A Geological and Geophysical Investigation of Ice Mountain Algific Talus, Hampshire County, West Virginia

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A Geological and Geophysical Investigation of Ice Mountain Algific Talus, Hampshire County, West Virginia

Kevin M. Andrews

Thesis submitted to the Eberly College of Arts and Sciences at West Virginia University in partial fulfillment of the requirements for the degree of

Master of Science in Geology

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Morgantown, WV 2003

Keywords: periglacial, slope winds, terrain conductivity, paleorefugia, North River Mills

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Abstract

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Kevin M. Andrews

Ice Mountain is an algific talus slope located in Hampshire County, West Virginia, protected for its status as a biological refugium. Both the origin of the slope and the mechanism by which cold air is produced from the base of the slope have remained mysteries to local residents and tourists for many years. This research effort takes a scientific approach to exploring the mysteries of Ice Mountain. Bedrock and surficial geologic mapping provides an understanding of the physical environment of the slope area, while geophysical survey data provide clues to the subsurface of the slope. The algific talus consists of Oriskany Sandstone boulders sitting unconformably on steeply dipping Devonian Marcellus-Needmore Shale bedrock. The talus accumulation probably formed under periglacial conditions that existed during the Quaternary at Ice Mountain. Bedrock benches in the subsurface of the slope provide surfaces on which cooler air and water become trapped, resulting in frost and ice accumulation. Surface benches at the bottom of the slope are continuously cooled by gravity-driven, katabatic down slope winds. A comparison of the Ice Mountain algific slope with algific slopes in Iowa reveals that the slopes differ somewhat in structural makeup and airflow cycles, yet both sustain unusual cold environments supporting species typically limited to more northern or higher elevation sites.
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Introduction

The Ice Mountain Preserve is located on the northwest slope of Ice Mountain in Hampshire County, West Virginia (Figure 1). The preserve contains an algific (“algus faciunt” in Latin, which means “make cold” [Sletto, 1994]) talus (broken material [Sletto, 1994]) slope that is reported to discharge air of a relatively constant temperature of approximately 3 to 7° C (38 to 45°F) (Anonymous, 1991). The algific talus site (Figures 2a through 3b), nicknamed “Nature’s Ice Box” and “Nature’s Refrigerator” by local residents, was used by American Indians and early settlers for storage of perishable food items during warmer months of the year (Anonymous, 1991). Theories of the origin of the ice inside Ice Mountain range from “underground glaciers” to annual freeze-thaw cycles (Potomac Pennysaver, 1991). Many questions concerning the cold air production at Ice Mountain remain unanswered.

Ice Mountain talus has been called “the finest example of the algific talus slope ecosystem in eastern North America” (Anonymous, 1991). The cold air exiting the base of the talus at 235 m (770 ft) above sea level supports alpine vegetation that ordinarily flourishes at elevations of 900 to 1200 m (3000 to 4000 ft) (Anonymous, 1991). The algific talus creates a unique environment in which Appalachian, Canadian, and Arctic species grow in the same area (Anonymous, 1991). The site was purchased by the Nature Conservancy in 1989 for $300,000 (Anonymous, 1991) and established as a preserve because it is a biological refugium, a habitat that supports species not able to live elsewhere in the region (Nekola, 1999).
**Figure 1:** Location map of Ice Mountain within the Potomac Highlands of Hampshire County, West Virginia. Blue = higher elevations, brown = lower elevations. (Image source: West Virginia GIS Technical Center)
Figure 2a: Photo of algific vent area at the base of Ice Mountain talus, taken from point bar deposit northwest of Ice Mountain. Algific vents are indicated by white arrows. North River, visible in the foreground, flows north, to the left in the photo. (Photo: K. Andrews)

Figure 2b: Photo of Ice Mountain algific talus taken from the northwest. The largest area of talus is visible in the center of the photo. Algific vents, as indicated by the white arrows, are located in the vegetated area at the base of the talus. (Photo: K. Andrews)
Figure 3a: Photo of algific vent area along access trail at the base of the talus on the east bank of North River. White arrows indicate larger vent locations. (Photo: J.S. Kite)

Figure 3b: Photo of large algific vent at the base of the talus on February 2, 2002. Yellow field notebook for scale. (Photo: J.S. Kite)
The algific talus is a paleorefugium, meaning that it supports species from a fragmented relict of a once widespread ecosystem that existed under different climatic conditions (Nekola, 1999). Rare plants in the preserve include Bristly Rose (*Rosa acicularis*), Twinflower (*Linnaea borealis*), Bunchberry (*Cornus canadensis*), Appalachian Wood Fern (*Gymnocarpium appalachianum*), and Purple Virgin’s Bower (*Clematis verticillaris*) (Zimet, 2000). All of these plants ordinarily occur at much higher elevation. Bristly Rose is also common in Siberia, the Yukon, and other extreme northern climates (Zimet, 2000).

Understanding the internal make-up of the Ice Mountain algific slope is important to preservation of the air circulation mechanism in the slope and the biological refugium it supports. Research has been conducted on algific talus slopes in Iowa, but there has been very little scientific research concerning the source of the cold air at Ice Mountain. Research of this kind is necessary for conservationists to understand the effects that disturbance of different parts of the slope may have on the cold air production and ultimately the algific environment.

**Purpose**

The purpose of this study is to collect scientific observations from the Ice Mountain area in order to better understand the mechanisms responsible for the cold air production at the base of Ice Mountain. Scientific observations conducted at Ice Mountain for this study include bedrock and surficial geology mapping, very low frequency (VLF) electromagnetic and terrain conductivity geophysical surveys, and limited algific vent air temperature readings. Information published on algific slopes in Iowa provides clues to the circulation mechanism at Ice Mountain. The scientific observations from Ice Mountain, analyzed with the Iowa observations in mind, lead to a model for cold-air production from Ice Mountain that is based on physical principles and supported with empirical data.
Origin of the Algific Talus

Background information pertaining to the environment in which the talus at Ice Mountain originated is crucial for developing a full understanding of the algific slope. Most studies on algific talus have been conducted in the Paleozoic Plateau (“Driftless” Area) of Iowa and adjacent states, so this paper presents the Late Quaternary geological history of the “Driftless” Area as well as the Late Quaternary geological history of the Valley and Ridge of eastern North America. Although there are differences between the Iowa algific talus sites and the Ice Mountain talus site, both areas have undergone similar exposure to periglacial conditions, indirect climatic effects of Late Quaternary deglaciation, and the subsequent northward migration of boreal biota.

Late Quaternary History of the Paleozoic Plateau of Iowa (“Driftless” Area)

The Paleozoic Plateau (“Driftless” Area) of northeastern Iowa contains a mixture of steep slopes, varied slope aspects, bluffs, well-exposed rock outcrops, waterfalls, rapids, sinkholes, springs, and entrenched stream valleys (Hallberg et al., 1984). The bedrock geology and stratigraphy of the area are easily recognized due to very pronounced differential erosion (Hallberg et al., 1984). The area supports microclimates and glacial refugia developed 20 ka during the Wisconsinan glacial stage (Hallberg et al., 1984). The talus slopes and other forms of colluvium that blanket parts of the landscape are thought to have formed under intense periglacial conditions contemporaneous with the close proximity of the Laurentide Ice Sheet (Hallberg et al., 1984). The colluvial slopes were truncated by lateral migration and downcutting of streams as the Laurentide Ice Sheet retreated and melt water flowed through the area (Hallberg et al., 1984). Hallberg et al. (1984) find evidence that meltwater from the retreating Laurentide ice sheet influenced fluvial activity in the Paleozoic Plateau from 14 to 9.5 ka. Deep valley incision, related to the great influx of glacial melt water, caused truncation of colluvium
and collapse of large karst conduits (Hallberg et al., 1984). As the glaciers receded and the climate of the Paleozoic Plateau changed, tundra vegetation was replaced first by fir trees (Abies), and then by pines (Pinus), followed by yellow birches (Betula) and oaks (Quercus) (Sletto, 1994). The complex geomorphic events of the late Wisconsin in the Paleozoic Plateau ended around 10.8 to 9.5 ka with an episode of widespread stream incision (Hallberg et al., 1984).

The late Quaternary geologic history of the Paleozoic Plateau resulted in the formation of two types of relict habitats discovered and documented by Frest (1991). Algific talus slopes and maderate cliffs are both believed to have formed under periglacial conditions (Frest, 1991). Frest (1984) describes an algific talus slope as “a type of habitat developed over the entrances of small fissures and caves in which air circulation and groundwater infiltration from the surface produces more or less permanent underground ice whose incomplete melting produces a constant stream of moist, cool air which filters through a thin plant and litter cover over an extensive rock talus”. Frest (1991) considered only those sites in which ice persisted through August as algific. Seasonal variations in soil temperature of an Iowa algific slope are presented in Figure 4.
Maderate (cold water) cliffs are algific environments with very little talus accumulation (Frest, 1986). The term “maderate cliff” implies the presence of an actively dripping cold water system that is duplicating some portion of a habitat found in colder climates (Frest, 1986). Maderate cliffs tend to have sinkhole-and-fissure systems smaller than those of algific talus slopes (Frest, 1985). The cold airflow and humidity of maderate cliffs, as compared to algific talus slopes, is very subdued because of the lack of insulating talus clasts and the resulting limited accumulation of ice (Frest, 1985). The absence of talus accumulation at the base of a maderate cliff may be attributed to a lack of original accumulation of talus or to fluvial erosion of talus (Frest, 1986). Less than 30 good examples of maderate cliffs have been identified (Frest, 1991), whereas as many as 300 to 400 algific slopes are known in Iowa (Ostlie, 1989).
Late Quaternary History of Eastern North America

Clark and Mix (2002) present modeling data that suggests sea level was between 118 and 135 m (382 and 437 ft) lower than the present at the Last Glacial Maximum, prior to the warming that began 18 ka. Isotropic studies of Summit, Greenland, utilizing the GRIP ice core, show a gradual warming started around 18 ka, catalyzing vegetational response to the deglacial transition (Kneller and Peteet, 1993).

Clark (1968) indicates that parts of West Virginia, including the Ice Mountain area, experienced periglacial conditions during the Late Quaternary Wisconsinan glacial stage. In the Valley and Ridge physiographic province of eastern North America, boreal forest was established between 30° and 40° North latitude prior to 17.5 ka (Delcourt, 1982). A 60 to 100 km (37 to 62 mi) wide belt of tundra vegetation was present near the margin of the ice sheet (Delcourt, 1982). In addition, Maxwell and Davis (1972) report evidence of tundra vegetation as far as 300 km (186 mi) south of the ice sheet border in higher elevations along the crest of the Appalachians. Harsh climate associated with the ice sheets and upland tundra facilitated the development of patterned ground and block fields along the crest of the Appalachians southward to the Great Smoky Mountains (Delcourt, 1982). Greater exposure and an average vertical temperature gradient of 6.33°C/1000 m (3.5°F/1000 ft) (Miller, 1985), led to discontinuous permafrost in higher elevations along the Allegheny Plateau as far south as West Virginia and along the Blue Ridge Mountains to North Carolina and Tennessee (Delcourt and Delcourt, 1986).

Less severe climate in lower, less exposed elevations led to accelerated colluvial processes (Delcourt and Delcourt, 1986). The frequency and intensity of geomorphic disturbances were greater at 38° N latitude than they were at 36° N latitude, demonstrating the
effects of closer proximity to the ice sheets (Delcourt and Delcourt, 1986). The accelerated mass wasting and geomorphic disturbances may have reset plant succession in the affected areas, resulting in dominance of pioneering plants after climate warming indirectly associated with deglaciation (Delcourt and Delcourt, 1986).

Kneller and Peteet (1993) report evidence of an abrupt change in atmospheric and oceanic temperatures between 14 ka and 13 ka based on pollen records and depositional environment mapping in the ridge and valley of Virginia. Up until 12.5 ka, geomorphic processes and vegetation patterns were under a glacial climatic regime (Delcourt and Delcourt, 1986). The abrupt change in climate brought upon a change in the dominant geomorphic regime from colluvial to alluvial processes between 13 ka and 12 ka (Delcourt and Delcourt, 1986). As the ice sheet receded, alluvial processes were fueled by glacial meltwater flowing down the Mississippi to the Gulf coast, resulting in an overall dilution of seawater and rise in sea level (Delcourt, 1982). At 12.5 ka, boreal taxa began to move northward as the ice sheet retreated and the climate became less severe (Delcourt and Delcourt, 1986).

The Valley and Ridge physiographic province became a major corridor for northward-moving taxa (Delcourt and Delcourt, 1986). The shift in climatic and geomorphic regimes changed the vegetation of central eastern North America from the Pleistocene pattern of open patchwork tundra, open glades, and boreal forests to the Holocene pattern of closed, temperate, deciduous forest (Delcourt and Delcourt, 1986). Evidence from 12.5 ka to 10 ka suggests that the combination of climate change, change in geomorphic regime, and northward plant species migration resulted in vegetational disequilibria and dynamically changing landscapes (Delcourt and Delcourt, 1986).
By 10 ka, the ice sheet was virtually gone from the Great Lakes Region, leaving a glaciolacustrine environment in the deglaciated terrain (Delcourt, 1982). Sea level approached its modern position and the Southeast had a near-modern climate regime by approximately 5 ka, following mid Holocene vegetation changes due to increased warmth and aridity associated with the Hypsithermal interval (Delcourt, 1982). According to Delcourt (1982), extensive boreal forests disappeared from the Southeast, with the exception of some spruce and fir populations that were stranded and survived as refugia at higher elevations in the southern Appalachians (Delcourt, 1982). Delcourt (1982) also reports that favorable moist ravines and slope habitats in the southeast provide refugial areas for mesic deciduous forest taxa that would otherwise perish due to climate warming.

Areas disturbed by mass wasting during the full-glacial colluvial geomorphic regime resulted in open patches of exposed forest that were good areas for pioneer herb and shrub communities (Delcourt and Delcourt, 1986). Slopes with coarse mass wasting deposits had the potential to become “permanent” biological refugia for certain boreal vegetation left behind during the northward vegetation migration. The ecosystem of the algific talus at Ice Mountain is likely remnant of a widespread ecosystem in the Valley and Ridge that existed under a colder climate. If so, the Ice Mountain algific talus slope may have been functioning continuously since the Pleistocene.

**Previous Research on Algific Talus Slopes**

Most research on algific talus slopes has focused on the unique biota inhabiting the slopes. Ostlie (1989) states that the 300 to 400 known algific slopes are found only within the unglaciated Paleozoic Plateau (“Driftless Area”) of southeast Minnesota, northeast Iowa, southwest Wisconsin, and northwest Illinois. The Ice Mountain site in West Virginia can be added to the list.
Frest (1986) concluded that the algific sites in Iowa and surrounding areas originated from a combination of mechanical karst development, freeze-thaw cycles, and eroding meandering stream channels. Algific talus in Iowa tends to form on north-facing slopes where large amounts of ice accumulate in jointed bedrock exposed along major drainages (Frest, 1991). The periglacial conditions that formerly existed in the “Driftless Area” facilitated mechanical karst development in the form of ice wedging and slippage of large carbonate bedrock blocks along underlying shale and bentonite layers (Frest, 1986). Mechanical karst is different from more typical solution karst in that it lacks speleothems; has no extensive cave systems; and forms only small sinkholes, vents, and fissures (Frest, 1991). During the development of mechanical karst, the base of the massive overlying carbonate commonly rotates outward, forming small roofed caves and opening vertical to semi-vertical fissures (Bounk and Bettis, 1984). Bounk and Bettis (1984) report that some mechanical karst features are shaped in such a way as to trap cold winter air and hold it throughout the warm summer months. The open vertical fissures, which commonly connect to the roofed caves, result in sinkhole-forming slumps on the surface (Bounk and Bettis, 1984).

Algific talus environments examined by Frest (1991) tend to have massive, capping, cliff-forming units in which sinkhole structures develop following ice wedging and block rotation. A conceptual picture of an Iowa algific talus system is included in Figure 5.
Figure 5: Conceptual model of Iowa algific system. Air is reported to circulate between sinkholes in the upland and algific vents along the river. Air flow is reported to reverse direction from summer to winter, although very little data supporting this idea is available. (Image Source: Image used by permission from Cathy Henry, Iowa Department of Natural Resources)

The massive capping bedrock unit is underlain by a thinner carbonate unit from which the talus originates (Frest, 1991). The algific system is commonly floored by shale that impedes water flow and provides a slippage surface on which larger overlying blocks move (Frest, 1991). Algific slopes do not form under current climatic conditions and active down slope movement on existing slopes has decreased significantly. The algific system of the talus is sometimes stabilized only by the soil that has accumulated in and on them (Frest, 1991). The soil on top of the algific slope and the matrix material between the boulders consists of weathered talus sand, wind-blown silt, and organic material. Soil depths on algific sites surveyed by Frest (1991) are no more than 46 cm (18 in).
Mean annual soil temperatures in the Iowa algific slope environments range only from +10° C to -10° C (50°F to 14°F), with vent air humidity often around 80% at ground surface (Frest, 1991). Many hypotheses have been made concerning the processes responsible for the expulsion of cold air from algific talus. Most researchers agree that the cold air is produced on slopes in Iowa as a result of temperature and pressure differences within and outside of the talus; however, the exact mechanism of airflow is debatable.

The most popular mechanism for airflow appears to be one in which airflow reverses according to outside air temperature. Solution vadose cracks along bedding planes in carbonate bedrock are connected to mechanically developed sinkholes on the overlying upland (Frest, 1991). In the Spring, cold dense air seeps out of the bottom of the slope causing warmer air to enter the fissures and spaces between the talus clasts (Frest, 1991). As the new relatively warm air enters the slope, it is cooled as it passes over ice, becomes more dense, and eventually flows out the bottom of the slope. The intake of warm air and expulsion of cold air continues throughout the warmer months of the year and the ice within the slope is slowly depleted (Frest, 1991). Cold air flows out of the slope until outside air temperatures drop below freezing (Frest, 1991). At that time, the air inside the slope is actually warmer and less dense than the air outside (Frest, 1991). The result is a reverse in the direction of airflow within the slope (Frest, 1991). The warmer, less dense air inside the slope moves upward through the vertical fissures and out the sinkholes on the surface (Frest, 1991). Upward movement of the warmer air inside the talus is enhanced by cool, moist air infiltrating the base of the slope and pushing the warm air up (Frest, 1991). Cold air causes groundwater in the lower parts of the slope to freeze, accumulating ice (Frest, 1991). During the late winter months, when there is a large accumulation of ice within the talus, air circulation may slow significantly. Relatively warmer
air may continue to exit the sinkholes above the slope if outside temperatures are extremely low
(Frest, 1991). The algific talus, in effect, cools the slope area in the summer months and helps to
ameliorate the effects of bad winters (Frest, 1991). The preceding airflow reversal mechanism
has been documented by Frest (1991), but the author of this paper found very little data to
support the idea and no evidence to suggest that the mechanism occurs at Ice Mountain.

Core (1975) reported the cold air production at Ice Mountain to be a result of circulation
of air and water between the boulders of the talus slope. He concluded that cold winter air,
mixed with water from rain, snow and snowmelt, is able to form a large mass of interstitial ice
several feet below the surface. According to Core (1975), the ice accumulation becomes thick
enough that the upper layer of ice, coupled with the insulating effects of the talus clasts, acts as a
protective layer shielding the deeper ice from melting during the onset of warm weather. In the
summer, air circulation in the interstitial spaces of the slope allows outside air to enter the slope,
cool in contact with the ice, sink because of its higher density, and subsequently flow out of the
bottom of the slope (Core, 1975).

**Bedrock Stratigraphy**

The Ice Mountain talus is situated in the Valley and Ridge Province of the Potomac
Highlands of West Virginia (Figure 6). The area is characterized by long, narrow mountains
parallel to one another and trending in a northeast-southwest direction of approximately 030.
Mountains in this area are capped by resistant sandstone and valleys are underlain by less
resistant siltstones, shales, and carbonates (Lessing *et al.*, 1999).

From west to east, the study area is underlain by Devonian Marcellus Formation-
Needmore Shale under the North River valley and the west flank of Ice Mountain, near-vertical
cliffs of Devonian Oriskany Sandstone near the summit of the mountain, and Devonian-Silurian
Helderberg carbonates on the mountain’s east slope (Lessing *et al.*, 1999) (Figure 7).
Figure 6: Location of Ice Mountain talus in the Potomac Highlands of Hampshire County, West Virginia (Source: West Virginia GIS Technical Center)
The Marcellus Formation-Needmore Shale is 90 to 120 m (300 to 400 ft) thick in the area. The Marcellus section of the shale is black or dark gray, marine, fissile, and tends to break into small platy chips (Woodward and Price, 1943). The Needmore section of the shale, roughly equivalent to the Onondaga shale and Selinsgrove lower shale, is medium to dark gray or greenish-gray to brownish-black. The Needmore is structurally incompetent and weathers easily with a blocky or chunky fracture (Woodward and Price, 1943). As a result, Needmore bedrock tends to form low ground and is not usually well exposed due to a mantle of weathered material from the Oriskany Sandstone (Woodward and Price, 1943).
The Oriskany Sandstone is light gray, medium to coarse-grained sandstone representative of reworked beach sediment (Woodward and Price, 1943). The Oriskany contains conglomeratic zones and fossil molds predominantly of brachiopods (Woodward and Price, 1943). Woodward and Price (1943) state that the texture of the Oriskany Sandstone can vary significantly at a single locality. The contact between the Needmore Shale and the Oriskany Sandstone is marked by an obvious lithology change. However, the contact is usually not visible in surface outcrop due to differential resistance to erosion of the two rock types and a mantle of Oriskany colluvium (Woodward and Price, 1943). Due to the erosional resistance of the Oriskany Sandstone, many outcrops are ridge formers, commonly as steep cliffs that supply large amounts of sandstone blocks and boulders to slopes below (Woodward and Price, 1943). Raven Rocks, an overlook located on the southern tip of Ice Mountain, is an example of such an outcrop (Woodward and Price, 1943). Woodward and Price (1943) report the Oriskany Sandstone to be approximately 30-45 m (100-150 ft) thick in northeastern West Virginia. Stewart Dean (Personal Communication, 2001) reports the Oriskany to be no more than 36 m (120 ft) thick in the Ice Mountain area. Field mapping, as well as drill hole logs, examined by Woodward and Price (1943) suggests that the Oriskany tends to thin regularly rather than abruptly, most likely a result of beach deposition. Dean (2001) also reports that the Oriskany is thin enough to be breached in areas north of Ice Mountain where stream erosion has exposed the underlying Helderberg carbonates. The thickening and thinning of the Oriskany, as well as the presence of sandstones in the upper section of the underlying Helderberg Group, has been the cause of over-estimation of thickness and section measuring mistakes. The contact between the Oriskany Sandstone and the underlying Helderberg Group is gradational, also contributing to confusion about the lower
boundary of the Oriskany (Woodward and Price, 1943). Woodward and Price (1943) note some evidence to suggest that there is an unconformity between the Oriskany and Helderberg.

Woodward and Price (1943) suggest that a boundary can be discerned between the Oriskany and the upper sandstones of the Helderberg Group based on the darker color, more calcareous nature, and presence of glauconite in the Helderberg sandstones. Lessing et al. (1999) report that a 3 to 4 m (10 to 15 ft) thick sandstone, known as the Elbow Ridge sandstone, commonly occurs near the top of the Helderberg Group. The Elbow Ridge sandstone may be an equivalent to the upper Helderberg sandstones discussed by Woodward and Price (1943). In general, the Helderberg Group is mostly massive-bedded, gray limestone 120 to 200 m (400 to 650 ft) thick, with abundant marine fossils. The uppermost mapped carbonate unit in the Helderberg is known as the Port Jervis Limestone (Woodward and Price, 1943). The Port Jervis Limestone contains a dark gray cherty section near its top, and is hard to recognize as a limestone in western Hampshire County (Woodward and Price, 1943).

**Structural Geology**

Ice Mountain lies on the west side of the Timber Mountain anticline and is west of the North Mountain fault, which means it is on the Martinsburg allochthonous sheet (Figure 8).

Lessing et al. (1999) indicate the strike of the rocks in the vicinity of Ice Mountain to be approximately 030° to 035°. Lessing et al. (1999) show the contact between the Marcellus Formation-Needmore Shale and the Oriskany Sandstone to be at an elevation of approximately 260 m (860 ft) above sea level. The contact between the Oriskany Sandstone and the underlying Helderberg Group in the study area is mapped as coinciding almost exactly with the spine of the ridge. Elevation of the Oriskany-Helderberg contact varies from 360 to 375 m (1180 to 1230 ft) above sea level as the ridge line rises and falls (Lessing et al., 1999).
Lessing *et al.* (1999) do not indicate any near-surface faulting in the immediate vicinity of the Ice Mountain talus. A small segment of the North River (thrust) Fault is mapped north of the talus in the Mahantango Formation, cropping out parallel to strike and dipping to the southeast (Lessing *et al.*, 1999). The northwest thrust of the fault has moved older, deeper rocks closer to the surface; however, there is no evidence that the fault has any direct influence on the geology underlying the Ice Mountain talus.
Figure 8: Geologic cross-section illustrating the Timber Mountain anticline and deep, west-thrusting faults. (Source: Lessing, et al., 1999)

Dmt – Mahantango Formation  
Dmn – Marcellus-Needmore Shale  
Do – Oriskany Sandstone  
Dhl – Helderberg Group (top of DS)  
DS – Silurian System  
Smc – McKenzie Formation  
Srh – Rose Hill Formation  
St – Tuscarora Sandstone  
Oj – Juniata Formation  
Oo – Oswego Sandstone  
Om – Martinsburg Formation
Results of Bedrock and Surficial Geology Mapping

Bedrock Geology Mapping

Geologic mapping at 1:24,000 scale by Lessing et al. (1999) did not provide enough detail to address the geology of the very small area occupied by the algific talus or the role each rock type plays in the cold-air production process. Bedrock lithology and spatial relationships among the different lithologies in the slope area are likely to be very important to the production of cold air, so a more detailed examination of the bedrock geology was necessary.

Geologic mapping of the Ice Mountain talus area was carried out using a Brunton azimuth compass, a 30 m (100 ft) measuring tape, and a handheld Garmin 12-Channel Global Positioning System unit. The mapping covers a limited area centered around the algific vents and extending outward as was necessary to establish reliable geologic control.

Five days of geologic mapping during Fall 2001 revealed patterns similar to the Lessing et al. (1999) report, with some notable differences (Figure 9). The highly deformed Marcellus Formation – Needmore Shale appears to have a near vertical to vertical dip in most outcrops; however, due to its deformed nature, no dip angles are reported in this study. Similarly, Lessing et al. (1999) do not display any dip angles for the Marcellus Formation – Needmore Shale west of Ice Mountain.
Figure 9: Bedrock geology map of the Ice Mountain study area. GPS points indicate important mapping observations. Dip direction is northwest.
Figure 10: Geologic cross-section A-A’ illustrating bedrock geology beneath the main talus area.
Figure 11: Geologic cross-section B-B’ illustrating bedrock geology beneath colluvium and talus near northeastern extent of study area.
Outcrops of Needmore Shale occur in the area around the GPS station labeled SADDLE. The outcrops are visible near a small, man-made pond located in the saddle. However, due to the less resistant nature of the shale, much of the bedrock mapping was done by observing shale chips in residual soil. The most important outcrops of shale occur at points labeled 12 and SHS3 in Figure 9. The presence of shale at station 12, and the presence of shaley soil at station SHS3, requires the contact between the Marcellus Formation-Needmore Shale and Oriskany Sandstone to be mapped further to the east than reported by Lessing et al., (1999). The contact, which was mapped by Lessing et al., (1999) at an elevation of 265 m (860 ft) above sea level and by Tilton et al., (1927) at 231 m (750 ft) above sea level, actually occurs at an elevation of approximately 350 m (1150 ft) above sea level on the northwest side of Ice Mountain. This eastward shift of the contact confirms that virtually the entire aligific area (cold producing area) of the talus slope is underlain by shale.

Good outcrops of the Oriskany Sandstone occur where Hiett Run cuts through Ice Mountain, south of Raven Rocks. Lessing et al., (1999) show a 76° northwest dip in this area. Similarly, the average dip measurement recorded from Hiett Run in this study was 72° northwest. The dip measurements for the Oriskany Sandstone range from 70° northwest to vertical at Raven Rocks, which forms a cliff famous for its beautiful overlook. Just north of Raven Rocks, along the ridgeline of Ice Mountain, the Oriskany Sandstone cliffs are absent and a large area of talus covers the mountain from the Oriskany Sandstone-underlain ridge top to the Marcellus Formation-Needmore Shale-underlain bank of North River. The Oriskany Sandstone reappears as a very fractured, isolated outcrop of near-vertical (70° northwest to 90°) beds at OCLIFF. The Oriskany Sandstone crops out as a semi-continuous line of vertical to near-
vertical cliffs from OCLIFF northeast along the Ice Mountain ridge at least as far as the northeast limit of the study area.

The east side of Ice Mountain is underlain by Helderberg Group carbonates. Outcrops of Helderberg carbonates vary in fossil content and show minor solution features in fractured sections. The carbonates occur in a small outcrop labeled LS1 on the east side of Hiett Run water gap. No reliable dip measurement could be attained at LS1. Limestone crops out at station 20, just east of the ridge top. The dip of the limestone in this outcrop is 25° northwest, much less than the vertical dip of the overlying Oriskany Sandstone exposed at nearby Raven Rocks. Limestone with dips ranging from 24° to 53° northwest crops out along the southeast side of the ridge at elevations from 364 to 377 m (1180 to 1220 ft) above sea level. The more gentle and varying dip of the limestone, compared to the overlying Oriskany, suggests that the apparent limestone “outcrops” may be out of place, collapsed beds (Figure 12)

(no scale, conceptual use only)

Figure 12: Illustration of collapsed limestone bed as suggested by outcrop measurements collected on the southeast side of the Ice Mountain ridgeline. Profile line similar to A-A’ on Bedrock Geology Map.
Strike of the limestone outcrops is consistent with the overlying Oriskany, with the exception of locations 27 and 29 where strikes measured 000° and 015°, respectively. The apparent shift in strike direction may indicate the edge of a collapsed layer of limestone. The anomalous strike measurements at 27 and 29 may be evidence that limestone blocks in that area rotated counterclockwise (as viewed from the southeast) toward the collapse.

A geologic map and cross-sections constructed using data from this study (Figures 9-11) suggest the possible presence of a thin sandstone unit, known as the Elbow Ridge sandstone, in the upper part of the Helderberg Group. Lessing et al. (1999) noted the Elbow Ridge sandstone as a 3 to 4.5 m (10 to 15 ft) thick sandstone. Dean (personal communication, 2001) reports that the Oriskany Sandstone is no more than 37 m (120 ft) thick in the area of the Ice Mountain talus. Assuming a maximum 37 m (120 ft) thickness and a 70° dip for the Oriskany, geologic mapping and cross-sections suggest that portions of the Ice Mountain ridgeline near cross-section B-B’ are underlain by the upper section of the Helderberg, which may include the Elbow Ridge sandstone (Figures 9 and 11). Vertical beds of Oriskany crop out on the west slope of the ridge, forming steep cliffs well below the ridgeline. No indisputable outcrops of the Elbow Ridge sandstone were identified during field mapping, therefore the presence of the Elbow Ridge sandstone beneath the ridgetop in the northeastern-most part of the study area is based entirely on thickness and dip angle geometry.

**Surficial Geology Mapping**

Surficial geologic mapping was accomplished utilizing orthophotoquad images and field observations in order to evaluate spatial relations of geomorphic features associated with the preserve area. The surficial map includes delineation of the algific talus, colluvium and residuum areas; locations of cold-air vents; morphological features in the boulder-covered areas; and other
geomorphic features relevant to this study (Figure 13). Surficial mapping distinguished the unvegetated talus areas from more-vegetated, colluvial areas on Ice Mountain. Unvegetated areas covered with large sandstone boulders were mapped as talus. Vegetated areas covered with sandstone boulders and a layer of organic or detrital soil were mapped as colluvium. Field mapping and slope measurements indicate that the more vegetated colluvium on the west slope of Ice Mountain tends to be steeper and less stable than unvegetated talus. Typically, talus is composed of larger rocks and colluvium is composed of smaller rocks and unstable vegetation. Morphology of the surface of the talus and colluvium were examined under the assumption that features and topography changes at the surface of the boulders are indicative of underlying features. The main algific zone lies at the base of the large talus area, just above the east bank of the North River in the vicinity of the GPS station labeled BIG1. Cold-air producing vents occur along an access trail on the east bank of the river between station SIGN and station ICE. The algific vent area includes colluvium covered in a thin layer of soil and vegetation (including mosses, small flowering plants, and trees). Approximately 18 to 25 m (60-80 ft) vertically above the algific zone, linear depressions in the talus originate near the down-slope edge of a topographic bench (located at GPS station BENCH1). The tops of the depressions are spaced along the edge of the bench and coalesce above the algific zone. The linear depressions may be an indication of a subsurface drainage network that channels water into the area above the algific zone. The bench from which the linear depressions originate was investigated with very low frequency (VLF) and terrain conductivity geophysical methods and is discussed in the geophysics section.
Figure 13: Surficial geology map of Ice Mountain area.
A second area, north of the main algific zone, also contains distinct morphology on the surface of the colluvium. GPS Stations NB, EB, SB, and WB delineate a large bowl-shaped depression in the colluvium surface (Figures 13 and 14). The depression is visible on the 7.5 minute topographic map as a relatively flat area with two undulations in the contour lines, coinciding with the north and south edges of the bowl-shaped depression. Kite (2002) suggests that the north and south edges of the depression may be slope failure lobes, and that the west edge of the depression is a bar of boulders reworked by North River.

Figure 14: Bowl-shaped depressions on northern end of Ice Mountain study area
(Photo: J.S. Kite)

Another linear depression is located just east of the Ice Mountain ridgeline between stations 37 and 38 (Figure 13). The depression runs parallel to strike on the southeast side of the mountain between the top of the ridge and limestone outcrops discussed in the geology section,
directly opposite the main talus area. The origin of the depression is unknown. Results of a terrain conductivity survey conducted over the depression are discussed in the Geophysics section.

Examination of the Ice Mountain area on a U.S.G.S. 7.5 minute Quadrangle topographic map suggests that a meander bend (MB1) of the North River was once further eastward than it is today (Figure 15). North River abruptly changes direction and narrows near the algific zone where the sandstone talus is first encountered. Topographic contour patterns above the algific zone show slight undulations representing a hollow extending upward from the meander bend (MB1) to the break in the ridgeline at the top of Ice Mountain. The hollow may suggest the North River left a terrace on the west slope of Ice Mountain prior to colluviation. The undercutting and oversteepening effects of the meander bend (MB1) are likely to be the cause of the hollow feature.

During mass wasting associated with periglacial conditions, sandstone talus accumulated on the east bank of the meander bend. The evolution of the meander bend at the base of the talus is unclear. Two scenarios, detailed in the Discussion section, are presented in Figures 16a-16b. Truncation of the toe of the talus accumulation on the East bank of North River plays a major role in the formation of the unique algific environment; therefore, attempting to understand the evolution of the present geomorphology provides additional clues to the mystery of Ice Mountain.
Figure 15: Topographic map of Ice Mountain area. Slight undulations in the contour lines are evident on the northwest side of Ice Mountain, extending from the meander bend (MB1) to the ridge top. Curved red lines exaggerate area of contour undulations (Map Source: West Virginia GIS Technical Center)
Results of Algific Vent Air-Temperature Data

Limited air-temperature data for two algific vents at Ice Mountain (GPS points “Big One” and “Other One”) was obtained with standard alcohol thermometers kept in the vents by the Nature Conservancy (Figure 17). Air-temperature measurements in the vents were susceptible to error due to instrument malfunction, human observation error, and error associated with interaction between vent air temperature and ambient air temperature.

Although the accuracy of the measurements taken with these thermometers may be questionable, it was assumed that relative changes in vent air temperature over the time of data collection were representative.

The air temperature in the algific vents at Ice Mountain is not a constant 3.3°C (38°F), as reported in some newspaper articles. With values as high as 9.4°C (49°F) and as low as 31°C (-0.5°F), Figure 17 indicates an overall decrease in vent air temperature from late summer to mid-winter. Temperature fluctuations from summer to winter are attributed to seasonal variations in outside air temperatures as well as depletion of potential ice in the slope.
Periglacial conditions initiate colluviation.

Talus accumulation occurs on east bank of North River.

Talus prevents further erosion on cut bank of meander bend and river adjusts by moving west.

Down cutting results in present river level. Abandoned meander terrace is left filled with talus and elevated above present river level.

**Figure 16a:** Creation of abandoned meander terrace on west slope of Ice Mountain as observed along cross-section A-A’.
Colluviation Begins

Colluviation Begins

The River Downcuts And Naturally Migrates East, Truncating The Toe of the Slope and Exposing the Vents

Figure 16b: Alternative scenario for creation of abandoned meander terrace on west slope of Ice Mountain as observed along cross-section A-A’.
Figure 17: Limited algific vent air temperatures collected from two Ice Mountain vents.

Geophysical Investigations

Very Low Frequency (VLF) Survey: Background Information

The very low frequency (VLF) method is an electromagnetic induction (EMI) method used in this study to locate conductivity anomalies and identify conductivity trends within the interior of the slope. The VLF method is suitable for work on the talus slope because of its 0 m to greater than 200 m depth of investigation, which covers the range of depths associated with talus. The VLF instrument is portable and there is no need for probes to be placed in the ground.

VLF has been used at the United States Geological Survey Fractured Rock Research Site in Mirror Lake, New Hampshire, to locate and trace fracture zones in crystalline bedrock (Powers et al., 1999). The VLF technique relies on large-scale military transmitters as the primary source for electromagnetic (EM) waves. A typical military VLF transmitter consists of a vertical cable several hundred meters long emitting a 300 to 1000 kw signal at a frequency of 15 to 25 kHz (Telford et al., 1976; McNeill and Labson, 1987). The VLF transmitter creates an electromagnetic field in which the electric field component is vertical and the magnetic field component is horizontal. Magnetic field lines emitted by the transmitter ripple outward in a
concentric pattern (Telford, et al., 1976). The waves of this field can be considered plane waves as long as the VLF transmitting station is greater than 80 km (50 mi) from the study area (Powers et al., 1999). As the magnetic field passes through low resistivity (high conductivity) material in the subsurface, secondary currents are induced. The secondary currents create magnetic fields of their own that are out of phase with the primary field (Telford, et al., 1976).

The two conventional methods using VLF data for subsurface mapping are tilt angle and resistivity (Powers et al., 1999). The tilt angle method relies on secondary, out of phase waves created by subsurface features that can be distinguished from the primary, in phase waves of VLF transmitters (Powers et al., 1999). The resistivity (or signal intensity) method measures the apparent resistivity and the delay between receipt of the electric and magnetic fields of the VLF transmitter waves (Powers et al., 1999). Neither of the conventional methods were utilized in this study. Instead, a third, more simplified method was used at Ice Mountain.

For this study, variations in horizontal field intensity were measured over the study area. Raw VLF intensity measurements from Ice Mountain were corrected for drift between consecutive base-station measurements and plotted in the Surfer contouring software, enabling interpretation in conjunction with field mapping and terrain conductivity data. Drift in the signal intensity measurements was corrected relative to a drift curve. Base-station measurements of VLF signal intensity were taken periodically to determine the amount of nongeological variations in signal intensity during the survey. The VLF operator isolated the changes in signal related to subsurface anomalies by correcting the data with respect to natural signal drift. Drift correction measurements were made at a base-station over relatively short time intervals of approximately one hour. Drift between consecutive base-station measurements was assumed to
vary linearly. Field intensities measured between consecutive base-station measurements were increased or decreased in linear proportion to the time elapsed between drift measurements.

The raw VLF data are the amplitudes of the secondary magnetic field created by low-resistivity (high-conductivity) areas (Ackman, 1996). The exploration depth of the VLF survey is dependent on transmitter frequency and net ground resistivity. Exploration depth is approximated as two-thirds of the calculated skin depth. Skin depth is defined as:

$$500\sqrt{\frac{\rho}{f}}$$ (McNeill, 1990),

where $\rho$ is resistivity (Ω·m) and $f$ is frequency (Hz). Resistivities estimated from the terrain conductivity measurements suggest that VLF skin depth in the area is approximately 100 m (330 ft). Higher VLF readings are indicative of higher-amplitude, out-of-phase, secondary fields created by high-conductivity subsurface anomalies (Ackman, 1996). Studies show that “subsurface horizontal electric fields are responsible for virtually all VLF anomalies” (Ackman, 1996). One objective of the VLF survey was to collect many data points over a large area in a relatively short time in order to uncover anomalous areas warranting more detailed investigation.

**Results of VLF Survey**

Seven VLF transect lines were delineated on the west side of Ice Mountain in an area outlined by GPS stations GOVLF2, VLF7N, VLF7S, and L-1STP (Figure 18). Transect lines were delineated using a Garmin 12-Channel GPS unit, a Brunton compass, and a standard 100-foot measuring tape. The Cutler, Maine, VLF transmitting frequency was used as the primary in-phase VLF signal, resulting in a perpendicular relationship between the horizontal magnetic field vector and the strike of local bedrock structure. The azimuth of a line between Ice Mountain and the Cutler transmitter lies in a northeast direction, roughly along strike. A perpendicular relationship between VLF waves and geologic strike enhanced signal variation related to
lithologic contacts, while still showing signal variations related to across-strike subsurface features.

The transect lines vary slightly in length, the longest measuring 570 m (1850 ft) with a northeast-southwest azimuth of 049°. The VLF survey covered a predominantly talus-covered area of approximately 61,300 m² (660,000 ft²), beginning upslope of the main algific vent area at approximately 250 m (800 ft) above sea level, and extending to an approximate elevation of 320 m (1050 ft) above sea level. Relatively high VLF measurements define an anomaly 309 to 370 m (1000 to 1200 ft) southwest of VLF3N, parallel to the transect lines (Figure 19).

The positive VLF anomaly (Figure 19) occurs on, and down slope of a small topographic bench in the talus labeled BENCH1 (Figure 13), the same bench discussed in the Surficial Geology section. The high intensity anomaly is interpreted to be an area of moisture or water accumulation. The VLF anomaly is located upslope of the algific vent zone and corresponds with the area of coalescing linear depressions in the talus surface. Several EM-34 terrain conductivity profiles were surveyed across the VLF anomaly, as discussed in a later section.
Figure 18: VLF Survey Location Map showing approximate locations of all 7 transect lines. GPS points shown were used to delineate the outer corners of the VLF survey.
**Electromagnetic Conductivity (EM): Background Information**

Terrain conductivity is a useful method to employ at this site because it does not require direct injection of electrical current into the subsurface as required by the electrical resistivity method. Resistivity surveys provide greater flexibility in the electrode spacing and number of offsets that can be measured; however, resistivity was not used because proper electrode coupling necessary for the resistivity survey would be impossible to achieve in the talus.

Electrical conductivity is a measure of the ease with which an electrical current can flow through a material. It is the reciprocal of resistivity. The MKS system unit for conductivity is mho/m, meaning that a conductivity of one mho/m is equivalent to a resistivity of 1 ohm/m (McNeill, 1980). Conductivities are generally represented in millimhos/m.

The physical basis of the terrain conductivity survey with the Geonics EM-31 and EM-34 conductivity meters is similar to the basis of the VLF survey (Ackman, 1996). However, the VLF survey is conducted using only a receiver and involves measurement of strength or tilt.
angle of a signal generated by a stationary source well outside the study area. Terrain conductivity survey, on the other hand, incorporates coupled source and receiver coils that can be positioned at different coil spacings. Larger coil spacings provide greater depth of signal penetration (Figure 20) (McNeill, 1980). In addition, the coils can be oriented to produce horizontal or vertical electromagnetic dipole fields. The horizontal dipole response is affected by materials distributed over a wider horizontal area and shallower depth than the vertical dipole response (McNeill, 1980). Measurements of terrain conductivity made at different intercoil separations provide information about conductivity variation as a function of depth.

![Figure 20: Approximate depth of penetration associated with different coil spacing for the Geonics EM-34 terrain conductivity meter. Spacing 1, 2, and 3 represent 10, 20, and 40 m intercoil spacings, respectively. (Source: Benson and Glaccum, 1988)](image)

Terrain conductivity is based on the analysis of electromagnetic fields induced by current flow. An alternating current is circulated through a transmitter coil, creating a magnetic field that penetrates the subsurface and induces current flow in conductive subsurface intervals.
Subsurface current flow creates a secondary electromagnetic field (McNeill, 1980). Current flow induced in the receiver coil is the net result of electromagnetic fields generated by both the transmitter coil and the secondary subsurface fields (McNeill, 1980). Comparison of current flow in the transmitter receiver coils yields a measure of the subsurface conductivity (McNeill, 1980). Apparent conductivity is a weighted combination of the conductivities of all the subsurface layers in which current was induced. The subsurface distribution of conductivity variations giving rise to the apparent conductivity can be estimated through model calculations. Models are non-unique. Different distributions of subsurface conductivity can give rise to the same apparent conductivity; therefore subsurface knowledge is critical to the development of realistic conductivity models.

Electrical conductivity of soils and rocks is controlled primarily by moisture-filled pores within the matrix of the material (McNeill, 1980). Therefore, variations in subsurface conductivity are caused by changes in soil-moisture content, groundwater-specific conductance, depth of soil over bedrock, and thickness of soil and rock layers (Benson et al., 1982). Conductivity readings obtained from subsurface surveys are primarily dependent upon porosity of the rocks, degree of pore space saturation, concentration of dissolved electrolytes in the pore water, as well as pore-water temperature and phase state (McNeill, 1980). Clay-rich weathered material in interstitial spaces between boulders may have a large influence on conductivity readings. Liquid water content of the slope material varies with season and short-term weather events. Conductivity readings vary accordingly, with ice having a very low conductivity compared to water because of slow ion movement in ice (McNeill, 1980). Studies indicate a change in conductivity of 2.2%/°C (McNeill, 1980).
Because terrain conductivity surveys of algific talus have never been conducted, a comparison with surveys from similar environments is warranted. The subsurface of an algific talus slope is believed to have characteristics similar to permafrost layers. The definition of permafrost requires only that the “mean annual temperature of ground be less than 0° C (32° F) for several years” (McNeill, 1980). Some of the ice that builds up in the Ice Mountain talus, whether interstitial or massive, may persist all year as suggested by the constant expulsion of colder 3-7° C (38-45° F) air from the algific vents at the base of the slope. Therefore, some earth material in the talus is likely to remain below freezing throughout the year.

Permafrost research contributes two important ideas to the study of the algific talus, both of which involve the conductivity response of intermixed ice and soil layers. The first idea is that the conductivity of frozen sandy soil is less than the conductivity of frozen, clay-rich soil due to incomplete ice formation in the clay-rich soil (McNeill, 1980). At very low temperatures, ions in the clay orient nearby water molecules and prevent total ice formation (McNeill, 1980). The water that does not freeze collects more dissolved ions, resulting in higher conductivity values (McNeill, 1980). The second important idea contributed by permafrost research is that a massive body of permafrost produces a more abrupt change in resistivity (or conductivity) and a laminated permafrost produces a more gradual change in resistivity (or conductivity). Daniels et al., (1976) illustrate the resistivity response of a laminated body of permafrost and compare it to the resistivity response of a massive body of permafrost (Figure 21).
Data collection techniques for terrain conductivity included profiling and sounding measurements. Profiling consisted of taking conductivity measurements along a transect line using a single intercoil spacing (Benson et al., 1988). Performing a sounding consisted of making conductivity measurements at varying intercoil spacings centered around a single point (Benson et al., 1988). Terrain conductivity soundings were compiled along transect lines to create a cross-section of terrain conductivity variation. Continuous profile horizontal and vertical dipole conductivity measurements were made using the Geonics EM-31 terrain conductivity meter. A sounding consisted of measurements at three different coil separations obtained using the Geonics EM-34 terrain conductivity meter. Generally, the EM-31 and the EM-34 were used together to increase the depth range covered by the sounding. A combination of sounding and profiling was employed in this study.

**Results of EM-31 Surveys**

Initial EM-31 reconnaissance surveys were conducted in March 2001 near the main algific zone (Survey 1) and across a linear depression on the east side of the Ice Mountain ridgeline (Survey 2) (Figures 22a and 22b). These initial surveys revealed that the shallow
subsurface has nearly zero conductivity. Zero conductivity in the upper portions of the talus was suspected to result from air-filled pore spaces associated with void space between talus blocks. It was hypothesized that water drains down through the highly permeable air-filled pore spaces of the talus to a deeper aquitard, but no such aquitard was detected in the shallow subsurface near the algific zone using both horizontal and vertical dipole settings of the Geonics EM-31. This implied that no aquitard layer was present in the shallow subsurface near the algific zone.

The March 2001 EM-31 survey on the southern side of the main algific zone (Survey 1) indicates an area of high conductivity, identified as a small spring (GPS station SPRING) discharging into the North River near the southern edge of the slope (Figures 22a and 22b). EM-31 readings are consistently zero nearer the center of the talus-covered area and increase noticeably at the transition from talus to shaley soil near the edges of the talus.

The March 2001 EM-31 Profile Survey 2 was conducted to investigate conductivity in the shallow subsurface of a linear depression at the top of the talus (Figures 22a and 22b). Zero conductivity readings were also collected from Survey 2.

Vertical and horizontal dipole EM-31 conductivity readings taken in November and December 2001 along the topographic bench above the main algific zone (Bench 1) were all zero, indicating a lack of conductive material near the surface.

**Results of EM-34 Surveys**

Terrain conductivity soundings conducted with the EM-34 using 10, 20, and 40 m intercoil spacings provided increasingly deeper views of conductivity within the subsurface. An EM-34 survey was conducted on the topographic bench above the main algific zone in November and December 2001, prompted by the anomalous VLF readings in that area. Two EM-31 readings and six EM-34 readings were taken at six different locations spaced 31 m (100 ft) apart along the topographic bench in the talus (Figures 22a and 22b). Two additional
sounding locations, labeled S-7 and S-8 are 15 m (50 ft) horizontally upslope and down slope of sounding location S-3 (Figures 22a and 22b). Independent sounding data were collected on two different days to evaluate the reproducibility of the sounding data (Figures 23a and 23b).

An automated inverse modeling program, EMIX34, was initially used to estimate subsurface conductivity distributions from the sounding data. However, EMIX34 evaluates the best fit between the model and observed data based on percent (%) difference. Since the observations from Ice Mountain contained very low conductivities, small changes in calculated conductivities less than one millimho/m commonly resulted in 100% error. An alternative forward modeling program using mathematical relationships between apparent conductivity and the distribution of subsurface conductivities as described by McNeill (1980) was programmed in Excel (Tables 1 and 2). The match between model calculations and observations was evaluated using a root mean square error (RMS error) instead of a % error. The root mean square error provides a more accurate measure of error than the percent difference calculation used by EMIX34.
Figure 22a: Terrain conductivity survey location map indicating locations of EM-31 and EM-34 surveys, as well as two main algific vent areas and a small spring.
Figure 22b: Terrain conductivity survey location map on an aerial photography base map.
Figure 23a: EM-34 sounding results for 16 Nov 2001.

Figure 23b: EM-34 sounding results for 20 Dec 2001.

NOTE: Conductivity results from both days are similar. V and H signify vertical and horizontal dipole configurations, respectively; 10, 20, and 40 refer to the coil spacing, in meters, used for each measurement.
The basic equation used to compute apparent conductivity for either horizontal or vertical dipole configuration in a three-layer model is:

$$\sigma_a = \sigma_1[1-R(z_1)] + \sigma_2[R(z_1)-R(z_2)] + \sigma_3[R(z_2)]$$

(McNeill, 1980),

where $\sigma_a$ is apparent conductivity measured with the conductivity meter; $\sigma_1$, $\sigma_2$ and $\sigma_3$ are the conductivities of layers 1 through 3 respectively; and $R(z)$ is the response function. The horizontal response function is:

$$R_h(z) = (4z^2 + 1)^{1/2} - 2z$$

(McNeill, 1980),

where $z = \text{depth to the top of a layer divided by the intercoil spacing}$. The vertical response function is:

$$R_v(z) = 1/(4z^2 + 1)^{1/2}$$

(McNeill, 1980),

where $z = \text{depth to the top of a layer divided by the intercoil spacing}$. Root mean square (RMS) error is used to calculate the error between the observed and calculated (or modeled) and calculated (or modeled) conductivity values. The root mean square error equation is

$$\text{RMS} = (\sum (E)^2/n_t)^{1/2},$$

where $n_t$ is the total number of observations and $E$ is the error or difference between calculated and observed conductivity values.
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### Depth to top

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| 1 thru 3 | RMS error= 0.3469 mmhos/m |

**Table 1:** Excel spreadsheet used to model terrain conductivity data from soundings S-1 through S-3. Layer conductivities and depths to tops of layers obtained from modeling are illustrated in Figure 25. An RMS error of 0.35 suggests that the model is a reasonable scenario.
### Table 2: Excel spreadsheet used to model 3-layer terrain conductivity data from soundings S-4 through S-6. Layer conductivities and depths to tops of layers obtained from modeling are visually presented in Figure 25. An RMS error of 0.79 suggests that the model is a reasonable scenario.
The six EM soundings were divided into two groups, based on the similarity of measured conductivity values. Soundings S-1 through S-3 were modeled for a more shallow three-layer system and soundings S-4 through S-6 were modeled for a slightly deeper three-layer system with relatively greater average conductance. The characteristics of both of the three-layer systems were based on the conductivity pseudo-sections. Average RMS error values were calculated for each of the two groups of soundings. The Excel spreadsheet allowed the modeler to view the difference in measured and calculated conductivity values for each individual coil spacing and dipole configuration. Large differences in measured and calculated conductivity values identified observations that were poorly explained by the model.

Effective penetration depths calculated by EMIX34 enable conductivity data to be represented in profile form as conductivity pseudo-sections. Horizontal dipole and vertical dipole pseudo-sections are presented for the December 2001 survey (Figures 24a through 24d). Contouring of the pseudo-section data was done utilizing both Kriging and the modified Shepard’s method. Kriging is a geostatistical method using a weighted, moving average estimation of the data being contoured. The modified Shepard’s method is an inverse-distance weighted least-squares method, used because it helps eliminate “bulls-eye” effect around individual data points.

Contouring of the EM-34 horizontal dipole measurements from December 2001 indicate 3 or 4 isolated, high conductivity areas at an apparent depth of approximately 10 m (30 ft) using the Kriging contour method (Figure 24a). The modified Shepard’s method produces a smoother, continuous, more realistic anomaly. The conductivity anomaly is likely to be associated with the talus block/shale bedrock interface.
Vertical dipole measurements for soundings S-1 through S-6 (Figure 24b) indicate an area of high conductivity at a depth of 10 to 40 m (32 to 130 ft), beginning below sounding S-4 and concentrated below soundings S-5 and S-6. The higher magnitude, deeper conductivity anomaly beneath S-4 through S-6 is likely a result of more extensively weathered and fractured bedrock allowing for increased moisture content. A small spring located down slope of S-4 and S-5 during the March 2001 EM-31 survey (Figures 22a and 22b) also suggests the presence of increased moisture in the area.

Two additional soundings (S-7 and S-8) form a second transect line perpendicular to S-1 through S-6, with S-3 at the intersection (Figure 24c and 24d). S-7 is 15 m (50 ft) horizontally upslope of S-3 and S-8 is 15 m (50 ft) horizontally down slope of S-3. Both vertical and horizontal dipole measurements show an increase in conductivity in the down slope direction (Figures 24c through 24d). The horizontal dipole measurements suggest a nonconductive layer below S-3 (Figure 24c), which may be evidence of a more impermeable bedrock surface. Horizontal and vertical dipole measurements from S-8, located down slope of the topographic bench in the talus, do not indicate such an impermeable surface. The presence of the impermeable surface below S-3 and the lack of impermeable surface below S-8 may indicate the edge of a subsurface bedrock bench.
Figure 24a: Conductivity pseudo-sections of horizontal dipole soundings surveyed on December 20, 2001. Pseudo-section A displays data using the Kriging contour method, pseudo-section B utilizes the modified Shepard’s contour method, and pseudo-section C displays a generalized representation of the shallow conductivity anomaly. The Kriging method creates “bulls-eye” patterns around individual data points. The modified Shepard’s method reduces the “bulls-eye” pattern.
Figure 24b: Pseudo-section representation of vertical dipole conductivities surveyed on December 20, 2001, utilizing two different contour methods. Results obtained using the A) Kriging contour method, B) modified Shepard’s contour method, and C) generalized representation of deep conductivity anomaly. Modified Shepard’s method is used to reduce “bulls-eye” effect of Kriging.
Figure 24c: Pseudo-section representation of horizontal dipole conductivities for 3-sounding transect line running across contour. A) Kriging contour method, B) modified Shepard’s method, and C) general interpretation of the shape of the conductivity anomaly. The soundings are S-7, S-3, and S-8 from upslope to down slope. S-3 is also part of the previous transect line.
Figure 24d: Pseudo-section representation of vertical dipole conductivities for 3-sounding transect line running across contour. A) Kriging contour method, B) modified Shepard’s method, and C) general interpretation of the shape of the conductivity anomaly. The soundings are S-7, S-3, and S-8 from upslope to down slope. S-3 is also part of the 6-sounding transect line.
**EM Modeling Results**

Figures 24a and 24b suggest three subsurface layers beneath S-1, S-2, and S-3 with relatively low conductivity values compared to that observed in soundings S-4, S-5, and S-6. From top to bottom, the three layers in S-1 through S-3 are believed to consist of a low conductivity zone of talus blocks, a high conductivity weathered boundary zone, and basal low conductivity shale bedrock. The conductivity readings at soundings S-1, S-2, and S-3 are likely a result of a single shallow, laterally extensive conductive zone. The higher conductivity readings at greater depth in soundings S-4, S-5, and S-6 are likely a result of more extensively fractured bedrock with an increased moisture content. The 3-layer system beneath S-4 through S-6 is hypothesized to include, compositionally, the same three layers present beneath S-1 through S-3. However, modeling suggests that the nonconductive talus layer and the conductive talus/bedrock boundary layer are both thicker beneath S-4 through S-6 than beneath S-1 through S-3. Modeling also suggests that the talus/bedrock boundary layer and the basal shale bedrock beneath S-4 through S-6 are more conductive than the same layers beneath S-1 through S-3.

Layer thicknesses for modeling purposes were based on the pseudo-sections presented in figures 24a through 24c. The pseudo-sections were based on effective penetration depths as used by the EMIX34 modeling program; therefore, the depths and thicknesses are only approximations and cannot be interpreted as completely accurate. The pseudo-sections provide an estimate of layer thickness and depth as a starting point for modeling.

Conductivity values for the modeled layers were based on previous electromagnetic studies, as well as empirical data collected at Ice Mountain. The conductivity of the unconsolidated, highly permeable talus is set at 0 mmhos/m, based on readings obtained with the Geonics EM-31. Conductivity of the shale bedrock is modeled at values from 0.5 to 1.0
mmhos/m, based on a conductivity range of 0.5 to 50 mmhos/m for consolidated shale (McNeill, 1980). Shale has a relatively low porosity (1-10%) and low conductivity unless extremely weathered or fractured (McNeill, 1980); therefore, values from the lower end of the conductivity range for consolidated shale were used. A conductivity range of 3 to 11 mmhos/m was used to model the conductive boundary between talus and shale bedrock. Higher conductivity values were used for the boundary zone because of the permeability contrast between talus and shale. The talus/shale bedrock boundary zone is believed to be weathered, loosely consolidated to consolidated, sandy material, which is reported by McNeill (1980) to have a porosity of 40 to 75%. Higher conductivity for the talus/bedrock boundary is dependent upon moisture content, which is variable throughout the year.

The Excel spreadsheets used to model the two 3-layer systems (Tables 1 and 2) are based on the McNeill equations. These show layer conductances, depth to top of layers, calculation of z values, calculation of vertical and horizontal response functions, measured and calculated apparent conductivities, and RMS error associated with the model. The relatively low RMS error calculated for the 3-layer scenarios suggests that the hypothesized models are reasonable estimations of the subsurface. In both 3-layer models, the RMS error was less than the average measured conductivity readings. Layer conductivities and thicknesses, as produced by modeling, are illustrated in Figure 25. The interpretation is one of many possible scenarios that yield reasonably low RMS error during modeling. However, the scenarios were based on a combination of many types of observations, in addition to terrain conductivity. Modeling confirmed that the hypothesized scenarios are consistent with the empirical geophysical data within a reasonable margin of error.
Figure 25: Illustration of layer conductivities and layer thicknesses obtained from modeling using Excel spreadsheet based on McNeill terrain conductivity equations.
Discussion

The right combination of geology, geomorphology, and location has created an algific environment at Ice Mountain in Hampshire County. Many factors are believed to contribute to the unique area. Collection and analysis of scientific data has resulted in a fact-based solution to the cold air phenomena at Ice Mountain. Observations of various aspects of the slope have been analyzed collectively to reach the following logical solution.

Geological and Geophysical observations suggest that the subsurface beneath the talus blocks upslope of the main algific area includes a topographical bench onto which relatively cooler air flows and collects. The bench is also a likely area for some moisture accumulation due to its relatively gentle gradient. The geomorphology of the slope area suggests that the meander in the North River below the algific zone has truncated the toe of the talus. The specific evolution of the topographic bench and truncation of the toe of the slope is unclear. One possibility is that the river was once located farther to the east. While located farther east, the river cut into the vertically dipping shale bedrock, creating steeper topography upslope and a topographic flat in the riverbed (Figure 16a). As periglacial activities and stream erosion occurred, Oriskany Sandstone boulders collected on the east bank of the river. The accumulation of boulders acted as natural armor for the riverbank. Similarly, Morris et al. (2000) report that boulders from colluviation armor the outside of meander bends in the Cheat Narrows and the New River Gorge. As a result, North River could no longer erode the bank at the rate needed to control its former sinuosity. The river reacted to the armoring by migrating westward. The convex slope created on the west slope of Ice Mountain next to the meander bend was covered with boulders derived from mass wasting under periglacial conditions (Figure 16a). A second theory for the evolution of the topographic bench and truncation of the toe of the talus slope is that North River naturally migrated westward, away from the base of Ice Mountain, prior
to periglacial conditions. As the river naturally migrated westward, it left an abandoned flat, which was covered by talus in periglacial times. At some point the river begin to cut down and migrate eastward again, creating an abandoned terrace and truncating the toe of the talus accumulation (Figure 16b). The topographic “bench” in the surficial morphology of the slope marks the upper most limit of the largest talus pieces. The larger size of the talus clasts on the bench, and in the area between the bench and the algific zone, results in higher permeability than elsewhere on the slope. The high permeability of the large boulders provides the conduits through which air and water circulate through the slope. In this way, the boulder bench is playing the role of the overlying plateau of the Iowa algific sites.

Precipitation and runoff from the area above the algific zone is funneled into and collected in lower portions of the talus on top of the impermeable shale bedrock, as suggested by the moisture sensitive VLF and terrain conductivity anomalies. Throughout the year, the coldest air of the time flows down the mountain and into the interstitial spaces of the talus, accumulating and slowly seeping from the algific vents. In colder months of the year, water collected on the impermeable bedrock surface and within the talus in areas where cold air is trapped may freeze. At such times, ice is likely to accumulate from the shale bedrock surface upward into the overlying framework of sandstone talus boulders. The slope’s northwest-facing aspect keeps it sheltered from the sun and the sandstone boulders insulate, possibly allowing ice to exist year-round. In the warmer months, relatively warm air and water infiltrate the talus boulders, increasing the temperature of the interstitial talus air and slowly melting any possible ice reservoir. Throughout the Summer, the temperature of the air in the algific vents increases with increasing outside air temperature, leading to depletion of potential ice buildup. Each year, the
onset of colder outside air temperatures and relatively colder air infiltrating the talus creates colder temperatures at the algific vents and the cycle begins again.

**Airflow and Ice Accumulation at Ice Mountain**

The following discussion of airflow and ice accumulation is based on extremely limited empirical airflow data. Basic concepts of physics and meteorology were analyzed and combined with physical delineation of the Ice Mountain slope, as well as general characteristics of airflow associated with hill slopes, to formulate a hypothesis. The author is hopeful that the work completed in this thesis will serve as a basis for further study of the Ice Mountain algific talus.

Airflow patterns within the algific slope are a direct result of wind patterns that commonly occur on mountain slopes. Local wind patterns, known as “slope winds”, are documented by Matthes (1911) and Reifsnyder (1980). Reifsnyder (1980) defines “slope winds” as “winds that blow upslope in daytime and down slope at night”. Slope winds are a result of the differential heating and cooling of the surface of the Earth and are most prevalent under clear skies and relatively weak large-scale atmospheric pressure gradients (Reifsnyder, 1980). The slope winds are evidence of heat in transit by way of “mass movement of molecules from one place to another”, a process known as convection (Giancoli, 1998).

Nighttime slope winds result in an accumulation of cold air in the valley bottom (Reifsnyder, 1980). At night, the surfaces of the unvegetated talus areas (specific heat of 0.21 cal/g/°C) cool due to release of radiant heat in the absence of the Sun (Reifsnyder, 1980). Air nearest the cooling unvegetated talus surface becomes denser than air above, resulting in airflow down slope (Reifsnyder, 1980). The fate of the down slope flowing air is dependent upon the relative temperature of the air already present in the interstitial spaces of the talus. If the down slope flowing air is warmer than the interstitial talus air, the down slope flowing air will ride on
top of the interstitial air creating stratified layers. However, if the down slope winds are cooler than the interstitial talus air, the interstitial talus air is replaced by the cooler down slope flowing air. The latter situation may occur frequently in the Summer when the interstitial talus air has been warmed by high Summer daytime temperatures and is replaced by air from a cool Summer night. Replacement of interstitial talus air may also occur during very cold winter temperatures when very cold down slope flowing air enters the talus. Both situations ensure that the coldest available air remains trapped in the interstitial spaces of the talus.

Sinking of cooler air during nighttime slope winds occurs more rapidly on steep slopes than on gentle slopes (Reifsnyder, 1980). “Air avalanches” may also occur when a pool of cold air develops on a “bench or topographic basin” (Reifsnyder, 1980). When the pool of cold air on the bench becomes too deep, the dense air flows over the side of the bench and continues down slope (Reifsnyder, 1980). “Small-scale cold air ponds and thermal belts will develop on benches that sit above the valley bottom. The coldest temperatures will occur at the very bottom of the valley, and the secondary cold air ponds will not be quite so cold” (Reifsnyder, 1980).

The topographic bench found during surficial geology mapping of Ice Mountain talus is likely to be a site upon which cold air pools form during nighttime slope winds. Furthermore, the void spaces in the talus between the topographic bench and the algific vent zone are likely to act as a cold air “sponge”, collecting cool nighttime slope winds from the large unvegetated talus area above. VLF and terrain conductivity surveys indicated an anomalously high conductivity zone in the area on and below the topographic bench, upslope of the algific vents. The high conductivity indicates a concentration of moisture in that area. The combination of cold air and moisture in the area just upslope of the algific vents suggests high probability of condensation and/or ice accumulation in that area.
Daytime slope winds occur due to warming of the Earth’s surface by the Sun, either by direct sunlight or indirect radiation (Reifsnyder, 1980). Warming of the Earth’s surface at Ice Mountain occurs more easily in the unvegetated talus areas. Air near the hot surface of the talus becomes warmer and less dense than the surrounding air. The warmer, less dense air rises vertically (Reifsnyder, 1980) but may also flow along the surface of the hot talus if constricted by cooler air from above (Matthes, 1911). The location and northwest facing position of the slope is such that the unvegetated talus areas are not often exposed to direct sunlight. Thus, daytime slope winds are likely more subdued in comparison to nighttime slope winds at Ice Mountain.

The apparent accumulation of ice in the talus of the algific zone can be attributed to air mass advection, the movement of a mass of air from one area to another (Moran and Morgan, 1994). As an air mass moves over a surface, it can be altered by the characteristics of that surface (Moran and Morgan, 1994). Cool air flowing down-slope on Ice Mountain enters the more permeable talus near the bench where the clasts are largest. As the air infiltrates between clasts, it creates a cold air pool upslope of the algific vents. If the air is saturated at a temperature above 0°C (32°F) it is at the dew point and condensation as liquid will occur, releasing 2260 kJ/kg (600 cal/g) of energy associated with the latent heat of vaporization (Giancoli, 1998). However, if saturation occurs at or below 0°C (32°F), water vapor changes directly into solid ice and frost occurs. Formation of frost releases energy to the talus blocks equal to the sum of the latent heat of fusion (333 kJ/kg or 80 cal/g) and the latent heat of vaporization (2260 kJ/kg or 600 cal/g) (Total = 2593 kJ/kg or 680 cal/g) (Moran and Morgan, 1994). However, the addition of energy to the talus may be somewhat equalized by the insulating effects of the talus in the form of nearly continuous radiational cooling (Moran and
Morgan, 1994). If cold air infiltrates the slope and becomes saturated at the frost point, ice may accumulate in the talus. As a result, there is potential for a reservoir of ice to form in the interstitial spaces between boulders, beginning at the shale bedrock/talus boundary. The ice reservoir is insolated by the thick accumulation of talus and shielded from the Sun due to the northwest facing orientation of the slope. During the summer, relatively warm air flows into the more permeable sections of the slope. As the relatively warm air flows over the reservoir of ice and/or relatively colder air, it is cooled, becomes denser and eventually flows out of the small vents at the bottom of the slope. The interaction between the warmer infiltrating air and the cooler reservoir cools the air while simultaneously increasing the temperature of the reservoir. Throughout the summer, warm air is cooled and any ice reservoir is slowly depleted. Limited temperature readings for the air in the algific vents at Ice Mountain show a gradual increase throughout warmer months of the year, suggesting a depletion of the potential ice reservoir or an overall warming of the interstitial talus air. Temperature readings also indicate a decrease in air temperature for the vents as the outside air temperature decreases. When the outside air temperatures drop below freezing, the temperature of the interstitial air decreases and ice accumulation is possible within the talus.

**Comparison with Iowa Slopes**

The following comparisons demonstrate similarities and differences between the Ice Mountain algific environment and Iowa algific environments. The two environments, located in significantly different areas of the country, evolved through time by natural processes to become similar habitats.

There are many similarities between the Ice Mountain algific talus and the algific slopes in Iowa. Algific slopes in both regions have vents at their base from which cold air is expelled. The cold air at the base of the slopes creates fragile environments, classified as paleorefugia, in
both regions. The cold air expelled from the slopes in both regions flows through large boulders and rock debris accumulated as a result of periglacial conditions, stream erosion, and oversteepening. Cold air in both slope environments is thought to be at least partly dependent on ice formation from water that collects on a more impermeable subsurface unit beneath the boulder matrix. The Marcellus-Needmore shale bedrock acts as the impermeable unit under the Ice Mountain slope, whereas thin beds of clay and bentonite within carbonate units act as impermeable units in Iowa. The Ice Mountain slope and most of the slopes in Iowa are north-facing slopes, a characteristic that helps shield accumulated ice from long exposure to direct sunlight.

There are many similarities between the two slope environments, but many specific differences exist. Ice Mountain algific slope is located in an area of very steeply dipping strata, while the rocks in the Iowa algific areas dip very gently. The Ice Mountain slope includes sandstone boulders consisting of quartz grains and quartz or calcite cement. Conversely, the Iowa slopes are found in dolomite and limestone areas. Although slopes in both environments expel cold air during the warmer months of the year, there is no evidence to suggest that the Ice Mountain slope reverses its airflow in the winter to aid in re-accumulation of ice, as is believed in the case of the Iowa slopes. Both environments rely on the high permeability of natural settings; however, the origins of the settings are drastically different. The Iowa slopes occur where mechanical karst development has created cracks, crevices, and some talus in the bedrock, through which air and water flow. In contrast, the flow of air and water in the Ice Mountain algific slope is a result of the permeability of the matrix of accumulated talus boulders. The Iowa slopes are connected to karst sinkholes at the top of overlying plateaus, which allow for circulation of air and water between the highland and the toe of the slope. A carbonate unit
crops out near the ridgetop above the Ice Mountain slope; however, geologic mapping suggests that the carbonate unit dips at a steep angle and is much too deep to have any interaction with the toe of the slope where the algific vents are located.

**Conclusion**

Whether in Iowa or West Virginia, algific environments are very vulnerable due to a dependency on year-long cold temperatures. Removal of the talus material or damaging of the thin veneer of soil in the algific zones would allow for rapid draining of cold air from the talus, as well as loss of insulation of the accumulated ice or cold air pools. Without ice or cold air pools, the temperature differences that drive the cold airflow out of the vents would be much less and the unique vegetation in front of the vents could no longer survive. Once the slope habitat is destroyed, it is almost impossible to fully restore (Iowa State University Extension, 1999).

Grazing animals can do damage to the slopes, but the most common damage to algific slopes is done by humans in the form of ice removal, timber harvesting, quarrying, road building, herbicide or pesticide spraying, and even hiking (Iowa State University Extension, 1999).

Proper management of algific environment areas is vital to their survival. The locations of many of the slopes in Iowa are kept secret to avoid overuse and ice removal damage. The plants that grow in the vent areas of the slopes are totally dependent upon the cold airflow from the vents. Alteration of a vent opening can result in redirection or depletion of the cold airflow, and potential death of plants in the abandoned area. The vent area is the most obvious vulnerable area of algific slopes; however, the permeable material through which cold air flows to the vent area must also be a conservation priority. Airflow in algific slopes requires both inflow and outflow to maintain the system. The permeable talus and rock above the algific vents, as well as the vegetation surrounding the talus areas, is also very important. Excessive removal of talus blocks from the area where ice accumulation occurs may result in increased depletion rates for
the ice throughout the year, causing a potentially disastrous increase in air temperature at the vents. Classification of algific talus slopes as paleorefugia implies they have existed for a long time. Man’s influences, as well as some natural influences, have resulted in the relatively recent destruction of many algific slopes and the associated environments. With proper management, good protection techniques, and minimum natural climate change, the remaining slopes will survive and continue to serve as natural wonders of the world.

Management of the algific slope environment at Ice Mountain has been, and continues to be, successful. The West Virginia Nature Conservancy protects the algific zone by limiting the amount of human access to the area. Guided tours, trail signs, and well-maintained trails provide both enjoyment and conservation of the slope. The Nature Conservancy also allows important scientific research to be done on the slope environment and its inhabitants.

This study focused on the origin and physical characteristics of the Ice Mountain algific slope, while briefly providing an introduction to the processes that drive the cold air production. Further research should be focused on airflow and temperature variations with respect to different algific vents, various areas of the slope, and trends associated with seasonal changes. The author intends this paper to provide information about algific slopes that can be used for future management and research opportunities.
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<td>Helderberg Limestone outcrop location with anomalous strike measurement.</td>
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<td>29</td>
<td>716183.59</td>
<td>Helderberg Limestone outcrop location with anomalous strike measurement.</td>
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<tr>
<td>37</td>
<td>715883.65</td>
<td>Northeast end of linear depression in talus on east side of Ice Mountain ridgeline.</td>
</tr>
<tr>
<td>38</td>
<td>715866.09</td>
<td>Southwest end of linear depression in talus on east side of Ice Mountain ridgeline.</td>
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<tr>
<td>BENCH1</td>
<td>715742.72</td>
<td>Central location of topographic bench visible in talus above the main algific zone. Location of larger algific vent near center of main algific zone. Limited vent air temperature data was collected from this vent.</td>
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<tr>
<td>BIG1</td>
<td>715739.48</td>
<td>Eastern edge of bowl-shaped depression formed by talus.</td>
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<tr>
<td>EB</td>
<td>716024.12</td>
<td>Northwest corner of VLF survey area and northeast end of VLF transect line 1. Approximate location of northern extent of algific vent area.</td>
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<tr>
<td>GOVLF2</td>
<td>715919.67</td>
<td>Southwest corner of VLF survey area and southwest end of VLF transect line 1. Helderberg Limestone outcrop location.</td>
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<tr>
<td>ICE</td>
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<td>Middle of bowl-shaped depression formed by talus.</td>
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<td>L-1STP</td>
<td>715551.51</td>
<td>Northeastern extent of algific vent area.</td>
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<tr>
<td>LS1</td>
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<td>715998.24</td>
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</tr>
<tr>
<td>NB</td>
<td>716002.86</td>
<td>Northern edge of bowl-shaped depression formed by talus.</td>
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<td>OCLIFF</td>
<td>716041.88</td>
<td>Oriskany Sandstone outcrop location.</td>
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<td>715843.91</td>
<td>Algific vent location from which limited vent air temperature measurements were collected. EM-34 sounding location</td>
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<tr>
<td>S-1</td>
<td>715777.92</td>
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<td>715761.87</td>
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<td>S-3</td>
<td>715741.32</td>
<td>EM-34 sounding location</td>
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<td>S-4</td>
<td>715721.22</td>
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<td>S-8</td>
<td>715733.02</td>
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<td>SADDLE</td>
<td>715377.43</td>
<td>Needmore Shale outcrop location. Southern edge of bowl-shaped depression formed by talus.</td>
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<tr>
<td>SB</td>
<td>715987.15</td>
<td>Needmore Shale soil observed at this location and used to map Needmore Shale/Oriskany Sandstone contact. Location of approximate southern extent of algific vents.</td>
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<td>SHS3</td>
<td>716180.44</td>
<td>Location of discharge point of small spring found by EM-31 survey.</td>
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<td>SIGN</td>
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<tr>
<td>SPRING</td>
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<td>VLF7N</td>
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<td>Western edge of bowl-shaped depression formed by talus.</td>
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<td>VLF7S</td>
<td>715604.37</td>
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<td>WB</td>
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### Temperature

(°F)

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**Appendix B:** Algific Vent Air Temperature Observations (see Figure 17).