THE GEOLOGY OF CLARKSVILLE CAVE, ALBANY COUNTY, NEW YORK

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SECTION A: Modification of Pre-Woodfordian Caves by Glacial Meltwater Invasion in East-Central New York

ABSTRACT

Periods of high glacial meltwater have altered some preglacial cave-passage configurations. Floodwater and fossil karst features, whose formation cannot be explained based on available water from the surrounding watershed, are found superposed on actively forming cave passages. These features may be recognized through correlation of watershed boundaries, peak-runoff observations through a cave system, the presence of anomalous in-cave and surface features, and the geomorphic interpretation of the area in question. Knowledge of minimum rates of karstification may be used to infer climatic conditions, making possible the reconstruction of the hydrology associated with deglaciation.

Clarksville Cave, situated in the hamlet of Clarksville, New York, provides an excellent example of invasion by Wisconsinan meltwater on a pre-Woodfordian cave system. Vadose development of a major part of the explored cave has occurred preferentially aslant a thrust-fault ramp, often along a calcite bed/limestone contact created by pressure solution. Other fault-related features include slickensides, extension veins, fault-bend folds, stylolites and the repeated basal Onondaga Limestone and impermeable Schoharie Formation thrust below the Onondaga Limestone stratigraphic column. An imbricate thrust east of the cave has upthrown the Esopus Shale against the Onondaga Limestone, forcing the development of an inefficient resurgence at the baselevel Mill Pond.

During the Wisconsinan glacial stage, subglacial meltwater formed a series of now abandoned bedrock channels and paleogorges that, due in part to topographic controls, found outlets along and over the flank of the Helderberg Escarpment. Some of this meltwater was pirated into Clarksville Cave where inefficient outlets resulted in the formation of higher in-cave "intermittent phreatic" levels not controlled by the thrust fault. These levels abruptly truncate and grade to lower vadose passages. The character of these upper levels, the paleogorge and related caves, and elevated paleo-insurgence points correlate with described alpine karst settings.

PHYSICAL SETTING

Clarksville Cave is nestled under the flank of a low wooded ridge virtually in the center of the hamlet of Clarksville, New York (Fig. 1). It is formed in the lower subunits of the Devonian Onondaga Limestone that were deposited approximately 380 million years ago. Its large passage size, up to 15 feet high and 40 feet wide, complete with multiple levels, makes it unique among other, usually smaller, Onondaga caves.

The Clarksville area lies at an elevation of 600 to 800 feet msl. It is situated within the foothills of the Helderberg Plateau, a part of the Appalachian Plateau physiographic province. Meyerhoff (1972) attributed the present day drainage pattern of this region to the normal erosive processes of stream adjustment to structure. The Helderberg Plateau has been modified by stream incision, physical weathering, glacial and postglacial erosion, and deposition during the Cenozoic era (Dineen, 1987).

In Garver, J.I., and Smith, J.A. (editors), Field Trips for the 67th annual meeting of the New York State Geological Association, Union College, Schenectady NY, 1995, p. 251-273.

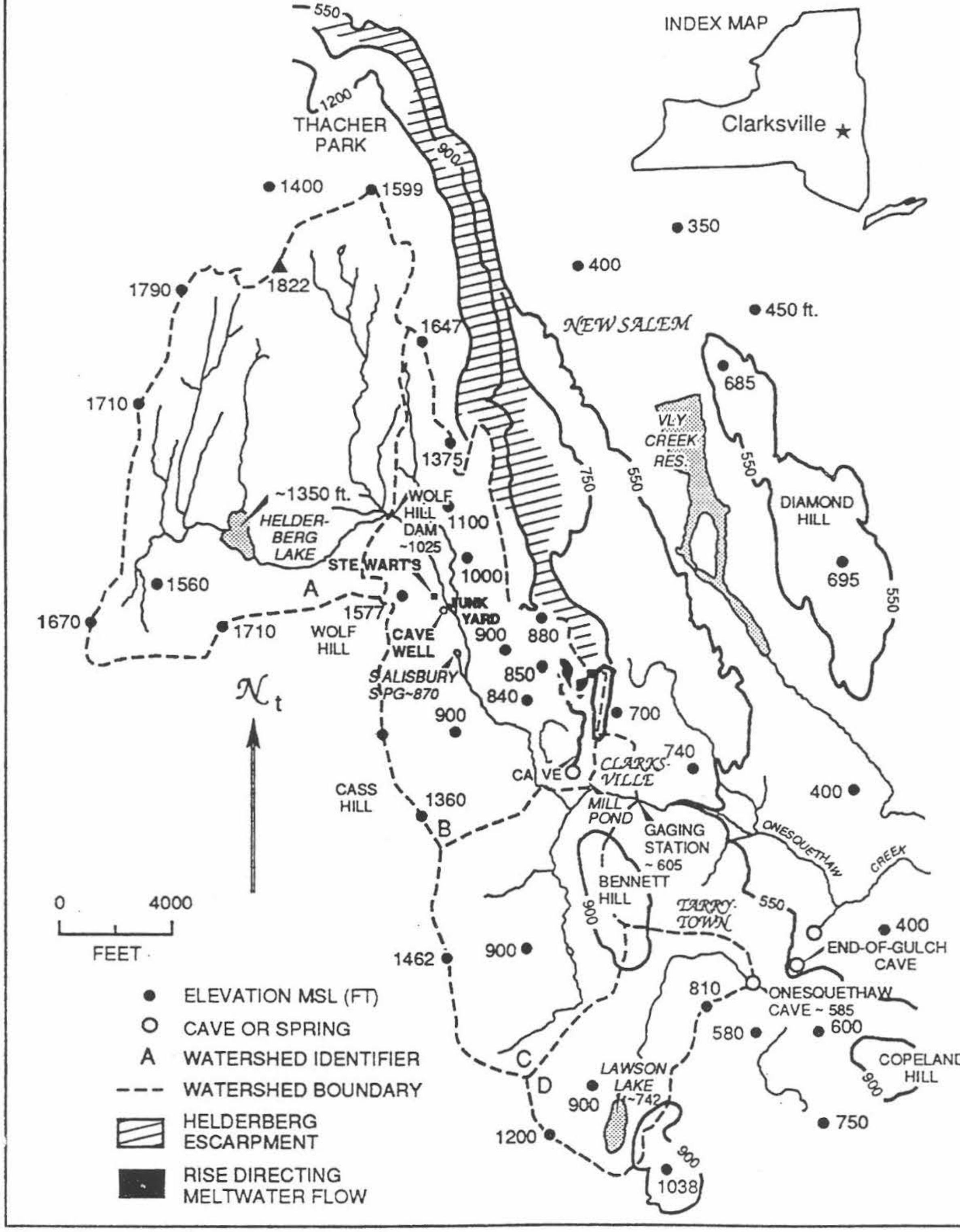

FIGURE 1: TOPOGRAPHY, DRAINAGE BASINS, AND SELECTED FEATURES ALONG THE HELDERBERG ESCARPMENT, ALBANY COUNTY, NEW YORK. WATERSHED BOUNDARIES BASED ON TOPOGRAPHY, GEOLOGY, AND DYE TRACES. CONTOUR INTERVALS VARIABLE FOR CLARITY.

Dineen (1987) has determined that present-day drainage trends in the Hudson Valley were established before the Wisconsinan glaciation, sometime prior to 70,000 years ago. Glacial striations in two locations near Clarksville further indicate that today's drainage was in place prior to inundation by Woodfordian ice. The direction of glacial movement was almost exactly north-south (S13°W), with a maximum ice thickness on the order of one mile about 22,000 years before present (Dineen, pers. comm.). Late Pleistocene drainage along the Onesquethaw Creek was probably little different from what it is today. Evidence presented in this paper argues for pre-Woodfordian cave development.

STRUCTURAL SETTING

The hamlets of Clarksville, Tarrytown and Feura Bush have all been subjected to extensive faulting. Marshak (1986), Marshak and Engelder (1987), and Cassie (1990) discuss structural deformation within parts of the Hudson Valley Fold-Thrust Belt (HVB). The HVB extends roughly from Kingston to Albany, New York, extending to a maximum of 20 kilometers east and west of the Hudson River (Marshak and others, 1986). The deformation may have occurred during the Acadian (Cassie, 1990) or Alleghanian orogenies (Geiser and Engelder, 1983), or during Mount Marion deposition (Murphy and others, 1980).

Faulting and deformation of the Esopus Shale, Schoharie Formation, and Onondaga Limestone, throughout the Clarksville area, may represent the farthest northwestern exposure of the Hudson Valley Fold-Thrust Belt. The extensive structural deformation present throughout Clarksville and the previously documented southern parts of the HVB are characteristic of deformation of sedimentary rock under relatively low pressure and temperature conditions (Marshak and Engelder, 1985). Mapping of the structural or bedrock geology in the area, both on the surface and in the cave, reveals that faulting in the Clarksville area is characteristic of either an imbricate thrust zone or a duplex.

At least six elongate ridges, trending north-south approximately along strike of the faults, are unevenly spaced throughout the Clarksville area. They often exhibit extensive fault-bend folding, slickensides, and in places an anticlinal structure. These limestone ridges, which are underlain by one or more basal thrust faults, can be mapped for distances of up to one mile. One such deformed ridge, situated at the eastern end of Clarksville, has been breached by Onesquethaw Creek. Perhaps the most prominent example is found in the upper Onesquethaw Creek gorge.

The upper Onesquethaw Creek gorge exhibits the best out-of-cave exposure of the repeated basal Onondaga Limestone and the impermeable Schoharie Formation (a quartz rich limestone) thrust below the Onondaga Limestone stratigraphic column. Here much of the bed of Onesquethaw Creek is guided by fault-zone features. The thrust-fault ramp, associated thick calcite bed, and fault-bend folds are the same as those along which Clarksville Cave has developed, except that they are farther south along strike.

STRUCTURAL FEATURES INFLUENCING GROUNDWATER FLOW IN THE KARST AQUIFER

Faulting in and immediately east of Clarksville Cave has resulted in thrusting, deformation, and upward movement of impermeable bedrock units underlying the Onondaga Limestone (Schoharie Formation and Esopus Shale) into a position that makes the eastern escape of groundwater impossible. The gentle southwesterly dip of the bedrock of the Mill Pond aquifer (Fig. 1) fails to direct all subsurface flow in this direction. Instead, significantly higher surface topography to the southwest (e.g., Wolf Hill and Cass Hill) retards dissolution in this direction, in favor of the 1.3° apparent dip between Wolf Hill Dam and the base-level discharge point at Mill Pond. Tracer studies generally verify this predicted flow path, at least during periods of low discharge. However, tracer studies also document an unexpected easterly diversion of moderate- to high-discharge waters through Pauley Avenue in Clarksville Cave. This is significantly farther north than the Mill Pond. This easterly deflection of floodwaters may occur in response to an inefficient outlet and conduit leading to the Mill Pond.

Pauley Avenue floodwaters flow easterly until they become perched on a thin bed of impermeable Schoharie Formation that has been thrust below the basal, or lower, non-cherty subunit of the Onondaga Limestone. Cave diver John Schweyen (pers. comm.) reports the presence of the Schoharie Formation overlying the lower non-cherty subunit of the Onondaga Limestone approximately 700 feet west of the north-south trending Clarksville Cave. This number reflects a minimum westerly displacement of beds above the fault ramp. Floodwaters remain perched, flowing down the apparent dip of the Schoharie Formation, until they encounter a fractured zone along a more steeply inclined part of the fault ramp. Here, subsurface water is deflected sharply to the south and aslant the strike and dip of the inclined fault plane, with the possible localized exception of following a horse for 200 feet north of the Lake Room.

Pirated surface water must rise at the Mill Pond, because the impermeable Esopus Shale is thrust vertically upward against the cavernous Onondaga Limestone. The leading edge of this upthrown shale formation, an imbricate thrust sheet separate from the fault zone that Clarksville Cave formed along, trends roughly north-south (Fig. 2). The Esopus Shale and thrust-fault-induced fault-bend folds in the Onondaga Limestone, present slightly west of the upthrown Esopus, ultimately form a wall or barrier to easterly karstic groundwater flow for a distance that is well in excess of one-half mile.

However, this geologic barrier has only retarded eastern groundwater movement in two locations: (1) east of the known parts of the Waterfall Passage and (2) at the Mill Pond spring resurgence. Formation of most of the north-south oriented cave occurred preferentially aslant an inclined ramp of a thrust fault. Deformation along this thrust plane has produced a fault zone with at least three easily discerned, slickensided surfaces. Although separate, they occur within a few feet of one another. In much of the cave, this fault ramp is accented by one or more prominent calcite beds, often accompanied by a zone of stylolites and calcite-filled extension veins. This calcite bed is podiform in shape, with a central thickness ranging up to eight inches. Whereas the calcite forms a continuous bed of variable thickness aslant the strike of the thrust fault, it is most pronounced in the Gregory Section of the cave, where the vadose part of the cave is steeply inclined along the fault ramp. The ledge forming the outlet of the Waterfall Passage is the same buff brown to black, weathered Schoharie Formation, with underlying calcite bed, as seen in the upper Onesquethaw Creek gorge. The thicker zones of the calcite bed, where dissolution and crystallization are greatest, coincide with the more inclined segments of the thrust ramp, where the stress was highest. Occasional remnant calcite blocks, up to eight inches in thickness, in the Ward's Section of the cave provide the only evidence of the former presence of the thick calcite bed.

Ramsay (1980) provides evidence that similar "extension veins are formed by an accretionary process involving the formation of a narrow fracture followed by the filling of the open space by crystalline material, a mechanism termed crack-seal." Such stress-induced chemical transfer, or pressure solution of materials, seems to be relatively common (Ramsay, 1980). The characteristic crack-seal mechanism of repeated tectonic stress (Ramsay, 1980) is best illustrated in Onesquethaw Cave, situated 2 miles southeast of Clarksville, where calcite infilling aslant the ramp of a thrust fault reaches a maximum thickness of 27 inches. Here, insight into the fault style and repeated activation in the area is suggested by the presence of multiple calcite-vein infilling events along the fault ramp.

Successive cracks often occur along vein-matrix contacts of a previously sealed crack system, because this is mechanically the weakest surface in the rock (Ramsay, 1980). Fractures have been found to increase towards major faults. The higher the fracture frequency, the higher the percentage of calcite-filled fractures (Carrio-Schaffhauser and Gaviglio, 1990). It is a combination of this mechanically weaker calcite bed/Onondaga limestone boundary and related fault partings, all present along this inclined thrust ramp, that have served to orient the north-south segment of Clarksville Cave. Further structural

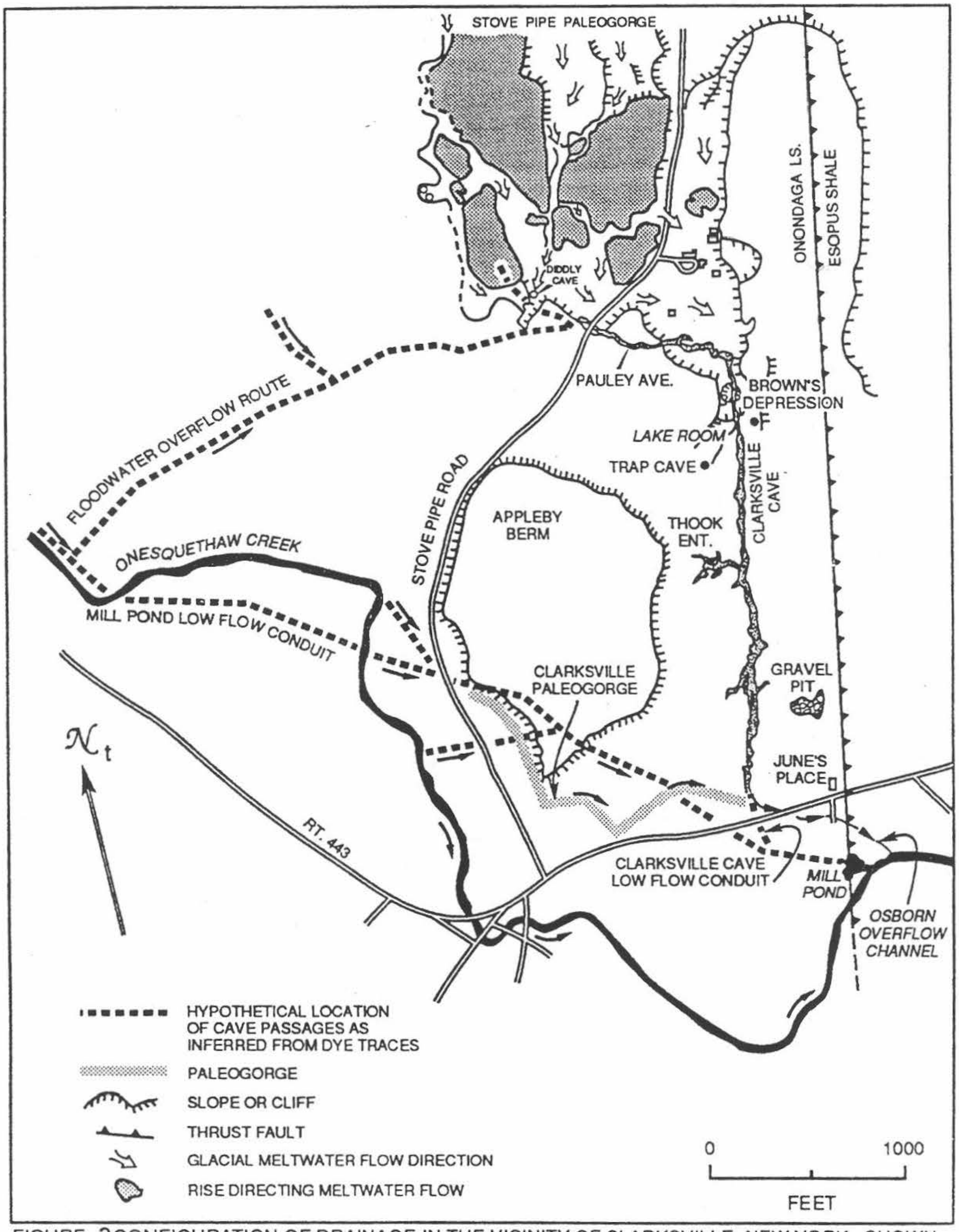

FIGURE 2 CONFIGURATION OF DRAINAGE IN THE VICINITY OF CLARKSVILLE, NEW YORK. SHOWN ARE CLARKSVILLE CAVE AND PRESENT AND PAST ROUTES OF FLOW. DYE TRACES ARE OFTEN IMPORTANT FOR DETERMINING SUBSURFACE FLOW ROUTES ESTABLISHED IN FORMER GEOLOGIC PERIODS.

and hydraulic control of cave-forming waters may also be locally attributed to perching on a fault-thinned Schoharie Formation. Similarly, much of Onesquethaw Cave has developed down and along the mechanically weaker vein-matrix contact. Both caves exhibit characteristic fault-bend folds, stylolites, and extension veins adjacent to the prominent thrust plane.

Of major importance to the development of both caves was Cenozoic structural deformation which provided a preferential solutional pathway along the inclined surface of a fault ramp. A steep hydraulic gradient was thus set up between infiltrating waters and their resurgence points along fault ramps. These faults may then be considered as both negative and positive influences on groundwater flow and cavern development: negative in the sense that downward dissolution did not readily penetrate far below the fault zone (Kastning, 1977 and 1984), and positive in the sense that almost the complete trend of the caves follows a structurally weakened zone of increased permeability.

INTERMITTENT PHREATIC PASSAGES

Meltwater invasion of pre-Woodfordian passages in Clarksville Cave occurred during glaciation, significantly enlarging the cave and its tributary conduits within the aquifer. Observation of the degree of flooding within the cave during major storm and runoff events reveals that only the lowest levels of the cave carry water along the fault zone. Two abandoned upper-level passages, both with relatively consistent ceiling elevations, were identified via a leveling survey. The level of these passages is determined by the relative uniformity of their ceiling heights. The highest of these two upper-level passages extends from the Lake Room, through the Big Room, until its truncation in the Pixie Passages immediately above the Corkscrew (*see* detailed map of Clarksville Cave in Fig. 3: Section B). This 714-foot level can roughly be characterized as the meandering upper level of Perry Avenue. In places, the lower ceiling elevation is controlled by chert beds. The 698-foot level extends from the Bathtub Passage through Upper Cook Avenue, where the passage is truncated by breakdown, and where flow had been diverted down a steeply dipping, fault-plane-controlled tube leading to Lower Cook Avenue.

These large upper-level passages are generally high and dry, and are sometimes tubular in cross-section, suggesting a water-table formation; they undulate upward and downward as is typical of phreatic passages, and have formed with only limited influence from the fault plane. However, they lack the low hydraulic gradient typical of phreatic or water-table origin, are discontinuous in size and extent, and truncate suddenly or become much reduced in cross-sectional area, grading systematically to active pre-Woodfordian vadose passages. The relatively smaller drain size at the down-gradient end of these passages, compared with the cross-sectional area of these floodwater conduits, resulted in temporary phreatic conditions within the cave, quite dissimilar from the conditions of normal phreatic water. Palmer (1991) documents similar floodwater formation of conduits behind local passage impediments such as collapse debris, insoluble beds, or sediment fill, where aggressive water results in rapid passage enlargement. Formation of these "intermittent phreatic" floodwater conduits, coincident with the direct influx of large quantities of subglacial meltwater, occurred under alpine karst conditions.

A third solutionally developed upper level is also identifiable in the cave at 739 feet msl. This discontinuous level is represented in only a short segment of the cave proximal to the Root Room (northeast of the Lost Rock Hammer Room) and some nearby domes. It clearly represents the maximum flood level attained in the cave. Solution domes at the 739-foot level are characteristic flood water-injection features. Like the two lower abandoned levels, it is systematically graded to the actively forming lowest level. These three abandoned levels are interpreted as reflective of different glacial discharges (perhaps seasonal fluctuations) coursing through the cave, rather than different time-based developmental stages. Variable discharges, perhaps influenced by variable outlet efficiencies and climatic conditions during glaciation, are inferred for formation of the levels.

There is no obvious mechanism present today that would explain both the elevation of these upper-level passages and their configuration, that exhibits little fault control. Other than the passage alignment, fault control appears to be a significant developmental factor only in the low-discharge lower-level passages. A ^{14}C date (27,350 $\pm$ 750 yrs BP) obtained from a wood rat bone sample excavated from upgradient Diddly Cave provides evidence for 1) a den site used in pre-Woodfordian time; and 2) pre-Woodfordian cave development (Steadman, pers. comm.). Further argument for pre-Woodfordian origin of the vadose-level passages in the cave stems from the recognition that the mean annual precipitation and the size of the Mill Pond watershed has probably not changed significantly in the last 10,000 years. Funk (1989), through the interpretation of archaeological sites, established that climatic conditions within the last 10,000 years were at times either dryer or similar to that of today. Thus, the availability of the significant discharges necessary to form the upper-level passages was not there postglacially.

These upper-level passages and related higher solution features on ceilings (upward to 739 feet) indicate that they are younger than many lower passages, having formed in response to aggressive glacial floodwaters behind an inefficient outlet. It is hypothesized that the lower-level Clarksville Cave passages served as a natural *in situ* drainage system for glacial meltwaters underneath warm-based Wisconsinan ice. Higher-level cave passages *(e.g.,* the 698- and 714-foot levels) formed in response to the massive influx of subglacial meltwaters behind inefficient, perhaps partially ice-blocked outlets. Similarly, the formation of high-ceiling solution domes, anastomoses, pendants, spongework-like dissolution, and diversion passages may be attributed to floodwater invasion. The gradation of upper-level conduits tributary to the lower levels of the cave, from relict meltwater infiltration points, also lends supportive evidence for a pre-Woodfordian origin of the linear (N12°E) Clarksville Cave passage.

RELICT KARST

A number of relict karst features are present both proximal to Clarksville Cave and to the northwest within the same watershed. These include a number of small shafts and caves *(e.g.,* Trap Cave, North and Thook entrances) that receive only minor amounts of direct meteoric or snowmelt infiltration today. The most important relict karst feature is the Stove Pipe Paleogorge (Fig. 2). This abandoned rockcut gorge grades directly into the pre-Woodfordian Clarksville Cave via the North Entrance, Brown's Depression area, Trap Cave, and the Thook Entrance. Its channel is well defined for most of its course. The morphology of upper reaches of the gorge is characteristic of an ice-marginal meltwater channel with small-scale hanging valleys, rather than a well-graded streambed which would be expected of a former channel of the Onesquethaw Creek. Sugden and John (1976) describe the ice-flow dynamics which cause favorable formation of drainage routes in bedrock versus ice. In some places, the channel configuration is such that only large quantities of water would have been capable of filling the channel sufficiently high enough to overflow into sub-parallel channels. This paleogorge is sharply truncated to the north by a downwardly sloping limestone cliff. A negligible catchment area is present (Fig. 1), certainly too small to carry any significant quantities of water or sediment into the cave as suggested by thick sediment banks and upper-level phreatic passages.

A second paleogorge, the Clarksville Paleogorge, proximal to Osborn Cave, may represent either a pre-Woodfordian drainage route of the Onesquethaw Creek or a channel carved around the Appleby Berm by glacial meltwaters. Gorges of this nature can form in a relatively short time if sufficient abrasive material is carried through it. Von Engeln (1911) documents the rapid formation of a rockcut marginal gorge at the outlet of the Hidden glacier in the Yakutat Bay Region of Alaska.

Brown's Depression (739 ft msl) is an important location in that it received significant paleo-streamflow from the northwest (Fig. 2). Stove Pipe Paleogorge streamflow, originating from a vast subglacial watershed to the north and northwest, incised a channel into the Onondaga Limestone from the northeast until it reached the Hunter's Fissure Cave and Diddly Cave area. Here, this paleo-streamflow was responsible for the formation of these caves. From Hunter's Fissure Cave the paleo-streamflow spread out to the southeast over the gently undulating topography. However, its course was partially constrained by the elevationally higher surface topography to the west, north and east. Thus, much of the Stove Pipe Paleogorge streamflow was funneled southeast into the Brown's Depression/North Entrance (above Lake Room) area, where it entered Clarksville Cave.

During periods of low- to moderate-glacial discharge, meltwaters converged proximal to Brown's Depression where much of the southeastern discharge was retarded from flowing east by a low north-south trending limestone ridge (747 ft. msl). These meltwaters were pirated into Clarksville Cave through the Diddly Cave and Brown's Depression/North Entrance areas. Diddly Cave was recently dug open, increasing the length from 5 to 550 plus feet. A dive push, a short distance into the cave, led to a master conduit which is hydrologically linked to Clarksville Cave. The exposed limestone pavement, coupled with the steep hydraulic gradient present between insurgence and resurgence points, provided a favorable avenue for subsurface piracy.

High-discharge meltwater, with a glacial hydrostatic head, encountered the Clarksville Cave ridge and sought the most efficient outlet, flowing southward and over the ridge barrier. However, the even higher surveyed elevations of a number of abandoned surface stream infiltration points and their conduits leading to Clarksville Cave provide information on the large magnitude of subglacial discharge necessary for their formation. Some of these features include Trap Cave (753 ft. msl), the Thook Entrance (766 ft. msl), and a deep solutionally-enlarged joint near the Ward's Entrance (761 ft. msl) [excavation recently exposed cave passage at its base]. During periods of high glacial discharge, meltwaters probably flowed both within and outside the Stove Pipe Paleogorge channel. This meltwater splayed outward around, and possibly over, the Appleby Berm (Fig. 2). The alignment of the Thook Entrance passages and the Pixie Passages suggests that meltwater coursing around both sides of the Appleby Berm sought to enter the pre-Woodfordian Clarksville Cave through the most direct pathway. Meltwater thus entered the Clarksville and Stove Pipe paleogorges, encountering the Gregory Entrance to the cave, the Ward's Entrance, the Thook Entrance, Trap Cave, the North Entrance, Brown's Depression, and the jointed pavement above the cave.

SEDIMENT FILL

Thick deposits of sediment in the cave provide direct evidence of the quantity of material carried down the Stove Pipe Paleogorge by glacial meltwaters. The point of entry of this material was largely through the Brown's Depression/North Entrance area. The finding of sediments in the newly discovered northwestern segment of the cave (Pauley Avenue) also argues for input via Hunter's Fissure and Diddly Cave. Thick remnant sediments reveal that at least the Ward's Section of the cave was once sediment filled. Because the physical opening of these sediment input points is believed to have been formed by subglacial meltwaters, a sedimentation, passage infilling, and re-excavation history may be constructed.

The thickness of the sedimentary column in contact with bedrock suggests that significant cave enlargement had occurred prior to sediment infilling, possibly during Illinoian and/or Kansan times. The basal deposits on bedrock include imbricate shale-clast-rich sediments with small cobbles, indicative of rapid infilling. Once much of the cave became filled, floodwaters stagnated, leaving their signature in finely laminated sand, silt, and clay layers. These layers are seen near the ceiling in the Ward's Section and are interlayered with courser sediments in the Gregory Section (to at least the 714-foot level). Sediment deposition may have occurred during pre-Woodfordian time, with re-excavation during Woodfordian or post-glacial times. Lack of significant sediment cover and fill in paleo-channels and abandoned insurgence points supports this hypothesis.

Partial plugging by sediment of Clarksville Cave's overflow outlets may have contributed to the degree of backflooding and upward passage development in the cave. Evidence is found for this in large glacial cobbles cemented in a clay matrix now terminating The Hidden Room. It is possible that sediments washing down the Clarksville Paleogorge partially or totally blocked the Gregory and Osborn Entrance overflow outlets for a period of time prior to being washed free again. Alternately, these outlets may have formed as a floodwater modification behind the inefficient Clarksville Cave low-flow outlet.

GLACIAL GEOLOGY

Two and possibly three glaciations are documented as far south as Corinth, New York (LaFleur, 1991, unpublished report). Approximately 14,700 yrs BP, the Wisconsinan ice sheet receded from the Helderberg Plateau (DeSimone and LaFleur, 1985). Dineen (1986) gives an extrapolated bog-bottom date of 15,060 $\pm$1,000 yrs BP for the Great Bear Swamp situated somewhat west of Clarksville. This date further confirms the timing of the deglaciation of the Clarksville area. DeSimone and LaFleur (1985) provide a date of approximately 14,700 yrs BP for the recession of the Pine Swamp ice front from the Clarksville area. They depict the ice front as a lobe or tongue projecting southward to Stuyvesant, New York, with Clarksville situated along the southwestern flank of the ice margin.

Dineen (1986) documents ice thinning during glacial stagnation over the Helderberg Escarpment. Large quantities of meltwater flowed southward proximal to the southwestern flank of the Hudson Champlain Lobe of the Schoharie ice margin. Dineen describes deposition of sediments in multiple meltwater tunnels under stagnant ice. It thus appears that deglaciation from the Stove Pipe Road area was characterized by a southward thinning ice cover, with a southward meltwater flow direction. Free-surface flow was probably present in the paleogorge prior to the final retreat of Woodfordian ice from the Clarksville area.

IMPLICATIONS OF WISCONSINAN CLIMATES

Of even greater importance than the physical presence of the Stove Pipe Paleogorge is its relationship to Hunter's Fissure and Diddly Cave and the implication this has on interpretation of Wisconsinan, and possibly Illinoian and Kansan climate. Hunter's Fissure and Diddly Cave formed along the abandoned Stove Pipe Paleogorge. The presence of small scallop wavelengths in joint-controlled Diddly Cave indicates rapid streamflow along the base of one or more Wisconsinan ice sheets. Rounded stream cobbles in walking-sized passages in Diddly Cave provide clear evidence that a large stream once flowed through the paleogorge. The recent finding of bones of a varying hare and the extinct passenger pigeon within clay deposits in the cave provide important scientific information on these species' recolonization following deglaciation (e.g., Layer 1: 4,350 $\pm$ 60 yrs BP; Layer 2: 9,040 $\pm$ 70 to 10,470 $\pm$ 60 yrs BP; Steadman, pers. comm.).

The Clarksville Paleogorge has probably not had a stream in it for the last 14,700 years, coincident with retreat of Woodfordian ice in the mid-Hudson Valley (DeSimone and LaFleur, 1985). Furthermore, Wisconsinan ice had retreated from the lower Hudson Valley 15,000 or 16,000 years ago (Connally and Sirkin, 1986; Dineen, 1986), with the ice front retreating to the St. Lawrence Valley by 13,000 years ago. Therefore, the maximum time frame for possible ice front stagnation in the Clarksville area during active deglaciation is on the order of 1,000 years.

Solutional cave formation will occur only where a pre-existing network of integrated openings connects the recharge and discharge areas (Palmer, 1991). This is a process that requires a minimum of 10,000 years (Palmer, 1984; Dreybrodt, 1987, 1990; Palmer, 1991) before passageways obtain sufficient size for human entry. Additional passage cross-sectional size requires additional time. A shorter time period, on the order of 5,000 years, may be possible depending on joint widths present in the bedrock prior to infiltration by glacial meltwaters. Thus, subglacial meltwaters apparently were not only responsible for the formation of Diddly Cave, but must have flowed for a minimum of 5,000 to 10,000 years in order for the cave to form. Because the maximum amount of time the retreating ice front could possibly have stayed in the Clarksville area was on the order of 1,000 years, it follows that Diddly Cave, Hunter's Fissure Cave, the Brown's Depression area and other southern infiltration points were receiving meltwater from below warm-based glacial ice for at least 5,000 to 10,000 years. It is likely that temperate climatic conditions were present during the early and later Wisconsinan.

MELTWATER FEATURES IN OTHER NEW YORK STATE CAVES

Many New York State caves need to be re-examined for evidence of glacial meltwater modification. Several caves in east-central New York exhibit features characteristic of meltwater invasion. Examples include Skull, Knox, Ella Armstrong, McFail's, Howe Caverns, Single X, Schoharie, Gage, Onesquethaw, and Surprise (Mystery) caves, all of which have one or more passage segments superposed above passages receiving Holocene peak floodwaters. For example, the upper levels of Skull Cave have aragonite speleothems that are incapable of surviving floodwater invasion. Similarly, caves such as Skull, Knox, and Ella Armstrong have watershed sizes too small to account for the volume of water necessary to form their observed vertical and areal extent. Other caves, such as McFail's and Surprise, carry underfit streams, yet exhibit anomalously large passage sizes. An artificially enlarged subglacial watershed would have been capable of providing the necessary recharge. Related fossil karst includes abandoned and sometimes glacial-debris-covered sinkholes and shafts which once served as significant infiltration points, but now serve only to focus localized drainage into the epikarst, funnelling runoff and infiltrating meteoric waters to deeper conduit flow routes. Anomalous in-cave features such as abandoned pits and multiple level, ungraded passages may also reflect meltwater invasion (e.g., Surprise Cave). Similarly, significant cave development proximal to the headwaters of a drainage basin *(e.g.,* Gage Caverns and Phoebe Pit) may also reflect meltwater invasion from an expanded ice-sheet watershed. Other relict caves, swallow holes and solution conduits such as Knox, Salamander, several Saugerties-area caves, and Joralemon's (this volume) are now abandoned and largely waterfree. Their derangement from active drainage patterns may portray development during a previous interglacial period or, more likely, may be a result of modification by glacial meltwaters.

The characterization of cave modification by glacial meltwater invasion poses many exciting geomorphic questions for researchers in New York State. Speleothem- and sediment-dating techniques may shed light on karstic evolution and modification through multiple glacial periods. A complete geomorphic interpretation must include an assessment of flow conditions and geologic features in a defined watershed, both on the surface and in the subsurface.

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<u>SECTION B:</u> Flow Characteristics and Scallop-Forming Hydraulics within the Mill Pond Karst Basin, East-Central New York

ABSTRACT

This is a study of the hydrology of Clarksville Cave and the headwaters of the Onesquethaw Creek, situated in the hamlet of Clarksville, New York, specifically the Mill Pond karst basin. During most of the hydrologic year, water entering that part of the watershed that is downstream of the Wolf Hill Dam is pirated into the Onondaga Limestone. Tracer tests and in-cave stream gaging indicate that extreme conduit conditions are present in the aquifer, with a maximum water velocity on the order of 5.3 km/hr.

It has been hypothesized that a submerged conduit must lie covered by breakdown blocks at the cave's northern terminus. Having established a known peak flow, a modified version of the Darcy-Weisbach equation was used to accurately calculate the minimum diameter of this conduit. Knowledge of the structural geology throughout the watershed, coupled with a detailed leveling survey in the cave, permitted reasonable estimates to be made for the two unknowns in the equation. A submerged conduit was subsequently opened and explored.

Scalloped cave walls are present in Perry Avenue (Fig. 3) at a key stream gaging location. Backflooding occurs behind inefficient passage constrictions a short distance downstream of, but not up to, this station. Evidence exists that documents that only long return-interval flood stages cause backflooding to this station. This situation permits a reasonable estimate of the maximum discharge and flow velocity responsible for scallop formation. Scallop wavelengths were measured below the elevation of peak floodwaters. By inputting measured values for discharge and flow velocity into published equations, it was possible to back calculate scallop Reynold's numbers that favorably correlate with measured flow velocities and discharges. A possible revision of the scallop Reynold's number is suggested when it is utilized in the determination of paleoflow velocities. It also appears that scallop wavelength is partially determined by the properties of the rock comprising the walls of the conduit.

LOCATION AND WATERSHED BOUNDARIES

A broad carbonate aquifer is present in the Clarksville area. Its boundaries extend north and northwest of the Mill Pond, situated less than 120 meters south of the restaurant, June's Place *(see* Fig. 1: Section A). The farthest boundary of the Mill Pond karst basin lies about 3.9 kilometers to the northwest, proximal to the Wolf Hill Dam on the Onesquethaw Creek. The elevation of the basin ranges from 1822 feet msl atop the Helderberg escarpment to approximately 645 feet msl at the Mill Pond. The boundaries of the catchment basin are depicted in bold dashed lines. These boundaries were defined through the use of low-altitude stereo aerial-photography, U.S.G.S. topographic maps, tracer studies and, in places, detailed structural geologic mapping.

The Mill Pond watershed may be subdivided into two parts: A) that part of the watershed located upstream of the Wolf Hill Dam (1,245 hectares), and B) that part of the watershed located downstream of the Wolf Hill Dam (829 hectares). The downstream part of the Mill Pond watershed exhibits features characteristic of karst terranes. These include sinking streams, limited surface drainage, solutionally enlarged joints, sinkholes, and the Clarksville/Diddly cave system. Structural deformation throughout the region has resulted in extensive jointing and faulting, providing solutional pathways for infiltrating waters.

PIRACY OF ONESQUETHAW CREEK WATERS

Most of the Onesquethaw Creek downstream from the Wolf Hill Dam and upstream of the Mill Pond is a losing stream, with a substantial amount of surface flow lost to solutionally enlarged joints in the streambed. During most of the hydrologic year little or no surface flow occurs in the area downstream of the Wolf Hill Dam, located nearly on the Marcellus Shale/Onondaga Limestone contact. Subsurface piracy of water into the Onondaga Limestone below the Wolf Hill Dam occurs via numerous joints in the streambed. The water briefly surfaces at the Salisbury Spring, only to again sink into joints in the stream bed. The volume of water flowing in the streambed and the relative efficiency of the often partially sediment-choked joints governs the distance water may be found flowing on the surface downstream from the Wolf Hill Dam and the Salisbury Spring. The greater the discharge of the stream, the farther its flow is capable of traveling prior to complete subsurface piracy. During periods of low or moderate discharge, all Onesquethaw Creek surface flow is pirated into the karst network prior to where the bed of the Onesquethaw Creek passes beneath Rt. 443 *(see* Fig. 2: Section A). Only large storm and

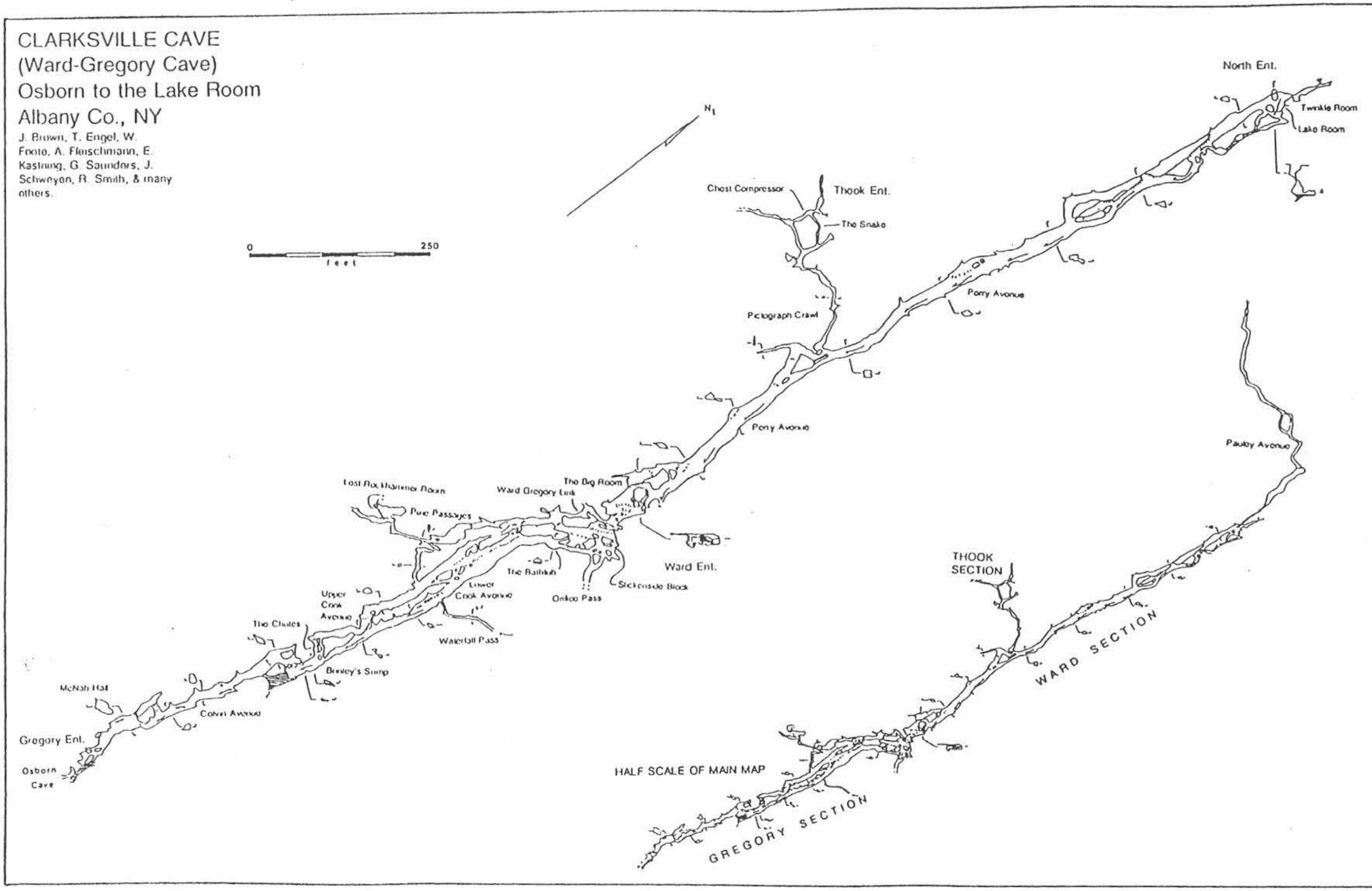

Figure 3 Map of Clarksville Cave, Albany County, New York. Notes: 1) This map is based on map of Kastning (1975). New passages added are Pauley Avenue, the Pinch Passage, the Thook Section, the Orifice Passage, and part of the Bathtub Feeder. 2) Stream flow is perennial, but varies from a low of 2 gpm to in excess of 60,000 gpm. 3) The Gregory Entrance, Brinley's Sump, and the stream access immediately east of McNab Hall are subject to flooding. During periods of extreme flood, the Bathtub and the downstream end of Perry Avenue also sump shut. 4) From the Lake Room south, the entire cave is Grade 5 except the Pinch Passage (grade unknown) and the Orifice Passage (Grade 2). Upstream of the Lake Room the underwater section is Grade 3; the rest as far as the Loop is Grade 5. Beyond the loop it is Grade 1.

snowmelt events generate enough surface flow in the watershed to cause the Onesquethaw Creek to flow throughout its course. This represents a very small part of the hydrologic year. Surface-stream flow is short-lived even after major storm events.

TRACER TESTS

A series of uranine-tracer tests have permitted partial delineation under varying conditions of discharge of the subsurface flow paths throughout the Mill Pond drainage basin. Uranine is a non-toxic tracer frequently used in karst investigations (Smart, 1984). It was injected into various joints in the Onesquethaw Creek streambed that were pirating water. Activated-carbon detection bugs were placed at all likely resurgence points, collected later, and chemically elutriated with Smart solution (Quinlan, 1986).

Tracer testing has revealed that dye injections from 3.2 km upstream of the Mill Pond resurgence remain perched above the Upper Cherty Subunit for at least 1.6 km before breaching chert beds that overlie the lower, more massive, non-cherty subunits. Water pirated into the Onesquethaw Creek streambed immediately downstream of the Wolf Hill Dam and upstream of Rt. 85 remains in one or more subsurface conduits, until surfacing briefly at the Salisbury Spring, only to again sink into joints in the stream bed downstream. The Salisbury Spring is located on the western side of Rt. 443, approximately 0.6 km southeast of Rt. 85. It is set back some distance from the road. It is likely that piracy of the Salisbury Spring discharge into the bed of the Onesquethaw Creek is roughly coincident with the point at which this water breaches the Upper Cherty Subunit of the Onondaga Limestone. Thus, one major tributary conduit to the system is likely to become physically impassable within 1.4 km northwest of the Lake Room in Clarksville Cave (Fig. 2). However, stream gaging and tracer studies indicate the presence of a second low-flow conduit entering the known parts of Clarksville Cave from the large, heavily jointed watershed to the north-northwest.

All subsurface flow resurges at the Mill Pond. The relative inefficiency of the outlet of the Mill Pond conduit may be due to structural problems resulting from the upward thrusting of the impermeable Esopus Shale against the cave-bearing Onondaga Limestone *(see* Section A). The presence of impermeable Esopus Shale in the bed of the Onesquethaw Creek at and immediately downstream of the Mill Pond forces all subsurface flow from the karst aquifer to surface at the Mill Pond. This author established a gaging station downstream of this point *(see* Fig. 1: Section A).

Tracer tests and discharge measurements throughout the watershed indicate that during periods of low discharge, pirated Onesquethaw Creek waters do not travel through Clarksville Cave. Surface and subsurface stream gaging and tracer tests establish the intersection of the pirated Onesquethaw Creek low-flow conduit with the Clarksville Cave low-flow drainage conduit to be located between the southern end of the cave and the Mill Pond *(see* Fig. 2: Section A). Hereafter, the conduit that resurges at the Mill Pond and is physically separate from the Clarksville Cave conduit north of Osborn Cave, is referred to as the Mill Pond low-flow conduit. Although the exact elevation of the lowest drain point in Clarksville Cave remains to be surveyed, it lies slightly below an elevation of 660 ft msl. The hypothesized flow routes of unentered parts of the network are portrayed in Figure 2 of Section A.

Tracer tests verify that after a certain critical subsurface discharge is reached, coincident with piracy of increasingly greater amounts of surface flow into the subsurface conduit system, the efficiency of the Mill Pond low-flow conduit is exceeded and surplus water is shunted to the Lake Room in Clarksville Cave. The Mill Pond low-flow conduit utilized today, which bypasses Clarksville Cave, may be the original flow route, with the flow route leading to the Lake Room (via Pauley Avenue) forming as a floodwater-overflow route. Alternatively, the flowpath to the Lake Room may represent the original subsurface flow route that was later abandoned due to further stream piracy, possibly coincident with lowering of the regional base level. Under this genetic interpretation, diversion of waters from the Mill Pond low-flow conduit to Pauley Avenue would occur behind an immature drain. During periods of base flow, it appears that only water from the north-northwestern part of the Mill Pond watershed rises in the Lake Room. Moderate and high discharge in the subsurface causes a significant backup of water behind the Mill Pond low-flow conduit, resulting in large overflows to the Lake Room. The greater the flow in the Onesquethaw Creek, the more water is lost through joints in the streambed, and the greater is the discharge that appears in the Lake Room.

From the Lake Room the water flows south through the cave where some of it joins, in a tributary manner, the Mill Pond low-flow conduit somewhere between the downstream end of the cave and the Mill Pond (*see* Fig. 2: Section A). As the discharge of floodwaters within Clarksville Cave increases, the hydraulic efficiency of the branched conduit leading to the Mill Pond is exceeded. The remaining water that cannot be handled by Clarksville Cave's low-flow subsurface conduit and the Mill Pond low-flow conduit backs up within the cave as temporary storage. After a critical flow on the order of 2.7 cfs is reached, excess floodwaters are discharged along the Osborn Cave overflow route (677 ft msl) to the surface. Osborn Cave is situated directly south of the Gregory entrance and is physically connected to Clarksville Cave by a water filled conduit. Figure 2 of Section A shows the Osborn overflow channel, that sometimes carries large quantities of water.

KARST BASIN CHARACTERIZATION

Tracer tests conducted in parts of the Mill Pond aquifer reveal that all subsurface waters reappear or resurge at the Mill Pond. During periods of low flow, all surface and groundwater downstream of the Salisbury Spring and upstream of the bridge crossing Rt. 443 discharge through conduits in the carbonate aquifer at the Mill Pond. During periods of moderate to high-subsurface discharge, part of the subsurface flow is shunted through Clarksville Cave. All flow throughout the Mill Pond watershed thus surfaces either in the Mill Pond or Osborn Cave, where, for much of the year, it comprises the headwaters of continuous surface flow of the Onesquethaw Creek. At times this flow is supplemented by water from the Clarksville South Road and western Bennett Hill Road sub-watersheds.

Water in the Onesquethaw Creek, from that part of the watershed upstream of Wolf Hill Dam that is not artificially diverted to the Vly Creek Reservoir, also sinks into the subsurface downstream of the Wolf Hill Dam. Much of the flow in the karst network originates as diffuse infiltration outside the Onesquethaw Creek Corridor. Virtually all meteoric water and snowmelt contacting the heavily jointed, generally thin-soil-mantled limestone pavement within the Mill Pond watershed is pirated into subterranean limestone conduits. Geologically, water entering the soluble Onondaga Limestone must stay within it because it is underlain by approximately 1 m of the Schoharie Formation (a quartzitic limestone) and approximately 30 m of impermeable Esopus Shale.

Physically unentered segments of the conduit network may be envisioned as being similar to a tree, where all branches coalesce downstream toward the trunk. Palmer (1991) describes such branchwork caves as the most common type. Water infiltrating from different segments of the aquifer's recharge area converges as higher-order passages that decrease in number and generally increase in size in the downstream direction. It is likely that the large northwestern part of the Mill Pond aquifer is branchwork in nature, with many tributaries coalescing downstream toward larger, master passages. It is also likely that segments of the conduit system directly underlie the bed of Onesquethaw Creek, whereas others extend far to the northwest. Still other segments must enter from the northwest where runoff from the Marcellus and Hamilton beds of Wolf and Cass hills sinks near the Onondaga Limestone contact and is rapidly pirated into the system. In the early 1990s, exploration via the newly opened Diddly Cave entrance yielded approximately 0.5 km of large stream passages extending north into the Mill Pond karst basin. These passages are branchwork in character, and if connected to Clarksville Cave would bring the cave's length to greater than 2 km. The dashed lines on Figure 2 of Section A portray a simplified version of the hypothesized configuration of conduits in the eastern end of the system.

SUBSURFACE TRAVEL TIMES

The combined flow from stream losses and diffuse fracture infiltration is documented as moving very rapidly through the karst system. Although effort has not been made to absolutely quantify the rate of subsurface flow in the aquifer, the timing of two tracer tests provides some insight on the situation. Under moderate flow conditions present on February 23, 1990, uranine tracer was injected into a joint in the bed of the Onesquethaw Creek, 3.2 km northwest of the Mill Pond. At this time, all surface flow in the upper reaches of the Onesquethaw Creek was being pirated into this joint. Tracer-detection bugs were collected from Clarksville Cave at 4:00 p.m. on February 24, 1990, about 22 hours after the tracer injection. All were positive for uranine. Thus, a subsurface groundwater transit time in excess of 150 meters per hour was documented.

A similar trace was conducted in October 1988 under low-flow conditions. In this instance, the tracer injection and sinking of the stream occurred farther northwest than during the above trace. In this second test the tracer-detection bug was removed from a location proximal to the Mill Pond 27 hours after tracer injection. After elutriation, the detection bug was positive for uranine. A subsurface groundwater transit time in excess of 120 meters per hour was documented for low-flow conditions.

In contrast, during a time of peak flow within the aquifer (March 15, 1986 at 1:45 a.m.), the discharge and velocity of flow within Clarksville Cave were measured. The velocity was recorded as 1.48 meters per second. This equates to 5,328 meters per hour (5.3 kms/hr) and may be considered as indicative of the peak velocity of potential groundwater movement within the aquifer and of extreme conduit conditions. At times of peak flow, groundwater may move from end to end through the karst aquifer, a distance of approximately 3.9 km, in less than one hour.

The rapid hydraulic response to significant precipitation or snowmelt within the watershed has been repeatedly documented with stream hydrographs both in Clarksville Cave and in the Onesquethaw Creek. Subsurface conduit flow in the Mill Pond aquifer is roughly analogous to open channel flow in a surface stream. A thin soil-moisture bank over much of the watershed's limestone pavement further permits rapid infiltration of meteoric waters and snowmelt, thus bolstering subsurface transit times. Flood pulses throughout the karstified system are flashy, providing evidence of mature conduit development. Rapid flow characteristics present within the 2,074-hectare Mill Pond watershed, especially that part downstream of the Wolf Hill Dam, make it and the Onesquethaw Creek extremely sensitive to infiltration of contaminants.

During much of the hydrologic year, discharge from the Mill Pond acts as the sole source of water to the upper reaches of the Onesquethaw Creek. During periods of base flow this discharge has been gaged at less than 0.1 cfs. The recent zoning of land central to the carbonate aquifer as rural commercial may have severe effects on both the aquifer and Onesquethaw Creek if untreated waste streams or septic infiltration are permitted (Rubin, 1990b, 1992).

IN-CAVE AND ONESQUETHAW CREEK FLOW CALCULATIONS

Measurements of discharge and streamflow velocity have been made periodically in Clarksville Cave since 1983. Over 99% of the water flowing through Clarksville Cave rises in the Lake Room. This water has been gaged during both low and high flow at discharges ranging between 0.002 and 111 cfs.

A maximum water depth of 63.5 cm was measured during the storm of March 15, 1986. Discussions with Ed Gregory revealed that the flood discharge component in the cave, associated with the 1938 failure of the Helderberg Lake Dam, was significantly greater than the above maximum-gaged amount. Gregory reported that floodwaters were ponded to an elevation of approximately 719 feet msl, a short distance down the entrance slope inside the Ward's Entrance. The elevation of the cave passage in upstream Perry Avenue, approximately 11 meters south of the Lake Room, lies between 715 and 708 feet msl, thus indicating that all of Perry Avenue was flooded during this event. Confirmation of this flood level, and possibly another in 1903, is manifested in a thick mud film covering historic names and dates chiseled near the passage ceiling.

A gaging station was established in the Onesquethaw Creek (*see* Fig. 1: Section A) in order to examine the relationship between in-cave discharge and surface-watershed discharge. This was monitored twice daily for 15 months, more frequently during flood events, and periodically for 4 years thereafter during major runoff events. Stream discharge was gaged at 13 different stages. Curvilinear regression was then utilized to establish a series of multi-order equations that could be used to correlate stage height with discharge. The greatest discharge recorded for the Onesquethaw Creek during the course of this study was approximately 1337 cfs. This occurred on March 15, 1986 at 3:00 am following heavy rains ($\approx$7.0 cm) on a 38-centimeter snow pack. Temperatures up to 40° F accompanied the coastal storm of March 13-14, 1986. Daily monitoring of stream stage in Clarksville Cave for the same 15-month period revealed that a direct correlation exists between this discharge and that in the Onesquethaw Creek. Approximately 8 percent of flood-peak discharge in the Onesquethaw Creek flows through Clarksville Cave.

Knowledge of expected flood-return intervals and their magnitude in the Mill Pond karst basin was found to be essential to both the interpretation of how abandoned upper-level passages in Clarksville Cave formed and an understanding of the dynamics controlling scallop formation. The limited data for statistical comparison among hydrologic years in the Mill Pond karst basin necessitated examination of another roughly comparable basin in order to assess flood-return intervals. The farthest headwater gaging station on Schoharie Creek at Prattsville was selected. Many inherent differences occur between the basins, notably elevation, geology, regolith thickness, size, and location. The Prattsville and Onesquethaw Creek gaging stations are approximately 48 kilometers apart. However, the Prattsville and Mill Pond watersheds are comparable under conditions of a saturated soil-moisture bank, high runoff, and similar storm systems. Eighty-two years of data were examined at the Prattsville, New York station.

A Log-Pearson-Type-III and Gumbel-distribution statistical comparison of historic peak flow of Schoharie Creek gaging data with this study's hydrograph information for the Onesquethaw Creek indicates that the largest Onesquethaw Creek peak of record (March 15, 1986) has a return interval on the order of 30 to 47 years (Fig. 4). This corresponds to a Prattsville hydrologic-year peak discharge of 54,900 cfs. Thus, if 40 years was the expected flood return interval, 25 floods of this magnitude could be expected every 1000 years. Reconstruction of the 1903 peak discharge at Prattsville (approximately 63,000 cfs), the highest on record, further reveals that a cave discharge well in excess of 111 cfs may also have occurred in 1903. The 1903 discharge, three standard deviations greater than the mean-annual Prattsville peak flow, has a predicted flood return interval on the order of 47 to 90 years. Although the magnitude of this flood was larger than the 1986 flood, it probably was not as great as during the dam-failure flood in 1938. These infrequent storm or runoff events reasonably represent a near-maximum quantity of water available in the watershed under ideal, thin-soil-mantled, rapid-infiltration conditions. Therefore, it is difficult to explain the "intermittent phreatic" upper-level passages in Clarksville Cave without a substantially greater quantity of water. A subglacially enlarged watershed, as discussed in Section A, appears to be the only viable explanation.

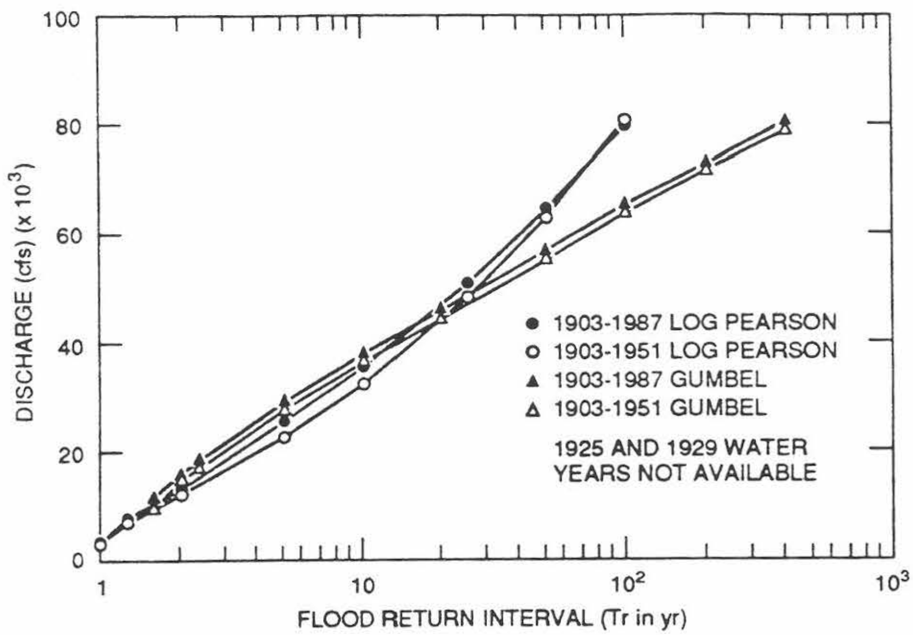

Figure 4: Schoharie Creek at Prattsville, New York. Log-Pearson type III and Gumbel distributions utilizing historic water year peak flow data. The flood return interval for the largest Onesquethaw Creek peak discharge of record correlated to the Prattsville March 15, 1986 flood of 54,000 cfs. The range of two methods shows a Tr of 30 to 47 years. Similar comparisons may be made for low flows in geologically comparable basins. Analyses of this type can be useful tools in predicting flow velocities, contaminant arrival times, and contaminant dilutions. Knowledge of peak or base flow return periods can sometimes be correlated with water chemistry results to help assess chemical loading both in the karst system and to stream receptors.

LAKE ROOM SUBMERGED CONDUIT

Measured, statistically predicted, and inferred (e.g. mud-covered historic names and dates) high discharges rising from the Lake Room indicated that an obscured conduit was present that must be capable of transmitting large discharges. It was thus hypothesized (Rubin, 1989) that a submerged conduit must lie covered by breakdown blocks, below the water surface in the Lake Room. By making a number of reasonable assumptions, it was possible to calculate the minimum diameter of an assumed circular conduit capable of discharging a given flow. A modified version of the Darcy-Weisbach equation

$$r = \left(\frac{Q}{\pi \sqrt{\frac{4g}{f}} \sqrt{\frac{\Delta h}{L}}} \right)^{2/5}$$

was successfully utilized to examine the size of the, until recently, unentered upstream segments of cave conduit, north of known parts of the cave. Calculations were confined to a circular conduit capable of discharging between 111 and 222 cfs (Q). The latter value was considered a reasonable approximation for the 1938 dam-failure discharge. A friction factor (f) of 0.1 was used. The two unknowns in the equation were the change in head (elevation of water upstream of the lake versus the elevation of the lake, Δh) and the length of flooded passage upstream of the lake (L) during flood events. A wide range of values of 1.5 to 30 meters, and 6 to 1219 meters were tested, respectively, for these unknowns. Although some of the values tested were likely to be extreme in nature, they were selected based on knowledge of the structural geology within the watershed, coupled with a detailed leveling survey throughout the cave. It was believed possible that significant backflooding might be occurring behind the Lake Room breakdown.

Insertion in the modified Darcy-Weisbach equation of a reasonable range of values for the change in head and the length of flooded passage suggested that the minimum diameter of a circular-conduit tributary to the Lake Room is between 0.6 and 2.4 meters. Recent excavation and penetration of a formerly blocked and water-filled conduit extending north and west of the Lake Room verified the calculations (Rubin, 1990a). The length of the water-filled passage was found to be 61 meters. The actual Δh value is probably no more than 3 meters. The smallest diameter found in these newly discovered passages was 1.4 meters. Maximum cross-sectional area found in the approximately 366 meters of conduit beyond the Lake Room that have been entered thus far is on the order of 8 square meters. These passage dimensions attest to mature conduit development in the carbonate aquifer within the catchment basin.

The assumed friction factor of 0.1 was found to accurately reflect the flow conditions through the Lake Room breakdown. Approximately two vertical meters of clean washed angular breakdown, interspersed with minor quantities of rounded glacial cobbles, were excavated. The heterogeneous mixture of breakdown blocks ranged in size from several centimeters in length, width and height to approximately one meter. The water's approach angle, toward the lake surface, rises at approximately 30 degrees for the last 3 meters before reaching an irregular constriction (1.2 meters by 0.5 meters) in the breakdown. Prior to the last 3 meters, the submerged conduit is generally horizontal. The maximum conduit depth below the surface of the lake was found to be approximately 4.3 meters.

SCALLOP-FORMING HYDRAULICS

Phases of this study focused on defining the Mill Pond karst basin, the relationship between flow in the carbonate aquifer versus that in Clarksville Cave, the expected return interval of peak discharge inside and outside Clarksville Cave, and flow conditions peculiar to Clarksville Cave. Specifically, a range of stream discharges and velocities were measured in an air-filled segment of linear passage and rectangular cross section. Water depth was recorded, as well as scallop wavelengths within the zone of the 30- to 47-year flood-return interval.

Paleoflow information that researchers hope to reconstruct, based on scallop wavelengths and dimensions of an abandoned passage, is either empirically measured or reasonably constrained. Measurements in Clarksville Cave permitted a cave-specific evaluation of Blumberg and Curl's (1974) scallop Reynold's number.

One potential problem with characterization of the physical conditions under which scallops form is defining the discharge, or range thereof, responsible for scallop development. It was possible to define minimum and maximum discharge limits leading to scallop formation at a key stream-gaging location in Perry Avenue. Here, cave walls in a fossiliferous sparite are scalloped. Streamflow across the width of the cobble floor does not become deep enough, or of sufficient discharge to form scallops until the water is approximately 18 centimeters deep. The 30 to 47-year flood (111 cfs) of March 15, 1986 resulted in a stream depth of 63 centimeters, but decayed in 34 hours to 10 cfs with a stream depth of less than 13 centimeters.

Backflooding occurs behind inefficient passage constrictions at the Big Room, approximately 120 meters downstream of this same key stream-gaging location. The level of backflooded waters, as measured on March 15, 1986 at the stream's surface, was only 48 centimeters lower in elevation than ponded water at the Perry Avenue gaging station. A small additional discharge amount, such as that probable in 1903, or certainly in the 1938 flood, would have substantially reduced the stream velocity here and its ability to form scallops. Thus, the 111 cfs measured on March 15, 1986 represents a value that is close to the maximum possible for discharge capable of forming scallops at the gaging station. This situation allows for a reasonable estimate of the maximum discharge and flow velocity responsible for scallop formation.

Statistical analysis of Clarksville Cave flood-return intervals indicates that cave discharges in excess of two standard deviations about the mean-annual peak discharge may be sufficiently short-lived and of infrequent recurrence to form the observed scallops. The frequency of shorter term flood intervals is greater, and perhaps it is these events which are recorded as scallops rather than very short duration, high-discharge long-return-interval floods. Although the shorter-return-interval floods are also relatively short-lived, it may be the combined contact time of water with bedrock of many similar magnitude floods that is of importance. The relationship among stream depth, discharge, and flood-return interval in the watershed, as partially indicated in Table 1 below, suggests that scallop formation in Clarksville Cave may occur during flood intervals that range between one and two standard deviations of the mean-annual peak discharge.

Channel Width (m)	Stream Depth (cm)	Discharge (cfs)	Velocity (cm/sec)
a) 3.3	18.3	14	68.6
b) 3.3	27.2	30	96.5
c) 3.3	63.4	111	147.8

Table 1: Three flow conditions in Perry Avenue, Clarksville Cave.

Scallop wavelengths were measured below the elevation of peak floodwaters. By inputting measured discharge and flow velocity numbers into published equations, it was possible to back-calculate scallop Reynold's numbers that favorably correlated with measured flow velocities and discharges. This procedure involved measuring scallops, streamflow, and stream velocity and examining the likely range of scallop-forming conditions utilizing published equations. For the rectangular Perry Avenue conduit:

The Sauter mean was used to calculate mean scallop wavelengths of scallop groups within 63.5 cm of the cave floor:

$$L_{32} = \frac{\sum li^3}{\sum li^2}$$

(Curl, 1974)

A range of in-cave flow conditions was examined. The three flow conditions presented in Table 1 bracket the minimum and maximum stream discharge and velocity believed to be responsible for scallop formation at the key Perry Avenue gaging station.

"The scallop Reynold's number, N^*_R based on friction velocity, is a universal constant for scallop formation and was determined from model experiments (Blumberg and Curl, 1974) to have the numerical value $N^*_R = 2200$" (White, 1988).

Scallop formation is controlled, in part, by a dimensionless Reynolds number:

$$N_R = \frac{vL_{32}\rho}{\eta}$$

where: v = mean velocity of fluid flowing past scallop in cm/sec
L = mean scallop length in cm
ρ = density of fluid $\approx$ 1.0 gm/cm for 5°C and 10°C
η = fluid viscosity $\approx$ 0.015 gm/cm/sec for 5°C
and $\approx$ 0.013 gm/cm/sec for 10°C

Thus, examining the specific flow conditions in a), b), and c) above *(see* Table 1) using L_{32} = 7.49 cm and η = 0.015 gm/cm/sec., a range of site specific Reynold's numbers was obtained:

a) N_R = 34,254
b) N_R = 48,186
c) N_R = 73,801

Curl (1974) provides the limiting geometry for a rectangular cave passage:

$$N_R = N^*_R \left[2.5 \left(\ln \frac{D}{2L_{32}} - 1 \right) + B_L \right]$$

By inserting the range of N_R's above into Curl's equation, we can examine N^*_R, the scallop Reynold's number based on a range of actual flow velocities:

a) N^*_R = 2,341
b) N^*_R = 3,293
c) N^*_R = 4,337

Blumberg and Curl (1974) derived a universal constant for the scallop Reynold's number, based on plaster model studies, of 2200. The wall material subject to scallop formation may influence the value of the scallop Reynold's number. Different types of surfaces, like limestone, ice, and plaster, may respond differently to water scour. Based on this study, it appears that N^*_R may actually not be a constant, but instead may best be characterized by a range of values. These values, based on this cave-specific study of a rectangular conduit, appear to be from 1 to 2 times the accepted constant. Empirical observation of the flow dynamics in Clarksville Cave, coupled with a characterization of flood-return intervals within the catchment basin, suggest that a scallop Reynold's number on the order of 3300 better approximates New York State cave-specific conditions. A scallop Reynold's number of 3300 was used for the determination of paleoflows at the Hollyhock Hollow Sanctuary and Joralemon Park (this volume).

It is possible that constants in accepted equations may lead to an underestimate of paleo or recent flow velocities and discharges. It should be noted that B_L, another constant in the Reynold's number equation (which deals with wall roughness) was accepted on face value. Further studies of the actual hydrologic conditions in which scallops form are warranted.

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ROAD LOG

CLARKSVILLE CAVE

Total Miles	Miles From Last Point	Route Description
0.0	0.0	Take the exit from the New York State Thruway at Interchange 22 (Selkirk). Proceed through the tollbooth, turn right, and proceed south on Route 144.
0.4	0.4	Junction of Route 144 with Route 396 to Selkirk. Turn right and proceed west through Selkirk on Route 396.
6.4	6.0	Junction of Route 396 with Route 102.
13.2	6.8	Continue west on Route 396. Route 396 becomes Route 301. Continue, crossing the Onesquethaw Creek in the hamlet of Clarksville, until the end of Route 301. Junction of Route 301 with Route 443. Turn left and proceed west on Route 443.
13.3	0.1	STOP #1 - Turn right into parking area for June's Place restaurant. Proceed to **upper** parking area behind June's. Clarksville Cave is a popular caving location in the Albany area. Always cave with experienced cavers, dress warmly, wear a hardhat with a chin strap, and carry three sources of light (one preferably mounted on your helmet to free your hands for climbing). Do not attempt to pass through Brinley's Sump when there is little or no air. Under these flow conditions, the Gregory Entrance is sumped shut and a through trip is not possible. Failure to follow standard safety procedures has resulted in numerous rescues, mobilization of multiple rescue teams, ambulances, and the press.