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Geophysical Analysis of Seasonal Montandon Gravel Ridge Water Table Fluctuation and Moisture Gradient Variation due to Storm Events

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GEOPHYSICAL ANALYSIS OF SEASONAL MONTANDON GRAVEL RIDGE WATER TABLE FLUCTUATION AND MOISTURE GRADIENT VARIATION DUE TO STORM EVENTS

By

Anne E. Strader

Class of 2010

A Thesis Submitted to the Honors Council for Honors in Geology

5/10/2010

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Senior Thesis

GEOPHYSICAL ANALYSIS OF SEASONAL MONTANDON GRAVEL RIDGE WATER TABLE FLUCTUATION AND MOISTURE GRADIENT VARIATION DUE TO STORM EVENTS

Anne E. Strader
Class of 2010
Dr. Robert Jacob, Advisor
Contents
List of Figures .................................................................................................................................................. 3
List of Tables .................................................................................................................................................... 5
Abstract .......................................................................................................................................................... 6
Acknowledgements ........................................................................................................................................ 7
Introduction .................................................................................................................................................. 8
Background Information on Wetland Research Site ..................................................................................... 9
Location of Research Site within LTHR ....................................................................................................... 15
Purpose .......................................................................................................................................................... 18
Methods ......................................................................................................................................................... 20
Precipitation: Data Collection Methods ....................................................................................................... 22
Seismic: Data Collection Methods ................................................................................................................ 23
Seismic: Data Interpretation Methods ........................................................................................................... 30
Seismic: Results ............................................................................................................................................ 35
Seismic: Interpretations .................................................................................................................................. 38
DC Resistivity: Data Collection Methods ..................................................................................................... 41
DC Resistivity: Data Interpretation Methods .................................................................................................. 43
DC Resistivity Results .................................................................................................................................... 47
DC Resistivity: Interpretations ....................................................................................................................... 51
Ground Penetrating Radar (GPR): Data Collection Methods ....................................................................... 57
GPR: Data Interpretation Methods ............................................................................................................... 63
GPR: Results .................................................................................................................................................. 65
Interpretations ................................................................................................................................................ 72
Combined Data Interpretations .................................................................................................................... 74
Conclusions .................................................................................................................................................... 75
Future Work ..................................................................................................................................................... 76
References Cited ............................................................................................................................................. 78
List of Figures

Figure 1: General map of the Montandon LTHR (11)

Figure 2: Plot of 1992 water table elevation data from 4/92 to 10/92 (13)

Figure 3: Cross section of wells near LTHR research site with water table elevations (17)

Figure 4: Geologic stratigraphic column with additional layer of dune sand (18)

Figure 5: Precipitation and Susquehanna River discharge data from 4/2009 to 10/2009 (24)

Figure 6: Diagram of Snell’s Law (27)

Figure 7: Diagram of seismic data collection alignment during 5/2009 and 8/2009 (28)

Figure 8: Diagram of seismic wave types (30)

Figure 9: Seismic data from Seistronix program (32)

Figure 10: First break arrival time data (forward and reverse profiles) for 8/15/2009 (33)

Figure 11: Plot of all seismic data (forward and reverse profiles) (37)

Figure 12: Plot of reverse profile seismic data for all seismic lines (38)

Figure 13: Figure of seismic interfaces, velocities, and layer thicknesses for 5/19/2009 and 8/15/2009 relative to LTHR geology (41)

Figure 14: Diagram of DC resistivity Wenner array setup (42)

Figure 15: Apparent resistivity plot for 4/20/09 (44)

Figure 16: 3-layer resistivity curve and parameters for 8/14/2009 (45)

Figure 17: 4-layer resistivity curve and parameters for 8/14/2009 (47)

Figure 18: Plot of all apparent resistivity data during research period (50)

Figure 19: Plot of resistivity profile data (52)

Figure 20: Finalized, 5-layer resistivity curve for 8/14/2009 (55)
Figure 21: Resistivity plot for two shallowest layers over entire research period (57)

Figure 22: Individual rain event resistivity plot for two shallowest layers (58)

Figure 23: Diagram of GPR CMP setup (61)

Figure 24: Plot of GPR velocity correlation with moisture content (62)

Figure 25: Example of GPR CMP data (63)

Figure 26: Example of GPR profile data (64)

Figure 27: Plot of (Tx-Rx)^2 vs (time)^2 for shallowest reflected phase from 5/20/2009 CMP (65)

Figure 28: GPR 200 MHz CMP data for 5/20/2009 (69)

Figure 29: Plot of Tx-Rx vs time for shallowest reflected phase for 5/20/2009 (70)

Figure 30: GPR 200 MHz profile data for 8/12/2009 (72)

Figure 31: Plot of GPR velocities and depths for third reflected phase over entire research period (74)

Figure 32: Plot of GPR velocities and depths for third reflected phase during individual rain events (75)
List of Tables

Table 1: Seismic velocities for saturated and unsaturated common geologic materials (25)

Table 2: Seismic refraction calculations from equations and REFRACT (39)

Table 3: All apparent resistivity sounding data (49)

Table 4: Apparent resistivity profile data (51)

Table 5: Most commonly used GPR parameters (67)

Table 6: GPR velocities and depths for the third reflected phase throughout the research period (71)
Abstract

The purpose of this research project is to continue exploring the Montandon Long-Term Hydrologic Research Site (LTHR) by using multiple geophysical methods to obtain more accurate and precise information regarding subsurface hydrologic properties of a local gravel ridge, which are important to both the health of surrounding ecosystems and local agriculture. Through using non-invasive geophysical methods such as seismic refraction, Direct Current resistivity and ground penetrating radar (GPR) instead of invasive methods such as borehole drilling which displace sediment and may alter water flow, data collection is less likely to bias the data itself. In addition to imaging the gravel ridge subsurface, another important research purpose is to observe how both water table elevation and the moisture gradient (moisture content of the unsaturated zone) change over a seasonal time period and directly after storm events. The combination of three types of data collection allows the strengths of each method combine together and provide a relatively strongly supported conclusions compared to previous research.

Precipitation and geophysical data suggest that an overall increase in precipitation during the summer months causes a sharp decrease in subsurface resistivity within the unsaturated zone. GPR velocity data indicate significant immediate increase in moisture content within the shallow vadose zone (< 1m), suggesting that rain water was infiltrating into the shallow subsurface. Furthermore, the combination of resistivity and GPR results suggest that the decreased resistivity within the shallow layers is due to increased ion content within groundwater. This is unexpected as rainwater is assumed to have a DC resistivity value of
3.33*10^5 ohm-m. These results may suggest that ions within the sediment must be incorporated into the infiltrating water.

**Acknowledgements**

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Introduction

Subsurface hydrodynamic processes have direct geologic and ecologic impacts that influence environmental preservation and effective conservation of natural resources. Soil water content is expected to change due to infiltration from precipitation events, drainage to the water table (the boundary between the saturated and unsaturated zone) and evapotranspiration (Farmani 2008). Precipitation that infiltrates into the subsurface will increase the soil water content (SWC) in the unsaturated zone but also has the potential to increase the water table elevation. Infiltration rates depend on subsurface geology and rainfall duration and quantity (Jury and Horton, 2004). Therefore, SWC as well as the water table elevation are expected to fluctuate seasonally and may even increase due to individual rainstorms. Monitoring changes in SWC is crucial for farmers as it dictates the availability of water for crop roots and other plants. In addition, the elevation of the water table may affect the direction and amount of water flow and subsequently water resources in areas such as wetlands and rivers.

Newly infiltrated rainwater, which contains minimal to no ions, may either dilute ion concentration of the groundwater (including water in unsaturated and saturated zones), or may increase dissolved ion concentration by remobilizing trapped water containing ions. The health of wetland ecosystems that are supported by the groundwater thus may be affected by either increases or decreases in ion concentrations (Rein 2003) or by a decrease in the elevation of the water table. Changes in water table elevation and infiltration rates, over a seasonal time period and days after storm events, also can provide insight regarding the hydraulic gradient, which
has direct implications for water flow direction and therefore the health of surrounding ecosystems.

**Background Information on Wetland Research Site**

**General Site Characteristics:** The Montandon Long-Term Hydrologic Research site (LTHR) comprises an area of approximately one square kilometer, bordered by the West Branch of the Susquehanna River to the west, a riparian wetland to the east, Route 45 to the north, and a gravel driveway to the south (Figure 1). The research site is an excellent location to investigate seasonal and short-term precipitation effects in the unsaturated zone and is located 1.5 miles east of Bucknell University. The LTHR is characterized by terrace remnants from Pleistocene glaciations events overlain by alluvial and aeolian sediments.

The topography, shown as contour lines in Figure 1, is relatively flat along the western portion of the LTHR. The eastern portion of the LTHR is characterized by a wetland, the northern portion is the location of ongoing sand mining activities, and the southern portion contains agricultural fields. There is a series of wells drilled throughout the LTHR and shown in Figure 1.

**Geologic Information on Site:** The subsurface geology of the Montandon LTHR is composed of sedimentary bedrock (consolidated sediment) overlain by unconsolidated sediment. The bedrock underlying the LTHR is composed of Silurian and Devonian age (approximately 400 Million Year Old) siltstone, shale and limestone. The unconsolidated material above the bedrock was primarily produced as a result of Pleistocene glaciations. Glacial retreat resulted in large glaciofluvial outwash deposits containing mostly coarse grained sediment (gravels, coarse sands and minimal quantities of fine-grained sands) which provided
the main source of material transported by the Susquehanna River approximately 10,000 years ago (Smith, 1991). Glacial retreat resulted in a coarse-grained sand and gravel deposit at the LTHR referred to as the Binghamton Terrace. After the deposition of gravelly-sand, dunes were deposited at the LTHR consisting of a fine grained sand material. The terrace and dune deposits have been affected by surface processes, such as soil development or erosion by the river, thus resulting in the modern topography, land use and terrace. The Binghamton Terrace is referred to as the gravel ridge and is indicated in Figure 1 by the increase in elevation from west to east. The primary dunes are located in the southeastern portion of Figure 1 as shown by the relatively high elevation and slightly steeper slopes in Figure 1. To the north, relatively flat topography indicates a younger terrace and the wetland to the east (Peltier, 1949). To the east, the wetland overlies an early Holocene paleochannel of the West Branch Susquehanna River (Kochel, 1994).
Specific Site Characteristics from Previous Invasive Studies – Seasonal: Numerous groundwater studies have been conducted over the last twenty years within the Montandon
LTHR. Studies conducted by Anderson (1992), Valigorsky (1993), and Ressler (1998), utilized invasive techniques, specifically collecting water table elevation data from a series of wells shown in Figure 1, to observe groundwater flow within the saturated zone. These groundwater studies at the Montandon LTHR have shown not only fluctuations in water table elevation from season to season but also changes in the flow direction within the saturated zone (Ressler, 1998). The most significant changes in groundwater flow directions were driven by fluctuations in the water table elevations beneath the highest ground-surface elevations between the marsh and the Susquehanna River (Figure 1).

**Specific Site Characteristics from Previous Invasive Studies – Rain Events:** Anderson (1992) observed short-term water table fluctuations through measuring elevation levels at seventeen wells located near or on the gravel ridge from June 2, 1991 to December 1, 1991 every three or four days. These data revealed that the hydraulic gradient changed due to water table fluctuations over the course of six months (Figure 2). Valigorsky (1993) contributed further to water table fluctuation research in two ways, first by collecting water table elevation data every day during the summer of 1992 and secondly by investigating the possibility of a groundwater divide along the gravel ridge. In addition, the water table fluctuations were larger in 1992 than in 1991 because there was more precipitation during the summer of 1992 than 1991 (Valigorsky 1993). These rainstorms strongly correlated with observed water table fluctuations, indicating that rainstorms directly caused the water table level to rise. Furthermore, the observations from nearby wells show that hydraulic gradient, i.e. the groundwater flow, reversed directions during this research period, causing most groundwater to flow away from the wetland instead of an even flow division in the east and west directions.
Therefore, there is a significant connection between amounts of precipitation and the water table gradient over seasonal time periods.

**Figure 2.** Depth to the water table recorded in Well 8 from early May to early October, 1992. An increase in depth to the water table is equivalent to a decrease in elevation of the water table.

**Summary of Previous Invasive Studies:** These previous studies indicate that the water table follows the ground surface topography across the LTHR, which forms a groundwater divide beneath the highest elevation – the gravel ridge as seen in Figure 3. Additionally, these studies indicate an increase in depth to the water table, which may result from a lack of rain sustaining the elevation of the W.B. Susquehanna River or infiltrating into the subsurface, and would also cause a decrease in SWC in the unsaturated zone. Another possibility is that increased depth to the water table is a result of an increase in evapotranspiration, the combined effect of evaporation and absorption of water by plants, which would also cause a
decrease in SWC in the unsaturated zone. While invasive techniques provide a great deal of information on subsurface conditions, the information is extremely localized to a relatively small area surrounding the borehole. In comparison, data from non-invasive geophysical study conducted at one specific location provide subsurface information averaged over a wider area. Additionally and most importantly, these invasive studies did not monitor changes in the SWC.

**Specific Site Characteristics from Previous Non-Invasive Studies:** Non-invasive geophysical studies have the potential to provide information over a more expansive area than is feasible from an invasive study and have been used to determine the bedrock structure across the transects of wells (Bock, 1992) and overall electrical resistivity structure of the gravel ridge (Byler, 2009). The capillary fringe is a phenomenon where water is trapped slightly above the water table by electrostatic forces (Fetter 2001) and is detected by geophysical methods but not through well drilling. Thus, there are expected to be slight discrepancies between the calculated depths to the water table depending on methodology. The subsurface properties of the connection between the gravel ridge and the wetland were investigated by Jacob (1997a) using both seismic refraction and electrical resistivity. While these studies indicated the efficiency of collecting geophysical data over large distances to observe lateral changes in subsurface structure, they did not monitor the dynamic nature of the subsurface. Jacob (2006) provides an example of a non-invasive geophysical study using GPR measurements to observe hydrodynamic processes occurring in the unsaturated subsurface. While this work took place at a research site in New England, the precision of GPR measurements apply to other research locations. The precision of GPR velocity was found to be ± 0.003 m/ns and the precision of depth estimates at an average depth of 4 m were ±0.2 m. While previous invasive studies
determine water table elevations at specific points through test boring, geophysical
measurements provide estimates of subsurface conditions that are spatially averaged over a
larger volume than a single borehole and have the potential to provide information regarding
changes over a distance, thus providing a more complete subsurface interpretation compared
to borehole data (Burger 2006).

The non-invasive geophysical studies conducted at the Montandon LTHR by Bock (1992)
and Jacob (1997a) have only utilized one or two data collection methods, specifically seismic
and possibly electrical resistivity to determine water table and bedrock elevations across the
southern portion of the LTHR. Non-invasive studies including EM-31 and GPR have also been
conducted to locate potential locations within the Montandon wetland and floodplain for well
drilling (personal communication Hayes, 2010). Jacob also used GPR common midpoint
sounding (CMP) data to estimate the volumetric water content of the shallow subsurface due
to a rain event. These studies’ effectiveness in determining subsurface properties might be
further improved through the systematic application of multiple geophysical methods repeated
at the same location over time periods similar to those used in the invasive studies of Anderson

**Location of Research Site within LTHR**

Previous studies have shown the water table to fluctuate the most within the thickest
layers of unconsolidated sediment. Therefore, the research site was selected to be on the
unconsolidated sediment near the highest elevations at the LTHR as shown in Figure 1. In
addition, the location was chosen at a sufficient distance from any well in order to minimize
interference of water flow in the unsaturated zone as well as interfere with geophysical signals, but close enough to reliable compare previous water table data with current results. Because current research concentrates on water table elevation variation and the moisture gradient over periods of time, a consistent research location is imperative to observing changes specifically due to time and precipitation events rather than lateral changes in subsurface properties. This project is expected to detect changes in hydrologic properties within the shallow subsurface on the gravel ridge within a 30x30m field site, shown in Figure 1.

A cross section through wells located north of the highest elevation of the LTHR is shown in Figure 3. Although this figure displays subsurface geologic and hydrologic interpretations from the nearby boreholes, it does not include the dune sand layer at the surface of the research location. At this location, the upper 7m of the surface is shown in Figure 3 and is characterized by a relatively thin (depth 0 to ~1m) lens-shaped layer of dune sand which overlies a thicker (depth ~1m to ~6m) layer of sand and gravel. Helderberg limestone bedrock is at the base of the unconsolidated subsurface, directly underlying gravelly-sand layers. Figure 4 displays geologic data with the additional dune sand layer at the research site.
Figure 3. A cross section of the Montandon floodplain displays an interpretation of the gravel ridge subsurface based on previously collected test boring and geophysical data (Ressler, 1998). Water table elevations at wells (Ressler) extending along the cross section are indicated by blue dots.
Purpose

This study differs from many previous studies at the LTHR in that it exclusively uses non-invasive methods, multiple geophysical methods instead of relying on one or two, and involves collecting several sets of data over time at a specific location. The use of geophysical methods allows for continuous information over a larger area compared to data acquired from a single well. Non-invasive methods are not only less spatially confined than invasive methods, but are
less likely to impede water flow and infiltration, which potentially could obscure the data which are the focus of this research project. Combining multiple geophysical methods provides not only different properties of the subsurface but also information at different subsurface depths. This benefit of multiple methods minimizes the probability of mistaking instrument error for conclusive evidence (Jacob and Hermance, 2005). Also, instead of collecting data over long distances (> 100 m) such as Bock (1992) or Jacob (1997b), research is conducted on a small traverse (< 50 m) on the gravel ridge in order to maximize temporal resolution, allowing for multiple data collection methods to be collected within approximately an hour after a rainstorm. The most significant difference for this geophysical study compared to previous studies is the principal zone of interest; the previous studies have been focused within the saturated zone, whereas this research observes geophysical characteristics predominately of the unsaturated zone.

The primary goal of this research is to observe variations in the SWC throughout seasonal time periods and determine how individual rainstorm events immediately change the SWC. The secondary goal of this research is to observe seasonal and short-term variations in the water table elevation due to rainstorms with non-invasive methods. There have been documented changes in water table throughout most LTHR wells. However, the research site is located away from the wells to avoid any potential interference or borehole induced changes in the unsaturated zone, and to determine whether or not similar water table fluctuations occur where wells are not located. Additionally the location was chosen at high elevation to enhance the expected changes observed in each geophysical method through the likelihood of greater water table fluctuations compared to changes within low-lying topography.
Methods
This study uses three different non-invasive geophysical methods – Seismic refraction, DC Resistivity and GPR (ground penetrating radar) - in order to provide a variety of precise information about hydrologic properties of the gravel ridge subsurface. Seismic refraction data can provide general information about acoustic wave velocity over a relatively broad range of subsurface depths. However, in this study it will be used for the upper 10 m of the subsurface. The DC resistivity method provides the resistivity of the subsurface at a variety of depths, and GPR can provide electromagnetic wave velocity within the shallow subsurface, specifically for this study < 3 m.

**Combined Geophysical Methods:** Combined data from each of these methods is helpful in inferring interface depths that represent geologic or hydrologic changes. For example, a water table boundary displays an increase in seismic velocity from the unsaturated zone to the saturated zone as well as a particularly large decrease in resistivity. This means that seismic refraction data can be used to calculate the depth to the water table as well as bedrock and other interfaces where an increase in acoustic wave velocity occurs. Similarly, resistivity can be used to locate the thickness of the unsaturated zone (i.e. depth to water table) as well as bedrock elevations and potentially observe significant changes in SWC (Burger et al., 2006). Like GPR and seismic data, changes in resistivity should be immediate and apparent. Sudden variations in resistivity should also closely reflect changes in velocity of radio waves and sound waves from GPR and seismic data. For instance, a sharp increase or decrease in resistivity may be correlated with changes in seismic and GPR velocities indicating increased sediment moisture content in that changes in relative ion concentration in groundwater may be inferred.
The velocity of electromagnetic waves used in the GPR method is significantly dependant on the water content in the unsaturated zone and is capable of providing more detail than either seismic refraction or resistivity (Annan, 2005). Additionally, GPR reflections provide depth to changes in the SWC variations in the unsaturated zone. If seismic data is collected and indicates multiple possible subsurface layers based on calculated refracted wave velocities, resistivity data can be compared to subsurface layers observed by the seismic method to determine which layer is likely the water table. The depth to interfaces observed through GPR may correspond to thicknesses of layers observed from the DC resistivity method as well as correlation between the two methods indicating increased SWC in the unsaturated zone. Correlations between all three methods will increase the confidence of the findings.

The relatively high frequency of GPR yields more precise results compared to seismic data, however, GPR is primarily focused on the subsurface to a depth of approximately 5m. At the 95% confidence limit, GPR velocity precision may be precise to ± 0.003 m/ns (Jacob and Hermance, 2004), while the thicknesses of detectable layers range from 2cm at 1GHz to 20 cm at 100MHz (Burger, 2006), corresponding to resulting wavelengths for a velocity of 0.8 x 10^8 m/s. Seismic waves range from 10-100 Hz, resulting in relatively longer wavelengths. For instance, a seismic velocity of 1500 m/s (reasonable for saturated sand) results in a wavelength of 15 m for a 100 Hz wave. Therefore, the seismic method is particularly useful for measuring seismic velocities at relatively deep interfaces that are not expected to undergo changes in elevation or geologic composition, such as the limestone bedrock at the Montandon field site. Because the seismic velocity of saturated sand is constant through time, the seismic method is also effective in measuring the depth to the water table.
The following sections lay out the precipitation data to which the geophysical data are compared and then outline field procedures and parameters for each geophysical method, then results and interpretations. Interpretations for sections focusing on an individual geophysical method are made solely based on the data for that particular method. After these sections, interpretations combining data from all three geophysical methods are discussed, illustrating how the use of multiple types of geophysical data provides more detailed and verifiable information about subsurface structure and changes in hydrologic properties within the subsurface than one or two types of geophysical data.

**Precipitation: Data Collection Methods**

The amount and timing of precipitation events is obviously vital to putting the geophysical interpretations in context. Precipitation was recorded by the Weather Underground website (http://www.wunderground.com) from late spring to early fall.

Precipitation data has been taken within a few miles of Lewisburg, PA, which is approximately eight miles south of the research location, the location with available data closest to the Montandon gravel ridge. Overall, most precipitation events within the study period (late spring and late summer – early fall 2009) occurred during late May and late August, with no major rain events prior to data collection that might have obscured the geophysical data. Figure 5 shows the daily precipitation occurring during the study period between May and August, 2009 as well as the days when each type of geophysical data used in this study was collected. Although there were several significant rain events (> 1 in.) during the research period, there were no significant dry periods (< .1 in.) between early May and late October of 2009. One of these rain events was distributed through three days, and was characterized by 0.92 inches of rain on
5/14/09, 0.22 inches on 5/16/09 and ending with 0.57 inches on 5/17/09. The precipitation data precede significant increases in the W.B. Susquehanna River discharge at Lewisburg, Pennsylvania. Kochel (1996) suggested that river levels may influence water table elevation.

Figure 5. This graph displays inches of precipitation and data collection dates from 4/1/09 to 10/3/09 (www.wunderground.com), as well as Susquehanna River height in Lewisburg, PA (USGS). The hydrograph indicates precipitation quantities while the individual points indicate which types of geophysical data were collected on specific days.

**Seismic: Data Collection Methods**

**Seismic Method Background:** Seismic signals are low frequency signals from 1 Hz to 1 kHz, which allows seismic signals to propagate further with little loss. Thus seismic data provide a method to analyze relatively deep subsurface areas compared to other methods through correlating seismic velocities with potential subsurface materials (Table 1). Seismic signals
typically travel faster with depth; hence refraction analysis may be used to interpret the subsurface material properties (Burger et al., 2006).

**Table 1.** Seismic velocities for common consolidated and unconsolidated geologic materials (Jacob 2009).

<table>
<thead>
<tr>
<th>Material</th>
<th>Unsaturated</th>
<th>Water-saturated</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sand</td>
<td>200-1000</td>
<td>900-2000</td>
</tr>
<tr>
<td>Sandy-gravel</td>
<td>400-600</td>
<td>900-1600</td>
</tr>
<tr>
<td>Clay</td>
<td>700-1200</td>
<td>1100-2500</td>
</tr>
<tr>
<td>Alluvium</td>
<td>400-900</td>
<td>1000-2000</td>
</tr>
<tr>
<td>Soil</td>
<td>320-450</td>
<td>1000-1800</td>
</tr>
<tr>
<td>Weathered bedrock</td>
<td>300-900</td>
<td>1200-1800</td>
</tr>
<tr>
<td>Granite</td>
<td>4200-5500</td>
<td>5000-6500</td>
</tr>
<tr>
<td>Basalt</td>
<td>5500-6200</td>
<td></td>
</tr>
<tr>
<td>Sandstone</td>
<td>2500-5100</td>
<td>3000-5500</td>
</tr>
<tr>
<td>Limestone</td>
<td>3300-6200</td>
<td></td>
</tr>
<tr>
<td>Metamorphic rocks</td>
<td>3000-6500</td>
<td></td>
</tr>
<tr>
<td>Andesite</td>
<td>5000</td>
<td></td>
</tr>
<tr>
<td>Shale</td>
<td>3700-5000</td>
<td>5300</td>
</tr>
<tr>
<td>Quartzite</td>
<td>3000-5400</td>
<td></td>
</tr>
</tbody>
</table>

As previous geophysical studies have shown (Bock, 1992 and Jacob, 1997a), the seismic method is expected to provide water table and bedrock elevations. Seismic data record the amplitudes of acoustic waves arriving at the ground surface as a function of time after propagating throughout the ground as a result of a seismic source, for example the strike of a weight against the ground surface. These waves propagate away from the source in all directions and may be illustrated using raypaths, which are approximate paths perpendicular to wavefronts. One such raypath travels parallel to the ground surface but within the subsurface,
and is referred to as a direct phase. These rays will also travel into the subsurface and reach any subsurface interface where the seismic properties change on either side of the interface (such as the water table). According to Snell’s law, there are two raypaths that seismic signals can travel after being incident on an interface, 1) reflection and 2) transmitted or refracted (Figure 6). When the material below the interface has a higher seismic velocity than the overlying material, as is the case at the water table, there is a single incident angle that will cause the raypaths to refract parallel to the interface and propagate at a velocity of the material below the interface. This raypath is called the refracted path and is shown in Figure 6 where the incident angle leads to the critical refraction may be calculated Snell’s law:

\[
\sin(\theta_i)/\sin(\theta_r) = V_1/V_2, \quad \text{Eq 1}
\]

where \(V_2\) is the seismic velocity of the upper material in Figure 6 and \(V_2\) is the velocity of the next deeper subsurface layer (Figure 6). Refracted rays may be used to determine multiple interface depths and layer velocities, as well as estimate interface dip angles. These interfaces potentially represent changes in lithology or in hydrologic properties. \(\theta_i\), the incident angle, represents the angle between the line normal to the interface and the direction of the wave path traveling from the source to the interface, while \(\theta_r\), the refracted angle, represents the angle between the line normal to the interface and the refracted wave path below the interface. At the critical angle \(\theta_{ic}\), which is equal to \(\theta_r\), \(\sin(\theta_{ic}) = V_2/V_3\), thus in order for the angle of refraction to be detected for both interfaces, it is necessary that \(V_1 < V_2 < V_3\).

In order to record the seismic signal arrival times at increasing distances from the source, an array of geophones are arranged linearly from the source (Figure 7). Spacing and number of
geophones used in an array dictate the precision and maximum depth of the investigation. An increase in the distance between the source and the geophones will result in a greater depth of investigation, but will also decrease the precision of the entire analysis.

If the seismic velocity of a layer is less than the overlying layer, the wave does not become critically refracted and the refraction method cannot be used to infer layer velocity or interface depth for that particular region or for deeper sections of the subsurface.

![Diagram](image)

**Figure 6.** This diagram illustrates the position of the incident angle (indicated as $\theta_1$) and the refraction angle (indicated as $\theta_2$) relative to the interface between the two uppermost layers.
Seismic Refraction Field Procedure: The procedure for collecting seismic refraction data at the research area began by attaching either 12 or 24 geophones to seismic cable which is connected to a seismograph (the RAS-24 by Seistronix and field laptop). The computer records magnitude and direction of particle motion at the geophone caused by the seismic wave propagating from the source. The seismic source, a 10 or 16 pound sledgehammer, was struck on a metal plate on one side of the geophone array, and the seismograph recorded the forward seismic profile (Figure 7). This was followed by placing the seismic source on the other side of the geophone array, striking the plate and recording the reverse seismic profile. The distance between the source and the nearest geophone was consistent for each forward and reverse shot, as well as the distance between the geophones in the array for lines collected during August 2009 and the line collected during May 2009 using a 5m geophone-source offset. For May 2009 data using a 0.5 m source-geophone distance, the geophone nearest to the source was moved 0.5 m closer to the nearest geophone. For the second seismic data set, collected on 8/15/2009, the number of geophones was doubled from 12 to 24, over the same distance as the previous profiles, and the source was aligned 1 m from the first geophone (Figure 7).
**Figure 7 (see previous page).** Seismic line parameter variation from 5/19/2009 to 8/15/2009 for the forward direction source, where the source is located on the western side of the seismic line.

Horizontal interfaces in the subsurface will produce equal arrival times observed as seismic energy propagates in both forward and reverse profiles (Burger et al, 2006). Seismic signal arrival times at an increasing distance from the source then provide the velocity of the subsurface material. Dipping interfaces in the subsurface result in travel times which are consistently less in one direction (forward or reverse) for the refracted energy from that interface than the travel times for the opposite direction. For a forward profile, where the source is located in its original location, the wave arrival time is expected to be zero at zero offset; the direct signal arrivals were sufficiently close to zero to assume this. The combination of data from the forward and reverse profiles is then used to calculate the velocity of the subsurface material, the dip of subsurface interfaces, as well as the depth to the interface at either edge. Seismic data were collected along a line centered on the gravel ridge on 5/19/2009 and 8/15/2009, in order to determine fluctuations in water table elevation as well as general seismic velocity structure for the subsurface at the research area.
Figure 8. Direct, refracted, and reflected seismic raypaths are shown propagating through an idealized subsurface overlying a homogeneous halfspace (an area where the deeper interface boundary is not visible). The wave arrivals are displayed on a seismogram, shown to the lower right of the figure (Hermance, 2004).
Seismic: Data Interpretation Methods

Using the traveltimes of seismic signals recorded at each geophone, layer velocities and interface depths were calculated for the seismic refraction data collected on 19 May and 15 August of 2009. An example of a seismogram collected for each seismic source is provided in Figure 9, which is from the software program called RAS24 from Seistronix.
Figure 9. The RAS-24 computer program displays seismic signals detected for a reverse profile (east to west), centered on the Montandon gravel ridge peak and collected on 5/19/09. The first breaks are indicated by the horizontal lines above where each signal begins.
Arrival times for each geophone were chosen by selecting the first positive (right relative to increasing time) break, or where the amplitude changed from earlier traveltimes. With travel time (in milliseconds) along the y-axis (displaying seconds in Figure 9) and geophone horizontal offset from the source (in meters) along the x-axis (displaying geophone number), each first break is plotted, ideally yielding a set of linear slopes related to either the direct wave or refracted raypaths (Figure 10).

Figure 10. First breaks from forward (west to east) and reverse (east to west) seismic profiles display three changes in slope, indicating three subsurface layers have been detected.
The first slope, extending from the origin, the direct phase, was used to determine seismic velocity of the material comprising the surface layer of the gravel ridge. The velocity was calculated by using linear regression to fit a line to the points comprising the line containing the steepest slope, calculating the slope, then taking its inverse. The resulting velocity was compared to common seismic velocities (Table 1) for a variety of geologic materials to infer possible subsurface composition. A decrease in slope with increasing offset in the arrival times of the first break picks indicated an increase in seismic velocity, and therefore an interface — a significant change in geologic or hydrologic composition.

The depth of the interface on either side of the seismic line along the east-west profile may be calculated through determining the critical incident angle:

\[ \Theta_{ic} = \left( \sin^{-1}(V_1 m_d) + \sin^{-1}(V_1 m_u) \right) / 2 \]  
Eq. 2

where \( V_1 \) is the velocity of the shallowest layer, \( m_d \) is the slope of the second refractor calculated from the forward profile, and \( m_u \) is the slope of the second refractor calculated from the reverse profile (Burger 2006). The distance of the line extending from the surface perpendicular to the first interface is then necessary in order to calculate interface depth on either side of the profile:

\[ T_{id} = \left( 2j_d \cos \Theta_{ic} \right) / V_1 \]  
Eq. 3

and

\[ T_{iu} = \left( 2j_u \cos \Theta_{ic} \right) / V_1 \]  
Eq. 4

where \( T_{id} \) and \( T_{iu} \) are the intercept times at the source location for forward and reverse profiles, respectively, and \( j_d \) and \( j_u \) are distances along lines drawn perpendicular to the interface from
the source to the first interface for the forward and reverse profiles, respectively. The dip angle is necessary in order to calculate interface depth on either side of the profile, and is determined as follows:

$$\beta = (\sin^{-1}(V_1m_u) - \sin^{-1}(V_1m_d))/2.$$  \hspace{1cm} \text{Eq. 5}

The interface depths on either side of the profile are therefore:

$$h_d = j_d/\cos\beta$$  \hspace{1cm} \text{Eq. 6}

$$h_u = j_u/\cos\beta,$$  \hspace{1cm} \text{Eq. 7}

where $h_d$ is the interface depth below the source for the forward profile and $h_u$ is the interface depth below the source for the reverse profile. The reader is referred to Burger et al. (2006) for further explanation.

These equations may be used to calculate the subsurface structure by hand or another possible method, which is particularly useful to determine subsurface interface depths and layer velocities for multiple layers, is to use the software REFRACT (Burger, 2006). Depending on how many interfaces are apparent in seismic line data, model parameters including the number of layers, estimated velocities and apparent dip angles are entered, along with travel times for the forward and reverse profiles. The model refraction slopes are matched to those from the data in order to infer velocities, dip angles and directions for each apparent interface (Figure 10). The velocities (see Table 2) may be used to infer possible subsurface lithology and an indication of SWC.
**Seismic: Results**

A total of two sets of seismic data were collected during the research period. One set of seismic profile data was collected on 5/19/09, which consisted of two reversed seismic profiles with 12 geophones spaced at an interval of 2m. Figure 7 shows the geophones aligned east-west on top of the gravel ridge, centered at the highest elevation within the research area. The first reversed profile was collected with the source 0.5 m away from the first geophone in both the forward and reverse directions. The second reversed profile was collected with the source 5 m away from the first geophone in order to increase the depth range of seismic refraction data while not decreasing the resolution of the seismic refraction data by increasing the geophone interval. The first breaks for each geophone were determined visually, which was a result of the arrival time for the fastest seismic raypath. While 2m spacing between geophones and an offset of 0.5m between the source and the first geophone proved sufficient to detect the interface between the surface layer and its underlying layer, in order to detect a third subsurface layer the 5 m offset source from the first geophone was necessary (Figure 7).

In order to determine if the subsurface interfaces observed in the May data changed, thus suggesting a hydrologic boundary, a reversed profile was collected on 8/15/09 along the same east-west line and centered in the same position at the highest elevation as shown in Figure 7. There were three differences in the August setup, 1) twenty-four geophones were used, 2) the geophones were spaced 1m apart and 3) the offset between the source and the first geophone was 1m. Figure 11 displays first break time picks plotted against the offset between each geophone and the source for the forward and reverse profiles for all seismic data:
The velocity of the surface material was 305 m/s during May and 246 m/s during August, with a consistent depth to the lower-bounding interface of 2.6 m. The second shallowest layer had a seismic velocity of 701 m/s during May and 889 m/s during August, with a depth to the lower-bounding interface of 5.46 m during May and 7.25 m during August (Table 2), calculated by adding together layer thicknesses. Because some seismic data from the forward profile displayed an unexplained time delay, velocities calculated from equations (Burger, 2006) utilized only the velocities from the reversed profiles (Figure 12).

**Figure 11.** The direct and first refracted phases are displayed for the forward and reverse profiles for the seismic lines collected on 5/19/09 and 8/15/09.
Figure 12. Reversed profile arrival times and slopes for 5/19/2009 and 8/15/2009.
Table 2. Seismic velocities and layer thicknesses calculated for each set of seismic data, through using seismic refraction equations (Burger, et al) or the program REFRACT. Units for velocity are m/s and for layer thickness m. Due to the possibility of multiple dipping interfaces, the velocity of the third layer was solely calculated by REFRACT. The layer thickness refers to the depth to V1; the depths to V2 and V3 required REFRACT to determine. These will be presented in the interpretations.

<table>
<thead>
<tr>
<th>Date</th>
<th>V1 (m/s)</th>
<th>V2 (m/s)</th>
<th>Depth to V1 (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>5/19/2009 (0.5 m)</td>
<td>252</td>
<td>708</td>
<td>2.26</td>
</tr>
<tr>
<td>5/19/2009 (5 m)</td>
<td>400</td>
<td>735</td>
<td>2.27</td>
</tr>
<tr>
<td>8/15/2009 (1 m)</td>
<td>260</td>
<td>902</td>
<td>2.63</td>
</tr>
</tbody>
</table>

Seismic Refraction Calculations (from REFRACT program in Burger et al)

<table>
<thead>
<tr>
<th>Date</th>
<th>V1 (m/s)</th>
<th>V2 (m/s)</th>
<th>V3 (m/s)</th>
<th>Depth to V1 (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>5/19/2009 (0.5 m)</td>
<td>261</td>
<td>706</td>
<td>1394</td>
<td>2.8</td>
</tr>
<tr>
<td>5/19/2009 (5 m)</td>
<td>308</td>
<td>653</td>
<td>1329</td>
<td>2.44</td>
</tr>
<tr>
<td>8/15/2009 (1 m)</td>
<td>232</td>
<td>875</td>
<td>1233</td>
<td>2.47</td>
</tr>
</tbody>
</table>

Seismic: Interpretations

The purpose of the seismic data was to locate the depth to the water table. The velocity of the deepest detected layer was 1362 m/s during May and 1233 m/s during August, suggesting clayey gravel and saturated gravelly sand. The seismic velocities of the surface layer, calculated using the seismic refraction method, are within the range of velocities for unsaturated sand (200-1000 m/s) and the relatively low values within the range suggest that the sand has very low moisture content. The increase in average velocity for the second layer from May to August suggests an increase in moisture content between 5/19/2009 and 8/15/2009, but the velocity remains within the range of unsaturated sand. Average seismic
velocities for the third layer, correspond with velocity ranges for saturated sand and gravel (900-1600 m/s), saturated sand (900-2000 m/s) (Bonner and Schock, 1981).

Seismic velocities and layer depths are not consistent with previous seismic refraction interpretations. Jacob (1997a) observed two interfaces along an E-W traverse of the wetland, located east of this study’s research site, at depths of 0.9 m and 3.7 m. Velocities from the surface layer to the third deepest layer are 205, 1370, and 5284 m/s. Because bedrock velocity is greater than 5000 m/s in the previous study, seismic velocities from current research indicate that the bedrock was not located through use of the seismic method. The differences in interpretations are primarily due to significant layer thickness (greater than one meter) variation over the distance between the line in this study and the line from previous research (Jacob 1997a).

Previous well log data from Well 8, not including a 1m surface layer of dune sand, provides evidence for a thin (<1m) layer of clayey gravel overlain by a 5m layer of sandy gravel (Figure 4). The first interface for both data collections does not vary significantly; it is likely that the location of the interface corresponds with the change in geologic material from dune sand to the sandy gravel of the Pleistocene terrace. The thickness calculated for the second layer varies to a greater degree and correspond with geologic layer thicknesses provided by previous well log data. If the second interface is interpreted to be the water table location, the increase in velocity of the second layer from 5/19/2009 to 8/15/2009 may be due to increased moisture in the unsaturated zone above the water table. This increased moisture may be caused by the rainfall immediately before the data collection. The velocity of the third layer is likely greater
for 8/15/2009 than for 5/19/2009 because the material below the second deepest interface is a saturated clayey gravel on 8/15/2009, which has a higher seismic velocity than a saturated sandy gravel, which was the material below where the water table was located on 5/19/2009 as shown in Figure 13.

<table>
<thead>
<tr>
<th></th>
<th>5/19/2009</th>
<th>8/15/2009</th>
</tr>
</thead>
<tbody>
<tr>
<td>dune sand</td>
<td></td>
<td></td>
</tr>
<tr>
<td>silty sand</td>
<td>V1 = 305 m/s</td>
<td>V1 = 246 m/s</td>
</tr>
<tr>
<td>sandy gravel</td>
<td>V2 = 701 m/s</td>
<td>V2 = 889 m/s</td>
</tr>
<tr>
<td>potential water table</td>
<td>V3 = 1233 m/s</td>
<td></td>
</tr>
<tr>
<td>clayey gravel</td>
<td></td>
<td></td>
</tr>
<tr>
<td>sandy gravel</td>
<td></td>
<td>V3 = 1362 m/s</td>
</tr>
<tr>
<td>weathered limestone bedrock</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

![Image](image)

**Figure 13.** The seismic velocities and interface depths for data collected on 5/19/2009 and 8/15/2009 are shown with respect to subsurface lithologies determined from previous well log data from Well 8.

Seismic data from this research project do not contain the depth to bedrock because the length of the line and/or number of geophones used for both days of data collection were insufficient to detect the change in the traveltime slope indicative of bedrock. Jacob (1997a)
determined the bedrock velocity near the current research site to be between 3500 and 5000 m/s, indicating that the third layer detected by this study likely is not limestone bedrock, but instead unconsolidated clayey and sandy gravel.

**DC Resistivity: Data Collection Methods**

Resistivity data is collected to determine the electrical resistivity of the subsurface. Natural ground water tends to carry ions which conduct electricity efficiently, so a significantly increased proportion of water within a layer causes the medium to be less electrically resistant – in other words, saturated sediments have a lower resistivity than unsaturated sediments. There are multiple ways to collect resistivity, for practical purposes it is advantageous to use four electrodes, wires and a resistivity meter where the resistivity meter is at the center and then space electrodes equally in a line away from the center. The two outer electrodes place current into the subsurface and the two “inner electrodes” (see Figure 14) measure the voltage change caused by the current electrodes. From these values, mutual resistance can be calculated, and the distance between any electrodes (“a-spacings”) may be incorporated to calculate resistivity as a function of volume of the subsurface represented by the current filament in Figure 14.
Several DC resistivity soundings, centered at the crest of the Montandon gravel ridge, were collected during mid-late May and late August. The a-spacings of most soundings collected during the late spring range from 0.5 to 10 m. In order to determine the resistivity of bedrock and depth to bedrock, some soundings extend to 50 m a-spacing. Because bedrock location and bedrock resistivity are not expected to change over a seasonal time period (as any changes to the bedrock are expected to occur over a longer time than months), few soundings contain a-spacings greater than 20 m. In order to potentially increase apparent resistivity precision for measurements taken at small a-spacings and hence shallow depths, a-spacings of 0.6, 0.7, 0.8 and 0.9 m were added to resistivity soundings collected during August and October. Because accuracy of the mutual resistance measurements decreases as a-spacing decreases,
five readings have been taken for each a-spacing, including all a-spacings for most soundings, collected in August and the sounding collected in October.

**DC Resistivity: Data Interpretation Methods**

Apparent resistivity is calculated by the Sting resistivity meter using the equation

$$\rho = 2\pi R a,$$

where “\(\rho\)” equals apparent resistivity, “\(R\)” equals mutual resistance, and “\(a\)” equals a-spacing length in a Wenner array (Figure 14). Due to observed instrument error in earlier resistivity measurements, particularly within measurements from a-spacings < 1m, five readings per a-spacing were taken for later resistivity soundings. Weighted averages were calculated for resistivities for each a-spacing by subtracting the percent errors from 1% (percent errors equal to or greater than 1% were considered invalid and could almost always be traced to incorrect array configuration or instrument malfunction). The averaged apparent resistivity values were plotted along the y-axis against a-spacing values along the x-axis on a log-log plot (Figure 15).

A significant increase or decrease in apparent resistivity indicates the presence of an interface, while the resistivity of each layer is estimated by reading the apparent resistivity value that most accurately represents a constant resistivity. The first resistivity value approximately represents the resistivity for the surface layer, to a depth of 0.5 m. Depth to an interface can be estimated by measuring the a-spacing for which a sharp change in resistivity
occurs. However, these estimations are relatively rough and it is necessary to determine layer thicknesses and resistivities more accurately using the DC Wenner Excel program (Figure 16).

![4/20 Montandon Gravel Ridge DC Resistivity Sounding](image)

**Figure 15.** Apparent resistivity, collected on 4/20/09 from a sounding, is plotted against a-spacing on a log-log plot. In this resistivity curve, three layers are visible.

The number of layers is estimated and entered into the program, along with estimates of resistivities and thicknesses for each layer. The data points are then entered and the program generates a resistivity curve according to the estimated parameters, which are adjusted accordingly to accurately fit the experimental data.
Because one single combination of resistivities and layer thicknesses that results in an accurate resistivity model with respect to the experimental data does not exist, it is necessary to implement parameter restrictions that correctly reflect general known properties of the
Montandon gravel ridge. For instance, the resistivity of saturated sand is not expected to change because the moisture content of saturated material would not vary significantly; therefore resistivity for the saturated layer is kept consistent for all resistivity models. Likewise, the resistivity of limestone bedrock is not expected to change within the relatively short time frame of the research project, thus it is also kept consistent for all resistivity models.

Occasionally, a spike in resistivity is observed which may or may not be an outlier. In the case where only one data point appears to not fit within a smooth resistivity curve, two resistivity models are generated; one with the minimum possible number of layers while omitting the extraneous data point, and one with additional layers to include all points. These resistivity models are compared to resistivity data from other soundings to infer which model is most accurate or provides the most complete description of the subsurface (Figure 17).

<table>
<thead>
<tr>
<th>Title: DC Wenner Response of N-layered medium</th>
<th>Resistivity (ohm-m)</th>
<th>Thickness (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>4 (Number of layers)</td>
<td>200</td>
<td>0.6</td>
</tr>
<tr>
<td></td>
<td>1000</td>
<td>4.8</td>
</tr>
<tr>
<td></td>
<td>150</td>
<td>3</td>
</tr>
<tr>
<td></td>
<td>260</td>
<td></td>
</tr>
</tbody>
</table>

Figure 17. Resistivity curve for the same data collected as the previous figure, but with an additional layer in order to increase goodness of fit.
DC Resistivity Results

Table 3 displays apparent resistivity data collected throughout the entire research period. The resistivity of bedrock was determined to be approximately 270 ohm-m, while the depth to bedrock was interpreted to be around 10-15 m based on apparent resistivity alone.
Table 3. Apparent resistivity values for soundings collected during late spring, late summer and early fall.

<table>
<thead>
<tr>
<th></th>
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<th></th>
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<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>a-spacing (m)</td>
<td>0.5</td>
<td>0.6</td>
<td>0.7</td>
<td>0.8</td>
<td>0.9</td>
<td>1.0</td>
<td>1.2</td>
<td>1.5</td>
<td>2.0</td>
<td>2.5</td>
<td>3.0</td>
<td>3.5</td>
<td>4.0</td>
<td>4.5</td>
<td>5.0</td>
</tr>
<tr>
<td>20-Apr</td>
<td>453.1</td>
<td>482.9</td>
<td>506.2</td>
<td>441</td>
<td>386.7</td>
<td>407</td>
<td>370.9</td>
<td>420.7</td>
<td>240.2</td>
<td>228.3</td>
<td>256.5</td>
<td>331.0</td>
<td>369.1</td>
<td>353.4</td>
<td>250.8</td>
</tr>
</tbody>
</table>
Apparent resistivities for each day of data collection were plotted on a log-log plot to directly observe resistivity variation (Figure 18).

![DC Resistivity on the Gravel Ridge](image)

**Figure 18.** Apparent resistivity curves for soundings collected during late spring, late summer and early fall.

Additionally, a resistivity profile, using a 10m a-spacing, 3m step size and extending from 60m east of the Montandon gravel ridge peak to 60m west of the peak, was collected in order to
determine the relative apparent resistivity on the ridge peak compared to surrounding subsurface resistivities (Table 4, Figure 19).

**Table 4.** Apparent resistivities for the resistivity profile collected on 5/12/09.

**Montandon Gravel Ridge Resistivity Profile (3m step size, 10m a-spacing)**

<table>
<thead>
<tr>
<th>Distance from Center (m)</th>
<th>Apparent Resistivity (Ωm)</th>
<th>Distance from Center (m)</th>
<th>Apparent Resistivity (Ωm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>-60</td>
<td>443.8</td>
<td>3</td>
<td>633.7</td>
</tr>
<tr>
<td>-57</td>
<td>448.0</td>
<td>6</td>
<td>733.9</td>
</tr>
<tr>
<td>-54</td>
<td>457.2</td>
<td>9</td>
<td>809.3</td>
</tr>
<tr>
<td>-51</td>
<td>458.3</td>
<td>12</td>
<td>742.6</td>
</tr>
<tr>
<td>-48</td>
<td>458.7</td>
<td>15</td>
<td>647.0</td>
</tr>
<tr>
<td>-45</td>
<td>456.5</td>
<td>18</td>
<td>613.9</td>
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<tr>
<td>-42</td>
<td>476.0</td>
<td>21</td>
<td>640.6</td>
</tr>
<tr>
<td>-39</td>
<td>547.4</td>
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<td>648.7</td>
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<tr>
<td>-36</td>
<td>508.6</td>
<td>27</td>
<td>659.4</td>
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<td>-33</td>
<td>515.0</td>
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<td>-30</td>
<td>529.3</td>
<td>33</td>
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<td>602.1</td>
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<td>652.2</td>
<td>57</td>
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<td>-3</td>
<td>604.1</td>
<td>60</td>
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<tr>
<td>0</td>
<td>620.4</td>
<td></td>
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</tbody>
</table>
Resistivity and depth values have been estimated based on obtaining a resistivity curve that best fits the data for each sounding. Resistivity curves are based upon gravel ridge geology inferred from borehole data, with an addition of approximately one meter of dune sand to compensate for geological differences between the locations of the highest elevation and the nearest well. A-spacings selected for earlier resistivity soundings, up to 50 m, were chosen in order to detect the water table and bedrock. These soundings (Figure 18) displayed a converging apparent resistivity of approximately 300 ohm-m at an a-spacing of 25-30m, corresponding to a resistivity model with a bedrock depth of approximately 10.5 m at 440 ohm-m. Additionally, soundings collected during late spring and early summer displayed an

**Figure 19.** Apparent resistivity profile aligned east-west and centered on the gravel ridge with 10m a-spacing and 3m step size.

**DC Resistivity: Interpretations**

Resistivity and depth values have been estimated based on obtaining a resistivity curve that best fits the data for each sounding. Resistivity curves are based upon gravel ridge geology inferred from borehole data, with an addition of approximately one meter of dune sand to compensate for geological differences between the locations of the highest elevation and the nearest well. A-spacings selected for earlier resistivity soundings, up to 50 m, were chosen in order to detect the water table and bedrock. These soundings (Figure 18) displayed a converging apparent resistivity of approximately 300 ohm-m at an a-spacing of 25-30m, corresponding to a resistivity model with a bedrock depth of approximately 10.5 m at 440 ohm-m. Additionally, soundings collected during late spring and early summer displayed an
apparent resistivity of approximately 700 ohm-m at a 7 m depth, indicative of saturated sand and gravel (Burger et al., 2006). Because the resistivity of saturated sand is expected to remain constant, it is likely that the shallower layer resistivity convergence represents the location of the water table. Evidence for this is supported by later soundings collected during late summer, where the apparent resistivity values are within 20 ohm-m of each other but the a-spacing of the apparent resistivity shifts from 7 m to 10 m. The decrease in apparent resistivity at a-spacings smaller than the interpreted water table is potentially due to decreased precipitation from early to late summer, and the similarity of these values suggests that within these time periods, there have not been any immediate changes in water table depth caused by individual precipitation events. Therefore, saturated sand and bedrock resistivities have been kept constant in resulting resistivity models (Figure 20).
Originally, relatively small a-spacings included 0.5, 1, and 1.5 m alignments. However, in order to increase resolution at shallower depths, 0.6, 0.7, 0.8 and 0.9 m a-spacings were added in soundings collected during late summer. Apparent resistivities for these a-spacings displayed significant fluctuation on a daily time scale (Figure 18). For example, during 8/12/09 to 8/14/09, a time range with no significant precipitation, there were observable changes in resistivity at depths of 0.5 to 1 m (Table 3). During 8/12/09, resistivity dropped approximately 70 ohm-m at 0.6-0.7 m a-spacing relative to 0.5 and 0.8 m depths, while during 8/13/09 resistivity decreased at slightly deeper depths of 0.9 m by the same amount. The first decrease in resistivity during 8/14/09, of 4.3 ohm-m, occurred at a depth slightly greater than 1 m. The previous rainstorm occurred on 8/9/09 with 1.53 inches of precipitation (Figure 5, www.wunderground.com), which may have caused the decrease in resistivity, suggesting that the resistivity change is due to a decrease in SWC due to length of time since the previous rainstorm.

Apparent resistivity and layer thickness values were calculated for input models of subsurface layers using an Excel resistivity program (DC Wen from Hermance, 2004). The apparent resistivities of saturated sand and bedrock were held constant for all models and were
440 ohm-m and 250 ohm-m, respectively. The depth to bedrock, within a 10-15 m range based on the location of resistivity convergence was also held constant for all models and correlated with bedrock depth ranges previous seismic studies (Jacob 1997a, Bock 1992) and borehole data. The depth to bedrock was based upon previously inferred subsurface geology near the research site.

According to resistivity curves calculated from apparent resistivity data, there is an overall increase in resistivity of the unsaturated zone between May and October (Figure 21), though the existence of higher frequency fluctuations in resistivity is possible during times when data are not available. From May to August, unsaturated zone resistivity decreases by approximately 50 ohm-m. However, this change is smaller than what would be expected for a significant change in geologic or hydrologic subsurface properties (100s of ohm-m). Because the standard deviation of resistivity readings does not exceed 2 ohm-m for any a-spacing value, we cannot conclude that error is the primary cause of variation in resistivity between May and August. One possible explanation is that, as the subsurface moisture content decreases (Figure 5), ions are remobilized, increasing groundwater ion concentration as precipitation infiltrates throughout the unsaturated zone. The increase in resistivity in October may possibly be explained by ion demobilization through moisture loss as the amount of precipitation decreases. These data do not prove that the change in resistivity between May and August is significant; rather, they reliably indicate an increase in resistivity between May and October. However, from DC resistivity data alone it is impossible to infer whether the drop in resistivity is caused by an increase or decrease in moisture content. An increase in moisture content may free ions released during sediment weathering, thus increasing dissolved ion concentration and lowering
resistivity. Alternatively, a decrease in resistivity may also occur when the SWC decreases due to drainage to the water table or evapotranspiration.

Figure 21. This graph displays the resistivities of the two shallowest layers in the resistivity model over a seasonal time period, from May 2009 to October 2009.

Summary of DC Resistivity Interpretations: Resistivity models showed a temporary, small decrease in resistivity due to storm events within the two shallowest layers, comprising dune sand and underlying sandy gravel. 0.92 inches of precipitation on May 14 resulted in a decrease of 50 ohm-m, then a gradual increase of 100 ohm-m throughout the week (Figure 22). However, because these changes are lower than what is expected for sudden addition of water to the shallow vadose zone, the data are inconclusive in proving how vadose zone resistivity changes due to individual storm events.
Ground Penetrating Radar (GPR): Data Collection Methods

**General GPR Background:** GPR radiates a high-frequency electromagnetic (EM) pulse from a transmitting antenna and then records the arrival of the high-frequency EM pulse at a receiving antenna – typically two separate GPR antennas – transmitter and receiver. The GPR pulse radiates outward from the transmitter and travels through both the atmosphere and the near-surface material (see Figure 23) at a velocity based on the electrical properties (dielectric and conductive) of the material, for example, the velocity of a GPR signal in air is 0.299792 m/ns. The velocity of a GPR signal propagating through subsurface is inversely related to the water content of the subsurface as shown in Figure 24, and typically ranges from 0.06 m/ns for a wetter subsurface material to 0.12 m/ns for a drier subsurface material (Annan, 2006).

*Figure 22.* This graph displays the resistivities of the two shallowest layers in the resistivity model from 5/12/2009 to 5/20/2009, due to a precipitation event on 5/14/2009.
The GPR antenna frequency dictates both depth and resolution within the shallow subsurface (Annan, 2005). The antennas used for this research include 100, 200 and 500 MHz, as the higher frequencies contain maximum resolution but are limited to shallower depths, while 200 and 100MHz frequencies are capable of measuring traveltimes at greater depths than higher frequencies. Therefore, data collected using 500 MHz GPR antennas are particularly useful in determining levels of moisture gradient variation within the soil zone (less than 2m depth) while the lower frequency antennas are expected to provide information on moisture changes below the soil zone – below 1m. Combining GPR data from multiple frequencies provides maximum resolution for a variety of depths within the shallow subsurface. Because GPR signals typically travel at slower velocities as depth increases, reflection analysis is used to calculate velocities and depths of observed interfaces (Bohidar and Hermance, 2001). Examples of GPR signals collected from CMPs and profiles are shown in Figures 25 and 26, respectively.

Seismic data collection is similar to GPR data collection except that it propagates sound waves instead of radio waves. Combining precise shallow (<3m) subsurface velocities with the correlation between sediment moisture and GPR velocity provides hydrologic information within the shallow vadose zone that cannot be determined by lower-frequency methods such as seismic or by DC resistivity.

**GPR Velocity and Thickness Method – CMP**: A common midpoint sounding (CMP) is used to determine depth to subsurface interfaces and subsurface material velocity where subsurface velocities change due to soil moisture conditions as well as uncertainty for observing the same interfaces while material properties are changing. Transmitter and receiver antennae are aligned linearly so that the edges of the antennas are parallel to each other (Figure 23). The
transmitter and receiver antennas are longer in one direction than the other, as indicated in Figure 23, and may be aligned so that the longer side is perpendicular to the line to collect a transverse electric (TE) sounding, or that the longer sides are parallel to the line to collect a transverse magnetic (TM) sounding. A TE sounding polarizes an incident wave normal to the plane of propagation, while a TM sounding polarizes an incident wave parallel to the plane. The electric and magnetic waves are perpendicular to each other as well as perpendicular to the direction of energy transfer. Therefore, during a TM sounding the magnetic wave is perpendicular to the direction of wave propagation, while during a TE sounding the electric wave is perpendicular to the wave propagation direction. Each antenna is located the same distance away from a chosen midpoint, and the midpoint stays consistent throughout the sounding (Figure 23). A step size, the amount the offset between the transmitter and receiver is increased between travel time measurements, is chosen to maximize measurement resolution for the selected frequency as well as select a reasonable time window for future interpretation. During each measurement readings are taken according to a preset sampling rate. Higher sampling rates are necessary for high-frequency waves in order to accurately represent the wavelength over a period of time. Using standard relationships, velocities measured through collecting CMPs can be used to determine the subsurface water content (Figure 24).
Figure 23. Diagram of electromagnetic waves traveling through subsurface from transmitter to receiver. (Jacob)
Lateral Variation in GPR Signals: A profile is collected to observe travel times at a consistent offset between the transmitter and receiver, while the midpoint is moved at a consistent step size along a line. A single CMP provides depth and velocity based on a reflection in the subsurface; however, it is impossible to determine the source of an observed reflected wave, which could be produced by an interface or a point scatter. If a point scatter is located directly underneath the gravel ridge peak, the associated depth and velocity measured by a GPR CMP does not provide accurate information regarding the overall subsurface structure and moisture content for that particular depth. Consistent traveltime observed along a significant amount of horizontal offset indicates an interface in a profile, while a hyperbolic

Figure 24. This graph depicts the Topp relationship between GPR velocity and water content, which may be used to determine water content at various depths of the gravel ridge.
curve extending from a specific point in the subsurface suggests the reflection originates from a point scatter that is located near the midpoint used to collect the CMP.

Figure 25. This example of data collected from a CMP displays the direct air and direct ground signals, as well as at least one reflected wave, which indicates the presence of an interface (after Jacob, 2008).
GPR: Data Interpretation Methods

GPR CMP Sounding Interpretation Methods: Unlike the seismic method where refraction is used to determine seismic velocities and depths to interfaces, the reflection method is typically used with GPR in order to infer similar information. For each CMP, travel time (in nanoseconds) is plotted on the y-axis against Tx-Rx offset (in meters). The first arrival of a GPR signal is the direct air phase (Figure 25), which appears linear and has the shallowest slope of all observed signals, with a consistent velocity range of approximately 0.29-0.33 m/ns. The other phase whose arrival time increases linearly with offset is the direct ground phase,

Figure 26. This example of a profile displays the direct air phase, a reflecting phase, and several point scatters (after Jacob, 2008).
which displays a relatively steep slope compared to the direct air phase. Velocities for the
direct air and direct ground phases are calculated by picking first breaks and taking the inverse
of their respective slopes. The velocity of the direct ground wave represents electromagnetic
wave velocity through the surface layer. Because signal loss is significant for an
electromagnetic wave traveling through a conductive medium and the EM pulse is directed into
the ground, often the direct ground and direct air signals may appear weak or invisible. Arriving
after the direct air and direct ground phases, reflected phases are displayed as half-parabolas.
The arrival times squared for reflected paths of waves incident on the ground surface increase
linearly with squared Tx-Rx offset. Thus, the travel times and Tx-Rx offsets of first breaks for
reflected phases are squared and plotted on a graph using linear regression to determine
normal moveout velocity of the reflected phase (Figure 27).

Figure 27. The square of Tx-Rx offset is plotted against the square of travel time to determine the slope
from which the velocity of the reflected wave is calculated.
The velocity of a reflected wave is then calculated by taking the inverse of the square root of the resulting slope, while the depth to an interface is approximated by dividing the square root of the two-way traveltime at zero offset (the y-intercept) by two and then multiplying by the velocity. For particularly weak signals, especially apparent in 500 MHz CMPs, SEC gain is added to maximize the ability to select arrival times for each reflected signal.

**GPR Profile Interpretation Methods:** Profiles are plotted similarly to CMPs using the program Ekko View Deluxe, where point scatters may be observed – or not (Figure 26). The depth of the point scatter, which generates a full parabolic signal, is calculated similarly to calculating the depth to an interface using a reflected phase. By comparing the depth of the point scatter to the depth of observed interfaces combined with its horizontal location, it can be inferred whether a reflection phase observed in the corresponding CMP is due to a point scatter or interface.

**GPR: Results**

During the late spring, late summer and early fall several sets of GPR CMP sounding and profile data were collected. Initially these data were collected using three different frequencies: 100, 200 and 500 MHz, in order to maximize resolution at a variety of subsurface depths. Parameters such as sampling rate, step size, and Tx-Rx offset were adjusted to maximize measurement precision as data collection progressed. Additionally, the 100 MHz CMP soundings did not provide deeper reflections. Therefore, CMP soundings and profiles were only collected for 200 and 500 MHz during late summer and early fall data collections (Table 5).
Table 5. These parameters were most commonly used when collecting GPR CMP sounding and profile data, and provided maximum precision for each frequency out of the parameters used throughout data collection.

GPR Parameters for Profile and CMP Sounding Measurements

<table>
<thead>
<tr>
<th></th>
<th>Begin Tx-Rx Offset (m)</th>
<th>End Tx-Rx Offset (m)</th>
<th>Sampling Rate (ns)</th>
<th>Step Size (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>100 MHz Profile</td>
<td>2.0</td>
<td>N/A</td>
<td>0.2</td>
<td>0.5</td>
</tr>
<tr>
<td>100 MHz CMP</td>
<td>1.1</td>
<td>12.1</td>
<td>0.2</td>
<td>0.2</td>
</tr>
<tr>
<td>200 MHz Profile</td>
<td>2.0</td>
<td>N/A</td>
<td>0.1</td>
<td>0.1</td>
</tr>
<tr>
<td>200 MHz Sounding</td>
<td>0.40</td>
<td>10.0</td>
<td>0.1</td>
<td>0.1</td>
</tr>
<tr>
<td>500 MHz Profile</td>
<td>0.25</td>
<td>N/A</td>
<td>0.1</td>
<td>0.05</td>
</tr>
<tr>
<td>500 MHz CMP</td>
<td>0.30</td>
<td>4.0</td>
<td>0.1</td>
<td>0.05</td>
</tr>
</tbody>
</table>

For most GPR soundings collected at 100 and 200 MHz, the direct air phase and three reflected phases are visible, whereas for soundings collected at 500 MHz, only the direct air and the two shallowest reflected phases are visible. The clearest signal is the direct air (Figure 28), because it is the earliest signal to arrive at any offset, and the direct air velocity ranges from 0.29-0.33 m/ns. The next clearest signal in 100 and 200 MHz soundings is the third reflected phase (Figure 28), where velocities can be determined relatively precisely compared to the first two reflections. Error in picking arrival times depends primarily on signal clarity; therefore, time picks for the third reflection exhibit an error of ± 0.05 ns, while error in time picks for the first and second reflections increases to ± 0.5 ns. Additionally, there is little to no signal interference between the third reflected phase and other reflected phases, whereas destructive
interference between the first and second reflected phases is observed in approximately a third of all soundings. The data do not clearly distinguish the direct ground phase from the first reflection (Figure 29), because for half the soundings the velocities calculated for a reflection and for a linear arrival are equal, and the intercept does not consistently occur at zero or a significant offset away from zero. Nine CMPs display possible direct ground phases, while thirteen soundings display potential reflected phases; therefore, the data do not decisively indicate which phase is displayed. Although all reflected phases display changes in velocity due to individual rain events, velocities for the third reflected phase correlate most closely with precipitation times. For example, directly after significant precipitation occurring on 5/14/2009, the observed velocity (measured on 5/15/09) of the third reflected phase is approximately 0.082 m/ns, as is expected for saturated sand (Topp et al, 1982), at a depth of 2.34 m. Data collected during 5/16/09 indicate a slight increase in velocity at a doubled depth to 0.094 m/ns, corresponding to a change in volumetric water content from 0.28 to 0.20 (Topp et al, 1982). Table 6 displays velocities and interface depths for the third reflected phase.
Figure 28. 200 MHz CMP sounding data collected 5/20/09 (step size = 0.1m, Tx-Rx offset ranges from 0.4m to 10m, sampling rate = 0.1 ns).
Figure 29. Phase arrival times are plotted against Tx-Rx offset for the CMP sounding data from Figure 27. The resulting data points are linearly aligned, but there is significant offset from zero, which would be the expected intercept for the direct ground phase.
Table 6. GPR velocities and depths of the third reflected interface from May 2009 to October 2009.

GPR Velocities and Interface Depths for the Third Reflected Phase

<table>
<thead>
<tr>
<th>Third Reflected Phase Velocity (m/ns)</th>
<th>Depth to Third Interface (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>11-May 0.085</td>
<td>2.162444506</td>
</tr>
<tr>
<td>12-May 0.085</td>
<td></td>
</tr>
<tr>
<td>13-May 0.085</td>
<td>2.340</td>
</tr>
<tr>
<td>15-May 0.082</td>
<td>2.335</td>
</tr>
<tr>
<td>16-May 0.094</td>
<td>2.740</td>
</tr>
<tr>
<td>17-May 0.088</td>
<td>2.653</td>
</tr>
<tr>
<td>18-May 0.085</td>
<td>2.368</td>
</tr>
<tr>
<td>19-May 0.086</td>
<td>2.586</td>
</tr>
<tr>
<td>20-May 0.090</td>
<td>2.521</td>
</tr>
<tr>
<td>21-May 0.091</td>
<td>2.630</td>
</tr>
<tr>
<td>12-Aug 0.101</td>
<td>2.296</td>
</tr>
<tr>
<td>13-Aug 0.094</td>
<td>2.450</td>
</tr>
<tr>
<td>14-Aug 0.094</td>
<td></td>
</tr>
<tr>
<td>18-Aug 0.106</td>
<td>2.574</td>
</tr>
<tr>
<td>24-Aug 0.118</td>
<td>2.957</td>
</tr>
</tbody>
</table>

The GPR profiles do not display any point scatters, indicating that reflected waves detected in the CMP soundings are due to reflections off of planar interfaces, instead of point sources or objects buried directly underneath the gravel ridge peak (Figure 30).
Figure 30. This 200 MHz profile was collected on 8/12/09 and displays shallow interfaces, but lacks the parabolic curves indicative of point scatters.
Interpretations

The high conductivity of the surface layer (3.70 mS/m) caused significant loss in electromagnetic signal. Therefore, visible reflected waves indicated relatively shallow interfaces at depths no greater than 3 m. Because the electromagnetic velocity for dry sand is 0.15 m/ns and the geologic composition of the surface layer is not predicted to vary over a day’s time span, it can be inferred that the increase in velocity is the result of a decrease in moisture content caused by infiltration beneath the shallowest 3m of the vadose zone. Only a few 100 MHz GPR data were collected during early summer due to the limited penetration depth, and none were collected during late summer and early fall. The limited penetration depth agrees with the low resistivity of the shallow subsurface layers.

Over a seasonal time period, the data show a linear increase in GPR velocities within the shallowest 3 m of the vadose zone from May 2009 to October 2009, though high-frequency fluctuations within times between days of data collection possibly exist. Precipitation data indicate an overall decrease in precipitation during this time period, so the GPR velocities are attributed to decreased moisture content (Figure 31).
Changes in relative water content due to individual storm events also affect the GPR velocities. 

Rain events on May 14 and 16, resulting in 0.92 and 0.79 inches of precipitation, respectively, are correlated with immediately lowered GPR velocities that gradually increase during days without precipitation (Figure 32). Data collected over both seasonal and daily time periods do not indicate any significant change in the depth of the third reflected interface due to changes in moisture content, which suggests that this is a geologic interface and not a hydrologic interface.

**Figure 31.** This graph displays GPR velocities and depths for the third reflected interface over a seasonal time period from May 2009 to October 2009.
**Combined Data Interpretations**

The combination of seismic, DC resistivity and GPR data indicates that the depth to the water table (or possibly where the capillary fringe begins) increased from May to August 2009 from 5.46 m to 7.25 m, and that the water table was located in either sandy gravel or clayey gravel, depending on the depth. Though the amount of precipitation received from May to October 2009 was unexpectedly higher compared to previous summers, the SWC decreased consistently from May to October, potentially due to increased evapotranspiration through crop growth. Ion concentration decreased from May to October 2009, though it remains unclear how ion concentration changed between May and August 2009.
GPR velocities display evidence for infiltration up to a depth of 3 m over a daily time period and decrease in apparent resistivity over the same time period directly after significant precipitation, which suggests that the increase in SWC observed in the GPR are leading to a decrease in resistivity. Through combining these data and models the effect of moisture infiltration on ground resistivity may be assessed. Therefore, the increase in moisture caused by approximately an inch of precipitation may be sufficient to mobilize trapped ions and result in an overall increase in dissolved ion concentration thus lowering resistivities (Rein, 2004). However, the resistivity data do not definitively indicate changes in vadose zone moisture content due to individual rain events. The precision of GPR data allows for evidence that the subsurface moisture content decreases over a seasonal time period and also displays an immediate increase and gradual decrease after significant amounts of precipitation.

Conclusions
The change in depth to the water table is supported by seismic data, which suggest an elevation decrease from 5.5 m to 7.5 m below the surface at the interface interpreted to be the water table. GPR data from all 100, 200 and 500 MHz frequencies provide information regarding infiltration up to depths of 3 m. Generally, moisture gradient variations are most evident directly after rain events as indicated by precipitation data (Figure 5) and corresponding GPR velocities (Figure 32). The depth range of clear GPR signals does not allow for interpretation regarding water table elevation changes, and the combined data do not indicate immediate changes in water table elevation due to individual rain events. However, reflected wave velocities for 100 MHz data occur at depths up to 3m. Therefore, 100 MHz data may
contain potentially useful information about relatively deeper subsurface velocities that would be useful to acquire during future research.

Combined resistivity and GPR data suggest that the observed seasonal increase in resistivity is due to decreased moisture content in the vadose zone. Directly after precipitation events, ions may be remobilized causing a brief decrease in resistivity, but the changes in resistivity are insufficient to form a conclusive interpretation over a daily time period. The amount of moisture in the unsaturated zone may be temporarily sufficient to increase the number of dissolved ions at a greater rate than they are being diluted by infiltrating rainwater, but over a seasonal time period moisture loss within the vadose zone decreases the dissolved ion concentration, increasing resistivity. The specific ions responsible for decreased resistivity cannot be determined by geophysical methods.

**Future Work**

Initially, 100 MHz GPR data appeared to have relatively low resolution that proved continued data collection for that particular frequency unnecessary. However, what first seemed to be a direct ground wave detected by 100 MHz antennas was actually a reflected phase with calculated depths deeper than those detected by higher frequency data. Therefore, future investigation using lower frequency GPR data is suggested in order to observe infiltration rates within deeper areas of the unsaturated zone.

Additionally, results linking increased moisture content with decreased resistivity suggest a need for further geochemical research near the Montandon gravel ridge, to
determine which ions are being catalyzed by precipitation. Ion concentration of groundwater may be useful in analyzing the stability of the Montandon floodplain and wetland ecosystems.
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