

### **3.2.4 Geology**

#### **Introduction**

In a landscape that has not been changed by human activities, the streams of a region reflect the climate, geology, and biology of that region. The Catskill high peak region has higher rain/snowfall amounts than the western and northern Catskills, and as a result, for a given watershed drainage area the streams are generally larger in the High Peaks than elsewhere in the Catskills (Miller and Davis, 2002). Likewise, the large amount of forest cover in the Catskills affects the amount of rain and snowfall that will run off the landscape to become streamflow, and therefore the shape and size (morphology) of the stream channel required to handle the amount of runoff (see Section 3.1.5 for more information about the role of riparian (streamside) vegetation). Similarly, the geology of the Catskill Mountains exerts a clear influence on the landscape and stream valley and channel morphology. This section describes the basic geology of the Catskill Mountains, what we know about the geology of Broadstreet Hollow, and finally, how this affects the stream channel morphology, or fluvial geomorphology, of the basin.

#### **3.2.4.1 Catskill Mountain Geology**

The Catskill Mountains are a dissected plateau of mostly flat-lying sedimentary rocks cut into by streams and ice flow over millions of years. The mountains are at the northeastern extreme of the Alleghany plateau, a physiographic province (a land area with fairly uniform physical characteristics) that extends from Tennessee along the western border of the Appalachians (Rich, 1935).

There are many descriptions of the boundaries of the Catskills (Rich, 1935; Thaler, 1996; Isachson et al, 2000). A useful definition is Rich's (1935) description of the escarpments that comprise this mountainous region: Northeastern Escarpment (Blackhead Range); Eastern Escarpment (Wall of Manitou); the Central Escarpment (Indian Head to Utsayantha); and the Southern Escarpment (Slide Mountain to Ashokan High Point) ([Figure 1](#)). Broadstreet Hollow is located along the Central Escarpment, draining the south slope of West Kill Mountain and part of North Dome Mountain. The geologic discussion that follows pertains to the Central Escarpment.

*Insert Figure 1. here*

#### *Bedrock Geology*

The bedrock of the central Catskill Mountains is comprised of conglomerate, sandstone, siltstone, mudstone and shale. The sediments that form the middle-to-late Devonian (390 to 360 million years ago) bedrock are interpreted to be deposits of a vast deltaic river system, often called the "Catskill Delta" deposits (Isachson et al, 2000). The sandstone and conglomerate are made up of river channel deposits, while the siltstones and shales are overbank and shallow fresh water deposits. The "red beds", or mudstones, are often paleosols (ancient soils) that record the presence of Devonian forests. The repeating sequence of deposits shows continuous aggradation (building up) of the channels and floodplains of a complex meandering (many bends and turns) river system.

The "Catskill Delta" deposits were buried beneath other sediment, then uplifted as a plateau during the Alleghanian Orogeny (mountain building, 330 to 250 mya). When bedrock buried at great depth is subjected to stress (e.g. related to plate tectonics such as mountain building events)

the rock can fold or fracture. In the case of the Catskill rock, intersecting sets of vertical fractures, called joints, formed during and after the Alleghanian Orogeny. The following eras eroded away the overlying rock, and streams cut multiple channels into the slowly rising plateau. The structure of the uplifted bedrock facilitated erosion and stream drainage development.

### *Surficial Geology*

The cyclic ice ages of the last 1.6 m.y. (Pleistocene Epoch) have left the latest indelible mark on the already incised landscape of the Catskills. Vast continental ice sheets and smaller local alpine and valley glaciers scoured the mountains. The last ice sheet reached maximum thickness over the Catskills during the Wisconsinan Stage of the Pleistocene, about 21,750 years ago (Isachsen, et al., 2000). This was a period of accelerated erosion from the abrasion (scraping of rock-on-rock) and bedrock quarrying by the flowing ice. Glacial erosion broke the rock down into an entrained mixture of fragments ranging in size from boulders to clay. This mixture of sediment was carried along by the ice and deposited as till (unsorted assemblage of glacial sediment) or as stratified “drift” if the sediment was subsequently sorted by melt-water streams. These glacial deposits filled in deep river ravines that drained the landscape before the glaciers advanced over the mountains.

As the ice thinned, the landscape was deglaciated – lobes of ice sheets melted back from the central Catskills, and from mountain glaciers formed on some of the newly exposed peaks (e.g. Hunter and West Kill Mountains). Meltwater along side the decaying ice left a complex array of

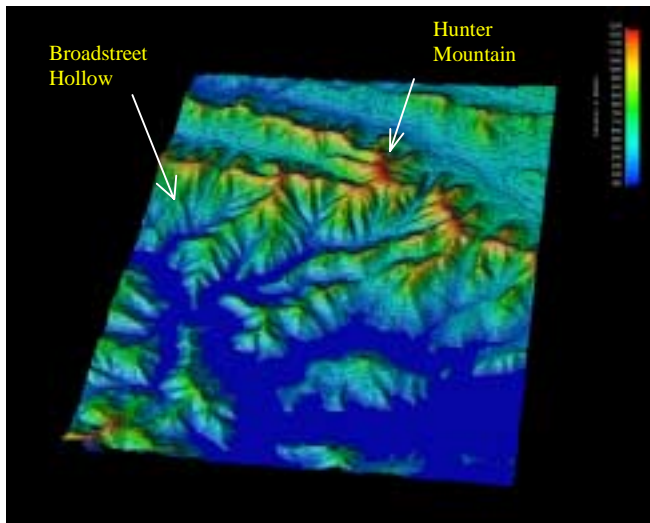


Figure 2. Model of glacial Lake Shandaken filling the Esopus Creek Valley. Water elevation is based on mapped lake deltas (created by Dominique Thongs, NYCDEP)

meltwater and ice-contact deposits, along with lake clay where moraines (deposits at glacial margins) and ice impounded water, forming glacial lakes that filled the valley floors (Figure 2). As climate fluctuated during the period of deglaciation, temporary re-advances of ice would leave till and other meltwater deposits on top of the earlier glacial material, resulting in the complex lateral and vertical distribution of glacial deposits observed today. For more detail on the glacial geology of the Catskills the reader is referred to Rich (1935), Cadwell (1986), and Titus (1996).

After the ice fully retreated north, rainfall-runoff returned as the predominant sculptor of the landscape. The Holocene (the last 12,000 years or so) has been a period of reclamation of the mountain landscape by forests and animals that together exert control on the stream valley and channel geomorphology. People then repopulated the landscape, and in the last 300 years humans have had the most dramatic effects on stream system morphology and stability.

### 3.2.4.2 Broadstreet Hollow Geology

The geology of Broadstreet Hollow valley has been included in several comprehensive investigations of Catskill bedrock and glacial geology (Rich, 1935; Cadwell, 1986; Willis and Bridge, 1988) as well as the subject of smaller investigations by NYCDEP. Also, whenever a water supply well is drilled the driller records a log of the underlying material encountered, layer by layer, providing additional data for interpreting the local geologic history. The following discussion is an abbreviated account of Broadstreet Hollow geology based on available information.

#### *Bedrock Geology*

Rickard (1975) mapped the bedrock geology of the area as part of the New York State Geological Survey Map and Chart Series (Figure 3). The lower portion of the valley, below approximately 1400 ft, is mapped as the Oneonta Formation. This formation is expected to be around 900 ft thick in the Broadstreet valley. Most of the water supply wells in the valley are probably withdrawing water from the fractured sandstones and siltstones of the Oneonta Formation.

The overlying Lower Walton Formation makes up most of the valley floor and walls, while the ridges are capped with the Upper Walton Formation. The Walton Formations consist of greater than 1,000 ft of red beds (shales and mudstones), gray sandstones and small amounts of gray shale (Fletcher, 1967). The uppermost beds are conglomeratic sandstones that grade upward into the overlying Slide Mountain Formation that is comprised of a yellowish-gray conglomerate.

*Insert Figure 3 here.*

#### *Surficial Geology*

Rock outcrops are rare along Broadstreet Hollow Road and even in the stream itself. The only rock observed along the road is an outcrop of the Oneonta Formation about 0.8 miles from Route 28. Similarly, a walk along the mainstem of the stream within the Management Area (approximately 3.5 miles of the main stream extending up from the Esopus Creek) reveals no evidence of bedrock in the valley bottom. This is because the valley is thick with glacial sediment, masking the bedrock geology.

Surficial geology (deposits and landforms) has been mapped by Rich (1935) and Cadwell (1986) (Figure 4). Significant glacial landforms in the valley include the moraines and delta that form some of the steep topography of portions of Jay Hand Hollow. The Jay Hand Hollow issues from the U-shaped notch of Mink Hollow (between West Kill ridge and North Dome), interpreted by Rich to be an important conduit of ice flow into Broadstreet valley. Along the mainstem Broadstreet Hollow stream, probable kame terraces and moraine features locally extend out from the valley bedrock wall, confining the lateral extent (available width) of the stream valley floor. This confinement occurs most notably just below the confluence of the two main streams (see also MU4) and the section between the Ulster/Greene County line and the David Merwin bridge (see also MU 6-12). The distinctive meander bends near the bottom of the hollow (see also MU12, 13 and 15) appear to be controlled by similar glacial deposits.

*Insert Figure 4 here.*

Along roadsides, hillsides and stream banks, the exposed geology is mostly glacial till, kame terrace (ice contact) deposits and Holocene alluvium (carried by post-glacial streams). In several places along the main stream, pure clay is observed in the stream channel and banks (Figure 5). The layered clay is “glacial lake clay” from minor ice-contact impoundments within the valley during deglaciation, and from a glacial lake that filled the Esopus valley as the ice sheet melted back to the Hudson Valley (Rich, 1935; Cadwell, 1986; Dineen, 1986).



Figure 5. Eroding glacial lake clay along the stream bank and valley wall, across from the Fischzang property (see also MU17, monitoring cross-section 3).



Figure 6. Eroding kame terrace (ice contact deposit) composed of silts, sands, gravels and boulders along the stream bank across from the Foss property (see also MU8, monitoring cross-sections 11, 12 and 13).

Ice contact deposits (e.g. kame terraces), composed of variably stratified (layered) silts, sands, gravels and boulders are exposed in mass-wasted slopes of some steep, terraced valley walls (Figure 6). Often, mass wasting is facilitated by the presence of localized clays within or underlying the more permeable ice contact deposits, forming water lubricated slip surfaces resulting in destabilized hillslopes. The glacial stratigraphy of the valley is complex, and the location and severity of stream network incision into it is influenced by this complexity.

Rich noted, back in 1935, exposures of lake clay veneered (covered in a thin layer) with till at the mouth of Jay Hand Hollow, which he interpreted as indicating considerable fluctuation of the ice margin (leading edge) in the valley while lake waters were present. He also noted slumping and land-sliding, attributable to the presence of “considerable” lake clay, along the entire portion of the valley below the confluence of Jay Hand

Hollow and the main Broadstreet Hollow (see also MU4 through 19). Similarly, during the course of the watershed assessment in 2001, and recent observation of an excavated foundation, laterally discontinuous exposures of lake clay were revealed to be in abrupt lateral contact with coarse alluvium (gravels to boulders) and/or glacial till. A preliminary review of the available well logs and geotechnical borings (Appendix Section 6.1) provides further evidence of both laterally extensive and highly localized clay deposits in a spatially complex arrangement. The laterally extensive clay deposits are probably attributable to glacial Lake Shandaken (Rich, 1935; Cadwell, 1987) while the localized clay deposits (lenses) resulted from isolated surface water

impoundments within the valley during deglaciation. In some places, subsequent re-advance of glacial ice has deformed and rearranged the layered clays and surrounding deposits.

The soils of Broadstreet Hollow are formed from glacial deposits and bedrock. The Ulster County Soil Survey provides maps and detailed descriptions of the soil types observed in the valley. A brief summary of the soil types and a copy of the soil maps are available at the Ulster County Soil and Water District Offices.

### **3.2.4.3 Applied Geology**

How does geology relate to management of the stream? Bedrock and surficial geology of a drainage basin (watershed) influence all aspects of the stream system, though bedrock geology plays a limited role in the Broadstreet Hollow management area. Bedrock geology plays a more significant role in the upper headwater reaches, where less active stream management is required (i.e., primarily in NY State owned lands).

Glacial geology sets the framework for the Broadstreet Hollow stream valley, controlling such characteristics as depth of alluvium (water worked sediments), sediment supply and stream channel slope and geometry. Understanding the geology of the Broadstreet Hollow valley can help identify causes of stream erosion and water quality problems as well as assist in prioritizing where future stabilization/restoration actions will be most useful. The discussion below is separated into four general categories that pertain to the stream system: (1) *hillslope process* – slope stability and sediment supply; (2) *groundwater and surface water hydrology* – base flow conditions and runoff characteristics; (3) *valley and channel morphology* – valley type and orientation, channel shape and size, stream bed form, and bank material; and (4) *water quality* – base flow chemistry and temperature; suspended sediment, and contamination.

#### *Hillslope process*

The stream within the designated management area (primarily the main stream mapped in the watershed assessment in 2001) is entirely incised into glacial deposits overlying bedrock, and the adjacent hillslopes are formed in glacial deposits. Where layered clay is present in the hillslopes, clay-sourced slope instability adjacent to the stream can cause stream bank and bottom instability. A prime example of slope instability in response to presence of extensive lake clay deposits is the story of what has occurred at the stream stability restoration demonstration project site, located above the confluence with the Jay Hand Hollow tributary (see also MU3, and Section 6.1 for project descriptions). Appendix in Section 6.1 contains the geotechnical (slope stability) report for the project site as well as a detailed description of the site geology/hydrogeology (Rubin, 1997).

Geologic borings (vertically drilled samples) and well logs (Appendix Section 6.1) reveal that clay is present and thick throughout the stream valley in multiple Management Units, and particularly so at the project site in MU3. The January 1996 flood event and subsequent emergency flood repair work removed the thin cobble and boulder stream bed armour overlying the soft, easily erodible clay – producing a short, sharp drop, or headcut, in the stream bed. The stream swiftly incised (cut downward) into the clay, which lowered the stream bed, putting greater stress on an already failure-prone hillslope along the north side of the stream. The hillslope failure was characterized by slumping within an active, well-lubricated, gravity slide

zone within and along the base of the thick glacial lake clays present in this location forming the valley wall (in places >30 ft thick). The resulting soil fault scarps (large, long cracks in the hillside) marked the progress of slumps of saturated clay toward and into the stream (Figure 7).

*Insert photo or Rubin sketch of fault scarps at project site.*

The clay also formed a confining layer causing artesian conditions (ground water under pressure, forced to the surface) in the underlying, more sandy material. This artesian condition extended under and into the stream bed, forming a “mud boil” in a zone where excess water pressure forced water and suspended clay out of the stream bed, resulting in a constant source of suspended sediment and turbidity, even at low flow conditions (Figure 8).



Figure 8. Artesian "mud boil", early summer, 2000, showing clay hillslope on right bank, mid-MU3. Stream flow is from right to left.

Similar soil fault scarps and associated slumps of varying age observed elsewhere in the basin show that this complex hydrogeologic/hillslope condition is pervasive and has been going on for a very long time within the Broadstreet Hollow basin (see eroding banks and hillslopes at monitoring cross-sections in MU5, 8, 14, 15 and 17). In fact, this phenomenon is observed in many of the Esopus Creek drainages (Rubin, 1997).

Eroding ice contact deposits, and secondarily, till, are significant localized sources of sediment supply to the stream. The poorly consolidated, variably-sorted ice contact deposits that form steep embankments along the eastern valley wall can readily erode, especially if clay is present near the toe of the slope (as in Management Unit 8; Figure 9). These deposits supply clay to boulder sized material. Bar development (large stream sediment deposits) and boulder clusters often form associated with these sediment sources.

### *Groundwater*

Where the stream flows through bedrock, base flow (the lowest stream flow, typically in summer or drought conditions, and primarily from ground water) is limited to discharge from the fracture network system (FNS). Base flow to streams throughout most of the Broadstreet Hollow valley is stored in the heterogeneous array of glacial deposits. The result is variable baseflow, water quality, and thermal refugia conditions for fish (protected cold water locations necessary for fish survival in hot summer



Figure 9. Glacial lake clay deposit at the toe of the slope across from the Foss property, monitoring cross-section 11 in MU8.

months).

Presence of clay near the surface, capped by coarse permeable deposits, causes groundwater to perch and flow near the surface with numerous springs, wetlands, and minor tributaries as a result. The presence of clay as a top confining layer also produces local artesian conditions (as noted above). When the clay has been breached (by fault formation or digging) groundwater can rise to, or near, the ground surface forming springs. There are many shallow “dug” wells in the valley that tap into these perched or shallow artesian “aquifers”.

#### *Valley and Channel Morphology*

Valley orientation is based on pre-glacial erosion controlled by the fractured bedrock. Most of the stream valleys draining the Central Escarpment are oriented NE-SW, bisecting the two predominant fracture orientations. Glacial landforms, such as kame terraces and moraine features (Rich, 1935) determine the valley floor confinement and cross-sectional configuration (Figure 10).

*Insert Figure 10 here (valley cross-section with labeled features)*

Modern stream deposits in the Catskill Mountains are principally derived from erosion of well-bedded sedimentary bedrock of the Catskills. As a result, stream clasts (sediment particles and classes) have a low sphericity (“roundness”), typically forming platy or disk particle shapes, which strongly influence the imbrication of the stream substrate (stacking of stream deposited sediment, often forming a pattern like fish scales) and the magnitude of flows required for mobilization.

The complex Pleistocene glaciation of the Broadstreet Hollow valley has significantly modified the landscape and left varying deposits of clay-rich to bouldery till, silts, sands, gravels, and cobbles of meltwater streams and ice-contact deposits, and easily erodible glacial lake clays. The result is often a significant downstream variation in sediment supply (amount and size of material) and channel boundary resistance (the “roughness” of the channel, and sediment resistance to moving in high stream flows). The architecture of the stream channel (size, shape, bed form) is influenced to a large extent by the surficial (surface) deposit material size. Variable distribution of particle size in stream eroded and mass-wasted glacial deposits (i.e., the different character of bouldery melt-out till vs. clay rich lodgement till or cobble-gravel meltwater deposits) results in variably distributed bouldery step-pool channels interspersed with cobbly riffle-pool channels.

#### *Water Quality*

During much of the year, stream water chemistry is dominated by groundwater base flow. Water quality and temperature is expected to vary, as a function of the material the groundwater is flowing through and the retention time (amount of time water is in contact with underground material before flowing to the surface as springs or as stream flow). At this time there are no surface water or groundwater chemistry data available to characterize the system. Where lake clays cause groundwater to perch beneath very permeable deposits on top, there may be limited treatment of septic leachate from older septic systems adjacent to the stream.

Exposed clay deposits in the stream bed, banks and eroding hillslopes provide source areas that episodically contribute high suspended sediment loads to the stream. Preliminary sediment load estimates for four 1996 storms indicate that the Broadstreet Hollow stream contributed 4%, 6%, 9%, and 1%, respectively of the total sediment load in the Esopus Creek measured at Coldbrook, having less than 5% of the total contributing drainage area (Rubin, 1997). Small exposures of this easily entrained material can significantly increase downstream turbidity. As noted before, under certain hydraulic conditions, “mud boils” can form in the channel bottom which literally pump clay into the stream even at low flow conditions. There were no known “mud boil” conditions in 1997, when storm turbidity measurements were made. There are no known mud boil present in the stream at the time of the preparation of this document.

### **Recommendations**

Given the importance of the geology on the stream morphology and water quality, a more detailed map of stream corridor geology is recommended for future planning and identification of potential stream instabilities. The map can be prepared by combining GPS recorded observations of geology along the stream corridor and available subsurface data.

## **3.3 Hydrology and Flood History**

### **Introduction**

The water that flows through the Broadstreet Hollow stream into the Esopus Creek reflects the integrated net effect of all the watershed characteristics that influence the hydrologic cycle. Such characteristics include the climate of the drainage basin (type and distribution patterns of precipitation), the geology (permeable or impermeable surfaces affecting timing and amount of runoff), and the vegetation (uptake of water by plants as well as protection against erosion). All these factors affect the timing and amount of streamflow in the stream. We refer to this variable timing and amount as the stream’s hydrologic regime. Understanding the hydrology of a drainage basin is important to the stream manager, since stream flow patterns affect aquatic habitat, flood behavior, recreational use, and water supply.

BSH Stream Statistics The streams in BSH are perennial streams, that is, they flow year round except in extreme drought conditions. The drainage pattern in the watershed is dendritic (branching tree-like form), typical of the Catskill Mountain watersheds.

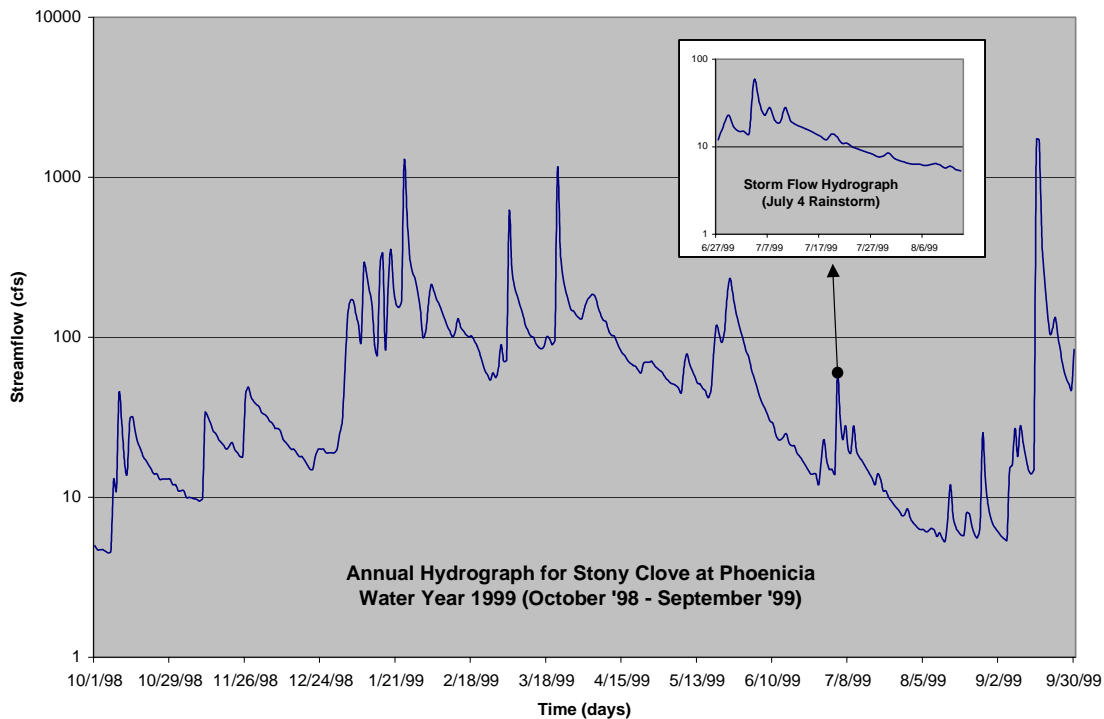
The BSH streams form a system of streams that can be classified by stream order. A stream order identifies the position in a hierarchy of tributaries occupied by a stream segment. Any clearly defined channel without tributaries is designated as a 1<sup>st</sup> order channel. Where two 1<sup>st</sup> order channels join they form a 2<sup>nd</sup> order channel; where two 2<sup>nd</sup> order channels join, a 3<sup>rd</sup> order channel is formed, and so on (Strahler, 1968). Most of the BSH watershed drains into small headwater 1<sup>st</sup> order channels. The main Broadstreet Hollow and Jay Hand Hollow are 2<sup>nd</sup> order streams, below their confluence is a 3<sup>rd</sup> order stream.

### **Streamflow**

Streams flow at many different levels from a small trickle during a dry summer to a raging torrent during rapid thaw of a thick snowpack. Streamflow varies on several temporal scales. Over the course of a year we can watch the stream swell and shrink with the seasons. Over the course of a summer storm (hours to days) or a spring thaw (days to weeks) we can also watch the

stream swell and shrink. Hydrologists use a hydrograph of a stream, a graphical representation of the streamflow over some period of time, to characterize the relationship between flow and timing. It is necessary to have a stream gage to monitor stream discharge to develop a hydrograph. Unfortunately BSH does not have a suitable gage to prepare a hydrograph for the stream. For illustrative purposes we can consider hydrographs for the adjacent basin of Stony Clove (Figure \_).

There are essentially two basic types of streamflow: storm flow and base flow. Storm flow appears in the channel in direct response to precipitation and/or snowmelt, whereas base flow sustains streamflow during interstorm or subfreezing periods. Figure \_ is an annual hydrograph that shows the peaks and lows of streamflow over the course of the year. The “rise and fall” of the peaks are associated with storm flows and the lows are generally associated with base flow conditions. A review of the hydrograph reveals that the 1998/1999 winter was a particularly wet period. Streamflows are also higher during this period since the vegetation is no longer taking up much of the flow as can be seen to occur during the summer months. The insert graph in Figure \_ is a close-up of one of the storm flow events in the Stony Clove watershed. At the end of June there was a small rainfall event that brought the streamflow up from the summer baseflow conditions. Storm flow receded back toward base flow conditions prior to the July 4, 1999 rain storm. The response of the streamflow to the July 4, 1999 rainstorm is rapid, presumably due to antecedent conditions from the previous rainfall. Minor rains followed but eventually the landscape drained and the streamflow returned to summer base flow conditions.



Storm flow reaches the channel as channel interception, overland flow, or subsurface storm flow. Channel interception is simply the precipitation that falls directly into the water that is already in a stream channel. It shows up on the storm water hydrograph right away and disappears as soon as the event is over. In a small watershed like BSH, it is a minor component of the streamflow. Overland flow, or surface runoff, is the portion of precipitation or snow-melt that runs off over the surface after the infiltrative capacity of the surface has been exceeded. Overland flow is generated more consistently on some parts of a watershed and at some times rather than at others. Relatively impermeable areas (exposed bedrock, frozen ground, clayey-soils) will generate more runoff to the stream than relatively permeable areas (deep, coarse soils) or well-vegetated areas. The role of overland flow in the BS watershed is variable, depending upon the time of year and location. Overland flow appears on the hydrograph pretty quickly and ends soon after the storm flow event. In the examples above overland flow and channel interception account for most of the sharp peaks. Subsurface storm flow, or interflow, comes from rain or snow melt that infiltrates the soil. It either flows rapidly through highly permeable portions of the soil or displaces existing water into a channel. Subsurface storm flow shows up on the hydrograph after the other two components and lags after as the flow in the stream declines back to base flow conditions. The distinction between base flow and subsurface storm flow is transitional.

The “bankfull discharge” is an important streamflow associated with storm flow events. This flow and its significance are discussed in greater detail in Section 3.1.2.

Base flow is the water that drains from the land to sustain streamflow during dry periods and between storm flows. The source is groundwater that flows through the unsaturated and saturated soils and bedrock adjacent to the stream.

### **BSH Flood History**

The flows that exceed stream channel capacity are called floods. Floods can range from minor overbank events to the raging torrents that wipe out bridges and carve new channels. Since there have been no historic stream discharge gages installed on the Broadstreet Hollow stream, we cannot report accurately on the major flood events in the basin. However, by evaluating the gage records at two nearby gages, and interviewing knowledgeable individuals from the area we can describe some major flood events and draw conclusions as to the nature of flooding in the valley.

The two nearby gages that have a sufficiently long enough record to evaluate flood frequency and occurrence are the Esopus Creek at Allaben gage and the Bushnellsville Creek at Shandaken gage. Table x. lists the floods recorded at these gages that exceed the 5 year recurrence interval, i.e. these are stream discharges that statistically could occur once every five years. A review of the dates shows that the recurrence interval concept can be misleading. During the 80’s four floods exceeding the “five year event” occurred within a seven-year span on the Esopus, while there were no five year, or greater, events during the entire decade of the 70’s. Another way to refer to these flood events is the likelihood of occurrence in any given year. A 5 year recurrence interval is the same as saying 20% of the years would have a 5 year flood. Thus there can be a cluster of years with large flood events and then a long period with only moderate flood events.

Table of Flood Flows at Nearby Gages that Exceed Five Year Recurrence Intervals<sup>3</sup>

<b>Esopus Creek at Allaben, NY (USGS Gage Number )</b>	
<b>Date</b>	<b>Flood Discharge (cfs)</b>
3/30/51	20,000
7/28/69	7,870
3/21/80	15,900
2/20/81	6,540
4/5/84	8,470
4/4/87	16,100
1/19/96	15,000
<b>Bushnellsville Creek at Shandaken, NY (USGS Gage Number )</b>	
<b>Date</b>	<b>Flood Discharge (cfs)</b>
11/25/50	1,350
10/15/55	1,830
3/21/80	845
4/5/84	896
4/4/87	1,000
1/19/96	996

Flooding occurs in response to excessive runoff associated with spring snowmelt, summer thunderstorms, fall hurricanes, and winter rain-on-snow events. Five of the seven major floods recorded at the Esopus Creek at Allaben station occurred in late winter/early spring and are presumably associated with major snow melt events from either spring thaw or rain-on-snow events. The largest recorded flood is a spring runoff event. A summer flood in 1969 and the flood of January 1996 are the two other large floods recorded at the gage. Three of the six major floods recorded at the Bushnellsville gage occurred during the spring and are coincident with three of the Esopus events. Two of the events (the two largest) are late fall events in the fifties and are associated with large tropical storms/hurricanes in the 1950s. The January 1996 flood was approximately a 10 yr RI flood on the Bushnellsville Creek. There have been many other floods that exceed the bankfull discharge event but are less than the 5 yr RI flood. A little less than half of all recorded flood events occur in the Spring. Approximately 10% of the flood events occur in the summer. The remainder of the floods is evenly split between winter and fall flood events.

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<sup>3</sup> Flood frequency statistics based on recorded peak flows through 1997.

Esopus Creek at Allaben, NY  
 5 yr RI flood: ~6,500 cfs  
 10 yr RI flood: ~9,500 cfs

Bushnellsville Creek at Shandaken, NY  
 5 yr RI flood: ~800 cfs  
 10 yr RI flood: ~1,000 cfs

From the review of the available data we can surmise that most of the events that are bankfull and greater will occur in the late winter/spring as the result of thaws and major rain-on-snow events. This is in large part due to the storage of available water as snow on the landscape, reduced infiltration capacity if the ground is still frozen (or partially so), and the fact that evapotranspiration from vegetation is not actively routing water back into the atmosphere. The major floods of the 1980s all occurred in this time period (late February – early April) most likely associated with considerable snow packs. For example, the largest of these floods occurred on April 4, 1987. Applications for stream disturbance permits typically peak following floods, as landowners attempt to repair the damage caused by the floods.

The 1990s were a time of moderate flood events in the vicinity of Broadstreet Hollow with the exception of the winter flood of January 19, 1996 which was similar in scale to the April 1987. Figure \_\_ shows the devastation of the January 1996 flood on the road in the upper reaches of Broadstreet Hollow. The Tropical Storm Floyd flood of September 1999 was typical of tropical storm events and the sometimes uneven distribution of precipitation associated with those storms. While the flooding in the Esopus drainages was typically less than a 5 year event, several drainages in the bordering Schoharie system had over a foot of precipitation in 24 hours with flooding that exceeded the 10 year event discharge. Summer thunderstorms are even more unevenly distributed across the Catskill Mountain landscape. The July 4, 1999 storm presented in Figure \_\_ only affected sub-regions of the Catskills. Flash flooding may be occurring in response to a sustained storm cell in some of the drainages that feed the Esopus Creek (e.g. Broadstreet Hollow), while only steady, lighter rainfall along the boundary of the storm results in a moderate increase in stream flow occurs elsewhere along the Esopus system.

The last three years (2000 – 2002) have been characterized by droughty conditions with intervening wet conditions. The high water events have typically been limited to bankfull events. It is hard to predict when the next 5 year (or greater) flood will occur in the Broadstreet Hollow. Given that the last large flood was back in 1996, chances are that another is due in the near future and that it will occur in the late winter/early spring.