A STUDY OF SOIL MOISTURE ACROSS THE DELMARVA PENINSULA FROM

2005-2008

by

Ryan Ippolito

A thesis submitted to the Faculty of the University of Delaware in partial fulfillment of the requirements for the degree of Master of Science in Geography

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ABSTRACT

The annual cycle and persistence of soil moisture is influenced by a multitude of factors including such variables as soil type, vegetation cover, weather patterns, and seasonal variability. Utilizing data derived from the Delaware Environmental Observing System (DEOS), a network of ground stations located throughout most of the Delmarva Peninsula, as well as remotely sensed observations using the NASA MODIS and AMSR-E satellite sensors, the spatial and temporal variability of soil moisture across the Delmarva Peninsula is determined. The vegetation indices and land surface temperatures provided by the MODIS satellite sensor provide an indication of soil moisture conditions at the land surface, and the AMSR-E satellite sensor gives a quantitative measure of soil moisture conditions at the top 1cm of soil. The remotely sensed data, used in conjunction with the DEOS ground-level observations of daily rainfall, maximum daily volumetric water content (30cm depth), and average daily air temperature provide insight into the conditions responsible for fluctuations in soil moisture conditions. The average soil moisture conditions at each ground station site were compared to one another in terms of soil texture and estimated field capacity. In general, it was found that soils with a higher percentage of sand show greater variability than those with a higher composition of silt and clay. In addition to soil texture, it was found that influences from topography, vegetation cover, and weather can also have a major impact on soil moisture conditions, particularly at the land surface. Using a statistical autocorrelation, soil moisture tends to persist no more than 1-2 weeks across the Delmarva Peninsula.
Chapter 1

INTRODUCTION

1.1 Importance of Soil Moisture

Soil moisture plays a vital role in the existence and stability of a number of important land processes governed by the water cycle. Some of these processes include an influence on mass and energy fluxes between the land surface and atmosphere by evapotranspiration, serving as a limiting factor for vegetative growth, and partially controlling precipitation runoff which can lead to flooding (Sandells et al., 2008). As a result, soil moisture is able to modify the development of weather patterns and precipitation locally, as well as assisting in the regulation of humidity levels.

The soil itself serves as a medium through which the land surface portion of the water budget is controlled. The amount of water plants have access to is directly related to the storage capacity and characteristics of the soil. Important occurrences such as droughts and floods are a function of a soil’s ability to hold moisture. Along with air temperature, humidity, and heat storage potential (a function of soil type and depth), soils are capable of regulating the evaporation rate of surface moisture, and also play a role in the governance of transpiration by plants. Essentially, the storage and transfer of mass and energy into and out of the land surface is managed either wholly or in part by soils.
Soils may also display information about past meteorological conditions, including the intensity and duration of certain extreme events such as droughts and floods. As Li et al. (2004) states, the most important component of meteorological memory, along with snow cover, is soil moisture. The amount of moisture within a sample of soil can offer insight into recent land surface temperatures as the top layer is most exposed to atmospheric fluxes. Since water-filled soil is more efficient at storing heat energy than a dry parcel of soil, the greater the amount of water present in the soil the more likely it is to maintain a consistent and stable overall temperature, as any change in temperature will occur at a much slower rate. As a result, land surface temperatures are dependent on variations in soil moisture.

Climate and hydrologic based land surface models require soil moisture data for precise and accurate simulations. Climate models, in general, do not capture the observed soil moisture variations when forced with either model-generated or observed meteorology. The accuracy of these models depends on adequate moisture and heat representation in the form of stored soil moisture and energy transfer in and out of the land surface (Li et al., 2004), although evapotranspiration and the percolation of moisture below the land surface represent the greatest modeling difficulties. As Robock et al., (2000: 1281) point out, “understanding and predicting variations of surface temperature, drought, and flood depend critically on knowledge of soil moisture variations, as do impacts of climate change and weather forecasting.”
1.2 Observational and Remote Sensing Soil Moisture Datasets

Several observational soil moisture datasets exist for various regions of the globe. However, because soil moisture data are generally not standardized, the measurements are of limited value without considerable initial processing (Robock et al., 2000). Rutgers University in New Brunswick, New Jersey has been collecting and compiling such data as part of the Global Soil Moisture Data Bank ([http://climate.envsci.rutgers.edu/soil_moisture](http://climate.envsci.rutgers.edu/soil_moisture)) for more than a decade to facilitate standardization of soil moisture observations. The Global Soil Moisture Data Bank is dedicated to the collection, dissemination, and analysis of soil moisture data from around the globe. Most of the data (600 stations) consist of \textit{in situ} gravimetric observations of soil moisture extending for at least 6 years and many for more than 15 years. The observed datasets are located primarily in China, Mongolia, the Former Soviet Union (including Ukraine), United States (Iowa and Illinois), and Boissy-Le-Châtel in north-central France.

Soil moisture data is most useful, as Lacava et al. (2006) contends, when it is available at regular and frequent intervals over the greatest area possible. Utilizing a network of ground monitoring stations is one of the most reliable methods for collecting soil moisture data. As network density increases the ability to capture spatial variability is improved.

Satellite remote sensing is an essential component of any comprehensive soil moisture monitoring effort. Remote sensing provides measurements of land surface
data over large areas where no ground stations exist and at regular intervals for extended periods.

Microwave (both passive and active) remote sensing is the preferred method used to detect soil moisture. The advantage of microwave remote sensing is with its relatively long wavelengths (1cm to 1m) there are fewer atmospheric particles (i.e., clouds, rain, dust, smog) capable of interfering with the signal. Microwaves are also able to penetrate the top few millimeters of the land surface allowing for detection of soil moisture. However, the natural emitted energy is smaller, requiring the use of longer wavelengths to decrease the image resolution but improve the signal detection. This creates a footprint ranging from tens of meters to tens of kilometers in size (Schneider et al., 2008), and makes the validation process with ground based data more difficult.

Visible and infrared remote sensing can be used for indirect soil moisture measurement and monitoring. However, these wavelengths are attenuated under cloudy weather conditions and underscore that the most promising method in the development of future global datasets comes from microwave remote sensing. Data from the Global Soil Moisture Data Bank data have been used to validate both passive and active microwave studies demonstrating the potential for microwave remote sensing of soil moisture in regions without snow or tall vegetation (Robock et al., 2000).

Additionally, there are no existing remotely sensed global soil moisture datasets. The Global Soil Moisture Data Bank does provide links to experimental
NOAA products (http://www.noaa.gov/) that use high-frequency passive microwave observations to produce a skin wetness product which is valuable for monitoring droughts and flood conditions. One such product sensor is the Special Sensor Microwave/Imager (SSM/I) located on the Polar Operational Environmental Satellite (POES), capable of making soil moisture observations once daily up to 70mm in depth with an accuracy of 1mm. SSM/I is a passive microwave sensor with 7 channels: 19 GHz (H, V), 22 GHz (V), 37 GHz (H, V), and 85 GHz (H, V), where H and V represent the horizontal and vertical polarizations, respectively (NOAA, 2009).

A more recent in situ observation network, the Delaware Environmental Observing System (DEOS) (http://www.deos.udel.edu), has ground-based observations to compare against remote sensed data. The DEOS network is a real-time system dedicated to monitoring environmental conditions, including soil moisture, and consists of a network of approximately 40 meteorological observation sites in and around Delaware (Legates et al., 2009). This dataset is the first of its kind in the eastern United States.

1.3 Goals and Objectives

The objective of this research is to investigate soil moisture variability across the Delmarva Peninsula using a combination of soil moisture observations from the DEOS network and remotely sensed soil moisture, land surface temperature, and vegetation index products derived from AMSR-E and MODIS satellite sensors. Remotely sensed data is essential to any monitoring program because of the
capabilities of satellites to capture soil moisture variability across large areas at high temporal resolutions. This research represents the first documented description of soil moisture in the Delmarva Region of the eastern United States.

As Robock et al. (2000) have pointed out, an observational dataset of *in situ* measurements is crucial for an accurate climatological analysis, model development and evaluation, and as a comparative basis for ground truth of remotely sensed products. DEOS is one such system, supplying real-time data every five minutes for each of its stations. It is the objective of this research to investigate the spatial and temporal variability of soil moisture and examine how derived remote sensing products compare to ground-level observations across the Delmarva Region.

### 1.4 Outline of Upcoming Chapters

The upcoming chapters will include a literature review focusing on some of the recent research involving soil moisture and remote sensing derived data, as well as details about the soil moisture monitoring capabilities of the DEOS. In addition, details involving each of the nine soil monitoring ground stations will be presented, including information about the soil characteristics and overall vegetative cover at each site. Local and regional maps will be incorporated to offer insight into the general topography and land cover of Delaware and the surrounding areas to provide a characterization of the land surface and identify the physical mechanisms influencing regional soil moisture variability.
Chapter 2

LITERATURE REVIEW

2.1 Background Information

2.1.1 Soil Moisture and the Hydrologic Cycle

Soil moisture is an important component of the hydrologic cycle, controlled by the complex interactions between soil, plants, and climate (Puma et al., 2005). A soil’s ability to absorb, store, manage and transfer water is a limiting factor in plant growth, including the survival and distribution of vegetation, climate and landscape stability, and variability in local weather. It influences such hydrological processes as infiltration, runoff, transpiration and evaporation (Demšar, 2004).

Not all soils are made equally however, and attributes such as composition and texture determine the sort of moisture-holding characteristics a soil displays. For instance, Demšar (2004) has found that the amount of moisture a parcel of soil holds is influenced by infiltration, soil particle size, chemistry, thickness of the soil layer, layering and local soil composition, vegetation cover, terrain roughness, topography, temperature and rainfall intensity. The research of Schmugge et al., 2002 concurs that in hydrology the primary set of state variables include land surface temperature, near-surface soil moisture, snow cover (or the water equivalent), water quality, landscape roughness, land use and vegetation cover. With so many variables at play controlling the hydrological cycle and its influence on soil moisture, it becomes clear to see how (and why) there is so much variation in the moisture handling abilities of soils.
Plants and vegetation have a large impact on a soil’s moisture managing abilities. While soil texture and composition are the primary components in terms of a soil’s moisture holding capacity, it is the local flora which uses the available moisture and essentially connects the subterranean world with the atmosphere above. The three most important hydrometeorological fluxes interconnecting the state variables, according to Schmugge et al. (2002) are soil evaporation, transpiration, and snowmelt runoff. Each of these processes are mechanisms responsible for moving moisture through the soil.

Puma et al. (2005) suggests that by having an understanding of soil moisture within the root zone (the region where active roots reside in the soil) an estimation of fundamental hydrological and atmospheric processes may be obtained. Furthermore, Puma et al. (2005) points out that plants are able to connect the soil to the atmosphere through their active roots, providing water transportation from the root zone to the atmosphere.

Gruhier et al. (2009) have found that soil moisture located near the surface has the greatest influence and control over water and energy exchanges at the soil-vegetation-atmosphere interface. It is at the land surface where the energy and moisture transfer between the ground and atmosphere (and vice versa) take place. Additionally, Robock et al. (2000) have also found that the partitioning of available energy at the ground surface into sensible and latent heat is controlled by near-surface soil moisture (as it is exchanged with the atmosphere). Since the land surface is the meeting point and contact surface between the land and atmosphere, it is logical to
assume that this is where much of the energy transition would occur. This process of energy partitioning at the ground surface is therefore what links the water and energy balances together through the moisture and temperature states of the soil (Robock et al., 2000).

By first analyzing the occurrences (in terms of the movement of moisture) at the land surface, the mechanisms responsible for causing the observed behavior (i.e., water movement) will eventually become apparent. Robock et al. (2000) states that in order to predict the reciprocal influence of land surface processes to weather and climate, an adequate knowledge of the distribution and linkage of soil moisture to evaporation and transpiration is essential.

Wigner et al. (2003) point out that the top 0-5 cm of the ground surface, also referred to as the skin surface, contain soil moisture which is important for estimating the delineation of precipitation between runoff and storage, and to compute a number of essential variables of the land surface energy and water budget. This finding is in agreement with Robock et al. (2000) and Gruhier et al. (2009). Moreover, Wigner et al. (2003) assert that the water content of the skin surface is an important variable needed to estimate the ratio between evaporation and the potential evaporation over bare soils. This confirms that while much of the energy required for soil moisture transport originates from above and below the soil surface, it is the top few centimeters of the soil surface itself that serves as the critical access point for this energy and moisture transfer to take place.
2.1.2 Observing / Measuring Land Surface Data

Soil moisture can typically be measured in three ways: *in situ* measurements (using stationary monitoring equipment), remote sensing, or estimated through hydrological modeling. It is possible to use just one method, although it is common practice to rely on at least two. In this study, *in situ* measurements will combine soil moisture estimates from fixed points with broader coverage (but less precise) remote sensing data across the state of Delaware.

The advantage of *in situ* data collection is that the sensor is actually located in the land surface providing very accurate soil moisture measurements, which are often taken continuously over small, fixed time increments (e.g., 5-minute increments). The drawback is that measurements can only be taken at locations where monitoring stations exist. Even with a relatively large network of monitoring stations, gaps will still exist forcing the use of either remote sensing, interpolation of available data or modeling techniques to estimate values beyond each station.

Remote sensing benefits from greater spatial coverage, consistent monitoring intervals (observations can vary from frequent – daily, to infrequent – monthly), and a multitude of available sensors (i.e., visible, infrared, microwave). While not as accurate (due to lower resolution) as *in situ* measurements, remote sensing still offers acceptable data quality without the spatial gaps associated with station monitoring networks. Still, the lack of high resolution detail makes remote sensing a poor stand-alone technique for measuring soil moisture, but an excellent addition to either modeling or *in situ* collection methods.
Demšar (2004) points out that the remote sensing of soil moisture using airborne observations of microwave emissions has been in existence for the last 30 years. The 1960’s and 1970’s introduced the world to some of the first space-borne satellite sensors including the Television Infrared Observational Satellite (TIROS I) launched on April 1, 1960 (NASA TIROS, 2010); Project Apollo, unveiled by the U.S. on May 25, 1961 (NASA Apollo, 2007); Nimbus 1, launched on August 28, 1964 (NASA Nimbus, 2009); and Landsat, launched on July 23, 1972 (NASA Landsat, 2010). Some current examples of measuring soil moisture from airborne sensors include experiments using the Electronically Scanned Thinned Array Radiometer (ESTAR), the Scanning Low Frequency Microwave Radiometer (SLFMR), and an active airborne Synthetic Aperture Radar AirSAR (Demšar, 2004).

Airborne and space-based remote sensing systems are used to observe and measure changes in soil and hydrological features across otherwise unobtainable landscapes. Measuring the solar energy reflected off the land surface can be useful in hydrology for snow mapping, vegetation/land cover, and water quality studies (Schmugge et al., 2002). However, anomalies may arise obscuring or negatively influencing the soil moisture values measured. Essentially, the reflected solar radiation is used for mapping the snow, vegetation, land cover and water quality, the thermal emission in the infrared spectrum for measuring surface temperature and the emission in the microwave range for soil moisture and snow studies (Demšar, 2004).

Climate/hydrology models are unique compared with the other two methods because modeling generally requires a priori knowledge of soil moisture conditions.
across the area of study before it can be utilized. Gruhier et al. (2009) point out that a quantitative soil moisture assessment is crucial for land surface modeling and understanding as well as for numerical weather prediction purposes. Therefore, modeling cannot be performed as a stand-alone method and must be used in combination with other collection techniques in order to yield any usable output data.

Similar to remote sensing, modeling is able to make estimates across an entire region without spatial gaps and estimate soil moisture conditions in the past, present and future. The disadvantage to modeling is that model estimates are only as good as the information used to initialize the model, and the physics considered within the model. The results of research by Puma et al. (2005) demonstrated that a simulation model, combined with known climate parameters and soil moisture from remote sensing measurements, allows accurate predictions of the mean of root-zone soil moisture for any averaging depth.

Gruhier et al. (2009) have found that due to high temporal and spatial variability of soil moisture, it is difficult to provide accurate quantitative information on soil moisture at regional and global scales. This is due in part to the lower resolution associated with a limited network of stations as well as microwave remote sensing (the primary method for remotely sensing soil moisture) techniques. It is important to remember that remote sensing is most useful when combined with \textit{in situ} measurements or modeling. When the satellite resolution is not high enough to garner the required accuracy, \textit{in situ} ground measurements can offer precision ground truth
information which can be assimilated in conjunction with the remotely sensed data to increase overall data quality.

The effect of soil moisture memory on climate can be significant. Seneviratne et al. (2006) suggest that soil moisture memory is one of the major “slow” drivers of the climate system, and that a detailed understanding of the processes controlling soil moisture memory is therefore necessary for accessing the predictability associated with soil moisture on sub-seasonal to seasonal time scales, and for characterizing important mechanisms impacting land-atmosphere interactions on these scales. DeLiberty and Legates (2008) also agree, stating that soil moisture memory can have profound implications, from a climatic perspective, for seasonal and long-term weather predictions. DeLiberty and Legates (2008) point out that periods of unusually heavy rainfall can produce a positive anomaly in soil moisture taking anywhere from weeks to months to dissipate through processes such as evapotranspiration.

Furthermore, similar timescales may apply to the dissipation of negative soil moisture anomalies in which the soil can “remember” the dry conditions (DeLiberty and Legates, 2008). This is why understanding the spatial and temporal nature of soil moisture is vital in determining the influence land surface processes have on climate (DeLiberty and Legates, 2008). In cases dealing with soil moisture memory, soil moisture modeling can be of utmost use in the identification of the spatial and temporal nature of the memory. Again, only well-designed and well-configured models which include the overall heterogeneity of the landscape are capable of performing these actions to any useful degree.
In their research, DeLiberty and Legates (2008) found that of the three land surface parameters they investigated (soil texture, soil depth, and vegetation type), only the depth of the soil profile had a significant impact on the persistence of soil moisture, with the memory of soil moisture increasing with increasing soil depth. Unexpectedly, vegetation type had a muted impact on the persistence of soil moisture (DeLiberty and Legates, 2008).

2.1.3 Variability: Large vs. Small Scale

The variability associated with soil moisture is great at any scale. Anderson and Croft (2009) have suggested that soil surface roughness is able to influence such characteristics as soil thermal properties, infiltration rate, surface runoff and susceptibility of soil to erosion. At the microscale (< 10 cm) soil surface roughness is defined by the structure (shape, size and stability) of aggregates which are dependent upon the texture (sand / silt / clay content) and organic matter content of the soil (Anderson and Croft, 2009). Soil texture is also important because it dictates the size of the pore spaces between the particles of soil, which directly influences the amount of moisture that a soil parcel can hold. Additionally, Anderson and Croft (2009) have found that spatial and temporal variations in soil surface roughness can result from natural or anthropogenic phenomena, including tillage, erosion, raindrop impact and physical crusting. Soil surface roughness can be detected at any scale, but is most prevalent at the smaller scales.
Soil moisture affects not only the vertical fluxes of energy and moisture, but also the horizontal fluxes of moisture, namely, runoff (Robock et al., 2000). DeLiberty and Legates (2008) have suggested that soil texture and depth are the characteristics which determine field capacity. Robock et al. (2000) state that the complex topography of natural landscapes, with spatially variable vegetation and soil types, and gravitational drainage and infiltration of water after heavy rains, is responsible for very small-scale spatial (tens of meters) and temporal (up to a few days) variability in the soil moisture field.

At large scales, where remote sensing is the primary method of soil moisture observation and data collection, weather and vegetation are the two most dominant factors in soil moisture variability. Small scale obstructions tend to be the result of variations within the soil itself, with topography, weather and vegetation having less impact. DeLiberty and Legates (2008) have found that at local scales, persistence appeared to be controlled more by the soil depth in comparison to soil texture and vegetation type. Several authors have also found that the accuracy of retrieved surface soil moisture values are highly scale-dependent, with poor agreement between estimated surface soil moisture and physical measurements at the field scale (Anderson and Croft, 2009). Anderson and Croft (2009) have shown, however, that results improved if field-scale results are averaged to a catchment-scale, reducing the impact of the poorly characterized, scale-dependent, smaller surface roughness elements over broader spatial extents.
2.2 Observational Measurements

2.2.1 Data Collection Techniques

The most accurate and reliable means of evaluating soil moisture is by taking *in situ* measurements. These measurements provide a point of comparison for remote sensing and computer modeling soil moisture measurements. Demšar (2004) has indicated three of the most commonly used methods for detecting and measuring soil moisture content are the thermogravimetric (gravimetric) method, neutron thermalization, and a group of methods based on the dielectric properties of soil.

The thermogravimetric method estimates the water content of the soil by taking into account the mass of the moisture in the soil. This is accomplished by weighing a sample of soil, drying it in an oven at 105°C for 24 hours, and subtracting the final dry weight from the initial weight. The mass of moisture is multiplied by 100 and then divided by the mass of the dry soil to give an estimated percentage of moisture by weight that existed within the soil. It is important to clear out as much organic material (i.e., sticks, plant material, insects, etc.) as possible prior to attempting any measurements.

The neutron method relies on hydrogen’s ability to slow down fast neutrons more efficiently than other substances, creating an opportunity for a sensor to detect the amount of moisture contained within the soil based on the degree of deceleration caused by the hydrogen molecules contained in the water. Demšar (2004) points out that most of the hydrogen present in any soil are contained in water molecules and
therefore the number of backscattered slow neutrons emitted from a radioactive source (i.e., active microwave sensor, ground monitoring station, etc.) directly corresponds to water content in the soil.

The dielectric method is based on the large differences inherent in the conductance properties of electricity through water (dielectric constant = 80) and that of a typical dry soil sample (dielectric constant < 5). In a mixture of water and dry soil, the resulting dielectric constant presents itself somewhere between these two extremes, thus offering a mechanism for detecting the water content in the soil (Demšar, 2004). Microwave satellite sensors can indirectly measure the dielectric constant of the soil due to its direct effect on microwave backscatter. Over the past 10 years, the spatial estimation of surface soil moisture has benefited from numerous advances in microwave remote sensing technologies (both active and passive) (Anderson and Croft, 2009).

The gravimetric and the dielectric methods are the two most reliable methods for estimating surface soil moisture (Demšar, 2004). The gravimetric method must be performed in a lab with field samples collected by hand, a task which can be quite demanding and time consuming. The dielectric method is more precisely and accurately accomplished using a network of ground stations, as opposed to remote sensing satellites (in situ vs. remote sensing), but this is also extremely costly. Therefore remote sensing is the prime method for observing and measuring soil moisture conditions when either in situ method proves practically or financially untenable. A combination of methods is preferred for best results.
While *in situ* methods are the most ideal for soil moisture data collection, remote sensing is a more viable and flexible alternative that is able to observe soil conditions at regional, continental, and even global scales. Active sensors can provide high-resolution data (on the order of tens of meters), making them a good choice for observing surface roughness, topography, and vegetation (Demšar, 2004). By comparison, Demšar (2004) also points out that passive sensors provide lower resolution data (on the order of tens of kilometers), which is more appropriate for meteorological and climate models on a global scale.

Various active and passive microwave sensors have been in use for more than a decade measuring land surface reflections and emissions. The following remote sensing microwave radiometers are in use today, as provided by Gruhier et al. (2009):

- **Advanced Microwave Scanning Radiometer (AMSR-E) on Earth Observing System (EOS) ([http://aqua.nasa.gov](http://aqua.nasa.gov))**
  - A passive microwave sensor located on the AQUA satellite.
  - Brightness temperatures measured at five frequencies from 6.9 to 89 GHz.
    - Soil moisture is measured via C-band (6.9 GHz) and X-band (10.7 GHz).
- **TRMM Microwave Imager (TMI) located on the Tropical Rainfall Measuring Mission (TRMM) ([http://trmm.gsfc.nasa.gov/overview_dir/tmi.html](http://trmm.gsfc.nasa.gov/overview_dir/tmi.html))**
  - A passive microwave sensor located on the TRMM satellite.
  - Microwave emissions are measured at five frequencies from 10.7 to 85.5 GHz.
Wind Scatterometer located on European Remote Sensing (ERS) satellites (http://earth.esa.int/ers/ws/)

- An active microwave sensor.
- Microwave measurements are made using C-band (5.3 GHz) via satellite ERS-1 and ERS-2.

Advanced Scatterometer (ASCAT) located on the Meteorological Operational satellite (METOP) (http://www.esa.int/esaLP/SEMBWEG23IE_LPmetop_0.html)

- An active microwave sensor with two sets of three antennae allowing for better resolution and near-global coverage in just 5 days, compared with the active microwave sensors of ERS-1 and ERS-2.
- Since 2008 real-time soil moisture measurements have been made using C-band (5.255 GHz).
- ERS/SCAT and METOP/ASCAT have provided the longest consistent and continuous global scale soil moisture data set since 1992 (Gruhier et al., 2009).

Soil Moisture and Ocean Salinity (SMOS) satellite of the European Space Agency (ESA) launched in 2009 (http://www.esa.int/esaLP/ESAMBA2VMOC_LPsmos_0.html)

- A passive microwave sensor.
- Measures soil moisture using L-band (1.4 GHz).
- SMOS carries the first-ever, polar-orbiting, space-borne, 2-D interferometric radiometer.
2.2.2 In Situ Datasets

Dr. Alan Robock of the Department of Environmental Science at Rutgers University has played an important role in the initiation of the Global Soil Moisture Data Bank (http://climate.envsci.rutgers.edu/soil_moisture/). The website includes a list of the current soil moisture data sets from around the world, including the U.S., Australia, Brazil, Sweden, Italy, and Russia/Ukraine. The data contained in the Global Soil Moisture Data Bank have been used in both passive and active microwave studies that have and continue to demonstrate the potential of microwave remote sensing of soil moisture in regions without snow or tall vegetation (Robock et al., 2000). Additional remotely sensed data is available on a number of web sites.

2.2.3 In Situ Limitations

A major disadvantage of in situ soil measurements is the resultant deformation of the soil surface profile (Anderson and Croft, 2009), not to mention the reduced resolution of soil moisture monitoring networks. Remote sensing can fix both of these
issues, but because the data is collected without making physical contact with the soil it must be validated for accuracy which underscore the need for \textit{in situ} measurements. This also applies to computer modeling, but with a proper soil moisture monitoring network in place these limitations can easily be overcome.

A method of interpolation known as “time-stability” has been used to characterize large fields with a minimized measurement effort (Schneider et al., 2008). The “time-stability” technique utilizes a small handful of carefully selected points from a larger sample volume to represent the average field soil moisture across the entire area. The benefits of data interpolation are many, including enhanced efficiency and practicality over traditional methods (i.e., measuring soil moisture by hand, constructing a network of soil moisture monitoring stations, etc.). When properly implemented, this technique has been shown to reduce the measurement effort in long-term studies (Schneider et al., 2008).

2.3 Remote Sensing Methods

2.3.1 Overview

In recent years, remote sensing has become an increasingly popular tool for the investigation of many current hydrological issues. The ability, through remote sensing, to observe large areas on the globe in a single pass has brought about significant improvements to hydrological research in the way of enhanced productivity and time. Furthermore, remote sensing applications typically allow for frequent and
repeated passes of study areas, offering greater dependability and freer access to regions of hydrological interest.

Remote sensing methods vary depending on the specific land surface data observed. If soil characteristics are known, thermal infrared remote sensing may be able to indirectly estimate soil moisture data. Passive and active microwave remote sensing are more successful methods for estimating soil moisture versus thermal infrared. Additionally, active / passive microwave remote sensing do not require any prior knowledge of soil characteristics in order to provide useful data. Radar remote sensing has also been used to estimate soil moisture levels, but has received mixed reviews due to the amount of signal correction required to make up for the extremely sensitive nature of the high frequency backscatter emanating from uneven soil surfaces (Santello Jr. et al., 2007). Coupling computer modeling with remote sensing creates the most promising and successful combination in which to accurately estimate soil moisture, whilst providing the greatest continual spatial and temporal consistency over time compared with remote sensing methods by themselves (Lacava et al., 2006).

Using remote sensing to measure soil moisture not only offers a record of spatial variability (despite making observations only at the ground surface) but of temporal variability as well, provided that observations are made at regular and frequent intervals on a recurring basis. Furthermore, the land surface can primarily be monitored in a number of ways using various remote sensing techniques. These techniques rely on observing land emissions and reflections using different
wavelengths from the electromagnetic spectrum, including visible (380-750 nm), thermal infrared (700 nm – 300 µm), and microwave (1 mm – 1 m).

2.3.2 Visible / Infrared

Utilizing light reflected in the visible (red) and near-infrared (NIR) spectrum is most ideal for observations of healthy green vegetation. The chlorophyll in plants tends to absorb incoming solar radiation in the photosynthetically active radiation (PAR) region (400-700 nm) of the electromagnetic spectrum. At the same time, they strongly reflect and scatter solar radiation in the near-infrared spectrum range (700-1100 nm). Live green vegetation generally appears dark in the visible (red) and relatively bright in the NIR imagery. The Normalized Difference Vegetation Index, or NDVI, is an effective method for calculating the ratio between reflected and incoming radiation. NDVI is calculated as follows:

\[
\text{NDVI} = \frac{\text{NIR} - \text{Red}}{\text{NIR} + \text{Red}} \quad (2.1)
\]

with NIR and red being the respective reflectance measurements in the near-infrared and red (visible) spectrums. Because the NDVI is a ratio, the values of reflected red, and NIR are based on a range of 0.0 – 1.0, with the NDVI value ranging between -1.0 and +1.0.

Since the atmosphere serves as a medium for the passing of light, only the signal is subjected to background interference. The most prominent forms of signal contamination come from cloud screening, atmospheric absorption, and other
miscellaneous scattering effects (from precipitation, dust, smog, water vapor, etc.). Indices based on visible and near-infrared signatures (such as NDVI) are especially sensitive to these distortive effects. As a result, NASA’s Moderate Resolution Imaging Spectroradiometer (MODIS) Science Team has improved upon the NDVI method by taking into account some of these atmospheric obstructions and creating the Enhanced Vegetation Index, or EVI (Weier and Herring, 1999).

Some of the benefits of EVI include, according to Weier and Herring (1999), corrections for distortions in light caused by air particulates and ground cover below the vegetation. Like the NDVI, however, the EVI is still subject to interference from clouds, aerosols, and glare from the sun. As for the ground cover itself, obstacles persist in the form of very dense or layered vegetation causing a weakening of the signal (transmission of direct, scattered, or reflected light from one leaf to another, then back to the satellite sensor). Additionally, Wigneron et al. (2003) have found that the penetration depth within the vegetation layer strongly decreases as frequency increases.

Thermal infrared (TIR) remote sensing is another option useful in observing land surface temperature. Both air and land surface temperature influence processes such as evaporation and transpiration and therefore, by extension, a soil’s ability to retain and transfer moisture, as well as the overall efficiency of each. Schmugge et al. (2002) points out that land surface temperature is essentially a result of the equilibrium balance achieved between the atmosphere, surface and subsurface soil, as well as the efficiency of the surface to transmit radiant energy into the atmosphere (surface
emissivity). Furthermore, the surface emissivity of soil is dependent on the overall composition, surface roughness, and physical parameters of the soil surface (Schmugge et al., 2002) including surface temperature, moisture content, vegetative growth, and topographical layout.

TIR remote sensing involves a sensor, either airborne or satellite-based, which measures the surface temperature (or brightness temperature) of the soil. Surface emissions must be estimated along with land surface temperatures in order to get an accurate observation. Atmospheric corrections must be made to help isolate the surface thermal emissions signals from radiances due to emission (in the atmosphere from suspended airborne particles, gases, clouds, and moisture, as well as various ground-based reflections). An effective surface temperature derived from all major sources of temperature variance (including soil depth and vegetation temperature) is required for an accurate representation of soil surface temperature (Wigneron et al., 2003). Once these corrections have been made, the signal can be considered as an estimate of the land surface temperature (Schmugge et al., 2002). A combination of TIR and EVI will be used to help determine a few of the physical parameters responsible for the soil moisture estimates observed in this study.

2.3.3 Microwave Overview

Microwave remote sensing is the preferred method for observing soil moisture over large expanses of land. Because this technique measures the dielectric constant of the soil, a more direct estimation of the amount of moisture present may be
obtained (up to a depth of a few millimeters) compared to TIR and EVI. The land surface may be observed using either active or passive microwave radiometry, as well as high- and low-frequency microwave bands to measure soil moisture in topographically diverse regions (i.e., changing vegetation cover, soil properties, and surface roughness), resulting in various soil penetration depths. Each of these methods are able to estimate soil moisture concentrations based on changes in the dielectric constant across the soil surface.

Large contrasting differences in the dielectric constant between water and dry minerals (soil) can best be sensed by using microwave radiometers in the L-band frequency (~1.4 GHz with a 21 cm wavelength). This allows for penetration of up to 0-5 cm into the top soil (approximately one fourth of the wavelength at this frequency). Additionally, Demšar (2004) has suggested that soil moisture data from the root zone (up to a depth of 30 cm) may be inferred based on surface soil moisture findings from various sources (i.e., ground data, computer modeling, remote sensing. The expected range of variation from soil emissions is from 0.95 F/m (farads per meter) for dry soils to 0.6 F/m for smooth wet soils (Schmugge et al., 2002). Furthermore, the greater the moisture quantity of the soil (and vegetation) the more pronounced are the emissions.

Gruhier et al. (2009) have found that the most efficient approach to characterizing soil moisture remotely, with minimum atmospheric interference, is with low-frequency microwave radiometry. Lower frequency / longer wavelength measurements tend to yield increased sampling depths along with greater noise
reduction from contributing sources such as vegetation and soil surface roughness (Schmugge et al., 2002). Vegetation interferes with the sensitivity of the soil moisture retrieval algorithm by masking the soil signal and adding microwave emissions of its own. The attenuation affect caused by vegetation increases as microwave frequencies rise, which is an important reason for using lower frequencies. In addition to microwave frequency and quantity of soil moisture, the strength of microwave emissions also depends on topography (i.e., soil surface, soil subsurface, elevation changes, etc.) according to Hornbuckle et al. (2006).

Soil moisture data estimations also rely heavily on the sensors themselves, including microwave sensor parameters such as incidence angle, frequency, and polarization. The incidence angle determines the extent to which soil surface roughness affects the soil emission signal. Anderson and Croft (2009) have found that high incidence angles (>45°) are able to discriminate well between smooth and rough areas, where the backscattered signal has an exponential dependence on soil surface roughness. Additionally, Anderson and Croft (2009) have mentioned the use of a multi-polarization, multi-angle, multi-frequency, or multi-temporal configuration to reduce conflicting signals provided by other soil surface parameters, as well for the inference of certain soil surface roughness characteristics.

According to Schmugge et al. (2002), microwave remote sensing offers four unique advantages over other spectral regions:
1) The atmosphere is effectively transparent providing all weather coverage in the decimeter ($1 \times 10^{-2}$ m) range of wavelengths (30 GHz frequency).

2) Vegetation is semitransparent allowing the observation of underlying surfaces.

3) The microwave measurement is strongly dependent on the dielectric properties of the target which for soil is a function of the amount of water present.

4) Measurement is independent of solar illumination which allows day or night observation.

2.3.4 Microwave (Active)

Active microwave radiometry involves radiating the ground surface and measuring the return signal. Soil surface roughness and vegetation can have adverse effects on the microwave emissions signal, making it more difficult to extract soil moisture information. The emissions signal must be corrected for both vegetation and soil surface roughness effects prior to the inversion from microwave brightness temperature to estimates of soil moisture (Schmugge et al., 2002). The benefit of active microwave remote sensing is attaining higher spatial resolutions compared with passive methods. This, according to Schmugge et al. (2002), can greatly contribute to our knowledge of spatial variation of soil moisture across the land surface.
Anderson and Croft (2009) have found that active microwave sensors are the only satellite sensors that are able to currently achieve the spatial resolution and coverage required for small-scale soil observations. The most common imaging active microwave system is Synthetic Aperture RADAR (SAR). In this configuration, microwave pulses are processed together to effectively simulate a very long aperture grid capable of very high spatial resolution of up to 1 meter (Anderson and Croft, 2009).

SAR observations are based on the dielectric method described earlier. With the increasing number of active SAR systems available today, it is now possible to map surface soil moisture at high temporal frequencies. New sensors such as TerraSAR-X (http://www.infoterra.de/terrasar-x.html) and Cosmo-SkyMed (http://www.telespazio.com/cosmo.html) also allow for increased temporal frequencies due to their shortened revisit intervals (Anderson and Croft, 2009). TerraSAR-X, for example, is able to measure within-field surface soil moisture variability at a spatial resolution of 1 meter while operating at X-band (31 mm wavelength) (Anderson and Croft, 2009). Traditionally, C-band (3.75 – 7.5 cm wavelength) RADAR has also been used to characterize SSM.

2.3.5 Microwave (Passive)

Passive microwave remote sensing is characterized by the measuring of microwave emissions from the ground surface occurring naturally without contribution from an artificial satellite source. The intensity of this emission is
expressed as a brightness temperature (similar to TIR), which is nonetheless vulnerable to interferences from the atmosphere, sky radiation, and the land surface (Schmugge et al., 2002). Sensors must be very sensitive in order to identify fluctuations in the dielectric constant without using an active microwave signal.

A major benefit of passive microwave remote sensing is the broad spatial coverage and high temporal resolution available from modern systems. However, spatial resolutions are poor compared to active microwave systems, offering up resolutions of 10-30 km (Anderson and Croft, 2009). Additionally, Anderson and Croft (2009) have found that the relationship between brightness temperature and surface soil moisture varies with differences in soil surface roughness, as well as vegetation biomass and soil texture. Despite the fact that microwave radiation is capable of penetrating vegetation canopies, a portion of the microwave signal emitted from the soil surface is ultimately altered by the vegetation.

A passive microwave satellite currently in use is the Advanced Microwave Scanning Radiometer for EOS (AMSR-E) (http://sharaku.eorc.jaxa.jp/AMSR/ov_amsre/index.html), which was launched in 2002, with algorithm updates (as of February 8, 2006) to reduce the range of measurement error from 5-10 km to 1-3 km. AMSR was located on the JAXA’s Advanced Earth Observing Satellite II (ADEOS-II) (http://www.jaxa.jp/projects/sat/adeos2/index_e.html) for 237 days in 2003 (from January through October). NASA’s AQUA satellite (http://aqua.nasa.gov/) currently carries another AMSR-E passive microwave radiometer as well.
2.3.6 Remote Sensing Limitations

General remote sensing limitations apply to microwave radiometry in addition to a few specific restrictions. Microwave remote sensing cannot replace ground-based methods because microwaves are only capable of penetrating the top few centimeters of the soil surface. As a result, in situ observation and collection techniques must be used in order to more thoroughly detect subsurface soil moisture conditions to the entire root zone. Furthermore, remote sensing has an inherent level of uncertainty contingent on variables such as satellite sensor type (i.e., active vs. passive), vegetation cover, soil surface roughness, topography, and soil type. Soil surface roughness in particular, according to Schmugge et al. (2002), will typically reduce the range of microwave response by roughly 10-20%, with up to a 50% reduction in sensitivity occurring in extreme situations. This reinforces the importance of including ground and/or computer modeling methods concurrently with microwave radiometry.

On the positive side, lower relative microwave frequencies are able to penetrate deeper into the soil surface, helping to bypass some of the surface roughness issues that can interfere with this type of remote sensing. Wigneron et al. (2003) has shown the results from numerous studies have demonstrated the immense potential of low-frequency observations at L-band (~1.4 GHz) for soil moisture retrievals. Conversely, high-frequency observations (above 6.6 GHz) have a reduced sensitivity to soil moisture, particularly when the water content exceeds ~1.5 kg/m², which
corresponds roughly to a leaf area index (LAI) of about two for crops (Wigneron et al., 2003).

The use of various microwave frequencies offers distinct perspectives on soil moisture attributes, and can be particularly insightful when used simultaneously. For instance, Wigneron et al. (2003) has found that the C-band (4-8 GHz) is most useful making soil moisture observations in semiarid regions with low levels of vegetation biomass. While C-band might provide an optimal soil surface penetration depth for a particular study, it can be limited by environmental factors such as excessive vegetation and the influence of atmosphere-related signal scattering. Alternatively, Wigneron, et al. (2003) discovered that satisfactory retrieval results can be obtained from dual-polarization C-band observations over agricultural test sites.

A satellite using single-frequency, single incidence angle, single-pass SAR data is not sufficient to accurately retrieve surface soil moisture information without a priori soil surface roughness information (Anderson and Croft, 2009). As Anderson and Croft (2009) have pointed out, some of the more recent SAR sensors have attempted to address this problem by operating at several frequencies, and multiple polarization and incidence angles, allowing soil moisture retrievals to be chosen as a function of selected parameters. This, therefore, minimizes the effects of other soil surface characteristics and permits a higher degree of precision, as well as accuracy, when analyzing the soil moisture conditions of any terrain. The most accurate method for surface soil moisture evaluation comes from a combination of high and low incidence angle measurements.
Schmugge et al. (2002) have suggested implementing a framework for combining multi-frequency remote sensing information (visible to microwave) for a more reliable estimation of vegetation and soil properties. However, since penetration depth depends on the frequency (approximately 3 cm at L-band and 1 cm at C-band) (Wigneron et al., 2003), using more than one frequency at a time will require multiple correction algorithms.
Chapter 3

SOIL MOISTURE DATA AND METHODOLOGY

3.1 Introduction

There are numerous methods available for the observation and measurement of soil moisture. The two primary methods utilized in this project are a network of land surface monitoring stations and remote sensing radiometry. In addition to quantifying soil moisture content, remote sensing also provides a direct quantitative measure of land surface temperature and vegetation health across the state of Delaware, including areas where DEOS stations do not currently reside. Vegetation is sensitive to changes in moisture conditions making it an important indicator of past and present soil moisture conditions. Land surface temperatures represent the energy state of the top soil surface, suggesting the present evaporative state of the land surface.

As previously suggested, remote sensing is best used in combination with ground truth observations to allow for the greatest possible land coverage extent without sacrificing accuracy of data. The nearly limitless field of view satellites provide (in multiple passes) makes them an ideal venue for collecting data continuously across time and space. This makes remote sensing an invaluable tool when it comes to studying land surface soil moisture.

Soil moisture is measured by DEOS station sensors as volumetric water content in 5-minute intervals at each site. The major drawback to this system is the limited number of stations from which observations are made. For this study, nine
stations were utilized throughout the state of Delaware (including one station in Maryland, on the border of northern Delaware and Maryland) to observe soil moisture variability in the northern, central, and southern areas of the state across a span of 3.5 years (January 2005 through June 2008). The benefits of these DEOS ground stations along with wide-range remotely sensed land observations combine to create an accurate database of soil moisture measurements across the state of Delaware enabling an investigation of soil moisture variability.

Remote sensing data were obtained through NASA’s Moderate Resolution Imaging Spectroradiometer (MODIS) website (http://modis.gsfc.nasa.gov/), providing the necessary information to view and map vegetation health and land surface temperature across the state of Delaware. Passive microwave data were obtained through NASA’s Advanced Microwave Scanning Radiometer – Earth Observing System (AMSR-E) via the National Snow and Ice Data Center’s website (http://nsidc.org/data/amsre/index.html), which provides direct soil moisture measurements of up to 1 cm in depth on average. The soil moisture observations provided by the AMSR-E satellite sensor offer the only quantitative measure of remotely sensed soil moisture conditions in this study. These observations will be used for direct comparison against those observations made by the nine DEOS station sensors.

MODIS is a key instrument aboard the Terra (EOS AM) and Aqua (EOS PM) satellites. The Terra satellite orbits around the Earth passing from north to south across the equator in the morning, while Aqua passes south to north over the equator
in the afternoon. Terra and Aqua satellites view the entire Earth’s surface every 1 to 2 days, acquiring data in 36 spectral bands, or groups of wavelengths. Bands 1-19 and 26 observe in the visible and near infrared range, while the remaining bands observe in the thermal range from 3 to 15 mm. Daylight reflection and day/night emission spectral imaging are provided for any point on the Earth every 1-2 days. The goal of these imaging systems is to improve our understanding of global dynamics and processes occurring on the land, in the oceans, and in the lower atmosphere. Moreover, the MODIS data is playing a vital role validating global climate data models.

MODIS’ new Enhanced Vegetation Index (EVI) product provides an indication of vegetation health that can be used in comparison to DEOS soil moisture measurements over time. The EVI is computed from atmospherically corrected bi-directional surface reflectances that have been masked for water, clouds, heavy aerosols, and cloud shadows. The index also uses the blue band to remove residual atmospheric contamination caused by smoke and sub-pixel thin clouds, along with minimizing canopy background variations and maintaining sensitivity over dense vegetation conditions. Lacking a 250m resolution blue band, the EVI algorithm uses the 500m blue band to correct for residual atmospheric effects, with negligible spatial artifacts. The global MOD13Q1 product was selected for this study, providing data every 16 days at 250-meter spatial resolution as a gridded level-3 product in the geographic latitude/longitude coordinate system, sinusoidal map projection. This dataset was specifically chosen as it provides the most frequent data updates at the
highest available spatial resolutions compared to the other MODIS vegetation index datasets.

There is a direct relationship between vegetation health and soil moisture. A dry soil layer cannot support healthy, green vegetation. The more moisture there is within the soil, the greater the potential for supporting healthy, green vegetation. Vegetation health also provides an indication of soil moisture conditions of the past and present throughout the area. Another important indicator of soil moisture is land surface temperature. A soil’s resistance to temperature change is directly related to the amount of moisture it holds. The greater the amount of moisture present, the more the soil is able to resist changes in temperature. Additionally, land surface temperatures may offer an indication of the soil evaporation and plant transpiration occurring at the surface, with higher temperatures indicating a decrease in the evapotranspiration at the land surface and lower temperatures indicating an increase in evapotranspiration at the land surface.

For this study, the level-3 MODIS global Land Surface Temperature (LST) and Emissivity 8-day data (MOD11A2) were selected. The LST data is composed from the daily 1-kilometer LST product (MOD11A1) and stored on a geographic latitude/longitude coordinate system, sinusoidal map projection as the average values of clear-sky LSTs during an 8-day period. This data is comprised of daytime and nighttime LSTs, along with quality assessment, observation times, view angles, percentage of clear sky days and nights, and emissivities estimated in Bands 31 and 32 from land cover types with a land temperature error margin of 1°K.
Soil moisture observations were made using the Advanced Microwave Scanning Radiometer - Earth Observing System (AMSR-E) instrument on the NASA Earth Observing System (EOS) Aqua satellite. Soil moisture values are estimated based on the Polarization Ratio (PR) of horizontal and vertical brightness temperatures collected by the sensor. A soil moisture algorithm is used to reduce or remove surface temperature effects at the specified frequency. The gridded Level-3 land surface product (AE_Land3) was chosen for this research as it provides soil moisture observations at a relatively low 6.9 GHz frequency (0.04m wavelength). Input brightness temperature data, corresponding to a 56 km mean spatial resolution, are resampled to a global cylindrical 25 km Equal-Area Scalable Earth Grid (EASE-Grid) cell spacing using a geographic latitude/longitude coordinate system, sinusoidal map projection. Data are stored in HDF-EOS format, and are available from 19 June 2002 to the present.

The Advanced Microwave Scanning Radiometer - Earth Observing System (AMSR-E) is a twelve-channel, six-frequency, passive-microwave radiometer system. It measures horizontally and vertically polarized brightness temperatures at 6.9 GHz (0.043m wavelength), 10.7 GHz (0.028m wavelength), 18.7 GHz (0.016m wavelength), 23.8 GHz (0.012m wavelength), 36.5 GHz (0.008m wavelength), and 89.0 GHz (0.003m wavelength). At an altitude of 705 km, it measures the upwelling scene brightness temperatures over a field of view ± 61° about the sub-satellite track, resulting in a swath width of 1445 km with scene measurements spaced at equal intervals of 10 km (5 km for the 89 GHz channels) along the scan.
3.2 Remote Sensing Derived Products

3.2.1 Vegetation

The health of ground-based vegetation can be determined by quantifying the amount of reflected red and near-infrared (NIR) energy. Reflected red energy decreases with plant development due to chlorophyll absorption within actively photosynthetic leaves, while reflected NIR energy will increase with plant development through scattering processes in healthy leaves. A simple ratio between corresponding red and NIR values can be represented by the following equation:

$$SR = \frac{\text{NIR}}{\text{red}}$$  \hspace{1cm} (3.1)

where red and NIR represent the spectral reflectance measurements acquired in the red and near-infrared regions, respectively. The normalized difference vegetation index (NDVI) is a comparison of red and near-infrared reflectance measurements used to minimize atmospheric interference. The greater the amount of healthy green vegetation detected by the sensor, the greater the NDVI value. Equation 3.4 shows the formula used to determine NDVI.

$$NDVI = \frac{(\text{NIR} - \text{red})}{(\text{NIR} + \text{red})}$$  \hspace{1cm} (3.2)

Some of the limitations that can result from various external influences include a) calibration and instrument characteristics, b) clouds and cloud shadows, c)
atmospheric effects due to variable aerosols, water vapor, and residual clouds, and d) sun-target-sensor geometric configurations and the resulting interactions of surface and atmospheric anisotropies (properties that differ based on the direction of measurement) on the angular dependent signal. Internal influences inherent to vegetated canopies which restrict the use and/or interpretation of vegetated indices include a) canopy background contamination in which the background reflected signal intimately mixes with the vegetation signal and influences the resulting vegetation index value, b) canopy background signals vary with soils, litter covers, snow, and surface wetness, and c) saturation problems whereby vegetation index values remain invariant to changes in the amount, type, condition of vegetation, normally associated with a saturated chlorophyll signal in densely vegetated canopies (Huete et al., 1999).

The atmosphere degrades the NDVI value by reducing the contrast between the red and NIR reflected signals. The red signal normally increases as a result of scattered, upwelling path radiance contributions from the atmosphere, while the NIR signal tends to decreases as a result of atmospheric attenuation associated with water vapor absorption. The net result is a drop in the NDVI signal and an underestimation of the amount of vegetation at the surface. The degradation in NDVI signal is also dependent on the aerosol content of the atmosphere, with turbid atmospheres resulting in the lowest NDVI signals.

A feedback-based approach known as the enhanced vegetation index (EVI) was developed by Liu and Huete in 1995 to correct for the canopy background and
atmospheric influences. This enhanced, soil and atmosphere resistant vegetation index is:

$$EVI = 2 \frac{(\rho_{nir} - \rho_{red})}{(L + \rho_{nir} + C_1 \rho_{red} + C_2 \rho_{blue})}$$ (3.3)

where $\rho$ is ‘surface’ directional reflectances, $L$ is a canopy background adjustment term, and $C_1$ and $C_2$ weigh the use of the blue channel in aerosol correction of the red channel (Huete et al., 1999).

Using the EVI, a more accurate representation of surface-level vegetation is possible compared to NDVI alone. Low-level vegetation is ideal because it reacts more quickly to fluctuations at the surface than tall dense vegetation. By removing as much interference from the atmosphere and large tree canopies as possible, a much clearer observation is made of surface-level vegetation and therefore a more accurate depiction of soil moisture conditions are possible.

### 3.2.2 Thermal IR

The thermal infrared bands record emissions using bands 31 and 32 (10.780-11.280 µm and 11.770-12.270 µm, respectively) with an instantaneous field of view (IFOV) of approximately 1km at nadir to retrieve surface emissivity and temperature. The MODIS instrument will view cold space and a full-aperture blackbody before and after viewing the Earth scene in order to achieve the calibration accuracy specification.
better than 1% absolute for thermal infrared bands (0.75% for band 20, 0.5% for bands 31 and 32). MODIS is particularly useful for the LST product because of its global coverage, radiometric resolution and dynamic ranges for a variety of land cover types, and high calibration accuracy in multiple thermal infrared bands designed for retrievals of soil surface temperature, LST and atmospheric properties. Specifically, band 26 is used for cirrus detection, thermal infrared bands 20, 22, 23, 29, 31-33 for correcting atmospheric effects and retrieving surface emissivity and temperature. Multiple bands in the mid-infrared range provide an opportunity to make accurate corrections of the solar radiation effects so that the solar radiation can be used as a TIR source for the purpose of retrieving surface emissivity in the mid-infrared range in the day/night MODIS LST method. The success of the LST algorithm depends on 1) accurately accounting for the atmospheric effects; 2) accurately accounting for the surface emissivity effects; and 3) the quality of TIR data including the stability of the spectral response function, signal-to-noise ratio, radiometric resolution, and calibration accuracy (Zhengming, 1999: 6-8).

For a given MODIS pixel, the split-window LST algorithm requires emissivities in bands 31 and 32.

\[
T_s = \left( A_1 + A_2 \frac{1-\varepsilon}{\varepsilon} + A_3 \frac{\Delta \varepsilon}{\varepsilon^2} \right) \frac{T_{31} + T_{32}}{2} + \left( B_1 + B_2 \frac{1-\varepsilon}{\varepsilon} + B_3 \frac{\Delta \varepsilon}{\varepsilon^2} \right) \frac{T_{31} - T_{32}}{2} + C
\] (3.4)
where $\Delta \varepsilon = (\varepsilon_{31} - \varepsilon_{32})$ and $\varepsilon = 0.5(\varepsilon_{31} + \varepsilon_{32})$ are the difference and mean of surface emissivities in MODIS band 31 and 32. $T_{31}$ and $T_{32}$ are the brightness temperatures in these two split-window bands. The coefficients $A_1, A_2, A_3, B_1, B_2, B_3, C$ are given by interpolation on asset of multi-dimensional look-up tables (LUT). The LUT were obtained by linear regression of the MODIS simulation data from radiative transfer calculations over a wide range of surface and atmospheric conditions (Akhoondzadeh, 2010).

### 3.2.3 Microwave

The objective of the Level-2A algorithm is to bring the Level-1A antenna temperatures to a set of common spatial resolutions using a set of weighted coefficients. The algorithm resamples Level-1A antenna temperatures and converts them to Level-2A brightness temperatures.

The resampled antenna temperature ($T_{ac}$) is defined as a weighted sum of observed antenna temperatures ($T_{ai}$):

$$T_{ac} = \sum_{i=1}^{N} a_i T_{ai}$$  \hspace{1cm} (3.5)

where $a_i$ is the weighting coefficients. Antenna temperature observations are corrected for cold-space spillover and cross-polarization effects to obtain brightness temperatures averaged over the normalized crossover-polarization antenna pattern.

The observed brightness temperatures ($T_{bi}$) are expressed as:
\[ T_{bl} = \int T_b(\rho) G_i(\rho) dA \] (3.6)

where \( T_b(\rho) \) is the brightness temperature at location \( \rho \), and \( G_i(\rho) \) is the antenna gain pattern corresponding to the specific observation.

Next, soil moisture is derived from brightness temperatures using a soil moisture algorithm. The soil moisture algorithm uses Polarization Ratios (PR), which are sometimes called normalized polarization differences of the AMSR-E channel brightness temperatures. PR is the difference between the vertical and horizontal brightness temperatures at a given frequency divided by their sum. The algorithm first computes a vegetation/roughness parameter \( g \) using PR 10.7 GHz and PR 18.7 GHz, plus three empirical coefficients. Soil moisture is then computed using departures of PR 10.7 GHz from a baseline value, plus four additional coefficients. The baseline values for PR 10.7 GHz are based on monthly minima at each grid cell over an annual cycle (Njoku and Chan, 2005).

AMSR-E measurements of soil moisture are directly sensitive only to the top 1 cm of soil averaged over approximately 60 km spatial extent. Significant uncertainty may therefore arise when these measurements are compared against point-derived in-situ data, due to differences in sampling depth and spatial extent between satellite and in-situ observations. The AMSR-E measurements of soil moisture are most accurate in areas of low vegetation. Attenuation from vegetation increases the retrieval error in
soil moisture with retrievals not performed in dense vegetation (Njoku et al., 2003). Areas of dense vegetation are represented by pixels displaying no data.

The retrieval algorithm does not explicitly model for the effects of topography, snow cover, clouds, and precipitation. Other potential error sources include anomalous inputs from bad radiometric data and low-level processing errors. The processing algorithm includes checks to identify these and other anomalies and assign appropriate flags (Njoku, 1999).

3.2.4 DEOS

Each DEOS ground station is fitted with a Campbell Scientific CS616-L Water Content Reflectometer, designed to measure the volumetric water content (vwc) of soils. Volumetric water content is a measure of the amount of water contained within a sample of soil measured by volume: where

\[
  vwc = \frac{\text{volume of moisture}}{\text{volume of soil}} \quad (3.7)
\]

Each sensor consists of two rods, each 300 mm long (3.2 mm diameter, 32 mm spacing) capable of receiving and transmitting electromagnetic signals. The water content information is derived from the probe sensitivity to the dielectric constant of the soil surrounding the probe rods. With the probes in the ground soil moisture measurements can be made from 0 to 30 cm in depth.
The fundamental principle for CS616 operation is that an electromagnetic pulse will propagate along the probe rods at a velocity that is dependent on the dielectric permittivity of the material surrounding the line. As water content increases, the propagation velocity decreases because polarization of water molecules takes time. The travel time of the applied signal along 2 times the rod length is essentially measured. The applied signal travels the length of the probe rods and is reflected from the rod ends traveling back to the probe head. A part of the circuit detects the reflection and triggers the next pulse. The probe output frequency or period is empirically related to water content using a calibration equation.

The accuracy specification for the volumetric water content measurement using the CS616 probes is based on laboratory measurements in a variety of soils and over the dry air to saturated water content range. The Water Content Reflectometer accuracy is $\pm 2.5\% vwc$ using standard calibration with bulk electrical conductivity $\leq 0.5 \text{ deciSiemen meter}^{-1} (\text{dS m}^{-1})$ and bulk density $\leq 1.55 \text{ g cm}^{-3}$ in measurements ranging from $0\% vwc$ to $50\% vwc$. The resolution of this sensor is better than $0.1\% vwc$. Variability between probes is $\pm 0.5\% vwc$ in dry soil and $\pm 1.5\% vwc$ in a typical saturated soil.

Water Content Reflectometer Data needed for CS616 calibration are the output period (microseconds) and an independently determined volumetric water content. From this data, the probe response to changing water content can be described by the following quadratic calibration equation:
\[ \theta_V(\tau) = C_0 + C_1 \times \tau + C_2 \times \tau^2 \]  
(3.8)

with \( \theta_V \) being the volumetric water content \((m^3 \cdot m^{-3})\), \( \tau \), the CS616 period (microseconds), and \( C_n \) the calibration coefficient \((n = 0..2)\).

The linear form is:

\[ \theta_V(\tau) = C_0 + C_1 \times \tau \]  
(3.9)

where \( C_0 \) represents the intercept, and \( C_1 \) the slope. In the typical water content range of about 10% to about 35% \( vwc \), the response can be described with slightly less accuracy by a linear calibration equation.

### 3.3 Ground Station Descriptions and Locations

Nine stations were chosen throughout the state of Delaware, including one in Maryland, in order to effectively capture the spatial variability of soil moisture across the Delmarva Peninsula. Each station was chosen based on its relative proximity to other stations in the network, attempting to adequately cover the entire state while still being able to observe the variability across local regions. Table 3.1 lists the nine DEOS stations including the geographical coordinates and elevation at each site:
Table 3.1. DEOS stations

<table>
<thead>
<tr>
<th>Location</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Elevation (ft)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Newark, DE</td>
<td>39° 40' N</td>
<td>75° 45' W</td>
<td>106</td>
</tr>
<tr>
<td>Dover, DE</td>
<td>39° 10' N</td>
<td>75° 36' W</td>
<td>51</td>
</tr>
<tr>
<td>Fair Hill, MD</td>
<td>39° 43' N</td>
<td>75° 50' W</td>
<td>150</td>
</tr>
<tr>
<td>Glasgow, DE</td>
<td>39° 36' N</td>
<td>75° 44' W</td>
<td>70</td>
</tr>
<tr>
<td>Harrington, DE</td>
<td>38° 55' N</td>
<td>75° 35' W</td>
<td>55</td>
</tr>
<tr>
<td>Laurel, DE</td>
<td>38° 32' N</td>
<td>75° 36' W</td>
<td>40</td>
</tr>
<tr>
<td>Wilmington, DE</td>
<td>39° 46' N</td>
<td>75° 33' W</td>
<td>270</td>
</tr>
<tr>
<td>Selbyville, DE</td>
<td>38° 28' N</td>
<td>75° 13' W</td>
<td>25</td>
</tr>
<tr>
<td>Viola, DE</td>
<td>39° 3' N</td>
<td>75° 34' W</td>
<td>55</td>
</tr>
</tbody>
</table>
3.4 Delaware Study Area

Figure 3.1. Map showing the State of Delaware and the DEOS soil moisture monitoring sites used in the study.

Delaware is made up of three counties (highlighted in green) including New Castle (top), Kent (middle), and Sussex (bottom), and is located on the east coast of the United States just south of New Jersey and Pennsylvania and just east of Maryland. The red dots on the map in Figure 3.1 highlight the locations of the nine DEOS stations utilized in this research study.
3.5 Methodology

In Situ

Soil samples were retrieved from each of the 9 sites at a depth of 5~10cm (approximately the depth of the DEOS soil sensors) during fall 2008. Rocks, sticks, roots, and insects were removed from each sample by hand before taking any measurements. The samples were weighed using a mechanical scale to find the mass (g) of the soil sample, measured to the nearest tenth of one gram. The mass of the soil sample was derived using the following formula:

\[
m_{\text{soil}} = (m_{\text{initial}} + m_{\text{container}}) - m_{\text{container}}, \tag{3.10}
\]

where \( m_{\text{container}} \) is the mass of the empty soil container, weighed prior to use.

Next, the samples were measured for volume using a 1000mL graduated cylinder. Due to unequal particle sizes and a fairly heterogeneous sample composition it was difficult to measure with consistent precision and accuracy. To ensure the best possible measurement, in addition to removing unnecessary detritus and inorganic substances, the graduated cylinders were covered and shaken to induce a more complete settling of the soils. Measurements were made to the nearest 10mL increment based on the number and extent of gaps observed in the soils. Once the mass and volume of the soil samples were determined, each sample was set to dry in a thermostatically controlled oven at a temperature of 105°C for 24 hours.
After the 24 hours passed each sample was again weighed to find the dry mass of the soil using the same technique described above. The volume of the dry soil was measured following the dry mass measurements using the same technique previously described. These values were used to determine the amount of moisture within each sample by mass and volume using the following formulas:

\[ m_{\text{moisture}} = m_{\text{initial}} - m_{\text{dry}} \]  \hspace{1cm} (3.11)

\[ v_{\text{moisture}} = v_{\text{initial}} - v_{\text{dry}} \]  \hspace{1cm} (3.12)

To determine the amount of water in each sample as a percent of the dry soil mass, the gravimetric water content, \( \theta_g \), was found:

\[ \theta_g = \frac{m_{\text{moisture}}}{m_{\text{dry}}} \times 100 \]  \hspace{1cm} (3.13)

The overall density of each sample was determined, otherwise known as the bulk density, using:

\[ \rho_{\text{bulk}} = \frac{m_{\text{dry}}}{v_{\text{dry}}} \]  \hspace{1cm} (3.14)

The volumetric water content, \( \theta_g \), was found by multiplying the bulk density by the gravimetric water content:
The acquired data were later used to determine the soil characteristics at each location. This information provided a more thorough analysis when comparing the DEOS ground station data with the remotely sensed imagery.

Next, each soil sample was separated into components of sand, silt, and clay, using a La Motte Soil Texture Unit (http://www.lamotte.com/pages/edu/1067.html). By finding the ratio of sand, silt, and clay in each sample a generalized texture was found at the DEOS sensor location approximately 10cm below the ground surface. Texture plays a major role in a soil’s ability to hold water, making it a critical component in identifying the variability between DEOS stations and across the entire Delmarva Peninsula.

The La Motte Soil Texture Unit separates the soils into sand, silt, and clay by volume, using test tubes, a texture dispersing reagent, and tap water. 15mL of soil are placed into a test tube, followed by 1mL of the texture dispersing reagent, and the rest of the tube is filled with tap water up to the 45mL line. After shaking up the sample for two minutes the particles immediately begin to settle. The larger and heavier sand particles settle first, requiring a measurement to be made after 30 seconds of settling. The silt particles settle after 30 minutes, and are measured from the top of the sand to the top of the silt. Clay particles require 24 hours to completely settle and are measured by subtracting the total sediment height (after 24 hours) from the initial
height of 15mL. Once these measurements were recorded for all nine stations, the percentage of sand, silt, and clay were calculated using the following equations.

\[
\%\text{Sand} = \frac{v_{\text{sand}}}{v_{\text{total}}} \times 100
\]

\[
\%\text{Silt} = \frac{v_{\text{silt}}}{v_{\text{total}}} \times 100
\]

\[
\%\text{Clay} = \frac{(v_{\text{initial}} - v_{\text{total}})}{v_{\text{total}}} \times 100
\]

where \(v_{\text{sand}}\), \(v_{\text{silt}}\), and \(v_{\text{clay}}\) represent the individual volumes of sand, silt, and clay per sample, and \(v_{\text{total}}\) represents the total additive volume of the three textural components. Having determined the ratio of sand, silt, and clay particles present in each soil sample, the soil texture was then found using the soil texture triangle in Figure 3.2.
From the percent sand, silt, and clay values it was possible to find the porosity of each soil sample using approximated values of soil particle density (see Table 3.2).

Table 3.2. Particle density (Hausenbuiller, 1981: 90).

<table>
<thead>
<tr>
<th>Textural Class</th>
<th>Particle Density (g/mL)</th>
<th>Average Density (g/mL)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sand</td>
<td>1.55</td>
<td>1.4167</td>
</tr>
<tr>
<td>Sandy Loam</td>
<td>1.40</td>
<td></td>
</tr>
<tr>
<td>Fine Sandy Loam</td>
<td>1.30</td>
<td></td>
</tr>
<tr>
<td>Loam</td>
<td>1.20</td>
<td>1.175</td>
</tr>
<tr>
<td>Silt Loam</td>
<td>1.15</td>
<td></td>
</tr>
<tr>
<td>Clay Loam</td>
<td>1.10</td>
<td>1.05</td>
</tr>
<tr>
<td>Clay Loam</td>
<td>1.05</td>
<td></td>
</tr>
<tr>
<td>Aggregated Clay</td>
<td>1.00</td>
<td></td>
</tr>
</tbody>
</table>
The average particle density was determined to accommodate the vast range of particle densities that could potentially exist within each sample. The sand density average consists of sand, sandy loam, and fine sandy loam; the silt density average is made up of loam and silt loam; and the clay density averaged is derived from the clay loam, clay, and aggregated clay particle densities. Once the averages were determined, the average particle density, \( \rho_{\text{particle}} \), was found for each soil sample using:

\[
\rho_{\text{particle}} = \rho_{\text{sand}} + \rho_{\text{silt}} + \rho_{\text{clay}}, \tag{3.19}
\]

where \( \rho_{\text{sand}}, \rho_{\text{silt}}, \) and \( \rho_{\text{clay}} \) represent the average particle density for sand, silt, and clay, respectively, and can be found using the following equations:

\[
\rho_{\text{sand}} = \frac{\text{(%Sand} \times 1.4167)}{100} \tag{3.20}
\]

\[
\rho_{\text{silt}} = \frac{\text{(%Silt} \times 1.175)}{100} \tag{3.21}
\]

\[
\rho_{\text{clay}} = \frac{\text{(%Clay} \times 1.05)}{100} \tag{3.22}
\]
Using the average particle density in conjunction with the bulk density determines the porosity of each sample. Porosity is a measure of the empty pore spaces between particles, and is important in assessing how much moisture a soil is capable of holding. The porosity, \( \phi \), of each sample is determined by:

\[
\phi = 1 - \frac{(\rho_{bulk})}{(\rho_{particle})}
\]  

(3.23)

where \( \rho_{bulk} \) is the bulk density, and \( \rho_{particle} \) is the average particle density for all particles.

The data shown in Table 3.3 describes the soil characteristics found at each site as well as the general vegetation cover observed. The soil texture was found using the soil texture triangle in Figure 3.2, while the soil type listed in Table 3.3 was determined using the Web Soil Survey (USDA, 2009). Table 3.4 displays the results of the \textit{in situ} soil sample analysis, describing the characteristics most prevalent of each soil. Of special note is the porosity value at the Selbyville location, which appears negative. Since porosity represents the empty space between soil particles, it must exist as a positive value greater than zero. This negative value was likely caused by debris, distorting the mass and volume and skewing the results.
Table 3.3. Soil and vegetation characteristics.

<table>
<thead>
<tr>
<th>Station</th>
<th>Location</th>
<th>Soil Type</th>
<th>Soil Texture</th>
<th>Vegetation Type</th>
</tr>
</thead>
<tbody>
<tr>
<td>Newark</td>
<td>Newark, DE</td>
<td>Delanco Silt Loam</td>
<td>Loam</td>
<td>Grass</td>
</tr>
<tr>
<td>Dover</td>
<td>Dover, DE</td>
<td>Fallsington Sandy Loam</td>
<td>Sandy Loam</td>
<td>Grass</td>
</tr>
<tr>
<td>Fair Hill</td>
<td>Fair Hill, MD</td>
<td>Manor Loam</td>
<td>Sandy Loam</td>
<td>Grass</td>
</tr>
<tr>
<td>Glasgow</td>
<td>Glasgow, DE</td>
<td>Pineyneck Loam</td>
<td>Loam</td>
<td>Grass</td>
</tr>
<tr>
<td>Harrington</td>
<td>Harrington, DE</td>
<td>Marshyhope Sandy Loam</td>
<td>Sandy Loam</td>
<td>Grass</td>
</tr>
<tr>
<td>Laurel</td>
<td>Laurel, DE</td>
<td>Glassboro Sandy Loam</td>
<td>Sandy Loam / Sandy Clay Loam</td>
<td>Grass</td>
</tr>
<tr>
<td>Wilmington</td>
<td>Wilmington-Porter, DE</td>
<td>Neshaminy-Montalto Silt Loam</td>
<td>Silt Loam</td>
<td>Grass</td>
</tr>
<tr>
<td>Selbyville</td>
<td>Selbyville, DE</td>
<td>Kiej Loamy Sand</td>
<td>Sandy Clay Loam</td>
<td>Grass</td>
</tr>
<tr>
<td>Viola</td>
<td>Viola, DE</td>
<td>Woodstown Loam</td>
<td>Sandy Loam</td>
<td>Grass</td>
</tr>
</tbody>
</table>

Six sites share a similar sandy loam texture, with four of those sites (Dover, Harrington, Laurel, and Selbyville) also sharing similar in situ volumetric water contents. These same four sites also exhibited similar porosities, with much more closely related values than any of the other five sites. Two other sites, sharing a similar loam texture, also shared similar in situ volumetric water contents. Porosity values were very closely related, more than any of the other seven sites. Altogether there are approximately three classes of soil texture present among the nine sites, with six having a sandy loam, two loam, and one a silt loam. These results demonstrate a strong link between a soil’s texture and its ability to hold moisture. As clay particles represent the smallest particle size, followed by silt and then sand as the largest, soils with a higher clay content are likely to hold more moisture than soils with a low clay content. The greater the number of particles in a soil sample, the more able the soil is to hold onto moisture.
Table 3.4. Summary of equation variables.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Name</th>
<th>Units</th>
</tr>
</thead>
<tbody>
<tr>
<td>$m_{soil}$</td>
<td>Soil Mass</td>
<td>g</td>
</tr>
<tr>
<td>$m_{moisture}$</td>
<td>Moisture Mass</td>
<td>g</td>
</tr>
<tr>
<td>$V_{moisture}$</td>
<td>Moisture Volume</td>
<td>mL</td>
</tr>
<tr>
<td>$m_{initial}$</td>
<td>Initial Soil Mass</td>
<td>g</td>
</tr>
<tr>
<td>$m_{dry}$</td>
<td>Dry Soil Mass</td>
<td>g</td>
</tr>
<tr>
<td>$V_{initial}$</td>
<td>Total Volume (initial)</td>
<td>mL</td>
</tr>
<tr>
<td>$V_{dry}$</td>
<td>Dry Soil Volume</td>
<td>mL</td>
</tr>
<tr>
<td>$\Theta_g$</td>
<td>Gravimetric Water Content</td>
<td>%</td>
</tr>
<tr>
<td>$P_{bulk}$</td>
<td>Bulk Density</td>
<td>mL</td>
</tr>
<tr>
<td>$\Theta_v$</td>
<td>Volumetric Water Content</td>
<td>%</td>
</tr>
<tr>
<td>$V_{sand}$</td>
<td>Sand Volume</td>
<td>mL</td>
</tr>
<tr>
<td>$V_{silt}$</td>
<td>Silt Volume</td>
<td>mL</td>
</tr>
<tr>
<td>$V_{clay}$</td>
<td>Clay Volume</td>
<td>mL</td>
</tr>
<tr>
<td>$V_{total}$</td>
<td>Total Volume (final)</td>
<td>mL</td>
</tr>
<tr>
<td>$\Phi$</td>
<td>Porosity</td>
<td>Ratio</td>
</tr>
<tr>
<td>$P_{sand}$</td>
<td>Average Sand Density</td>
<td>g/mL</td>
</tr>
<tr>
<td>$P_{silt}$</td>
<td>Average Silt Density</td>
<td>g/mL</td>
</tr>
<tr>
<td>$P_{clay}$</td>
<td>Average Clay Density</td>
<td>g/mL</td>
</tr>
<tr>
<td>$P_{particle}$</td>
<td>Average Particle Density</td>
<td>g/mL</td>
</tr>
</tbody>
</table>
### Table 3.5. *In situ* soil sample analysis.

<table>
<thead>
<tr>
<th>Site Location</th>
<th>Soil Sample Mass (g)</th>
<th>Dry Soil Mass (g)</th>
<th>Water Mass (g)</th>
<th>Water Mass (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Newark</td>
<td>296.8</td>
<td>240.8</td>
<td>56.0</td>
<td>18.87</td>
</tr>
<tr>
<td>Dover</td>
<td>251.7</td>
<td>221.6</td>
<td>30.1</td>
<td>11.96</td>
</tr>
<tr>
<td>Fair Hill</td>
<td>243.9</td>
<td>171.1</td>
<td>72.8</td>
<td>29.85</td>
</tr>
<tr>
<td>Glasgow</td>
<td>280.3</td>
<td>238.2</td>
<td>42.1</td>
<td>15.02</td>
</tr>
<tr>
<td>Harrington</td>
<td>295.2</td>
<td>262.5</td>
<td>32.7</td>
<td>11.08</td>
</tr>
<tr>
<td>Laurel</td>
<td>206.9</td>
<td>179.5</td>
<td>27.4</td>
<td>13.24</td>
</tr>
<tr>
<td>Wilmington-Porter</td>
<td>280.4</td>
<td>208.7</td>
<td>71.7</td>
<td>25.57</td>
</tr>
<tr>
<td>Selbyville</td>
<td>336.9</td>
<td>300.9</td>
<td>36.0</td>
<td>10.69</td>
</tr>
<tr>
<td>Viola</td>
<td>193.2</td>
<td>159.6</td>
<td>33.6</td>
<td>17.39</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Soil Sample Volume (mL)</th>
<th>Dry Volume (mL)</th>
<th>Water Volume (mL)</th>
<th>Water Volume (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Newark</td>
<td>290</td>
<td>230</td>
<td>60</td>
</tr>
<tr>
<td>Dover</td>
<td>210</td>
<td>220</td>
<td>-10</td>
</tr>
<tr>
<td>Fair Hill</td>
<td>280</td>
<td>190</td>
<td>90</td>
</tr>
<tr>
<td>Glasgow</td>
<td>230</td>
<td>230</td>
<td>0</td>
</tr>
<tr>
<td>Harrington</td>
<td>280</td>
<td>210</td>
<td>70</td>
</tr>
<tr>
<td>Laurel</td>
<td>220</td>
<td>150</td>
<td>70</td>
</tr>
<tr>
<td>Wilmington-Porter</td>
<td>280</td>
<td>230</td>
<td>50</td>
</tr>
<tr>
<td>Selbyville</td>
<td>260</td>
<td>220</td>
<td>40</td>
</tr>
<tr>
<td>Viola</td>
<td>190</td>
<td>140</td>
<td>50</td>
</tr>
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<table>
<thead>
<tr>
<th>Gravimetric Moisture (%)</th>
<th>VWC (%)</th>
<th>Bulk Density (g/mL)</th>
<th>Porosity</th>
</tr>
</thead>
<tbody>
<tr>
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<td>23.26</td>
<td>24.35</td>
<td>1.05</td>
</tr>
<tr>
<td>Dover</td>
<td>13.58</td>
<td>13.68</td>
<td>1.01</td>
</tr>
<tr>
<td>Fair Hill</td>
<td>42.55</td>
<td>38.32</td>
<td>0.90</td>
</tr>
<tr>
<td>Glasgow</td>
<td>17.67</td>
<td>18.30</td>
<td>1.04</td>
</tr>
<tr>
<td>Harrington</td>
<td>12.46</td>
<td>15.57</td>
<td>1.25</td>
</tr>
<tr>
<td>Laurel</td>
<td>15.26</td>
<td>18.22</td>
<td>1.20</td>
</tr>
<tr>
<td>Wilmington-Porter</td>
<td>34.36</td>
<td>31.17</td>
<td>0.91</td>
</tr>
<tr>
<td>Selbyville</td>
<td>11.96</td>
<td>16.36</td>
<td>1.37</td>
</tr>
<tr>
<td>Viola</td>
<td>21.05</td>
<td>24.00</td>
<td>1.14</td>
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</table>

<table>
<thead>
<tr>
<th>Sand (mL)</th>
<th>Silt (mL)</th>
<th>Clay (mL)</th>
<th>Total (mL)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Newark</td>
<td>6.00</td>
<td>6.00</td>
<td>5.00</td>
</tr>
<tr>
<td>Dover</td>
<td>8.50</td>
<td>3.50</td>
<td>5.00</td>
</tr>
<tr>
<td>Fair Hill</td>
<td>9.00</td>
<td>4.00</td>
<td>6.00</td>
</tr>
<tr>
<td>Glasgow</td>
<td>5.50</td>
<td>5.25</td>
<td>2.25</td>
</tr>
<tr>
<td>Harrington</td>
<td>9.75</td>
<td>3.00</td>
<td>2.25</td>
</tr>
<tr>
<td>Laurel</td>
<td>8.75</td>
<td>3.25</td>
<td>3.00</td>
</tr>
<tr>
<td>Wilmington-Porter</td>
<td>3.50</td>
<td>8.50</td>
<td>2.50</td>
</tr>
<tr>
<td>Selbyville</td>
<td>8.50</td>
<td>2.50</td>
<td>5.00</td>
</tr>
<tr>
<td>Viola</td>
<td>8.00</td>
<td>5.00</td>
<td>2.00</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Sand (%)</th>
<th>Silt (%)</th>
<th>Clay (%)</th>
<th>Total Soil (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Newark</td>
<td>41.38</td>
<td>41.38</td>
<td>17.24</td>
</tr>
<tr>
<td>Dover</td>
<td>58.62</td>
<td>24.14</td>
<td>17.24</td>
</tr>
<tr>
<td>Fair Hill</td>
<td>60.00</td>
<td>26.67</td>
<td>13.33</td>
</tr>
<tr>
<td>Glasgow</td>
<td>42.31</td>
<td>40.38</td>
<td>17.31</td>
</tr>
<tr>
<td>Harrington</td>
<td>65.00</td>
<td>20.00</td>
<td>15.00</td>
</tr>
<tr>
<td>Laurel</td>
<td>58.33</td>
<td>21.67</td>
<td>20.00</td>
</tr>
<tr>
<td>Wilmington-Porter</td>
<td>24.14</td>
<td>58.62</td>
<td>17.24</td>
</tr>
<tr>
<td>Selbyville</td>
<td>60.71</td>
<td>17.86</td>
<td>21.43</td>
</tr>
<tr>
<td>Viola</td>
<td>53.33</td>
<td>33.33</td>
<td>13.33</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Sand Particle Density (g/mL)</th>
<th>Silt Particle Density (g/mL)</th>
<th>Clay Particle Density (g/mL)</th>
<th>Average Particle Density (g/mL)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Newark</td>
<td>0.59</td>
<td>0.46</td>
<td>0.18</td>
</tr>
<tr>
<td>Dover</td>
<td>0.83</td>
<td>0.28</td>
<td>0.18</td>
</tr>
<tr>
<td>Fair Hill</td>
<td>0.85</td>
<td>0.31</td>
<td>0.14</td>
</tr>
<tr>
<td>Glasgow</td>
<td>0.60</td>
<td>0.47</td>
<td>0.18</td>
</tr>
<tr>
<td>Harrington</td>
<td>0.92</td>
<td>0.24</td>
<td>0.16</td>
</tr>
<tr>
<td>Laurel</td>
<td>0.83</td>
<td>0.25</td>
<td>0.21</td>
</tr>
<tr>
<td>Wilmington-Porter</td>
<td>0.34</td>
<td>0.69</td>
<td>0.18</td>
</tr>
<tr>
<td>Selbyville</td>
<td>0.86</td>
<td>0.21</td>
<td>0.23</td>
</tr>
<tr>
<td>Viola</td>
<td>0.76</td>
<td>0.39</td>
<td>0.14</td>
</tr>
</tbody>
</table>
In order to quantitatively measure the persistence of soil moisture over time, an autocorrelation analysis was performed for all sites from August 1st through September 30th for 2005, 2006 and 2007 (except Selbyville and Viola which were not implemented until 2008) using separate lags of 1-week, 2-week, and 4-week intervals.

The daily maximum volumetric water content recordings from each site were placed into daily data tables sorted by year. Monthly averages are included at the bottom of each table. Graphs were made of the maximum daily volumetric water content at each site for all years from 2005-2008 in which a DEOS station was present. A few of the sites are missing data (indicated by an “ND” signifying no data) due to problems with or the absence of instrument sensors. The station at Selbyville, for instance, was not erected until the end of 2007 which is the reason for a lack of data from 2005-2007. Along with graphing volumetric water content data, the average daily air temperature and total daily rainfall are presented at each site in order to visually display the relationships between these three variables. These data are included in Appendices A, B and C.

Monthly averages (e.g., January 2005, 2006, 2007, and 2008) of daily maximum volumetric water content were placed into separate tables sorted by station and graphed to show the variability of soil moisture. An annual soil moisture average was derived using the monthly averages at each site, as well as monthly averages spanning all years (i.e., January 2005-2008, February 2005-2008, etc.). These averages may serve as the basis for comparison between expected and observed data.
Graphs were made of the monthly averages at each site and are provided to demonstrate the annual variability in soil moisture at each location. Field capacity estimates of each site were based on an average of volumetric water content recordings immediately following heavy rain events throughout 2005-2008. The average field capacity value from each year is averaged into one final field capacity estimate at each site.

An autocorrelation analysis was used to determine the persistence of soil moisture over the course of one week, two weeks, and four weeks. The following equation was used in each autocorrelation analysis:

\[ r_l = \frac{C_{x, x-l}}{S_x}, \quad (3.24) \]

where \( r_l \) is the regression associated with each lag \( l \), \( C_x \) is the covariance between \( x \) (vwc) and \( x - l \) (time interval), and \( S_x \) is the variance in the dependent variable (vwc). The regression value can range from -1.0 to 1.0, with -1.0 signifying an inverse relationship between the vwc during the specified time interval, 1.0 signifying a direct relationship of the vwc during the specified time interval, and 0.0 representing an uncorrelated relationship (i.e., vwc changes over time are completely random). A T-test was performed in order to determine the approximate positive and negative regression values which can be considered as significantly correlated. The following equation was used in the T-test analysis:
\[ T = r \sqrt{\frac{N-2}{1-r^2}}, \]  

(3.25)

where \( T \) represents the number of standard deviations from the mean, \( r \) is the minimum regression required for a significant correlation, \( N - 2 \) represents the number of days under analysis (i.e., August and September = 61 days) minus 2 degrees of freedom (i.e., 61 - 2 = 59), and \( 1 - r^2 \) is the scaling parameter used to determine the distribution of \( T \). A \( T \) value of 1.96 was used, offering a 95% confidence level in the statistical significance of each lag. This also established the positive and negative \( r \) values at 0.2475 and -0.2475 respectively.

Remote Sensing Data Pre-Processing Methods

All remote sensing derived products were initially downloaded in HDF-EOS (Hierarchical Data Format - Earth Observing System) format from the NASA MODIS website (http://modis-land.gsfc.nasa.gov/vi.htm (vegetation indices) and http://modis-land.gsfc.nasa.gov/temp.htm (land surface temperature)) and from the National Snow and Ice Data Center (NSIDC) website (http://nsidc.org/data/ae_land3.html (soil moisture)), which is not directly compatible with ESRI Desktop ArcGIS ArcInfo (version 9.3.1) software. Used for mapping the geographic distribution of the DEOS stations and remotely sensed data. A program available on the NSIDC website (http://nsidc.org/data/hdfeos/hdf_to_binary.html) known as “hdp utility” was used to
convert the data into band-sequential (.bsq) files directly by GIS software with a separate header (.hdr) file.

Using ArcGIS “Extract Subdataset” tool located in ArcToolbox (ArcToolbox > Data Management Tools > Raster > Raster Processing > Extract Subdataset), the data layers were extracted from the main HDF-EOS file into an IMAGINE (.img) file, a format ArcGIS can read and analyze. Prior to mapping, the spatial reference of each image is defined and all data values are converted from raw digital numbers to their physical values (i.e. binary to Kelvin), using instructions found on the AMSR-E and MODIS websites. Additional layers were downloaded to provide spatial reference proximity for the image data including the U.S. Streams and Waterbodies (U.S. Fish & Wildlife Service, 2001), Municipal Boundaries (Mahaffie, 2004) and U.S. Counties (National Atlas of the United States, 2009). All final GIS layers were re-projected using the GCS North American 1983 coordinate system with an equal-area projection using the North American Datum 1983 in decimal degrees.

The soil moisture data collected by the AMSR-E sensor were acquired as part of the gridded Level-3 land surface datasets from the National Snow and Ice Data Center (NSIDC) (http://nsidc.org/index.html). Data are projected in a 25 km Equal-Area Scalable Earth Grid (EASE-Grid) cell spacing utilizing a global cylindrical equal-area projection.

All data are recorded twice: once during the satellite’s ascension and again during its descent, resulting in two sets of data layers (e.g., A_Soil_Moisture and D_Soil_Moisture). Either data layer will suffice, though for this project the
descension data (D_Soil_Moisture) were used. Pre-processing requirements were the most intensive with these data, requiring a number of steps in order to create a user-defined projection.

After extracting the soil moisture subdataset as described in the remote sensing pre-processing section above, the next step was to define the projection. Using the instructions located on the NSIDC website (http://nsidc.org/data/ease/gis.html), a user-defined projection was established starting with step 4, “Define the projection for the new data set”. In ArcGIS, the “Define Projection” tool located in ArcToolbox (ArcToolbox > Data Management Tools > Projections and Transformations > Define Projection) can be used to define the soil moisture subdataset projection.

The data values included in the soil moisture data layer were converted before they could be analyzed in the GIS. Using the “Times” tool located in ArcToolbox (ArcToolbox > Spatial Analyst Tools > Math > Times) all data values were multiplied by 0.001, thereby making the conversion from 16-bit binary numbers into useable density values (g/mL) representing the mass of moisture (grams) per volume of soil (milliliters).

Following the subdataset extractions of vegetation index (250m 16 days EVI) and land surface temperature (LST_Day_1km) data layers, the next step was to define the spatial reference. Both data layers were defined and projected to an equal-area map projection in the GCS Latitude/Longitude coordinate system in decimal degrees with the North American Datum 1983 using the Define Projection tool in ArcGIS (found in ArcToolbox > Data Management Tools > Projections and Transformations >
Toolset). The next step involves the conversion of data values to NDVI (-1 to +1) for the EVI layers, and to degrees Kelvin for the LST layers. Using the “Times” tool located in ArcToolbox (ArcToolbox > Spatial Analyst Tools > Math > Times), the EVI data layer needed to be multiplied by 0.0001 and the LST data layer by 0.02. After this preprocessing, both data layers were ready to be mapped and analyzed in ArcGIS.

DEOS observed soil moisture and remotely derived data were compared during late August into early September from 2005-2008 to determine the average conditions as well as the general variability of soil moisture across the Delmarva Peninsula. This time period was chosen for its rainfall variability with 2005 representing a period of drought, 2006 representing a period with heavy rain events, and 2007 representing a more moderate period, as well as relatively high plant transpiration activity. In situ and AMSR-E soil moisture data allow for direct comparison, while land surface temperatures and vegetation indices lend themselves to an indirect measure of soil moisture through the measurements of associated conditions such as land surface temperature and vegetation health. The direct comparison of all the remote sensing data with DEOS was accomplished by locating the nearest-pixel to each DEOS station within the GIS.
Chapter 4

OBSERVED SOIL MOISTURE CONDITIONS IN DELMARVA PENINSULA

4.1 Soil Moisture Conditions as Observed by DEOS

Graphs were made at each site comparing the maximum daily volumetric water content (vwc), daily rainfall, and average daily air temperature. These graphs extend from January 1, 2005 through June 30, 2008, when the data is available, and are separated by year. The average monthly volumetric water contents were included in these graphs in order to enhance trend visibility, but are also shown separately in an additional table and graph at each site.

In general, the graphs appear nearly identical with soils yielding the greatest volumetric water contents in the winter months (when evaporation and plant transpiration are very low). High average air temperatures result in a higher average rate of evapotranspiration, which causes greater fluctuation in the vwc particularly noticeable in the summer. As mentioned previously in chapter 3, soil texture plays a large role in the storage of moisture and provides some resistance to the evapotranspiration by vegetation.

Figure 4.1 represents the average conditions in Newark from January 1, 2005 to June 30, 2008, with Table 4.1 listing the average vwc for each month according to year as well as the average vwc for each month for all years. These averages give an idea of how vwc changes inter- and intra-annually. Note the dampened uni-modal curve that forms from January through December in every year, visible in the
averaged data (Figure 4.2) but less prevalent in the daily data in Figure 4.1. Both 2005 and 2007 display a maximum vwc during the winter months (November - February) and a minimum vwc during the summer months (June – September). An exception occurs in 2006 due to increased rainfall during July, and to a lesser extent June and August. There is a noticeable decrease in the vwc between each rain event during the first six months of 2008. The graph in Figure 4.2 represents the average vwc by month from 2005-2008 for Newark, DE. A curve is noticeable here although it is not especially pronounced.
Figure 4.1. Maximum daily vwc, daily rainfall, and average daily temperature in **Newark** from January 1, 2005 – June 30, 2008.
Table 4.1. Soil Moisture Monthly Averages (Maximum Daily Volumetric Water Content) - NEWARK

<table>
<thead>
<tr>
<th>Newark</th>
<th>2005</th>
<th>2006</th>
<th>2007</th>
<th>2008</th>
<th>Average VWC (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>30.91</td>
<td>30.23</td>
<td>31.71</td>
<td>29.26</td>
<td>31</td>
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<td>February</td>
<td>33.01</td>
<td>30.52</td>
<td>21.44</td>
<td>31.32</td>
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</tr>
<tr>
<td>March</td>
<td>31.40</td>
<td>26.63</td>
<td>31.55</td>
<td>30.06</td>
<td>30</td>
</tr>
<tr>
<td>April</td>
<td>30.47</td>
<td>28.81</td>
<td>31.66</td>
<td>28.18</td>
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</tr>
<tr>
<td>May</td>
<td>29.60</td>
<td>26.45</td>
<td>26.39</td>
<td>29.86</td>
<td>28</td>
</tr>
<tr>
<td>June</td>
<td>26.76</td>
<td>29.61</td>
<td>27.09</td>
<td>26.31</td>
<td>27</td>
</tr>
<tr>
<td>July</td>
<td>26.91</td>
<td>29.63</td>
<td>25.72</td>
<td>ND</td>
<td>27</td>
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<td>August</td>
<td>25.37</td>
<td>24.73</td>
<td>26.79</td>
<td>ND</td>
<td>26</td>
</tr>
<tr>
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<td>30.81</td>
<td>30.81</td>
<td>24.16</td>
<td>ND</td>
<td>29</td>
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<tr>
<td>October</td>
<td>31.70</td>
<td>31.70</td>
<td>27.79</td>
<td>ND</td>
<td>30</td>
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<tr>
<td>November</td>
<td>33.06</td>
<td>33.06</td>
<td>30.03</td>
<td>ND</td>
<td>32</td>
</tr>
<tr>
<td>December</td>
<td>29.43</td>
<td>29.65</td>
<td>32.62</td>
<td>ND</td>
<td>31</td>
</tr>
</tbody>
</table>

*Annual Average* | 29.95  | 29.32  | 28.08  | 29.17  | 29              |

*Note years with incomplete data.

Figure 4.2. Maximum daily VWC monthly averages; Newark, DE.
Table 4.2 shows the average vwc for each month according to year as well as the annual average vwc for Dover. The lack of visible trending of vwc in 2005 can be attributed to a severe lack of rainfall for two months as well as missing data. 2006 and 2007 both demonstrate strong highs and lows of the vwc during the winter and summer months, respectively, responding to large precipitation events and evapotranspiration demands. Soil moisture reserves are recharged as can be clearly seen in fall (e.g., in September 2006 and late October 2007). The first six months of 2008 also display a trend of gradually dipping vwc from February through the end of June. The graph in Figure 4.4 represents the average vwc by month from 2005-2008 for Dover. The curve is an excellent representation of the expected annual cycle of soil moisture, with the lowest values occurring at the time of maximum daily evapotranspiration in August attributing to a more pronounced annual cycle. A relatively low field capacity of 24 can be attributed to a soil containing a relatively high percentage of sand at 59%. This explains the dynamic range in annual vwc as the large sand particles are unable to maintain soil moisture levels as well as silt and clay.
Figure 4.3. Maximum daily VWC, daily rainfall, and average daily temperature in Dover from September 19, 2005 – June 30, 2008.
Table 4.2. Soil Moisture Monthly Averages (Maximum Daily Volumetric Water Content) – DOVER

<table>
<thead>
<tr>
<th>Dover</th>
<th>2005</th>
<th>2006</th>
<th>2007</th>
<th>2008</th>
<th>Average VWC (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>ND</td>
<td>26.92</td>
<td>24.63</td>
<td>23.98</td>
<td>25</td>
</tr>
<tr>
<td>February</td>
<td>ND</td>
<td>26.97</td>
<td>25.61</td>
<td>25.16</td>
<td>26</td>
</tr>
<tr>
<td>March</td>
<td>ND</td>
<td>21.30</td>
<td>24.98</td>
<td>22.50</td>
<td>23</td>
</tr>
<tr>
<td>April</td>
<td>ND</td>
<td>16.32</td>
<td>22.62</td>
<td>19.33</td>
<td>19</td>
</tr>
<tr>
<td>May</td>
<td>ND</td>
<td>12.29</td>
<td>11.84</td>
<td>20.22</td>
<td>15</td>
</tr>
<tr>
<td>June</td>
<td>ND</td>
<td>11.92</td>
<td>9.86</td>
<td>15.32</td>
<td>12</td>
</tr>
<tr>
<td>July</td>
<td>ND</td>
<td>17.12</td>
<td>8.35</td>
<td>ND</td>
<td>13</td>
</tr>
<tr>
<td>August</td>
<td>ND</td>
<td>9.14</td>
<td>8.62</td>
<td>ND</td>
<td>9</td>
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<tr>
<td>September</td>
<td>11.38</td>
<td>17.49</td>
<td>6.90</td>
<td>ND</td>
<td>12</td>
</tr>
<tr>
<td>October</td>
<td>20.57</td>
<td>21.69</td>
<td>10.17</td>
<td>ND</td>
<td>17</td>
</tr>
<tr>
<td>November</td>
<td>22.37</td>
<td>24.50</td>
<td>16.43</td>
<td>ND</td>
<td>21</td>
</tr>
<tr>
<td>December</td>
<td>27.29</td>
<td>22.52</td>
<td>22.36</td>
<td>ND</td>
<td>24</td>
</tr>
<tr>
<td><strong>Annual Average</strong></td>
<td><strong>20.40</strong></td>
<td><strong>19.01</strong></td>
<td><strong>16.03</strong></td>
<td><strong>21.08</strong></td>
<td><strong>20</strong></td>
</tr>
</tbody>
</table>

*Note years with incomplete data.

Figure 4.4. Maximum daily volumetric water content monthly averages; Dover, DE.
Table 4.3 shows the average vwc for each month according to year as well as the average vwc for each month for all years at Fair Hill. These monthly averages give an idea of how vwc changes from season to season and from year to year. Similar vwc trends occur throughout 2005 and 2007, with the minimum vwc taking place from June-September, and the maximum vwc taking place from November-March. Similarly, 2005-2007 exhibit a steep rise in vwc during the fall months, recovering from depleted soil moisture levels during the summer season. 2006 experienced periods of significantly heavy rainfall throughout the summer and fall, which limited lower soil moisture levels to several weeks in early and late summer versus the entire summer season. Despite these rain events, the vwc in 2006 achieves its lowest level in August and its highest peak in November. All three years (2005-2007) exhibit a strong curve in vwc levels annually with the summer months displaying the lowest vwc levels and the winter months showing the greatest. The vwc stays the same from January-April in 2008, with very little change taking place. There are two large vwc spikes during April 2008 without any increase in rainfall, signifying an error in either the recorded rainfall or vwc. The graph in Figure 4.6 represents the average vwc by month from 2005-2008 for Fair Hill, MD. The large range in vwc annually is likely attributable to the high ratio of sand at the site, which is unable to maintain moisture as effectively as silt and clay.
Figure 4.5. Maximum daily vwc, daily rainfall, and average daily temperature in Fair Hill from January 1, 2005 – June 30, 2008.
Table 4.3. Soil Moisture Monthly Averages (Maximum Daily Volumetric Water Content) – FAIR HILL

<table>
<thead>
<tr>
<th>Fair Hill, MD</th>
<th>2005</th>
<th>2006</th>
<th>2007</th>
<th>2008</th>
<th>Average VWC (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>40.35</td>
<td>40.86</td>
<td>42.02</td>
<td>40.23</td>
<td>41</td>
</tr>
<tr>
<td>February</td>
<td>41.86</td>
<td>40.84</td>
<td>35.98</td>
<td>41.18</td>
<td>40</td>
</tr>
<tr>
<td>March</td>
<td>43.75</td>
<td>37.20</td>
<td>41.95</td>
<td>41.15</td>
<td>41</td>
</tr>
<tr>
<td>April</td>
<td>36.62</td>
<td>36.67</td>
<td>40.84</td>
<td>40.48</td>
<td>39</td>
</tr>
<tr>
<td>May</td>
<td>29.95</td>
<td>27.28</td>
<td>36.37</td>
<td>36.72</td>
<td>33</td>
</tr>
<tr>
<td>June</td>
<td>23.88</td>
<td>35.38</td>
<td>25.66</td>
<td>27.72</td>
<td>28</td>
</tr>
<tr>
<td>July</td>
<td>29.59</td>
<td>36.97</td>
<td>20.82</td>
<td>ND</td>
<td>29</td>
</tr>
<tr>
<td>August</td>
<td>28.42</td>
<td>24.98</td>
<td>26.79</td>
<td>ND</td>
<td>27</td>
</tr>
<tr>
<td>September</td>
<td>21.22</td>
<td>38.13</td>
<td>19.65</td>
<td>ND</td>
<td>26</td>
</tr>
<tr>
<td>October</td>
<td>35.57</td>
<td>40.02</td>
<td>39.92</td>
<td>ND</td>
<td>39</td>
</tr>
<tr>
<td>November</td>
<td>39.13</td>
<td>43.33</td>
<td>39.02</td>
<td>ND</td>
<td>40</td>
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<tr>
<td>December</td>
<td>41.20</td>
<td>41.15</td>
<td>42.03</td>
<td>ND</td>
<td>41</td>
</tr>
<tr>
<td>Annual Average</td>
<td>34.30</td>
<td>36.90</td>
<td>34.25</td>
<td>37.91</td>
<td>37</td>
</tr>
</tbody>
</table>

*Note years with incomplete data.

Figure 4.6. Maximum daily volumetric water content monthly averages; Fair Hill, MD.
Table 4.4 shows the average vwc for each month according to year as well as the average vwc for each month from 2006-2008 for Glasgow. Due to a lack of data for all of 2005 as well as most of 2006 and half of 2008, the average vwc is insufficiently represented at this site and should be used with caution. The minimum vwc for 2007 occurs in July, with the maximum vwc occurring in December. This is followed by an overall gradual descent of vwc from the winter into the summer, and a gradual ascent of vwc taking place from winter into summer. Both 2006 and 2008 show similar trending in vwc levels across their respective time periods. The graph in Figure 4.8 represents the average vwc by month from 2006-2008 for Glasgow. While not a well-represented average, a strong annual cycle is still evident with minimum vwc in September and maximum vwc in December.
Figure 4.7. Maximum daily vwc, daily rainfall, and average daily temperature in Glasgow from October 13, 2006 – June 30, 2008.
Table 4.4. Soil Moisture Monthly Averages (Maximum Daily Volumetric Water Content) – GLASGOW

<table>
<thead>
<tr>
<th></th>
<th>2005</th>
<th>2006</th>
<th>2007</th>
<th>2008</th>
<th>Average VWC (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>ND</td>
<td>ND</td>
<td>30.73</td>
<td>29.25</td>
<td>30</td>
</tr>
<tr>
<td>February</td>
<td>ND</td>
<td>ND</td>
<td>21.46</td>
<td>32.93</td>
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<tr>
<td>March</td>
<td>ND</td>
<td>ND</td>
<td>30.30</td>
<td>29.40</td>
<td>30</td>
</tr>
<tr>
<td>April</td>
<td>ND</td>
<td>ND</td>
<td>28.54</td>
<td>25.05</td>
<td>27</td>
</tr>
<tr>
<td>May</td>
<td>ND</td>
<td>ND</td>
<td>18.06</td>
<td>25.75</td>
<td>22</td>
</tr>
<tr>
<td>June</td>
<td>ND</td>
<td>ND</td>
<td>13.70</td>
<td>19.35</td>
<td>17</td>
</tr>
<tr>
<td>July</td>
<td>ND</td>
<td>ND</td>
<td>13.64</td>
<td>ND</td>
<td>14</td>
</tr>
<tr>
<td>August</td>
<td>ND</td>
<td>ND</td>
<td>16.41</td>
<td>ND</td>
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<tr>
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<td>ND</td>
<td>ND</td>
<td>12.42</td>
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<td>12</td>
</tr>
<tr>
<td>October</td>
<td>ND</td>
<td>26.79</td>
<td>14.83</td>
<td>ND</td>
<td>21</td>
</tr>
<tr>
<td>November</td>
<td>ND</td>
<td>27.81</td>
<td>25.39</td>
<td>ND</td>
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<tr>
<td>December</td>
<td>ND</td>
<td>29.02</td>
<td>33.08</td>
<td>ND</td>
<td>31</td>
</tr>
<tr>
<td>Annual Average</td>
<td>ND</td>
<td>27.87</td>
<td>21.55</td>
<td>26.96</td>
<td>25</td>
</tr>
</tbody>
</table>

*Note years with incomplete data.

Figure 4.8. Maximum daily volumetric water content monthly averages; Glasgow, DE.
Table 4.5 shows the average vwc for each month according to year as well as the average vwc for each month from 2006-2008 for Harrington. The same downward trend in vwc from winter to summer is evident in all years (2006, 2007, and 2008), which is apparent at some of the other sites including Dover, Glasgow, and Wilmington. Due to a lack of data in May and September 2006, it cannot be stated with absolute certainty that the minimum vwc occurred during the month of May as shown in Table 4.5. Minimums for all three years take place within three months of each other, with maximums taking place in January and February for 2006 and 2007-2008, respectively. Additionally, all three years display a strong depletion in the vwc during April with soil moisture levels not recharging until October and November due to the large evapotranspiration demands in summer. The graph in Figure 4.10 represents the average vwc by month from 2006-2008 for Harrington. Even with data covering a span of just 30 months there is still a strong annual cycle in vwc levels taking place. Having the largest portion of sand (65%) of all nine sites, the dynamic annual range can be attributed to the lack of efficiency possessed by sand particles in maintaining soil moisture. Interestingly, the annual cycle at Harrington is similar to annual cycle observed in Glasgow, despite Glasgow having a lower overall percentage of sand (42%). The high ratio of sand is also likely responsible for the relatively low vwc levels taking place each year during the summer months.
Figure 4.9. Maximum daily vwc, daily rainfall, and average daily temperature in Harrington from January 1, 2006 – June 30, 2008.
Table 4.5. Soil Moisture Monthly Averages (Maximum Daily Volumetric Water Content) – **HARRINGTON**

<table>
<thead>
<tr>
<th>Months</th>
<th>2005</th>
<th>2006</th>
<th>2007</th>
<th>2008</th>
<th>Average VWC (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>ND</td>
<td>37.43</td>
<td>34.65</td>
<td>34.40</td>
<td>35</td>
</tr>
<tr>
<td>February</td>
<td>ND</td>
<td>37.15</td>
<td>35.87</td>
<td>35.25</td>
<td>36</td>
</tr>
<tr>
<td>March</td>
<td>ND</td>
<td>27.90</td>
<td>34.94</td>
<td>31.85</td>
<td>32</td>
</tr>
<tr>
<td>April</td>
<td>ND</td>
<td>20.38</td>
<td>31.46</td>
<td>23.26</td>
<td>25</td>
</tr>
<tr>
<td>May</td>
<td>ND</td>
<td>11.11</td>
<td>15.04</td>
<td>25.30</td>
<td>17</td>
</tr>
<tr>
<td>June</td>
<td>ND</td>
<td>19.54</td>
<td>9.95</td>
<td>20.85</td>
<td>17</td>
</tr>
<tr>
<td>July</td>
<td>ND</td>
<td>25.30</td>
<td>8.74</td>
<td>ND</td>
<td>17</td>
</tr>
<tr>
<td>August</td>
<td>ND</td>
<td>22.93</td>
<td>11.06</td>
<td>ND</td>
<td>17</td>
</tr>
<tr>
<td>September</td>
<td>ND</td>
<td>23.75</td>
<td>13.56</td>
<td>ND</td>
<td>19</td>
</tr>
<tr>
<td>October</td>
<td>ND</td>
<td>30.91</td>
<td>14.93</td>
<td>ND</td>
<td>23</td>
</tr>
<tr>
<td>November</td>
<td>ND</td>
<td>32.15</td>
<td>26.06</td>
<td>ND</td>
<td>29</td>
</tr>
<tr>
<td>December</td>
<td>ND</td>
<td>31.03</td>
<td>32.35</td>
<td>ND</td>
<td>32</td>
</tr>
<tr>
<td><strong>Annual Average</strong></td>
<td><strong>ND</strong></td>
<td><strong>26.63</strong></td>
<td><strong>22.38</strong></td>
<td><strong>28.49</strong></td>
<td><strong>27</strong></td>
</tr>
</tbody>
</table>

*Note years with incomplete data.

Figure 4.10. Maximum daily volumetric water content monthly averages; Harrington, DE.
Table 4.6 shows the average vwc for each month according to year as well as the average vwc for each month from 2005-2008 for Laurel. The graph in Figure 4.12 represents the average vwc by month from 2005-2008 for Laurel. Compared to the other sites, Laurel has the lowest annual average of vwc. However, the average maximum vwc (16%) is nearly three times as large as the average minimum vwc (6%), which shows that the annual soil moisture variability in Laurel is very high. This is caused by the differences in soil properties at this site. Laurel has the third lowest make-up of silt (22%) and the highest make-up of clay (20%) of all 9 sites. This composition is most similar to Harrington, and although the field capacities and average volumetric water contents differ they both share the same relationship between the vwc and field capacity. That is, when the average vwc is divided by the estimated field capacity at each site, Laurel and Harrington retain an estimated 84% and 87%, respectively, of the total field capacity on average. This suggests that the soil at Laurel has the potential to hold larger amounts of moisture relative to total field capacity. The large fluctuation in vwc annually can be attributed to the large overall make-up of sand particles (58%) which are unable to store soil moisture as effectively as silt and clay.
Figure 4.11. Maximum daily vwc, daily rainfall, and average daily temperature in *Laurel* from March 31, 2005 – June 30, 2008.
Table 4.6. Soil Moisture Monthly Averages (Maximum Daily Volumetric Water Content) – **LAUREL**

<table>
<thead>
<tr>
<th></th>
<th>2005</th>
<th>2006</th>
<th>2007</th>
<th>2008</th>
<th>Average VWC (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>ND</td>
<td>16.63</td>
<td>16.50</td>
<td>15.37</td>
<td>16</td>
</tr>
<tr>
<td>February</td>
<td>ND</td>
<td>16.22</td>
<td>16.90</td>
<td>15.60</td>
<td>16</td>
</tr>
<tr>
<td>March</td>
<td>17.20</td>
<td>12.90</td>
<td>16.34</td>
<td>14.18</td>
<td>15</td>
</tr>
<tr>
<td>April</td>
<td>12.09</td>
<td>11.26</td>
<td>14.21</td>
<td>12.59</td>
<td>13</td>
</tr>
<tr>
<td>May</td>
<td>11.50</td>
<td>6.25</td>
<td>7.75</td>
<td>13.27</td>
<td>10</td>
</tr>
<tr>
<td>June</td>
<td>12.67</td>
<td>13.72</td>
<td>4.37</td>
<td>9.65</td>
<td>10</td>
</tr>
<tr>
<td>July</td>
<td>11.64</td>
<td>12.11</td>
<td>3.46</td>
<td>ND</td>
<td>9</td>
</tr>
<tr>
<td>August</td>
<td>7.84</td>
<td>4.93</td>
<td>5.97</td>
<td>ND</td>
<td>6</td>
</tr>
<tr>
<td>September</td>
<td>3.89</td>
<td>14.09</td>
<td>4.95</td>
<td>ND</td>
<td>8</td>
</tr>
<tr>
<td>October</td>
<td>12.38</td>
<td>14.64</td>
<td>6.21</td>
<td>ND</td>
<td>11</td>
</tr>
<tr>
<td>November</td>
<td>14.62</td>
<td>16.04</td>
<td>12.66</td>
<td>ND</td>
<td>14</td>
</tr>
<tr>
<td>December</td>
<td>16.62</td>
<td>15.51</td>
<td>14.55</td>
<td>ND</td>
<td>16</td>
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<tr>
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<td>12.86</td>
<td>10.32</td>
<td>13.44</td>
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*Note years with incomplete data.*

Figure 4.12. Maximum daily volumetric water content monthly averages; Laurel, DE.
Table 4.7 shows the average vwc for each month according to year as well as the average vwc for each month from 2005-2008 for Wilmington. The curve in Figure 4.14 represents the average annual vwc during this time period. Similar vwc trends can be seen from January through December for 2005 and 2007, with the minimum and maximum vwc levels occurring in September and July, and February and December, respectively. This same curve can be seen in 2006 with the exception of the heavy rainfall events which occurred in the summer months causing a significant rise and larger vwc levels. A steadily falling vwc can be seen in 2008 from January through June, similar to the three previous years. The generally stable annual soil moisture conditions (2005-2007) can likely be attributed to the large percentage of silt and clay (75.86% combined) which are able to store moisture more effectively than sand.
Figure 4.13. Maximum daily vwc, daily rainfall, and average daily temperature in Wilmington from January 1, 2005 – June 30, 2008.
Table 4.7. Soil Moisture Monthly Averages (Maximum Daily Volumetric Water Content) – WILMINGTON

<table>
<thead>
<tr>
<th>Wilmington</th>
<th>2005</th>
<th>2006</th>
<th>2007</th>
<th>2008</th>
<th>Average VWC (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>36.94</td>
<td>36.14</td>
<td>36.79</td>
<td>48.76</td>
<td>40</td>
</tr>
<tr>
<td>February</td>
<td>38.22</td>
<td>34.74</td>
<td>34.85</td>
<td>50.75</td>
<td>40</td>
</tr>
<tr>
<td>March</td>
<td>37.60</td>
<td>31.44</td>
<td>36.95</td>
<td>48.53</td>
<td>39</td>
</tr>
<tr>
<td>April</td>
<td>35.83</td>
<td>30.01</td>
<td>36.86</td>
<td>40.78</td>
<td>36</td>
</tr>
<tr>
<td>May</td>
<td>31.00</td>
<td>25.22</td>
<td>27.08</td>
<td>38.51</td>
<td>30</td>
</tr>
<tr>
<td>June</td>
<td>23.78</td>
<td>30.76</td>
<td>22.79</td>
<td>28.75</td>
<td>27</td>
</tr>
<tr>
<td>July</td>
<td>22.45</td>
<td>32.51</td>
<td>15.95</td>
<td>ND</td>
<td>24</td>
</tr>
<tr>
<td>August</td>
<td>15.83</td>
<td>18.35</td>
<td>20.45</td>
<td>ND</td>
<td>18</td>
</tr>
<tr>
<td>September</td>
<td>9.73</td>
<td>33.36</td>
<td>19.30</td>
<td>ND</td>
<td>21</td>
</tr>
<tr>
<td>October</td>
<td>22.36</td>
<td>35.52</td>
<td>26.88</td>
<td>ND</td>
<td>28</td>
</tr>
<tr>
<td>November</td>
<td>28.77</td>
<td>36.70</td>
<td>37.52</td>
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<td>34.50</td>
<td>47.17</td>
<td>ND</td>
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<tr>
<td><strong>Annual Average</strong></td>
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<td><strong>31.60</strong></td>
<td><strong>30.22</strong></td>
<td><strong>42.68</strong></td>
<td><strong>35</strong></td>
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</tbody>
</table>

*Note years with incomplete data.

Figure 4.14. Maximum daily volumetric water content monthly averages; Wilmington, DE.
Table 4.8 shows the average monthly vwc for the first half of 2008 as well as an overall average of the vwc at the Selbyville site. More data is needed in order to provide an accurate annual average. The vwc remains relatively uniform throughout the first 5 months of 2008 before it begins to deplete in early June due to increasing evapotranspiration. Soil texture at Selbyville is a sandy clay loam, and similar to other sites with this soil texture, the vwc is relatively high compared to the estimated field capacity. The average vwc divided by the estimated field capacity showed a 91% average moisture yield, which, along with the steady rainfall, is why the vwc levels remain very consistent throughout January-May.

![Figure 4.15](image_url)

Figure 4.15. Maximum daily vwc, daily rainfall, and average daily temperature in Selbyville from January 1, 2008 – June 30, 2008.
Table 4.8. Soil Moisture Monthly Averages (Maximum Daily Volumetric Water Content) – SELBYVILLE

<table>
<thead>
<tr>
<th>Selbyville</th>
<th>2005</th>
<th>2006</th>
<th>2007</th>
<th>2008</th>
<th>Average VWC (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>31.07</td>
<td>31</td>
</tr>
<tr>
<td>February</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>33.18</td>
<td>33</td>
</tr>
<tr>
<td>March</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>30.37</td>
<td>30</td>
</tr>
<tr>
<td>April</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>28.94</td>
<td>29</td>
</tr>
<tr>
<td>May</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>28.14</td>
<td>28</td>
</tr>
<tr>
<td>June</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>23.10</td>
<td>23</td>
</tr>
<tr>
<td>July</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
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<tr>
<td>August</td>
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<td>September</td>
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<td>November</td>
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<td>December</td>
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<td>ND</td>
<td>ND</td>
<td>ND</td>
<td><strong>29.13</strong></td>
<td><strong>29</strong></td>
</tr>
</tbody>
</table>

*Note years with incomplete data.

Figure 4.16. Maximum daily volumetric water content monthly averages; Selbyville, DE.
Table 4.9 shows the average monthly vwc for Viola from the end of December 2007 through the first half of 2008. More data is needed in order to provide an accurate annual average. There is a general downward descent from December 2007 through June 2008 as shown in Figure 4.17. Note the impact large rain events have on the vwc, particularly the spike in April due to heavy rain events. Soil texture at this site is also a sandy loam, with vwc averaging an estimated 85% of total field capacity annually.

![Graph](image-url)

Figure 4.17. Maximum daily vwc, daily rainfall, and average daily temperature in Viola from December 20, 2007 – June 30, 2008.
Table 4.9. Soil Moisture Monthly Averages (Maximum Daily Volumetric Water Content) – VIOLA

<table>
<thead>
<tr>
<th></th>
<th>2005</th>
<th>2006</th>
<th>2007</th>
<th>2008</th>
<th>Average VWC (%)</th>
</tr>
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<td>31.14</td>
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*Note years with incomplete data.

Figure 4.18. Maximum daily volumetric water content monthly averages; Viola, DE.
By determining vwc measurements one day following a heavy rain event, the soil’s ability to hold moisture against the pull of gravity is obtained. The field capacity tables listed in Tables 4.10 – 4.18 give the volumetric water content measurements at each site from 2005 through 2008 one day following a heavy rain event. An average vwc was created for each year with a final average derived from the annual averages. This final average (the bolded and shaded value at the bottom right of each table) accurately represents the field capacity at each location.

Although the vwc tends to be higher during a heavy rain event, the result is not a field capacity measurement, but a saturated capacity measurement. Saturated capacity is significant when monitoring flood effects from areas susceptible to constant heavy rain. The field capacity represents the amount of moisture a soil is able to hold on to against the pull of gravity, once the excess moisture has been forced out through percolation. By waiting one day, a much closer estimation of a soil’s ability to hold moisture can be observed. This moisture represents the volume (minus the portion held tightly to soil particles) available to the vegetation.

Soil texture is important because, based on the composition of sand, silt, and clay, it regulates based on pore space the amount of moisture a soil is able to hold. Typically, the smaller particles (clay) are able to hold greater amounts of moisture than larger particles (sand). This is due to the increased pore space provided by the higher number of soil particles (silt and clay). In sand, the larger particles produce larger but fewer pore spaces, allowing moisture to infiltrate into the land surface more quickly. Silt and clay particles provide smaller pore spaces but in higher numbers than
sand particles which allows them to hold moisture more tightly than sand, and therefore fine soils are able to hold more moisture than coarse soils. Different combinations of sand, silt, and clay permit varying degrees of moisture retention.

Table 4.10. Field capacity estimation for **Newark**.

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<th>Newark</th>
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<th>% VWC 2005</th>
<th>% VWC 2006</th>
<th>% VWC 2006</th>
<th>% VWC 2007</th>
<th>% VWC 2007</th>
<th>% VWC 2008</th>
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**Average**: 33.5  32.7  32.9  34.4  33
Table 4.11. Field capacity estimation for Dover.

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Table 4.12. Field capacity estimation for Fair Hill.

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<th>% VWC</th>
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Table 4.13. Field capacity estimation for **Glasgow**.

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<td></td>
</tr>
<tr>
<td>Average</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>32.5</td>
<td>32</td>
<td></td>
</tr>
</tbody>
</table>

Table 4.18. Field capacity estimation for Viola.

<table>
<thead>
<tr>
<th>Viola</th>
<th>% VWC</th>
<th>2005</th>
<th>% VWC</th>
<th>2006</th>
<th>% VWC</th>
<th>2007</th>
<th>% VWC</th>
<th>2008</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>30.6</td>
<td>12-Jan</td>
<td></td>
<td>37.9</td>
<td>2-Feb</td>
<td>34.9</td>
<td>14-Feb</td>
<td></td>
</tr>
<tr>
<td></td>
<td>30</td>
<td>6-Mar</td>
<td>32.1</td>
<td>9-Mar</td>
<td></td>
<td>29.1</td>
<td>29-Apr</td>
<td></td>
</tr>
<tr>
<td></td>
<td>41.7</td>
<td>13-May</td>
<td>32.2</td>
<td>21-May</td>
<td></td>
<td>30</td>
<td>6-Jun</td>
<td></td>
</tr>
<tr>
<td>Average</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>33.2</td>
<td>33</td>
<td></td>
</tr>
</tbody>
</table>

The autocorrelation was conducted using 1-week, 2-week, and 4-week lag intervals for all sites from July 1<sup>st</sup> through October 31<sup>st</sup> for 2005, 2006, and 2007. T-test results suggest correlations which exceed 0.2475 can be considered significantly positive, and those which fall below -0.2475 can be considered significantly negative. Selbyville and Viola were excluded due to their lack of data during these time periods. The results are graphically shown in Figures 4.19 to 4.25 and tabularly in Tables 4.19 to 4.25 for each station.

Generally the 1-week lags provide the most significant positive autocorrelations. This implies that the amount of moisture within the soil is not likely
to change much over the course of one week. However, since evapotranspiration is more prevalent in the summer months the 1-week autocorrelations are still subject to much fluctuation. The 2-week lags are typically not as significant as the 1-week lags, and border between being significantly positive and completely random. The 4-week lags typically are not significant, and the negative values all border on being significantly negative and being completely random.

Each of the three years (2005, 2006, and 2007) from August into September experienced significantly different rain events. August to September 2005 was very dry, August to September 2006 was very wet, and August into September 2007 was near normal. The occurrence of precipitation varies from site to site as well, but the more consistent the weather is the more significant is the correlation of soil moisture over time.
Figure 4.19. Autocorrelation results for Newark, DE from July 1 – October 31.

Table 4.19. Autocorrelation results for Newark, DE from July 1 – October 31.
Figure 4.20. Autocorrelation results for Dover, DE from July 1 – October 31.

<table>
<thead>
<tr>
<th>Dover</th>
<th>Lag (1-week)</th>
<th>Lag (2-weeks)</th>
<th>Lag (4-weeks)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2005</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
</tr>
<tr>
<td>2006</td>
<td>0.65</td>
<td>0.37</td>
<td>-0.07</td>
</tr>
<tr>
<td>2007</td>
<td>0.12</td>
<td>0.01</td>
<td>-0.13</td>
</tr>
</tbody>
</table>

Table 4.20. Autocorrelation results for Dover, DE from July 1 – October 31.
Figure 4.21. Autocorrelation results for Fair Hill, MD from July 1 – October 31.

<table>
<thead>
<tr>
<th>Fair Hill, MD</th>
<th>Lag (1-week)</th>
<th>Lag (2-weeks)</th>
<th>Lag (4-weeks)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2005</td>
<td>0.53</td>
<td>0.29</td>
<td>-0.56</td>
</tr>
<tr>
<td>2006</td>
<td>0.51</td>
<td>0.19</td>
<td>-0.09</td>
</tr>
<tr>
<td>2007</td>
<td>0.31</td>
<td>0.48</td>
<td>-0.42</td>
</tr>
</tbody>
</table>

Table 4.21. Autocorrelation results for Fair Hill, MD from July 1 – October 31.
Figure 4.22. Autocorrelation results for Glasgow, DE from July 1 – October 31.

<table>
<thead>
<tr>
<th>Glasgow</th>
<th>Lag (1-week)</th>
<th>Lag (2-weeks)</th>
<th>Lag (4-weeks)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2005</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
</tr>
<tr>
<td>2006</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
</tr>
<tr>
<td>2007</td>
<td>0.15</td>
<td>0.06</td>
<td>-0.26</td>
</tr>
</tbody>
</table>

Table 4.22. Autocorrelation results for Glasgow, DE from July 1 – October 31.
Figure 4.23. Autocorrelation results for Harrington, DE from July 1 – October 31.

Table 4.23. Autocorrelation results for Harrington, DE from July 1 – October 31.
Figure 4.24. Autocorrelation results for Laurel, DE from July 1 – October 31.

<table>
<thead>
<tr>
<th>Laurel</th>
<th>Lag (1-week)</th>
<th>Lag (2-weeks)</th>
<th>Lag (4-weeks)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2005</td>
<td>0.66</td>
<td>0.46</td>
<td>-0.28</td>
</tr>
<tr>
<td>2006</td>
<td>0.49</td>
<td>0.28</td>
<td>-0.18</td>
</tr>
<tr>
<td>2007</td>
<td>0.32</td>
<td>-0.05</td>
<td>-0.11</td>
</tr>
</tbody>
</table>

Table 4.24. Autocorrelation results for Laurel, DE from July 1 – October 31.
Figure 4.25. Autocorrelation results for Wilmington, DE from July 1 – October 31.

<table>
<thead>
<tr>
<th>Wilmington</th>
<th>Lag (1-week)</th>
<th>Lag (2-weeks)</th>
<th>Lag (4-weeks)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2005</td>
<td>0.50</td>
<td>0.44</td>
<td>-0.16</td>
</tr>
<tr>
<td>2006</td>
<td>0.71</td>
<td>0.32</td>
<td>-0.23</td>
</tr>
<tr>
<td>2007</td>
<td>0.35</td>
<td>0.35</td>
<td>-0.05</td>
</tr>
</tbody>
</table>

Table 4.25. Autocorrelation results for Wilmington, DE from July 1 – October 31.
4.2 Soil Moisture Conditions Derived from Remote Sensing Data

The vegetation indices (Figures 4.26 – 4.28) utilize NASA’s Enhanced Vegetation Index algorithm (EVI), which is based on the same Normalized Difference Vegetation Index (NDVI) scale of -1 to +1 that represent the “greenness” of the vegetation at each site and across Delmarva. The closer the value is to +1 the healthier the vegetation is providing an indication of adequate soil moisture for vegetation. Healthy green vegetation is represented by values greater than 0.3 on this scale. These maps represent the overall vegetation greenness during a 16-day measurement cycle.

The surface temperature maps (Figures 4.29 – 4.31) represent the average daytime surface temperature over the course of an 8-day measurement cycle. Surface temperature is an indication of soil moisture content due to differences in the heat storing ability of a dry soil parcel versus a saturated soil parcel. The saturated parcel will be more resistant to temperature change and therefore should change temperature more slowly than a dry soil parcel which will fluctuate a great deal between day and night. Most vegetated soils will transpire at their potential rate given adequate temperatures and will cool near the surface (evaporative cooling). As these temperatures were all measured in the daytime the dry soil parcels should stand out as significantly warmer than the more moist and saturated soil parcels across the map.

The AMSR-E microwave derived soil moisture maps (Figures 4.32 – 4.34) represent the amount of soil moisture present in the top 1cm of soil at the land surface.
These values are much lower than the DEOS derived soil moisture values due to the variability in fluxes that occur at the land surface (atmosphere interface). Incoming solar radiation directly influences the soil evaporation and plant transpiration rates at the land surface (i.e., at the 1cm observation depth). The DEOS observations are made at a 30cm depth, which are influenced by the conduction of heat and not the latent and sensible heat fluxes occurring at the surface. The change in soil moisture conditions between the 1cm and 30cm layer depths provide insight into the way soil moisture varies with depth.
Figure 4.26. Vegetation Index (EVI) – “Drought” (August 29 – September 13, 2005).
Figure 4.27. Vegetation Index (EVI) – “Heavy Rain” (August 29 – September 13, 2006).
Figure 4.28. Vegetation Index (EVI) – “Rain” (August 13 – 28, 2007).
Figure 4.29. Surface Temperature (Kelvin) – “Drought” (August 29 – September 5, 2005).
Figure 4.30. Surface Temperature (Kelvin) – “Heavy Rain” (August 29 – September 5, 2006).
Figure 4.31. Surface Temperature (Kelvin) – “Rain” (August 21 – 28, 2007).
Figure 4.32. Microwave (Soil Moisture) – “Drought” (September 5, 2005).
Figure 4.33. Microwave (Soil Moisture) – “Heavy Rain” (September 2, 2006).
Figure 4.34. Microwave (Soil Moisture) – “Rain” (August 22, 2007).
DEOS-Remote Sensing Soil Moisture Comparison Table

Table 4.26 is a comparison of all pertinent data measured by DEOS and remote sensing derived soil moisture conditions from 2005-2007 for all sites. All DEOS data (average daily air temperature, max daily vwc, and daily rainfall) were recorded over a one-day time span each year, as were the remotely sensed soil moisture microwave data. Surface temperature was recorded over an 8-day time span, and EVI was recorded over a 16-day time span. Refer to Table 4.27 below for a complete list of the dates.

The main focus of this discussion centers around the DEOS-derived max daily %vwc and the passive microwave soil moisture data. While the DEOS sensor measures soil moisture at a 30cm depth, the AMSR-E soil moisture sensor is only capable of measuring the top 1cm of soil. The objective is not to determine the accuracy of one measuring technique over the other, but instead it is to create a holistic look at Delmarva soil moisture. Notice the very large range in data, which depicts the soil as being very dry at the surface yet moist at the 30cm depth.

The remaining data describes the conditions at each site at the time of the AMSR-E soil moisture estimates. Daily rainfall is the primary contributor to the increase in soil moisture, with air and surface temperatures acting primarily to decrease the amount of soil moisture through increased evapotranspiration, especially at the surface. Below the surface as moisture enters the root zone, it is the vegetation and percolation which most effectively remove moisture from the soil (held against the pull of gravity, while surface runoff and percolation partition excess moisture).
The conditions preceding the dates of observation are examined in greatest detail for the comparison. Typically a 2-week dry spell will yield low soil moisture values, even after a rain event. The opposite may not be true as the overall conditions surrounding each site play a major role in the ability of soil to retain its moisture. Table 4.26 lists the conditions at each site at the time of the soil moisture observations.
Table 4.26. Comparison of remotely sensed observations with DEOS.

<table>
<thead>
<tr>
<th></th>
<th>2005 (Drought) DEOS</th>
<th>2005 (Drought) RS</th>
<th>2006 (Heavy Rain) DEOS</th>
<th>2006 (Heavy Rain) RS</th>
<th>2007 (Rain) DEOS</th>
<th>2007 (Rain) RS</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Avg. Daily Air Temp (°C)</td>
<td>Max Daily VWC</td>
<td>Daily Rainfall (mm)</td>
<td>Max Daily VWC (%)</td>
<td>Avg. Daily Air Temp (°C)</td>
<td>Max Daily VWC</td>
</tr>
<tr>
<td>Newark</td>
<td>22.0</td>
<td>0.23</td>
<td>0.0</td>
<td>22.6</td>
<td>305.3</td>
<td>32.1</td>
</tr>
<tr>
<td>Dover</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>300.9</td>
<td>27.7</td>
</tr>
<tr>
<td>Fair Hill</td>
<td>19.0</td>
<td>0.24</td>
<td>0.0</td>
<td>23.7</td>
<td>300.0</td>
<td>30.6</td>
</tr>
<tr>
<td>Glasgow</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>299.2</td>
<td>26.1</td>
</tr>
<tr>
<td>Harrington</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>298.4</td>
<td>23.3</td>
</tr>
<tr>
<td>Laurel</td>
<td>19.6</td>
<td>0.05</td>
<td>0.0</td>
<td>4.5</td>
<td>297.8</td>
<td>22.3</td>
</tr>
<tr>
<td>Wilmington</td>
<td>22.0</td>
<td>0.09</td>
<td>0.0</td>
<td>9.2</td>
<td>294.8</td>
<td>21.0</td>
</tr>
<tr>
<td>Selbyville</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>304.4</td>
<td>31.4</td>
</tr>
<tr>
<td>Viola</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>306.2</td>
<td>29.7</td>
</tr>
</tbody>
</table>
Table 4.27. Notes for DEOS-Remote Sensing comparison table (Table 4.26).

<table>
<thead>
<tr>
<th>Notes:</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>ND</strong> = No Data</td>
</tr>
</tbody>
</table>

2005 = 9/5/2005
2006 = 9/2/2006
2007 = 8/22/2007

Surface Temperature was recorded on the following dates:
Aug 29 - Sep 5, 2005
Aug 29 - Sep 5, 2006
August 21 - 28, 2007

EVI values were recorded on the following dates:
Aug 29 - Sept 13, 2005
Aug 29 - Sept 13, 2006
Chapter 5

DISCUSSION OF THE SOIL MOISTURE CONDITIONS ACROSS DELMARVA

The temporal variability of soil moisture across the Delmarva Peninsula was first quantified through 5-minute DEOS observations (primarily volumetric water content and daily rainfall) from January 1, 2005 through June 30, 2008. It is expected that the volumetric water content (vwc) would spike during periods of rainfall and gradually diminish during periods of drought. The most prevalent factor affecting this trend were the soil conditions preceding rain events. The more time that passes without rain, especially during the summer months, the more rain is required to increase the vwc. A gradual decrease during dry spells was observed at most stations with a rapid jump in replenishing soil moisture in the fall.

In general, the average vwc at all sites for August 2005-2007 was roughly 50-60% the estimated field capacity, with a range of 32% to 80% for individual site averages. However, most of the sites demonstrated consistency in vwc averages for August, adjusting accordingly to each rainfall event. All sites displayed a higher vwc during the “rain” period in 2007 than during the “heavy rain” period in 2006. This is likely the result of long-term precipitation accumulation in the soil in 2007.

An annual vwc average was calculated for each site, derived from the averaging of all monthly averages from 2005-2008 into one average annual vwc value per site (Tables 4.1 through 4.9). This value, when compared against the estimated field capacity, shows that the persistence of soil moisture from 2005-2007 equates to
roughly 88% of the field capacity over 7 of the 9 sites (Selbyville and Viola were excluded due to insufficient data) with the annual average vwc for all sites ranging from approximately 81% to 98% of their estimated field capacities.

Additionally, the vwc content at each site tended to be higher during the colder winter months and lower during the warmer summer months. Because of the reduced incoming solar radiation caused by shorter days and dormant state of the majority of vegetation, the soil is able to maintain moisture content more easily during the cold season. As the seasons change and summer approaches, due to the longer days, increasing incoming solar energy, and the growth of vegetation, soil moisture is more likely to be absorbed by plant roots or transpired into the atmosphere.

There is a sharp contrast between soil moisture values derived from AMSR-E data and those taken from DEOS ground observations. The AMSR-E soil moisture (g/cm³) data compared to estimated field capacity showed all sites averaging approximately 38% of total estimated field capacity, with a range of 27% to 75% (it should be noted that the next highest value was 39%). This discrepancy exists between DEOS and AMSR-E soil moisture values due to the soil depth at which each was observed. Because AMSR-E microwave observations are only able to penetrate the top 1cm of soil, values are expected to be much lower than those of DEOS which measure a soil depth of 30cm. However, this discrepancy offers an excellent cross-section of the top 30cm of soil, with the top 1cm holding approximately 38% of the total estimated field capacity and the bottom 30cm holding nearly 88%, on average throughout the year. These percentages undoubtedly vary over time, but it is clear that
the top layer of soil is unable to store or maintain the same quantity of moisture as the deep layers of soil. This can be attributed to the fact that the land surface serves as the contact point between the surface and the atmosphere, where both radiative and non-radiation fluxes occur. The top surface is subject to soil moisture losses due to soil evaporation and transpiration while soil moisture at the 30cm depth is only subject to the transpiration by plants.

The autocorrelation results have shown that more than 82% of the time the 1-week lag produced a significantly positive correlation, with 14 out of 17 vwc measurements proving significant. The 2-week lag produced a significantly positive correlation more than 58% of the time, with 10 out of 17 comparisons proving significant. The 4-week lag produced significantly negative correlations less than 36% of the time, with no significant positive correlations occurring. The 1-week and 2-week lags produced no significant negative correlations.

The persistence of soil moisture is very strong over 1-week intervals, and substantially strong over 2-week intervals. By 3 weeks the persistence of soil moisture is likely to become more random with a mix of positive and negative correlations. By 4 weeks the persistence is completely random. Negative autocorrelations do not require as much consideration as positive autocorrelations, simply because negative autocorrelations very seldom occur. Transition periods in weather produce negative correlations such as when a large rain event ends a period of dry conditions. In areas without large fluctuations in weather soil moisture levels
should become more persistent. In general, the shorter the time-lag, the more persistent the soil moisture will be.

Newark

The comparison of data from the EVI with daily rainfall data measured by DEOS shows that the weeks leading up to the EVI measurement determine the health of the vegetation. Delaware underwent a 16-day drought from August 29 – September 13, 2005, during which time Newark maintained an average air temperature of 23° C and on September 5, 2005 yielded a vwc of 23%. During this period, from August 29 – September 13, 2005, the EVI was recorded at a healthy 0.52 for Newark, DE. In the 4 weeks preceding this drought, from August 1 – 28, 2005, the total rainfall measured 46.1mm. The EVI of 0.52 and vwc of 23% after weeks of reduced or non-existent rainfall demonstrates the persistence of soil moisture in Newark, DE (the vwc at Newark steadily declined during this time period, with %vwc initially starting in the mid to upper 20’s).

Similarly, despite 2006 being labeled as a period of heavy rain (151.9mm total rainfall from August 27 – September 9, 2006) the 4 weeks (29 days) preceding this rainfall experienced only 0.8mm total rainfall. The vwc fell from 30% on July 29, 2006 to 21% on August 26, 2006. At the time of heavy rainfall an EVI of 0.41 was recorded (August 29 – September 13, 2006) with the vwc replenishing to 39% on September 2, 2006. Soil moisture in the top 1cm was measured by the AMSR-E
sensor at 0.087 g/cm$^3$ on September 2, 2006, which is likely low due to the rapid infiltration of water into the soil depths brought on by weeks of drought.

August 2007 saw a steady influx of rain occurring at bi-weekly intervals. 97mm of rain fell during the entire month of August, all of it reaching the ground sometime between August 6 – 26, 2007. During this time period the vwc averaged 27%, with a recorded 30% on August 22, 2007. The EVI registered 0.38 from August 13 – 28, 2007, with daytime surface temperatures averaging 33° C from August 21 – 28, 2007.

Dover

Prior to the heavy rainfall, Dover experienced a 23-day drought in August 2006 in which only 4.5mm of rain fell from August 9 – 31, 2006. September 1 – 6, 2006 saw 123.9mm of rain, with a recorded vwc of 21% on September 2, 2006 and an overall rise from 11% to 22% from September 1 – 6, 2006. Soil moisture in the top 1cm of soil measured 0.095 g/cm$^3$, likely so low due to the preceding drought conditions. The EVI was recorded at 0.44 from August 29 – September 13, 2006, with the daytime surface temperature averaging 25° C from August 29 – September 5, 2006.

While August 2007 wasn’t exactly a drought, although it was very close supplying only 58.8mm of rainfall from August 1 – 21, 2007. During this time period the vwc averaged only 8%, and on August 22, 2007, the day following 47.2mm of rainfall (from August 19 – 22), the vwc was measured at a low 12%. The EVI
registered 0.57 for August 13 – 28, 2007, with daytime surface temperatures averaging 29° C from August 21 – 28, 2007. Soil moisture in the top 1cm of soil measured 0.096 g/cm³ on August 22, with the average air temperature reaching 18° C.

**Fair Hill**

Throughout the month of August 2005, Fair Hill experienced a total of 68.5mm of rainfall. By September 5, 2005, Fair Hill was in the midst of an 8-day (and counting) dry period. The vwc was measured by DEOS at 24% with an AMSR-E measured soil moisture of 0.117 g/cm³ in the top 1cm of soil. The daytime surface temperature averaged 27° C from August 29 – September 5, 2005, with an average air temperature of 19° C measured on September 5, 2005. The EVI was recorded at 0.63 from August 29 – September 13, 2005, a period marked by dry conditions (0mm rain during this time) with an average vwc of 24% ranging from 26% on August 29 to 22% on September 9, 2005.

During August 2006, a short period of heavy rain took place at the time of remote sensing observations. In the two weeks leading up to September 2, 2006, a one-day rainfall event of 55.8mm took place, with the vwc at Fair Hill fluctuating slightly from 23% on August 19, 2006 to 25% on September 1, 2006 with the lowest recorded vwc reaching 21% on August 26, 2006. On September 2, the vwc spiked up to 42% in response to the rainfall. The EVI was recorded at 0.58 from August 29 –
September 13, 2006 with soil moisture in the top 1cm of soil measuring 0.103 g/cm³ on September 2, 2006. The daytime surface temperature from August 29 – September 5, 2006 averaged 24° C with an average air temperature of 18° C measured on September 2, 2006.

August 2007 saw a steady influx of rain occurring almost daily. 101.1mm of rain fell during the entire month of August, with the majority of it reaching the ground sometime between August 5th and 22nd of 2007. During this time period the vwc averaged 27%, with a recorded 38% on August 22, 2007. The EVI registered 0.62 from August 13 – 28, 2007, with daytime surface temperatures averaging 28° C from August 21 – 28, 2007. Soil moisture in the top 1cm of soil measured 0.105 g/cm³ on August 22, 2007.

Glasgow

Harrington

Harrington mostly found itself with dry conditions throughout August 2007, with rainfall amounts totaling 48.4mm from August 1 – 31, 2007. The vwc averaged 9% from August 1 - 20, 2007, before it jumped slightly to 14% on August 22, 2007 due to a rain event lasting from August 19 – 22, 2007, producing 39.1mm of total rainfall. The daytime surface temperature averaged 31° C from August 21 – 28, 2007, with an EVI of 0.30 measured from August 13 – 28, 2007. Soil moisture in the top 1cm of soil measured 0.109 g/cm³ on August 22, 2007, with an average air temperature of 19°C.

Laurel

Throughout August 2005 (drought), Laurel experienced only 0.3mm of total rainfall. By September 5, 2005, Laurel had undergone a 29-day drought that did not end until September 18, 2005 when only 0.5mm of rain fell. The vwc was measured by DEOS at 5% with an AMSR-E measured soil moisture of 0.138 g/cm³ in the top 1cm of soil. The daytime surface temperature averaged 31° C from August 29 – September 5, 2005, with an average air temperature of 20° C measured on September 5, 2005. The EVI was recorded at 0.38 from August 29 – September 13, 2005. The average vwc from August 5, 2005 – September 5, 2005 was 7%.

August 2, 2006 – September 2, 2006 (heavy rain event) had a total recorded rainfall of 130.8mm, with 109.2mm of it falling between August 29, 2006 and September 2, 2006. 66.6mm of rain fell on September 2, 2006 which yielded a vwc
recording of 25%, jumping from 6% which was the average vwc observed from August 2, 2006 – September 2, 2006. The EVI was recorded at 0.24 from August 29 – September 13, 2006, with a soil moisture measurement of 0.106 g/cm³ in the top 1cm of soil on September 2, 2006. The average daytime surface temperature was 23°C from August 29 – September 5, 2006.

Laurel experienced only mild amounts of rainfall in August 2007 (normal rain event), with a total rainfall amount for the entire month of 97.4mm. Most of the rain fell between August 20 – 22, 2007, totaling 59.5mm during those 3 days. The average vwc for the month was 6%, with a vwc of 12% recorded on August 22, 2007. The EVI registered 0.39 from August 13 – 28, 2007, with daytime surface temperatures averaging 29° C from August 21 – 28, 2007. Soil moisture in the top 1cm of soil measured 0.102 g/cm³ on August 22, 2007, with an average air temperature of 20° C.

Wilmington

From August 5, 2005 through September 5, 2005 (drought) Wilmington experienced a total of 50.2mm of rainfall. By September 5, 2005, Wilmington was in the middle of an 8-day dry spell which would go on to last another 8 days (August 29 – September 13, 2005). The vwc was measured by DEOS at 9% on September 5, 2005 with an AMSR-E derived soil moisture of 0.097 g/cm³ in the top 1cm of soil. The daytime surface temperature averaged 31° C from August 29 – September 5, 2005, with an average air temperature of 22° C also measured on September 5, 2005. The EVI was recorded at 0.46 from August 29 – September 13, 2005.
Once again August 2006 (heavy rain event) was marked by a long period of drought before finally giving way to substantial rainfall at the very end of the month. A total of 67.1mm of rain fell during the entire month of August, with the majority of it falling from August 27 – 31, 2006. From August 27 – September 6, 2006 a total of 183.5mm of rain fell onto Wilmington causing a rise in vwc from 14% on August 27th to 39% on September 6th of 2006. 46.8mm of rain fell on September 2, 2006, yielding a vwc of 34%. The EVI was recorded at 0.43 from August 29 – September 13, 2006 with soil moisture in the top 1cm of soil measuring 0.106 g/cm³ on September 2, 2006. The daytime surface temperature from August 29 – September 5, 2006 averaged 25° C with an average air temperature of 18° C measured on September 2, 2006.

Wilmington experienced sporadic amounts of rainfall in August 2007 (normal rain event), totaling 115.1mm for the whole month. Much of the rain fell between August 19 – 22, 2007, totaling 56.8mm during those 4 days. The average vwc for the month was 20%, with a vwc of 28% recorded on August 22, 2007. The EVI registered 0.44 from August 13 – 28, 2007, with daytime surface temperatures averaging 30° C from August 21 – 28, 2007. Soil moisture in the top 1cm of soil measured 0.101 g/cm³ on August 22, 2007, with an average air temperature of 1° C.
Selbyville

EVI data for Selbyville suggests moderate overall vegetation health from August 29 – September 13, 2005 (drought) with a recorded value of 0.46. The average daytime surface temperature was measured at a moderate 28° C from August 29 – September 5, 2005, with soil moisture in the top 1cm of soil measured at a relatively low 0.096 g/cm³ on September 5, 2005. Vegetation health was measured at 0.47 using EVI from August 29 – September 13, 2006 (heavy rain). The average daytime surface temperature was measured at a low 23° C from August 29 – September 5, 2006, with soil moisture in the top 1cm of soil measured at a low 0.083 g/cm³ on September 2, 2006. For 2007 (normal rain event), overall vegetation health had improved to 0.60 as measured using the EVI from August 13 – 28, 2007. The average daytime surface temperature was measured at a moderate 30° C from August 21 – August 28, 2007, with soil moisture in the top 1cm of soil measured at a moderately low 0.095 g/cm³ on August 22, 2007.

Viola

Remote sensing data for Viola suggests excellent overall vegetation health from August 29 – September 13, 2005 (drought) with a recorded EVI of 0.71. The average daytime surface temperature was measured at a low 26° C from August 29 – September 5, 2005, with soil moisture in the top 1cm of soil measured at a low 0.092 g/cm³ on September 5, 2005. Vegetation health had dropped to a moderate 0.42 using EVI from August 29 – September 13, 2006 (heavy rain). The average daytime
surface temperature was measured at a low 24° C from August 29 – September 5, 2006, with soil moisture in the top 1cm of soil measured at a relatively moderate 0.95 g/cm³ on September 2, 2006. For 2007 (normal rain event), overall vegetation health had not changed much compared to the previous year as it was measured at 0.44 using the EVI from August 13 – 28, 2007. The average daytime surface temperature was measured at a relatively high 31° C from August 21 – August 28, 2007, with soil moisture in the top 1cm of soil measured at a moderately low 0.096 g/cm³ on August 22, 2007.

Analysis

The analysis performed at each site is meant to compare the conditions preceding each remote sensing measurement with the conditions at the time the measurements were taking place. Not all of the same data were included at each site, only those datasets with the greatest impact, both directly and indirectly, on the soil moisture conditions. Precipitation events preceding each remote sensing period (i.e., EVI) seemed to have the greatest impact on observed vegetation health. The land surface temperatures were reflective of precipitation events and soil moisture conditions at the time of observation. Additionally, the AMSR-E soil moisture observations were influenced the least by rain events preceding the one day measurement. This is because the moisture at the soil surface evaporates or percolates very rapidly due to the influx of radiative heat from the sun, and the pull of gravity.
Chapter 6

CONCLUSION

The overall trend in soil moisture fluctuations across the Delmarva Peninsula is closely related to the natural seasonal cycles and weather patterns occurring across the region. Soil moisture values vary significantly from place-to-place, though the increase and decrease in the amount of moisture present within the soil follows a unimodal curve with maximum and minimum levels occurring in the winter and summer, respectively. The soil moisture responds directly to rainfall with moisture reserves spiking during extreme precipitation events (such as the heavy rain events in August 2006). A gradual steady decline is also clearly evident when evapotranspiration demands increase and precipitation is limited.

The unimodal curve associated with annual soil moisture cycles is significant because it represents the relationship between soil moisture and evapotranspiration, whose rate is dependent on both air and ground surface temperatures. The high points in the curve are expected to occur during the winter months in which the air and ground temperatures are lower, the days are shorter, and vegetation is in a dormant state. This is when the soil moisture is much easier to maintain allowing for higher and more consistent vwc levels. The low points in the curve are expected to occur during the summer months in which the air and ground temperatures are at their annual peak, the days are longer, and vegetation is most active. During this time period soil moisture is constantly being lost due to soil evaporation, and plant transpiration.
Annual soil moisture cycles can vary depending on the severity, frequency, and duration of rain events, as well as overall air temperatures. Other factors may also influence annual soil moisture variability, such as local topography and soil composition and texture. The greenness and overall health of vegetation appears to be dependent on the severity, frequency, and duration events occurring in the week leading up to the observational periods. Drought conditions occurring in the weeks leading up to the remotely sensed vegetation measurements tend to produce less healthy vegetation even if a significant amount of rain falls during the 16-day observational time period (such as during August 2007 in Newark which experienced lower EVI values despite steady rainfall throughout the month). Additionally, when there is heavy rainfall in the weeks leading up to the 16-day observational period the vegetation appears to be green and healthy even when drought conditions persist during the period of observation.

According to the autocorrelation results the soil moisture at each of the 9 sites are nearly equally persistent over time. Soil texture plays a major role in each site’s ability to maintain moisture in the absence of regular and consistent rain events, as does land cover and the extent of the heterogeneity within each soil matrix. Persistence was also significantly greater during dry spells, when the volumetric water content of the soil gradually decreased over the course of days and weeks. In general, soil moisture conditions tended to persist for up to 2 weeks between August 1st and September 30th. That is, soil moisture conditions during this time tended to remain the
same or similar in spans of 1- to 2-week intervals. By 3 or 4 weeks, soil moisture conditions were much more random.

While the results of this study are based on data taken from only a 3.5 year study, a larger study period of a minimum of 5 years (preferably 10 years or more) would generate an even greater level of variability over the course of many more seasonal cycles, thereby offering a more substantial and definitive conclusion on soil moisture across the Delmarva Region. The number of monitoring stations used in this study for the most part proved to be spatially adequate in representing the northern half of the Delmarva Peninsula. An increase in the number of stations to the west, and especially to the south, of Delaware is recommended in order to spatially represent the entire Delmarva Peninsula.

The use of two microwave-based satellite sensors would be useful in observing the soil surface at varying depths. As was stated earlier, a microwave frequency of ~1.4 GHz is capable of penetrating the top soil to a depth of 5cm, which would be an excellent addition to the 6.9 GHz observational frequency used by AMSR-E. Using a 1cm and a 5cm observation depth, it would be possible to estimate more effectively the amount of moisture within the soil as it varies with depth. A temporal resolution of less than 1 week for the satellite data (EVI and LST) would also be ideal. Additionally, the number of AMSR-E observations should be increased during periods of rainfall and dry spells in order to capture in more detail the temporal variability of soil moisture at the land surface (i.e., top 1cm of soil). Because the land surface is so dynamic, it is important to be able to capture the rate and extent to which it stores and
transfers soil moisture. A climate model may allow for a better overall comparison of DEOS observations at the 30cm depth to the 1cm AMSR-E observations. Additional remotely sensed soil moisture images of the top 1cm could potential see what is controlling soil moisture at the surface.

For future studies, an increase in the number of DEOS stations within Delaware could possibly reveal even greater variability. The proximity of sites should also be taken into account, as they can be clustered to more easily identify local variables, spread out over a wide range to observe greater overall variability across the Delmarva Peninsula, or a combination of both. A more in-depth study of the soil properties at each site would help to answer some of the questions surrounding the differences evident from the average annual soil moisture cycles at each site. Adding additional years of data would also provide a better idea about the average annual soil moisture conditions at each site. Glasgow, Selbyville and Viola specifically would benefit from the use of additional data.
### Appendix A - MAXIMUM DAILY VOLUMETRIC WATER CONTENT

**Maximum Daily Volumetric Water Content (%) - NEWARK**

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*Average: 23.58, 25.16, 22.50, 19.33, 20.22, 15.32*
## Maximum Daily Volumetric Water Content (%)

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**Average**: 36.14, 34.74, 31.44, 30.01, 25.22, 30.76, 32.51, 18.85, 33.36, 35.52, 36.70, 34.50
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**Average:** 48.76 | 50.75 | 48.53 | 40.78 | 38.51 | 28.75
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*Average:* 31.07 33.18 30.37 28.94 28.14 23.10

163
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**Average**: 30.82, 31.26, 28.63, 25.69, 28.05, 23.49
## Appendix B - DEOS AVERAGE DAILY AIR TEMPERATURE DATA

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**Monthly Mean**

|          | -0.05 | 1.60 | 3.61 | 12.49 | 14.63 | 22.96 | 25.16 | 24.82 | 21.81 | 13.82 | 8.52 | 0.40 |

165
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**Monthly Mean**: 4.40, -1.70, 7.07, 10.21, 17.61, 22.44, 24.04, 24.05, 20.72, 17.20, 7.60, 3.39
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180
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**Monthly Mean**: 4.86, -0.96, 7.68, 10.89, 17.78, 22.41, 24.48, 24.52, 20.65, 17.59, 8.07, 4.39
### Average Daily Air Temperature (°C) – LAUREL

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**Average Daily Air Temperature (°C) – WILMINGTON**

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- October: 6.1
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## Daily Rainfall (mm) – GLASGOW

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**Monthly Mean**: 3.9, 1.7, 0.1, 2.4, 0.0, 3.3, 3.4, 0.9, 5.9, 5.0, 5.1, 1.8
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**Monthly Mean**: 3.9, 2.3, 3.0, 4.7, 1.0, 2.5, 1.2, 3.1, 0.0, 3.5, 1.4, 3.8
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126
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