is vertically displaced between the well and the mapped fault by 129 m. It is not clear if this vertical offset is related to a thrust fault, the presence of a breached anticline, or both.
- Given the presence of numerous ephemeral springs on the northeastern side of the lake, this fault is thought to provide pathways for recharge to the lake. An

area of diffuse perennial springs in the middle of the lakebed at a contact with the Juniata formation may also be related to this structure.

- Evidence for deep flowpaths to the northwest of the lake was found using electrical resistivity tomography, suggesting the existence of several localized areas of groundwater lake outflow at or below the bottom of the lake.
- Responding to the third hypothesis, the presence of a northwesterly fault through the deepest crevice as suggested by Cawley [1999] was not confirmed with ERT results, joint sampling, or the lineament analysis.

Karst:

- The uppermost section of the Reedsville-Trenton formation was confirmed to have significant carbonate content at Narrows, Virginia as first reported by Butts [1940], suggesting the possibility of the presence of carbonate content in the same formation at Mountain Lake as well.
- Concentrations of calcium and bicarbonate in the lake water when the lake levels were low in November 2010 indicate that lake water may be composed of a mix of water similar to well water from the Reedsville-Trenton formation, and water similar to that found in streams over siliciclastic rocks. This indicates that the lake may be fed by springs that are in contact with carbonates within the Reedsville-Trenton formation.
- Diffuse springs in the middle of the lakebed were observed to flow in the fall of 2010 even after all other springs surrounding the lake had ceased flowing. These springs may be connected to the fault to the east of the lake. The fault may provide flowpaths for water to make contact with carbonate rocks in the upper Reedsville-Trenton.
- Lower alkalinities measured in the past when the lake level was high indicate that the ratio of deeper groundwater flow to runoff and shallow springflow through siliciclastic rocks may change with lake levels. This suggests that increased inputs to the lake causing periods of higher lake levels come from increased precipitation, such as snowmelt or shallow springflow, rather than increased high-alkalinity groundwater.
- In response to the fourth hypothesis, chemical and structural evidence suggests that dissolution of calcium bicarbonate in the upper Reedsville-Trenton occurs in some pathways entering the lake and remains a possibility for pathways leaving the lake. Theories considering karst dissolution associated with collapse of the

overlying siliciclastic formations at the north end of the lake and/or the formation of discrete flowpaths exiting the lake to the northwest should not be discredited.

Additional Observations:

- No indications of lake outflows were found in the form of springs along the upper 1.5 km of Pond Drain during the course of this study.
- Well abstraction for hotel use does not have a significant impact on lake level, and use of the newest public supply well in an attempt to recharge the lake could, at best, replace only ten percent of the water lost to net subterranean lake outflow.
- No effects on lake or well hydrographs were observed following the mild regional seismic events that occurred during the study period.

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7. Appendices

A. Locations of Former Springs

Before water wells were drilled on the property, the water supply for Mountain Lake Hotel was taken from the lake itself and several springs. The location and state of the springs were described in a 1934 engineering inspection on file with the Virginia Department of

Environmental Quality [DEQ, 2010] . Major points are summarized here for future reference. Caretaker's Spring was located just west of the caretaker's house on the hill behind what is now the Activities Barn. East Spring (Dodsworth Spring) had dry-weather and wet-weather sections. Both were piped to the hotel and cottages. Cold Spring, located just below the Fitness Trail (old Bridle Path) approximately 400 m northeast of the hotel, was piped to the kitchen only and was intended to be the only source of drinking water. A pump station was located on the lake shore northeast of the hotel, and Meredith reports that "there is said to be a spring at the point where the pump is located, but this spring during normal times discharges below the water level in the lake." Testing in the summer of 1934 showed some bacteriological contamination, and the hotel had plans to discontinue using the lake as a source of water and instead to develop Buckeye Spring, the spring next to what is now the Maintenance Shed Well. The supply of ice for the hotel was taken each year from the lake and stored for the summer, though its purity was questioned by the investigating engineer.

B. Use of Environmental Tracers

In this study, apparent ages of water in two hotel wells were used to justify the use of previously collected well data and to consider the nature of flowpaths in the watershed. In this appendix, background and methods of age dating are discussed.

An environmental (also known as anthropogenic or historical) tracer is one that is already present as a result of human activity in such a way that its concentration fluctuations can be correlated with known dates. It is distinct from an artificial tracer, which is applied to the system by the investigator. The ideal environmental tracer would be present in equal atmospheric concentrations the world over, have a steadily changing rate through all time, be consistently soluble in precipitation independent of factors such as temperature, not react with any substance once it enters the ground surface, and be easily detectable. Unfortunately, these conditions are often not reality. Therefore, an understanding of each tracer's limitations guides the choice of an appropriate method.

The most commonly used historical tracer in groundwater studies for many years has been tritium (^3H), which was released at maximum concentration with thermonuclear testing in the late 1950's and 1960's. The point of maximum tritium concentration in a groundwater could be correlated with the date of that water's last contact with the atmosphere. However, with a half-life of 12.43 years and dispersion and diffusion effects, as time passes fixing tritium levels to a date requires increasingly more detailed vertical sampling, and is effectively no longer a good option in the southern hemisphere, where levels were 10 times smaller than North American levels to begin with. [Cook and Solomon, 1997b]

Production of industrial chlorofluorocarbons (CFC-11, CFC-12, and CFC-113) increased steadily from the 1950's to the 1990's, and unlike ^3H values that vary depending on initial proximity to testing sites, CFC concentrations have been similar worldwide at a given time. Given a known CFC solubility that is a function of recharge temperature and a measured CFC concentration in water, the resulting CFC partial pressure is correlated with known levels through time and a recharge date can be put on the sample. However, since the worldwide ban on CFC's in the 1990's, atmospheric CFC levels are dropping slowly and will continue to do so over the next several decades, thus introducing large uncertainties into CFC-dating groundwater that was recharged subsequent to the late 1990's. CFC-12, which was still the most distinguishable through the late 90's, still had a large uncertainty that is carried over from estimating recharge temperature. Due to sorption of CFC's in high-organic carbon environments, this method is not appropriate for dating flow from carbon-rich residential or industrial waste sites, although different types of CFC's sorb and degrade at different rates and a suite of measurements should be made in order to evaluate these effects. Corrections for the presence of excess air also have to be introduced. [Cook and Solomon, 1997a]

Since ^3H decays to ^3He , as soon as groundwater enters the saturated zone and the ^3He cannot escape back into the atmosphere, the ratio of ^3H to ^3He concentration can be used as an indicator of groundwater age. This indicator does not require detailed vertical sampling because the location of the bomb peak is not important. The ^3H clock starts when the water enters the ground and the ^3He clock starts when the water enters the saturated zone. If the residence time in the unsaturated zone is large, some error may be introduced. However, if the peak is also located, it is possible to calculate this residence time. Possible sources of error are gas stripping that may occur during sampling, sensitivity to initial ^3He solubility as a function of recharge temperature, excess air bubbles, and radiogenic production by alpha decay of U/Th series. These other sources must be subtracted out of the measured $^3\text{H}/^3\text{He}$; doing so leads to uncertainties of 10% or less when analyzing waters that are less than 15 years old, which makes it a good method to date younger groundwater. [Cook and Solomon, 1997a]

Sulfur hexafluoride is another compound that can be both natural and anthropogenic and is present in water in very low, but measurable, concentrations, and has been increasing at a constant rate for the past 35 years. A detection method has been developed that can resolve less than 0.01 femtomols per liter and it is hoped to be a good replacement for CFC dating for younger waters. [Plummer *et al.*, 2001]

Heavy stable oxygen isotopes are found in higher concentrations in rainwater that condensed at higher temperatures. Thus, seasonal variations in heavy oxygen isotope ratios present in rainwater can be carried into groundwater. If water has a residence time less than the seasonal changes in input, this signal can be discerned. Methods have been developed where even for longer residence times, the degree of signal dampening is itself used as a marker for residence times. [Plummer *et al.*, 2001]

The USGS sampled W1 and W2 for major ions, chlorofluorocarbons, sulfur hexafluoride, tritium/helium-3, and stable oxygen and hydrogen isotopes in 2000 as part of a statewide aquifer susceptibility study. The equations, corrections, and field methods used in the calculations of apparent age used in this study are described in [Nelms, 2000]. In both W1 and W2, concentrations of CFC-11, CFC-113, and tritium to hydrogen-3 ratios were used to obtain the apparent ages. Apparent age is defined as the time elapsed since water was isolated from the atmosphere and/or air in the unsaturated zone during recharge, and does not take into account effects of sorption, biodegradation, or mixing of tracers along the flowpath [Plummer *et al.*, 2001]. Though measured as part of the study, concentrations of SF_6 were found to be higher than water that is at equilibrium with modern American air, and thus the sample was considered contaminated and the water sample could not be dated using this method. The samples were also found to be contaminated with CFC-12. Oxygen isotopes would require frequent sampling that would show a seasonal signal, if one was still discernable, in order to age-date the water.

Environmental Tracer	Apparent age (years)	
	<i>Well 1</i>	<i>Well 2</i>
CFC-11	12.6 ±1.5	16.6 ±1.0
CFC-12	contaminated	contaminated
CFC-113	15.6 ±0.5	13.6 ±0.5
SF6	contaminated	contaminated
Tritium/Helium-3	15.3 ±0.2	error
AVERAGE	14.5	15.1

Table 6: Apparent ages of MLH well water

C. Datum for Full Lake

In this study, “full lake” is measured from the bottom of the rectangular notch in the spillway behind Newport House. An elevation of 1184 masl has been assigned to this point as a local datum for the purposes of this study, but its true elevation is uncertain. The USGS Newport quadrangle [USGS, 2009] places a benchmark at 1184.76 m (3887 ft.) near the north end of the lake, but the marker has not been located to the knowledge of this author by anyone who has worked on this problem. The true location of the Pond Drain stream as it exits the lake is some 30 m further north than its placement on the USGS map (Figure 10), although the USGS map could be showing an original stream location before the spillway and/or road were constructed. The USGS map shows a lake elevation of 1181.1 m (3875 ft), but the date and time of this level were not recorded. A 2-foot contour map made by an engineering firm on file at the Virginia Department of Environmental Quality shows an elevation near the spillway of 1182.01 m (3878 ft.), but the date of the document is unclear because it also shows a currently nonexistent road that appears only on older maps. No single altitude reading from the differential GPS could be trusted to provide a benchmark under the canopy at the edge of the lake. 1184.01 m was the level of the base of the spillway read from the bathymetric map, which was rounded to an even 1184.00 m and taken as a reasonable value for full lake based on the combined interpolation of the USGS Newport Quadrangle and the differential GPS data. All lake level measurements were referenced to points on the lakebed, which in turn were surveyed using the spillway as a local datum. The most significant source of error that derives from this problem is likely the calculation of the lake volume of the topmost cut-and-fill layer of the lake. If a water balance is ever calculated with a full lake, the lake perimeter could be surveyed with a differential GPS when canopy is absent and these values are substituted for the values taken from the Newport Quadrangle.

A prior measuring system put in place by Tom MacAvoy (Virginia Tech Entomology) marks intervals in feet below the spillway notch with sections of rebar. Some of these markers, however, have been moved and may need surveying and repositioning.

D. Measured Lake Depth

It should also be noted that the maximum depth of the lake as measured by this survey is only 29.33 meters. Previous reported depths have been measured with plumb lines, pressure gauges used by scuba divers, and sonar. Due to the very small area of the deepest holes, it is possible that the sonar did not reflect off the deepest parts of the crevices; it also possible that previous plumb line measurements suffered from some horizontal drift in the line, exaggerating the measurement; this problem was encountered in this study as well in plumb line measurements aimed at confirming sonar depths. It is also possible that sedimentation has caused infilling of deep crevices and the lake depth therefore does not remain constant over

time. Any additional depth in crevices, however, adds very little volume to the lake, and should not significantly affect storage calculations.

E. Lake Volume Calculations

In the event that lake volume calculations need to be reprocessed with more accurate perimeter data (as would be the case if the lake fills and overflows the spillway), an outline of the method for performing a water balance is given in Figure 49 when there is flow over the spillway. The steps used in ArcGIS® are depicted below.



Figure 49: Interpolation process in ArcGIS®

F. Barometric Correction of Pressure Transducers

Barometric compensation was required in this investigation because both pressure loggers, the In Situ and Solinst probes, measure absolute pressure. Both loggers require a barometric pressure sensor whose readings must be subtracted from the logger data to acquire the necessary water pressure. Two *methods* of barometric correction and three *sources* of barometric data were available, and comparisons were made to evaluate both methods and sources. The chosen method and sources of data for barometric correction are listed below for the benefit of future researchers using this equipment.

The Solinst-recommended procedure [Solinst, 2009] was used to manually compensate both the In-Situ and the Solinst transducers for barometric correction. Although an automatic procedure is available in the software for the In-Situ barometric logger, it was found not to be equivalent to the compensation recommended by Solinst. Since the Solinst procedure is well-documented and all other transducers used in the study are Solinst, all barometric compensation was processed as documented in the Solinst technical notes. It should be noted that care must be taken when compensating for the unpublished but constant zero-point offset in the In Situ equipment.

Barometric data sources were chosen for reliability, location, and sampling rate. Of the three sources of barometric data (Table 7), the In Situ BaroTroll data was used from May 2009-May 2010. An offset of 4 mmHg was required to make the In Situ data match the Solinst data during a period of overlap, and as the In Situ barometric logger says that the factory calibration has expired, the offset was applied to the In Situ barometric data. For dates after May 2010, the MLH CS100 data was used as the source of data, given the quality of the sensor, hourly sampling rate, and its proximity to the sites for which a barometric pressure correction was needed. Note that the MLH sensor returns pressure that is normalized to sea level, which needs to be transformed to absolute pressure using the pressure differential [CS, 2010] given by

$$dP = 1013.25 \left(1 - \left(1 - E / 44307.69231 \right)^{5.25328} \right)$$

where dP is the pressure differential in mbars and

E is elevation in meters.

Instrument	Location	Altitude	Measurement frequency	Start date	Reference
CS100	ML Hotel	1201 m	Hourly	May 2010	Normalized to sea level
In Situ BaroTroll	Lakeside	1185 m	Daily	May 2009	Absolute
CS100	MLBS	1173 m	Half-hourly	1994, but gaps in data exist	Absolute

Table 7: Sources of barometric data

G. Additional Resistivity Profiles

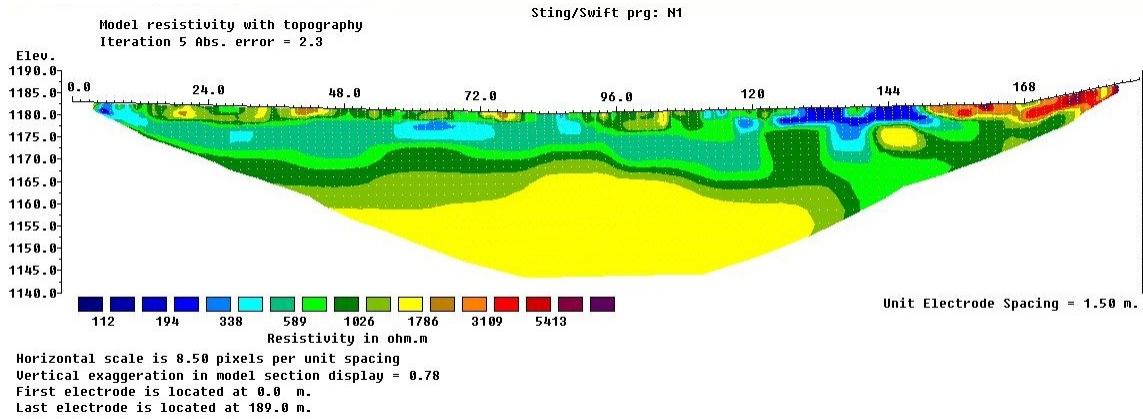


Figure 50: Resistivity profile N1

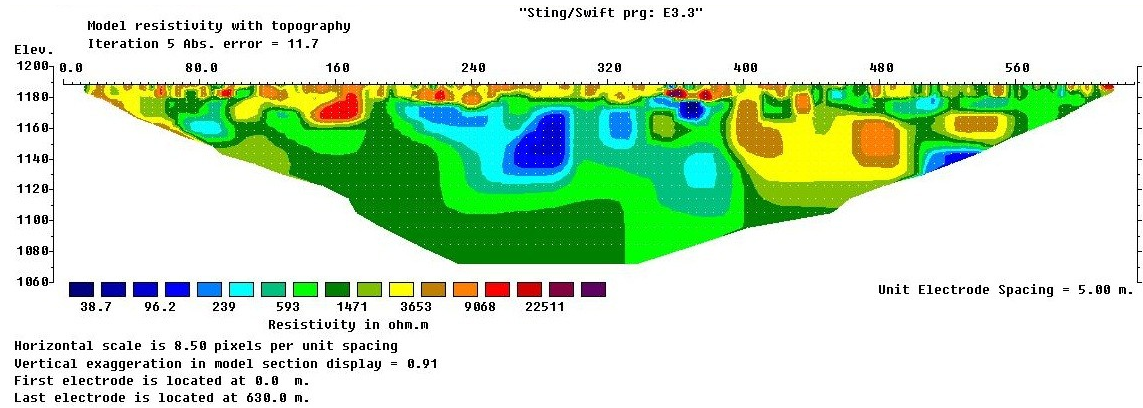


Figure 51: Resistivity profile E3

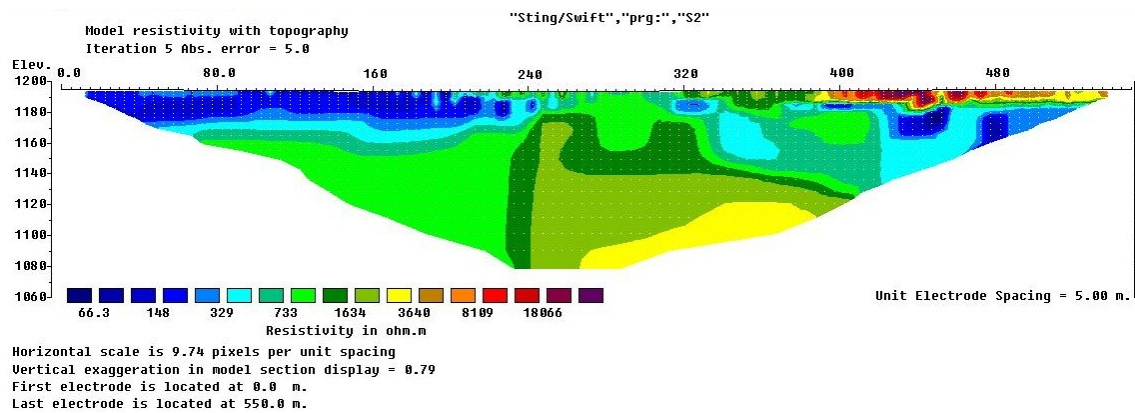


Figure 52: Resistivity profile S2

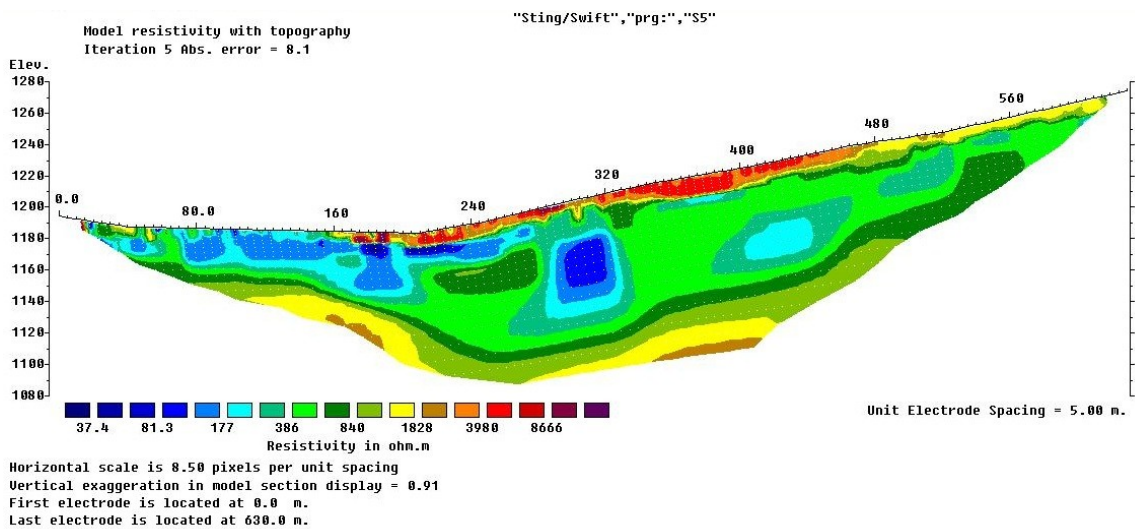


Figure 53: Resistivity profile S5

H. Chemical Sampling Data

Location	Abbr.	Date of collection	Charge balance* *(%)	pH lab (field)	Conductivity (μS/cm)	Temp. (C)
Well 1	W1	7/20/2000		6.2	87	10.1
Well 1	W1	5/9/2000	6.3	*	*	*
Well 2	W2	7/20/2000		6.5	129	9.2
Well 3	W3	5/9/2008	4.5	*	*	*
South Spring	SS	9/4/2010	13.9	6.2 (6.0)	25	17.5
East Spring	ES	9/4/2010	14.4	6.2 (5.6)	20	13.7
Lake, shallow	LS	9/4/2010	-4.4	7.1 (7.7)	57	22.9
Lake, deep	LD	9/25/2010	8.5	6.7	-	-
Horse Barn Well	HB	9/8/2010	27.7	4.0	34	11.5
Sartain Creek at 4000 ft.	S4000	11/6/2010	8.9	4.9	-	-
Sartain Creek at 3500 ft.	S3500	11/6/2010	-2.7	4.5	-	-
Sartain Creek at 3200 ft.	S3200	11/13/2010	-15.8	4.5	-	-
Pond Drain, Branch	PDBch	11/15/2010	19.9	5.8	-	-
Pond Drain,	PDBd	11/15/2010	17.1	6.0	-	-

Bridge	g					
Pond Drain, South	PDS	11/15/2010	12.7	6.0	-	-

-Not measured or reported.

*Measured as stability criteria only for well purging, not reported.

**Charge balances for samples with total dissolved solids (TDS) above 15 mg/L are acceptable at 4-6%; these correspond to the lake and most wells. Balances for samples with TDS below 15 mg/L are above desirable limits but given the low total dissolved solids are still acceptable at 13 to 20%. The Horse Barn Well balance was an exception to this with a low TDS but a high charge balance at 28%.

Table 8: Water samples