Piezocone Mapping, Groundwater Monitoring, and Flow Modeling in a Riverine Peatland: Implications for the Transport of Arsenic

by
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Submitted to the Department of Civil and Environmental Engineering in Partial Fulfillment of the Requirements for the Degree of Doctor of Philosophy in Civil and Environmental Engineering at the Massachusetts Institute of Technology

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Abstract

In the Aberjona Watershed of Woburn, Massachusetts, deposit-scale features of a contaminated riverine wetland are linked to the processes that formed the wetland as well as those that influence the interaction between surface water and wetland groundwater. The Aberjona River carried large quantities of arsenic starting as early as the late 1800s. The fate of riverborne arsenic within this wetland is of particular interest due to the former presence of municipal supply wells G & H, which are believed to have captured contaminated river infiltration between 1964 and 1979. Serious health effects have been attributed to exposure to water pumped from these wells, although the causative agent or agents have not been determined.

Detailed stratigraphic mapping of the wetland that separates the Aberjona River at its closest approach to Well H was conducted using a new wetland piezocone penetrometer. Maps revealed an ice-block depression filled with diatomaceous silt and overlain by several distinct peat strata, including sand layers from 5 to 150 cm in thickness. The post-glacial history of the wetland, further defined using pollen abundance and radiocarbon dates, includes a period of fresh water lacustrine deposition from 14,000 to 9,000 ybp, followed by the evolution from woody swamp to sedge meadow to cattail marsh. The distribution of arsenic in this soil profile indicates that arsenic both deposited on the former wetland surface in a sedimentary form, and some arsenic was transported to depth in groundwater. The influx of arsenic starting in the later 1800s may have been responsible for the decline of tussock sedge and the transition to cattails that corresponds with this horizon.

The mobilization of arsenic from the streambed to a sand layer within the wetland was observed during a dry period of river discharge. Up to 600 μg/l total arsenic was measured in porewater. The mechanisms for delivering and trapping of arsenic in wetland soils, and the potential for its mobilization and transport, are explored using soil
and groundwater arsenic data, a synthesis of the available data on the stability of arsenic minerals, and the simulation of groundwater flow within the wetland under various conditions. Results indicate that the arsenic contained in the shallow peat at the site is relatively immobile, whereas the arsenic contained in streambed sediments could be an important source to the wetland itself and to pumping wells located adjacent to the wetland. It is also likely that municipal pumping gradients were the driving force for delivery of arsenic to the lower peat by groundwater flow. The studied section of wetland carries a small share of the total expected induced infiltration due to pumping, indicating the existence of adjacent areas where the streambed is more directly connected to the underlying aquifer. Under the appropriate chemical conditions, arsenic mobilized from such areas could be transported to the wells.

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1. **Introduction**

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1.1. **Background**

Wetlands connect groundwater to surface water; this strategic hydrologic position makes them critical to the understanding of pollutant transport processes. Wetlands in discharge areas can be thought of both as sensitive receptors of groundwater pollution or as sophisticated treatment systems that confound our efforts to understand the natural groundwater flushing process. In recharge areas, physical and chemical processes occurring in wetlands influence the quality of surface waters that recharge the aquifer and may be pumped from nearby wells. Recently, increasing concern over the quality of groundwater has focused national and global attention on the critical function of wetlands in the hydrologic cycle and the threat posed to drinking water supplies by continued development and destruction of wetland areas (Mitchell, 1992).

Watershed scale studies typically model wetland peat deposits using bulk hydraulic and chemical parameters. However, the internal structure of such deposits can offer clues as to their origin, and can also have profound effects on the flow of groundwater and transport of solutes. In a wetland, subsurface variability can regulate the exchange of pollutants between surface and groundwater.

Wells G & H Peatland in the Aberjona Watershed of Woburn Massachusetts is not unlike other wetlands that form along the margins of rivers draining buried valley deposits, in terms of both its appearance and its largely unknown stratigraphy. At this site, the fate of pollutants and the interaction of the Aberjona River with the Aberjona Valley aquifer has particular import. The City of Woburn, desperate for additional water supplies and reluctant to buy water from the Metropolitan District Commission, first explored this aquifer in the 1960s. Industrial process water had been withdrawn from this aquifer for many years for use by the J. J. Riley Tannery. Water pumped from test wells in the area
had high chloride concentrations, and the nearby Riley tannery was cited by officials as
the probable source (Tarr, 1987). Nevertheless, the water was deemed fit for municipal
use, and Wells G & H were installed in 1964 in sand and gravel deposits referred to as
some of the most transmissive in the greater Boston area (Delaney and Gay, 1980;
Halberg and Pree, 1949).

As early as 1967, bacterial contamination required continuous chlorination of the well
water (Tarr, 1987). The high salinity and occasional bacterial contamination of the
Wells G & H water were typical indicators of surface water infiltration. The high organic
content of the water led water supply engineers to suspect wetland recharge sources
(Whitman and Howard, 1968). Despite frequent complaints of strong odor, foul taste,
cloudiness, and other discoloration, Wells G and H were used most months of the year
from 1964 and 1967, respectively, except for several years (1967, 1972 - 1975) when use
was limited to months of high demand (Dufresne and Henry, 1978; Murphy, 1986). In
1977, an additional test well drilled between Wells G and H yielded water of similar
quality, with high levels of hardness, sodium, iron, manganese, sulfate, chloride,
ammonia, and organic content (Tarr, 1987). In 1979, the discovery of a number of
empty solvent drums near the Aberjona River upstream of Wells G & H led to the first
tests of the well water for volatile organic chemicals (VOCs). Elevated levels of
trichloroethylene (TCE), as well as several other VOCs, were found in the water and the
wells were shut down.

Between 1964 and 1983, 21 cases of childhood leukemia occurred in Woburn (Lagakos et
al., 1984). The claimed cause of these illnesses, TCE contamination, was the subject of a
lawsuit filed by the families of the victims against Grace Chemical and Beatrice Foods,
two industries located within what was to become the Wells G & H Superfund Site. The
eventual fate of TCE and other VOCs released to the aquifer by a number of local
industries has not been determined. The lawsuit was settled in 1987, without any clear
resolution of the various points of view regarding the fate of TCE and its potential health effects in drinking water.

Recently, a Massachusetts Department of Public Health study concluded that increased exposure to Wells G & H water *in utero* was positively correlated with an increased likelihood of contracting childhood leukemia (Costas et al., 1996). Thus it appears that Wells G & H water is likely to have caused health effects. However, many possible causative agents exist.

The generally poor quality of the Aberjona River was well-documented as early as the 1870s. Studies conducted at MIT since 1990 have shown that the Aberjona River was heavily contaminated with arsenic as well as chromium and several other toxic metals since the late 1800s, and continues to carry contaminated sediments from its headwaters to the Upper Mystic Lake (Durant et al., 1990; Knox, 1991; Aurilio, 1992; Solo, 1995; Spiethoff and Hemond, 1996).

Years of study and millions of dollars have been spent by industry and the public to determine the fate of pollutants in the Wells G & H aquifer, while the wetland that separates the Aberjona River from the aquifer below has remained largely unexplored. Composed at places of wetland soils more than 8 m thick, the Aberjona River riverbed was referred to in 1973 as “leaky”, and, based on a pumping test conducted by the U.S. Geological Survey in 1985, the wetland soils were described as “loose and permeable”, implying significant flow through this wetland (Myette et al., 1987; Tarr, 1987). Based on the same pumping test, it has been concluded that more than half of the well recharge was derived from river infiltration. These conclusions, as well as the original observations of poor water quality at the wellhead, indicate that river - aquifer interactions could be crucial to understanding past and future water quality trends.
1.2. **Purpose and Goals**

At the Wells G & H site in Woburn, catastrophic human illness has been statistically associated with the consumption of groundwater pumped from municipal wells. Water quality indicators typical of surface water contamination are well-documented. The pumped water is believed to be derived to a great extent from the nearby Aberjona River, which also has a long history of contamination, including arsenic. The hypothesis that riverborne arsenic was transported to the water supply wells and caused adverse health effects is a logical one, and extremely difficult to address comprehensively. This research focuses on exploring water flow and arsenic transport in this and similar wetlands, using both novel and traditional techniques. An integrative approach is used to address this and related issues, exploiting various disciplines, including glacial geology, hydrology, geotechnical engineering, paleolimnology, and environmental geochemistry.

At the Wells G & H Wetland, the issue of water supply protection lends particular import to our efforts. As will become clear, however, the underlying questions that are addressed in the pursuit of this goal also have intrinsic importance and general relevance to numerous other sites.

- What are the scales of heterogeneity in wetlands, and how can the internal structure of wetlands be efficiently and reliably mapped?

- What processes formed the features of the Wells G & H Wetland, and how did post-colonial activities impact wetland development?

- What is the distribution of arsenic in wetland soils, and how can it be explained in the context of wetland and industrial history?
• How does wetland structure influence wetland hydrology, and in turn the mobility of arsenic?

• What role has this wetland played in the transport and distribution of arsenic in the watershed, and what role is it likely to play in the future?

1.3. Scope

The topics that define the scope of this work include instrumentation design and testing, geologic and paleolimnological reconstruction, arsenic geochemistry, and wetland groundwater monitoring and hydrologic modeling.

Chapters 2 and 3 describe the development of two pieces of equipment designed especially for the detection of subsurface features in wetland deposits: a portable piezocone penetrometer driver and a highly sensitive piezocone. Chapter 4 is a detailed investigation of the stratigraphy of the Wells G & H Wetland, using soil core and piezocone data. A method for wetland piezocone data interpretation is presented, and several stratigraphic cross-sections are constructed from these data.

Chapter 5 addresses the geologic, vegetational, industrial, and geochemical history of the wetland. The post-glacial and industrial history of the wetland is reconstructed based on stratigraphy, soil arsenic measurements, paleoecological analysis using pollen abundance, radiocarbon dates, and historical information concerning the industrial development of the Aberjona River. The modes of arsenic transport into and storage within the wetland are explored based on a synthesis of current knowledge of arsenic geochemistry.

Chapter 6 is a hydrologic study that investigates groundwater movements and the mobilization of arsenic over time scales spanning from days to months. Data are
synthesized from the monitoring of hydraulic head at a number of locations as well as measurements of arsenic concentration and several other water quality parameters. The hydraulic parameters of wetland soil layers are measured in the field and in the lab, and inferred from the calibration of a two-dimensional cross section numerical model. Based on these data, an assessment is made of the current and past potential for arsenic mobility in wetland porewater, as well as the importance of spatial heterogeneities and transient flow.
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2.1. **Introduction**

Wetland deposits are often located in groundwater discharge or recharge zones where highly three-dimensional flow regimes exist. These flow regimes, along with the chemistry of wetland sediments and pore waters, may control the exchange of solutes, including toxic contaminants, between groundwater and surface waters. Where this occurs, deposit scale heterogeneities of soil permeability and chemical characteristics of the wetland are great significance to modeling the transport and fate of such solutes.

Traditionally, deposit scale heterogeneities are ignored in watershed-scale modeling of groundwater flow and solute transport, most often because discrimination of deposit-scale heterogeneities is extremely laborious by conventional means. Networks of borings spaced closely enough to detect such heterogeneities are prohibitively expensive and otherwise impractical; while typical *in situ* hydraulic tests necessarily average the effects of the surrounding media. Geophysical methods, while providing continuous and very broad coverage, do not appear as yet to possess the resolution or discrimination necessary to identify features that may dominate the hydraulics and chemistry of transport between surface and groundwaters. As a result, most watershed scale models use averaged bulk parameters to represent wetland deposits as single geologic and hydrologic units.

The piezocone penetrometer is in principle well-suited to the mapping of small scale heterogeneities in wetland soil type. With resolution approximating one centimeter and essentially continuous vertical coverage, it is capable of detecting thin lenses of contrasting soils extending only several probe diameters (5 - 10 cm) in the horizontal direction (Robertson et al., 1986). Because it is relatively expeditious to operate (many profiles can be collected in a day), as well as only slightly destructive, dense spatial coverage is possible, such that the continuity of soil units can be verified over large
distances. Although soil models capable of specific \textit{a priori} soil identification are not fully developed, general correlations can be made between piezocone response and major soil types (sand, silt, clay) (Jones and Rust, 1982; Senesset and Janbu, 1984). In addition, site-specific correlations between piezocone response and soil type can be established that allow more precise classification of thicker strata (Aubeny, 1992).

While piezocone data have successfully been used to distinguish peat from other soft sediments, further discrimination of various types of peat within a profile is not commonly made, and the physical interpretation of piezocone data collected in layered peat has not been attempted (Robertson et al., 1986; Erwig, 1988; Olsen and Malone, 1988; Cheng-hou et al., 1990). Although a variety of clays have been extensively tested and classified using the piezocone, the unique and variable structure of peat makes analogies to other soft sediments difficult. Clearly a new interpretive framework is needed for analysis of piezocone records collected in peat.

An even more basic barrier to the development of the piezocone as a tool for wetland exploration has been the difficulty of driving the piezocone from the surface of a wetland. The piezocone is typically driven through terrestrial soils using a truck weighing five tons or more. Much lighter and more portable driving machinery is required for use on the wetland surface. In this paper we describe the development and testing of such machinery, and present limited data that demonstrate that the driver works and that enhanced piezocone technology can be useful in the investigation of wetland sediments.

2.2. Study Site

Motivation for the development of this piezocone driver is hydrogeochemical investigation of the Aberjona River Watershed northwest of Boston, Massachusetts. The
central Aberjona River Valley is filled with up to 50 meters of glacially deposited sands and gravels, thought to contain a number of ice-block depressions now filled with wetland sediments (Chute, 1959; Myette et al., 1987). This watershed has a long and well-documented industrial history with associated releases of various contaminants to surface and groundwaters (Durant et al., 1990; Aurilio, 1992; Aurilio et al., 1994). Of particular concern is the release of arsenic and chromium wastes in the headwaters of the watershed starting early in the century. Work by Spliethoff and Hemon (1994), as well as Solo (1995) suggests the subsequent transport of these and other metals in the Aberjona River and through the Wells G & H wetlands, which are underlain by two to eight meters of peat and other soft organic-rich soils. Former public water supply wells located adjacent to this wetland in the underlying glacial outwash sands and gravels received over half their recharge from the Aberjona River; hence the potential for transport of toxic contaminants through the wetland is of great interest (Hemon, 1995; Myette et al., 1987; Zeeb, 1994). Central to this study is the identification of high conductivity strata in the wetland that may have served as preferred flow paths between the river and wells.

2.3. **Materials and Methods**

2.3.1. **Design Criteria**

The portable driver was designed to push a standard piezocone to the full depth of the Wells G & H wetland deposits while causing minimal damage to the wetland surface. The design criteria for this piezocone penetrometer driver were:

1) Driving capacity: at least 4500 N (1000 lbs.) force.
2) Portability and operability: must be feasible to transport to site in van; must be possible to assemble, disassemble, and operate with one or two people on the soft wetland surface.

3) Must create minimal environmental disturbance.

4) Drive speed: approximately 2 cm per second +/- 10\% over the expected range of loadings.

5) Data acquisition: Data recording on three or more channels at various intervals from half a second to 60 seconds.

2.3.2. Design

The resulting design is a tripod-mounted driver powered by an electric motor that synchronously spins two high-efficiency ball screw drives (see Figures 1 and 2). The tripod is constructed of 32 mm (1 1/4 in) aluminum pipe, 10 cm (4 in) aluminum channel, Nu-Rail fittings (Hollaender Manufacturing, Cincinnati, OH), and several custom welded fittings. All vertical loads (upward and downward) are carried by the horizontal lower (triangular) frame; the superstructure of the tripod holds the drive frame in place, but carries none of the active driving load. The strength of the load-bearing structural members was designed for a factor of safety > 3 and less than 1 cm of flexure. The lower frame is anchored to brackets bolted into the bottom of 25 gallon steel drums (made by cutting a standard 55 gallon drum into halves). When pumped full of water, these drums provide the bulk of the downward reaction force necessary to advance the cone.
Figure 2-1. Piezocone driver tripod.
Figure 2-2. Detail of ball screw/clamp assembly used to drive piezocone drill rods.
Figure 2 shows the drive assembly, including gearing, ball screws, and drill rod clamp sized to hold standard 35 mm (1 3/8 in) 'EW' drill rod (ours is aluminum for portability). A 1/4 horsepower, permanent magnet, 12 VDC 1200 rpm motor (Pacific Scientific, Rockford, IL) was chosen to power the drive assembly because of its high startup torque, constant speed, and compact size. The motor is geared down to spin the ball screws at 300 rpm using 9.5 mm (3/8 in) pitch sprockets and chain. The ball screws (Warner Electric, S. Beloit, IL, model R075) are 19 mm (3/4 in) in diameter with a pitch of 5 mm (0.2 in), such that 300 rpm produces a downward velocity of 2.5 cm/sec (1 in/sec). High efficiency ball screws (> 95% efficiency) were necessary to avoid the large power losses (> 40%) of conventional screw drives, which would have necessitated a larger motor and more maintenance, and caused much greater forces to be carried by the drive assembly. The bottom of each ball screw is fastened to a large flange block bearing (Browning Manufacturing, Maysville, KY, model FB900) which transfers both the upward and downward loads to the lower frame. Much smaller bearings are used at the tops of the ball screws.

The piezocone is driven 1.5 m (5 ft) at a time, with pauses to add additional sections of drill rod and re-position the drive clamp. The motor, chain drive and ball screw assembly is capable of supplying a total downward force of approximately 7400 N (1650 lbs); the factor limiting downward force is the reaction weight of the tripod. This weight is supplied by 300 kg (650 lbs) of water ballast plus the weight of the hardware and approximately 4900 N (1100 lbs) with two people.

The driver is controlled by a switch box with a three-position switch (forward, off, reverse). The switch box is powered by a 12 Volt marine battery with a relay wired to a limit switch mounted at the base of the driver. This safety switch prevents the clamp
from contacting and damaging the chain drive if the motor is not shut off manually at the end of the drive.

Data are collected on a Campbell Scientific CR10 Data logger (Campbell Scientific, Logan, UT) providing 12 1/2 bit resolution over any of several user-selected voltage ranges. The datalogger's programmed operation is controlled and data are monitored in real time using a laptop computer. Data are stored intermittently in ASCII format in the data logger RAM, and uploaded periodically to the hard drive of the computer.

The vertical position of the piezocone is measured relative to the ground surface at the beginning and end of each drive to correct for small movements of the drive frame. During the drive, vertical location is tracked by counting the number of revolutions of the screw drive. Each half-revolution of the screw drive trips a momentary switch wired through a de-bouncing circuit to a pulse-counting channel in the data logger. The number of switch closures (converted to vertical distance) between samples is recorded along with each set of transducer readings. Because the screw drive has a pitch of 5 mm (0.2 in), the vertical position is (theoretically) known to the nearest 2.5 mm (0.1 in), although flexure of the drive frame and compressible ground surface typically limit accuracy to +/- 1 cm (0.4) in.

A standard piezocone (ASTM, 1986) has a 60 degree cone angle and a 10 cm² cross section. Many variations in pore pressure measurement geometry are found. The piezocone configuration used in our pilot test is shown in Figure 3. Pore pressure is measured at the tip with a cylindrical tip extension. Tip resistance is measured with a 50 kN load cell located underneath the standard 150 cm² area friction sleeve. The friction sleeve response was not recorded because it was not considered reliable or sensitive enough for this application. Table 1 summarizes the performance characteristics of this piezocone and data acquisition system.
Figure 2-3. Piezocone design used in this study.
Table 2-1. Specifications for piezocone driver, piezocone, and data acquisition

<table>
<thead>
<tr>
<th>Piezocone Driver</th>
<th>Nominal Speed</th>
<th>Vertical Position</th>
<th>Maximum Downpressure</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>2.5 cm./s.</td>
<td>+/- 1 cm. (0.4 in.)</td>
<td>~ 5 MPa (725 psi)</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Piezocone Measurement</th>
<th>Full scale/output</th>
<th>Analog-to-digital conversion (12.5 bit)</th>
<th>Measurement Resolution</th>
</tr>
</thead>
<tbody>
<tr>
<td>Penetration Resistance</td>
<td>50 MPa/25 mV (7250 psi)</td>
<td>resolution = 0.0033 mV on 25mV Scale</td>
<td>0.0073 MPa (1 psi)</td>
</tr>
<tr>
<td>Pore Pressure</td>
<td>1.38 MPa/100mV (200 psi)</td>
<td>noise = +/- 2 bits</td>
<td>0.000046 MPa (0.00066 psi)</td>
</tr>
</tbody>
</table>
2.4. Results: Pilot Test at Wells G & H Wetland

2.4.1. Driver Performance

The piezocone driver successfully drove the piezocone though the entire wetland thickness at speeds of 2 - 2.5 cm/sec. Because the pulse counter does not allow drive speed differences of less than 0.5 cm/sec to be resolved over short intervals, it was not possible to calculate the instantaneous drive speed variation more accurately. Drive speed variation is generally less than 25% over the entire profile, whereas drive speed measured within each of the two major soil types (soft sediments and sands) is more consistent, varying less than 12%.

The driver penetrated sand lenses up to 1.5 meters thick within the wetland sediments. Piezocone refusal occurred in the aquifer sands below the wetland sediments at a penetration resistance of 2.3 to 3.5 MPa (300 to 500 psi). Refusal was defined as the point at which the combined forces of tip resistance and wall friction approach the weight of the piezocone driver, lifting one of the legs off the wetland surface.

Several records collected in the Wells G & H wetland verify that the data collected are essentially noise-free and otherwise of excellent quality (Figures 4 and 5). The locations of these drives are indicated in Figure 6. Soil samples from borings located within 50 cm of the piezocone drive have been used to classify qualitatively the sediments in these profiles.
Fig. 2-4. Piezocone response for drive 93-11 compared to profiles of soil water fraction and organic fraction by dry weight. Water fraction is defined as mass water divided by total mass; organic fraction is defined as mass organic divided by dry mass. "Breaks" indicate locations of stops to add drill rods.
Fig. 2-5. Piezocone response for drive 93-16 compared to profiles of soil water fraction and organic fraction by dry weight. Water fraction is defined as mass water divided by total mass; organic fraction is defined as mass organic divided by dry mass. "Breaks" indicate locations of stops to add drill rods.
Figure 2-6. Project Study Area: The Wells G & H Wetland, in the central Aberjona River Valley.
2.4.2. Piezocone Data Records

Piezocone profile 93-11 (Figure 4) shows an upper peat stratum with several distinct peaks in tip resistance that correlate with positive penetration pore pressure deflections centered at approximately 12 and 50 cm. A transition zone between 75 and 150 cm below the surface separates the upper from the lower peat, the latter having a more uniform pore pressure signal with increased high frequency variations and a steeper average slope (approximately 0.024 MPa/m. or 0.0875 psi/in.). Soil samples identify the lower peat as having a consistently high water and organic content (> 90%). Below the lower peat is a stratum of organic silt with a smaller pore pressure slope (0.01 MPa/m. or 0.0375 psi/in.), only slightly above the hydrostatic line. A dense sandy stratum exists between 400 and 550 cm. This layer is correlated with the highest tip resistance encountered in the profile and negative penetration pore pressure.

In contrast, profile 93-16 (Figure 5) shows a sand stratum at 200 cm separating the upper and lower peats, and only thin sand interbeds near the base of the profile (450 - 500 cm). The upper and lower peats at this location have piezocone "signatures" similar to those in 93-11, showing the same distinct peaks in tip resistance in the upper peat, and changes in pore pressure response in the lower peat. The point resistance and pore pressure response in the lower organic silt also closely resemble the response at 93-11.

2.5. Conclusions

These data demonstrate that the performance of the piezocone driver is sufficient to produce high quality data from a standard piezocone, and that data from wetland soils are interpretable based on several basic soil types. Despite the relatively high variation in
drive speed the piezocone driver and data collection system perform as intended, distinguishing features that are well-correlated with the generalized soil types and distinctly different responses for peat, sand, and organic silts. Better performance might result from improved drive speed control (by using a more powerful motor, for instance).

Evidence of small scale heterogeneities in the vertical dimension is seen, including 5 - 10 cm thick resistant interbeds at the base of 93-16 and several continuous layers in the upper peat marked by peaks in pore pressure and tip resistance. High frequency variation in the penetration pore pressure for most of the soft strata may eventually be interpreted as sub-layers in these soils.

Significant changes in the soil profile over the horizontal scale of the site are also apparent. In the two profiles shown, the major sand strata detected in these profiles appear to be parts of separate, discontinuous deposits. The continuity of sand strata found throughout this wetland will be of primary importance in the interpretation of preferred flow paths and potential transport of contaminants.

2.6. **Additional Work**

These results and our review of the literature indicate the need for a piezocone better suited to profiling of wetland soils than the standard geotechnical piezocone. For wetland use a piezocone must have more sensitive point resistance and friction sleeve load cells to better discriminate sublayers in soft sediments and to differentiate structured organic soils from inorganic fine-grained sediments. Additionally, such a piezocone should incorporate pore pressure measurements at several locations so that the behavior of wetland soils under transient loading can be better understood.
References


Zeeb, P. J., 1994, “The Effects of Deposit Scale Heterogeneities on the Trapping and Transport of Toxic Metals in an Urban Riverine Peatland (poster), Spring Meeting, American Geophysical Uniton, Baltimore, MD.
3. A Five-Channel Piezocone Penetrometer for Wetland Soils

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3.1. Introduction

Wetland soils are notable for their physical and hydraulic heterogeneity, fine layering, unique structural properties (such as root mat development), and high sensitivity (weak, easily disturbed). As a consequence, such soils are particularly difficult to sample in an undisturbed manner or to test reliably *in-situ*, and may lead to highly variable and three dimensional groundwater flow. These characteristics, along with the important role that wetlands may play in controlling the interaction of surface and groundwaters, make these deposits both difficult and essential to understand.

The piezocone is a promising tool for the characterization of such deposits. The piezocon was developed from the cone penetrometer in the 1970's, and is generally recognized as a powerful soil profiling tool (Wissa et al., 1975; Baligh et al., 1981; Campanella and Robertson, 1988). Historically, its use has focused on the exploration of terrestrial and marine soils, with the goals of identifying layers of contrasting properties and estimating certain soil parameters, such as the undrained shear strength, overconsolidation ratio, and consolidation coefficient. A standard piezocone measures pore pressure, tip resistance, and frictional resistance while being driven at a rate of approximately 2 cm/sec through the soil profile (ASTM, 1986). Measurements are made continuously, generating a highly resolved record of soil response to penetration. Site-specific correlations can be made between piezocone response and soil type. General soil classifications are also available, but do not reliably distinguish among different soft soils (Jones and Rust 1982; Seneset and Janbu 1984). Direct measurement of soil hydraulic and other engineering properties is not achieved with this technique, due to the complex boundary conditions at the cone as well as soil non-linearities.
An important advantage of the piezocone is that it provides information over a wide range of spatial scales. A piezocone profile quantifies vertical heterogeneity at the 5 to 10 cm scale while traversing a virtually unlimited vertical distance. Because it creates only limited disturbance, profiles can be closely spaced in the horizontal dimension if necessary. In contrast, other common techniques used to gather stratigraphic information typically resolve either small scale or cover large scale features, but are correspondingly limited in either breadth of coverage or resolution. Laboratory tests suffer from sampling disturbance and are extremely labor intensive. Non-intrusive (geophysical) methods typically lack the resolution and specificity necessary to discriminate many features of interest, and present difficult inversion problems. Other in-situ tests (pumping tests, for instance) average the properties of a relatively large volume of soil, and are unable to detect or measure small heterogeneities.

Existing piezocones are not suitable for wetland investigation, primarily due to the lack of sensitivity in the tip resistance measurement; the typical tip resistance range is 0 - 50 MPa. Several other standard piezocone features are also non-optimal for wetland soils: 1) friction sleeve measurements are usually made in the unloading zone directly behind the cone face, resulting in poor sensitivity; and 2) pore pressure is measured in only one location, such that the cone is either insensitive to thin layers (when sensor is located on the shaft) or is without a means of making accurate pore pressure corrections to the sleeve friction measurement (when sensor is at the tip).

Despite these limitations, our initial wetland profiling using a standard piezocone yielded promising results. Piezocone response was well correlated with soil core data from a highly layered and heterogeneous site (Zeeb et al., 1996). The maximum tip resistance encountered within the wetland soils was 3 MPa. Although we did not measure sleeve friction due to the inconsistent performance of the friction sleeve, existing literature suggests that sleeve friction can be particularly helpful in distinguishing peat from other soft inorganic soils (Cheng-hou et al., 1990; Robertson et al., 1986; Erwig, 1988; Olsen
and Malone, 1988). The results of initial tests encouraged us to design a piezocone optimized for use in soft wetland soils.

The resulting piezocone, described here, was intended for several purposes: 1) delineating sub-strata within peat deposits; 2) distinguishing peat from other soft soils in the profile; 3) locating interbedded sand strata; and (4) testing existing soil classification models, or supporting the development of new ones. The portable piezocone driver needed to push the piezocone from the wetland surface is described elsewhere (Zeeb et al., 1996).

3.2. Piezocone design

3.2.1. Design Criteria

The piezocone design was governed by the following considerations:

- Only small tip resistances were expected (<3 MPa). This required the use of a particularly sensitive load cell, in combination with data acquisition hardware that can take advantage of the cell’s sensitivity.
- Sleeve friction forces were expected to be small (~0.1 MPa), but were considered important to measure. This required high sensitivity load measurement, as well as pore pressure measurement along the piezocone shaft to correct for the effect of differential pore pressures exerted on the ends of the friction sleeve.
- Penetration may be either undrained or partially drained, causing significant pore pressure excursions both above and below hydrostatic pressure. This condition also
made pore pressure measurements at more than one location desirable, especially for the collection and interpretation of dissipation data.

- Identification of thin soil layers is important. Measurement of pore pressure at the tip was therefore of great importance, as this location is best for detecting thin layers (which are significantly altered by the passage of the tip itself).

3.2.2. Measurements

Each piezocone sensor is discussed below. Because both transducer and data acquisition characteristics affect overall performance of the piezocone, characteristics of the data acquisition system as well as sensor choice and design are discussed. Figure 1 shows a cutaway drawing of the piezocone, and Table 1 summarizes the characteristics of the piezocone sensors and data acquisition system. Sampling frequency is 4 Hz on all channels, corresponding to a measurement spacing of approximately five to seven mm. This level of resolution provides for the possibility of soil characterization based on both the spatial frequency of variation of the piezocone measurements and their absolute magnitudes.

3.2.2.1. Pore Pressure

Pore pressure measurements are made at three different locations on the piezocone (Figure 1): at the tip, using a cylindrical tip extension (sintered stainless steel filter); approximately 10 cm (three cone diameters) behind the tip; and 1.65 cm above the top of the friction sleeve, or about ten cone diameters behind the tip. These locations enable the measurement of pore pressures in a low pressure-gradient zone near the sleeve ends, allowing for reasonably accurate pore pressure corrections to the sleeve friction measurement. In addition to providing this correction, multiple pore pressure measurements can provide information on the loading history of the soil (Sills et al.,
Figure 3-1. Schematic of piezocone penetrometer, showing cabling to analog-to-digital converter and laptop computer.
Table 3-1. Measurement resolution of piezocone/data-logging system.

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Penetration Resistance</td>
<td>7 MPa/7 mV (1000 psi)</td>
<td>resolution &lt; 0.001 mV on 50 mV Scale noise = +/- 0.5 μV</td>
<td>0.001 MPa (0.14 psi)</td>
<td>17 MPa (2433 psi)</td>
</tr>
<tr>
<td>Sleeve Friction</td>
<td>0.2 MPa/4.4 mV</td>
<td></td>
<td>0.000046 Mpa (0.0066 psi)</td>
<td>0.7 MPa (164 psi)</td>
</tr>
<tr>
<td>Pore Pressure at Tip</td>
<td>1.38 MPa/100mV (200 psi)</td>
<td>resolution = 2 x noise = 0.001 mV</td>
<td>0.000014 MPa (0.002 psi)</td>
<td>2.76 MPa (400 psi)</td>
</tr>
<tr>
<td>Pore Pressure on Shaft</td>
<td>1.38 MPa/10mV (200 psi)</td>
<td></td>
<td>0.00014 MPa (0.02 psi)</td>
<td>2.76 MPa (400 psi)</td>
</tr>
</tbody>
</table>
1988; Sully et al., 1988), and can help to distinguish among various types of soft soils (Juran and Tumay, 1989).

Pressure measurements in the shaft locations are not made through the typical annular porous element, but through a sintered stainless steel disk 31 mm thick and 6.2 mm in diameter. This element is pressed into a threaded plug which is machined to match the exterior of the cone shaft when threaded into its socket. Bond et al. (1991) successfully used a similar geometry for pore pressure measurements on an instrumented pile. This arrangement allows for a very small fluid cavity and fast response times, as well as convenient access to the porous element for field replacement in the event of surface desaturation. The annular geometry was avoided on the basis of earlier MIT testing experience which indicates that annular elements are sensitive to axial loadings and often produce erroneous pressure measurements as a result (Baligh and Azzouz, 1984). In addition, annular elements are not easily replaceable in the field and require very close tolerances to provide a good seal while avoiding transfer of penetration stresses to the filter.

The tip pressure transducer (Data Instruments, Acton, MA, Model AB/HP) has a full scale output of 100 mV at 1.38 MPa (200 psi); whereas the shaft pressure transducers (Cooper Instruments, Warrenton, VA, model psg-110) have a full scale output of 10 mV at the same pressure.

As emphasized by many investigators, rapid pore pressure response is critical to the generation of reliable piezocone data (Rand and Tumay, 1985; Juran and Tumay, 1989). In general, the response time of the pore pressure measurement system must be somewhat shorter than the time scale of the piezocone sampling process:
sampling time scale = (width of filter element)/(piezocone drive speed):

\[ \frac{0.5 \text{ cm}}{2.5 \text{ cm/s}} = 0.2 \text{ s} \]
required response time < 0.2 s

With a response time of less than 0.2 s, the piezocone is potentially capable of detecting the pore pressure signature of a soil layer as thin as several times the filter element width at standard driving speeds. Achieving such response times requires that the porous element connecting the pressure transducer to the soil pores be permeable enough to allow the finite volume exchange necessary, but with a large enough bubbling (air-entry) pressure such that the porous element remains saturated during use despite soil suction. In addition, the cavities and seals that connect the pressure transducer to the porous element be fully saturated, small in volume, and stiff (i.e. allow very little volume change with pressure variation). Finally, the transducer itself must also be stiff, requiring only a very small diaphragm deflection (volume change) to record a pressure change. These qualities minimize signal damping by requiring only the very slightest flow of water through the porous element to transmit the pressure signal. Because of water's low compressibility, transducer diaphragm deflection typically accounts for the largest proportion of flow through the porous element.

In our cone, the porous elements separating the pressure transducers from the soil pore space are made of sintered stainless steel having pore sizes in the range of 2 - 6 microns (Mott Metalurgical, Farmington, CT). Spare porous elements are de-aired and saturated under a vacuum in the lab; several spares are kept on hand in the field. The cavities between the porous elements and pressure transducers in our cone are small (0.1 cm³), and the o-ring seals are designed to create the maximum possible o-ring squeeze; the design volume of the o-ring grooves is only 3% larger than the o-ring volume.
The theoretical pore pressure response time scale in water can be calculated from the known pressure-volume characteristics of the transducer, the size of the water filled cavity behind the stone, and the effective hydraulic conductivity of the stone. In our calculation, the effect of water compression was linearized using a reference volume, and expansion of the metal cavity was not considered. The pressure response to a step change in pressure is exponential, and $T_{90}$, the time required for the pressure recorded to reach 90% of the value of a step is taken as a measure of the response time. Although the tip transducer system is more compliant than the shaft systems, it uses a larger porous area, such that the response times in water are similar for the two systems; calculated $T_{90}$ is about 0.004 seconds.

These response times are for measurements made in water. The response time for measurements made in-situ will be slowed by the process of soil consolidation in the vicinity of the porous stone. Kutter et al. (1990) have demonstrated the potential importance of nonuniform pore pressures near the stone using a finite element model. Earlier analytical solutions by Gibson (1963) and de Josselin de Jong (1953) consider soil consolidation around a spherical pore pressure transducer after isotropic step loading, and match Kutter's results well. The in-situ response times for our piezocone were estimated for the range of soil properties encountered at the test site using de Josselin de Jong's analytical solution. A full spherical space was assumed for the tip, and both half and full space solutions were calculated for the shaft; the proper solution is an average of these.

In the solution given by de Josselin de Jong and Gibson, response time is controlled by the compliance of the measurement system and the product of soil hydraulic conductivity ($k$) and compressibility ($m_v$). For our cone, values of $k \cdot m_v$ less than about $9 \times 10^{-11}$ $m^3/\text{sec} \cdot \text{kN}$ will cause the tip measurement response time to be limited by soil processes. The shaft transducer is slightly more sensitive to soil properties, and is affected at $k \cdot m_v$ less than $1.2 \times 10^{-10}$ $m^3/\text{sec} \cdot \text{kN}$. For most of the soil types at our test site, calculated soil response times are somewhat slowed by soil processes, but never by more than a factor of...
10, resulting in a maximum response time of approximately 0.04 sec. This response time is sufficiently fast to maintain the fine (0.5 cm) measurement scale of this piezocone.

3.2.2.2. Tip Resistance

The tip resistance load cell was designed to provide full scale output at 7 MPa (1000 psi), with sufficient strength to be undamaged by loads as much as 17 MPa. We used standard foil strain gauges to minimize zero shift; the gauges were configured in a double wheatstone bridge to provide both bending moment correction and temperature compensation. The output of this strain gauge/load cell is about 1 mV/MPa: with a high resolution data logger, a tip resistance resolution of 0.001 MPa (0.14 psi) is achieved.

3.2.2.3. Sleeve Friction

The sleeve friction load cell uses the same type of foil strain gauges used for the tip resistance cell, and has a resolution of $4.6 \times 10^{-5}$ MPa (0.0066 psi). The friction sleeve is positioned four tip diameters (14 cm) behind the tip, a location that increases measured friction loads by a factor of 2 - 4 over the conventional location immediately behind the cone face (Konrad, 1987). This arrangement causes the skin friction forces acting on the intervening portion of the shaft to be measured as part of the tip resistance. The resulting error is insignificant, however, as the typical measured sleeve friction rarely exceeds 0.01 MPa, in contrast to the minimum tip load of 0.2 MPa.

3.2.3. Pore Pressure Corrections

Hydrostatic pore pressure corrections to the tip resistance and sleeve friction measurements were measured in the lab using a pressure vessel. The tip correction is 31% (area ratio of 0.69), whereas the sleeve correction for a constant pressure along the shaft is - 1.1% of the measured pressure. The measured end area of the friction sleeve is
used to make corrections to the sleeve measurements; corrections are applied to the field
data by using a weighted average of the two shaft pressure measurements to estimate
actual pressure at the top and bottom of the friction sleeve.

The data acquisition system uses a low noise (< 0.001 mV), high resolution (dynamic: up
to 22 bit), eight channel analog-to-digital board (Strawberry Tree, Cupertino, CA)
controlled by a laptop computer running vendor-supplied software. The eight input
channels are used to record the output voltages of five sensors at a measurement
frequency of 4 Hz, two input voltages (one for the pressure transducers, one for the load
cells), and the current drawn by the piezocone driver motor. The latter measurement
documents motor operation and provides an approximate measure of the total driving
force applied to the drill rods. The pulse output of the piezocone depth recorder is
measured by a separate counting channel on the data acquisition board. This system
allows all measurements to be monitored and stored to the hard disk of the computer in
real time. All aspects of the data collection protocol including the measurement
frequency can be changed in real time.

3.2.4. Materials and construction

Piezocone components were fabricated from a selenium-alloyed stainless steel (304S) for
durability and machinability (see Figure 1). Standard configurations were used where
practicable to maximize comparability of results with other piezocone studies. The cone
tip has a standard 60 degree angle, with pore pressure measured through a porous
cylindrical extension, a proven and very durable design. The tip and sleeve friction load
cells are load-isolated in a standard configuration: loading of either cell has no effect on
the other. Each piezocone measurement is performed by a separate module with
identical threading and seals. This arrangement permits the selection of alternate pressure
and sleeve friction measurement points (e.g. changing the order, or adding or removing modules).

3.3. Field Testing

3.3.1. Site

Motivation for the development of this piezocone driver is hydrogeochemical investigation of the Aberjona River Watershed northwest of Boston, Massachusetts. The central Aberjona River Valley is filled with up to 50 meters of glacially deposited sands and gravels, thought to contain a number of ice-block depressions now filled with wetland sediments (Chute 1959; Myette et al. 1987). This watershed has a long and well documented industrial history with associated releases of various contaminants to surface and groundwaters (Aurilio 1992; Aurilio et al. 1994; Durant et al. 1990; Spiethoff and Hemond 1994). Of particular concern is the release of arsenic and chromium wastes in the headwaters of the watershed starting early in the century, and subsequent transport of these and other metals via the Aberjona River through the Wells G & H wetlands, which are underlain by two to eight meters of peat and other soft organic-rich soils. Former public water supply wells located adjacent to this wetland in the underlying glacial outwash sands and gravels received over half their recharge from the Aberjona River; hence the potential for transport of toxic contaminants through the wetland is of great interest (Hemond 1995; Myette et al. 1987; Solo 1995; Zeeb 1994). Central to this study is the identification of permeable strata in the wetland that may have served as preferred flow paths between the river and wells.
3.3.2. Piezocone performance

3.3.2.1 Pore pressure response

The pore pressure response of this piezocone was measured immediately before each piezocone drive. Under optimal conditions (pore pressure element fully saturated), the frequency response approached the sampling frequency of 20 Hz used in the test procedure. When degraded response due to porous element desaturation was noted, the porous element was replaced. Figure 2 shows the pore pressure response, measured in a field test chamber, with both partly and completely saturated porous elements. Desaturation of the pore pressure measurement system was typically limited to the outer layer of the porous element, based on the observation that replacement of unresponsive elements restores response.

In the wetland soils the frequency of penetration pore pressure variation was high enough to warrant filtering the data for ease of visual interpretation. Figure 3 shows typical pore pressure response at the tip and shaft locations, and the corresponding data filtered with a seven member moving triangular average. Because the drive speed is fairly constant at 2.2 cm/sec, the filter width is about 4 cm, varying only slightly with driving speed and therefore with tip resistance. These data clearly exhibit fine detail in the shaft pore pressure measurements. Tip pore pressure measurements do not show the same degree of high frequency variation, but still reveal detail on a scale of several centimeters. Because the calculated in situ response times for the two different pore pressure measurement locations are similar, differences in signal frequency reveal a fundamental difference in the characteristics of pore pressure variation at these two locations. The high frequency response of the shaft measurement may contain useful information on small scale vertical variations in the soil profile.
Figure 3-2a. Typical response of a fully saturated system.

Figure 3-2b. Typical response of a poorly saturated system.
Fig. 3-3. Comparison of raw and filtered pressure measurements made at the piezocone tip and along the shaft.
3.3.2.2. **Tip Resistance and Sleeve Friction Response**

Tip resistance and sleeve friction values reflect the average properties of a volume of soil ahead of and surrounding the probe. Although the absolute magnitude of high frequency variations observed in the tip resistance record is similar to those observed in the pore pressure record, the amplitude relative to the full scale tip resistance of 3 MPa is small. Nevertheless, the resolution of the tip resistance and sleeve friction measurements (0.001 MPa and 0.000046 MPa, respectively), combined with their fine spatial resolution (about 0.5 cm), is sufficient to discriminate among different layers within the soft soils. Figure 4 shows a portion of a piezocone record with pore pressure and tip resistance plotted at the same scale. This figure illustrates the similarity of the higher frequency variations in tip resistance and pore pressure. Measured tip resistance values in the wetland soils were 3 to 30 times greater than the measured pore pressures. The sleeve friction axis is scaled up by a factor of 100.

The maximum sleeve friction measured in these profiles is 0.02 MPa, a reasonable value compared to the design full scale range of 0.2 MPa. This allows resolution to 1 part in 435 in zones of maximum friction load, based on a resolution of 4.6 x 10^-5 MPa. Further improvement of sleeve friction resolution would require increased load cell sensitivity; however, small but unquantified errors in the pore pressure correction may then impose the practical limit on sleeve friction accuracy. If the pore pressure correction error is 5% of the correction, sleeve friction accuracy is limited to 4 x 10^-5 MPa at a measured pressure of 0.08 MPa, regardless of resolution.

Drift in the zero output of both the sleeve friction and tip resistance load cells occurs over both the long and short term. Long term drift may reflect the natural aging of the bonding agent used to secure the foil strain gages to the piezocone. Temperature effects may be superimposed on the drift pattern, although the strain gages are installed in a
Figure 3-4. Small scale variations in piezocone measurements. All pressure scales are the same except sleeve friction, which is expanded 10 times.
pattern that minimizes the effects of temperature change. We observed a roughly logarithmic trend in long term drift from July to December that is consistent with both the aging process and with seasonal changes. A marked decrease in scatter about this trend probably reflects decreasing diurnal temperature variations or decreasing contrast between ground and air temperatures over this time period. The impact of long term zero drift on piezocone data quality is minimal, as the load cell zero output values can be measured at the beginning of each drive and accounted for. From July to December of 1995, the tip resistance and sleeve friction load cells drifted $+3.5 \times 10^{-5}$ and $-4 \times 10^{-5}$ V/V, respectively, corresponding to 6 and -11% of full scale.

Daily zero drift never exceeded $+/1.6 \times 10^{-6}$ and $+/2 \times 10^{-6}$ V/V for the tip and sleeve load cells, or $+/-0.27$ and $+/-0.55$% of full scale, respectively. These values correspond to drift having an approximate magnitude of $+/0.63$% and $+/5.5$% of the typical measured maximum values of tip resistance and sleeve friction. The variable direction and magnitude of this daily drift, superimposed on the long term drift, suggests a different origin, which may be diurnal (and other less regular) temperature variations. For the tip resistance load cell, the impact on measurement accuracy is insignificant. Although the sleeve friction drift is partially corrected by re-zeroing the output at the beginning of each drive, error of as much as 5% of the measured maximum value may occur during a drive.

3.3.3. Comparison with Soil Core data

Two representative piezocone profiles, horizontally separated by 15 meters, are shown in Figures 5 and 6. The major soil strata previously identified in this wetland (peat, sand, diatomaceous earth) are well-correlated with the pore pressure, tip resistance, and sleeve friction responses of this piezocone. Furthermore, several distinct sub-layers within the peat can be identified from the detailed record.
Figure 3-5. Soil profile data for location 94-38. Fraction water is defined as mass of water divided by total mass. Fraction organic is defined as mass organic divided by dry mass. Diamond and triangle symbols ("Breaks") indicate locations in data set corresponding to stops to add drill rod. All data are corrected to the position of the piezocone tip.
Figure 3-6. Soil profile data for location 94-50. Fraction water is defined as mass of water divided by total mass. Fraction organic is defined as mass organic divided by dry mass. Diamond and triangle symbols ("Breaks") indicate locations in data set corresponding to stops to add drill rod. All data are corrected to the position of the piezocone tip.
It is important to note that vertical misalignments of as much as +/- 10 cms can exist between the piezocone and soil core profiles. These shifts are due to horizontal variability over the 30 - 50 cms that separate the two profiles, to vertical variations in wetland surface elevation, and to small errors in tracking the vertical position of the piezocone. Three major peat layers have been identified based on botanical classifications, including a fine grained woody swamp peat (P2), a sedge peat (P1c), and, at the surface, a peat composed mostly of *Typha* and *Phragmites* roots, rhizomes, and stems (P1a and P1b). The upper (non-woody) peat strata (P1) are highly layered, often with many thin sand interbeds, and contain many large pieces of relatively undecomposed plant material. Although piezocone response is highly variable in the upper peat, the piezocone records contain several coherent and consistent features. Notable is the presence, in both profiles, of two peaks in both tip resistance and sleeve load at 10 - 20 cm and at 50 - 70 cm depth which coincide with peaks in penetration pore pressure (see arrow notation on Figures 5 and 6). Several soil cores have revealed dense root mats that correlate with these peaks; it is not always possible to obtain representative samples in such layers.

The lower woody peat (P2) has a distinctive piezocone signature, characterized by elevated sleeve load (fine line in center plot) and high penetration pore pressures (heavy line in left hand plot). Characteristic peaks in these parameters are shown in figures 5 (170 to 200 cm) and 6 (170 to 230 cm). Sand lenses within the peat are readily identified by elevated tip resistance and low penetration pore pressure (see arrows on both figures). Thicker sand strata are the most obvious, having large, broad peaks of tip resistance.

The underlying diatomaceous earth has low and relatively constant values of both tip resistance and sleeve load. Penetration pore pressures in this soil fluctuate less than in the peat, and have an average slope of 0.016 MPa/m. Thin sand lenses located within the diatomaceous earth are quite variable in occurrence. The slight increase in friction and tip load correlated with a decrease in pore pressure at 450 cm in Figure 6 could be the
record of a sand lens; however, the only sand lens encountered in the corresponding soil core is located below 490 cm.

Piezocone refusal occurs when the combined forces of tip resistance and wall friction exceed the reaction weight of the tripod, lifting it off the wetland surface. Typically the tip resistance is near 3 MPa when this occurs, although it also depends on the depth of refusal. Samples of the top meter of the underlying soils reveal well-sorted medium sand, which are part of the underlying stratified drift deposits. Existing well logs from deeper borings verify that the depth of piezocone refusal occurs at the bottom of the wetland soils.

3.4. Conclusions

The piezocone design described in this paper meets its objectives of providing high resolution measurements of sleeve friction, tip resistance, and pore pressure in soft. Multiple pore pressure measurements allow correction of sleeve friction and tip load measurements. The design also meets the objective of fast pore pressure response, which provides an accurate record of penetration pore pressure in low conductivity soils, and, in theory, allows the detection of thin (on the order of 1 - 2 cm when driven at 2.5 cm/sec) soil layers.

The response of this piezocone correlates well with major soil strata found across the wetland site. Sub-layers within the more than 2 meters of peat found at the site are readily distinguished by their piezocone signatures. Sand lenses as thin as 10 cm are detected by the tip pressure transducer; thicker sandy strata generate elevated tip resistance and sleeve friction deflections as well. The underlying limnic sediments have distinctive tip and sleeve load signatures. Accurate sleeve friction measurements contribute significantly to this piezocone’s ability to distinguish among different types of soft soils.
This tool provides detailed stratigraphic information without requiring sampling or sample analysis, and therefore represents a valuable tool for the investigation of wetland stratigraphy. Detailed soil profiles can be gathered relatively quickly, at the rate of 30 to 60 meters per day, depending on depth and site access. The piezocone has a particular advantage in thick wetland/limnic deposits, where sampling the entire depth is very difficult. Only a few soil cores are necessary to establish soil correlations with the piezocone data.

3.5. **Further Work**

Experience with this piezocone has suggested additional areas where further development may be warranted. These include:

1) Sleeve Friction

The sleeve friction measurement has been problematic for many investigators for several reasons. First, the measured force is typically small compared to the tip resistance and pore pressures. Load cells that are robust enough to withstand large transient forces or oblique loads are typically not sensitive enough to provide high output and corresponding accuracy and resolution. Second, the o-ring seals in the standard friction sleeve are subject to the entry of soil during the driving process. Embedded soil can pre-load the sleeve or transmit part of the sleeve friction force to the walls of the piezocone rather than the sleeve friction load cell. The forces thus generated or relieved by the soil can be as large as the measured sleeve friction. The performance of the friction sleeve on this piezocone was judged to be adequate for discrimination of a number of different soil types, although further improvement is possible.
2) Miniaturization

A smaller piezocone could still be rugged enough for use in soft sediments, and would require less driving force. Driving the standard size piezocone used in this study requires a water-ballasted tripod driver (Zeeb et al., 1996) that can not be moved without pumping the water out. A smaller piezocone could be driven by lighter, more portable equipment, with a significant increase in efficiency. However, such a piezocone would complicate comparisons to data generated by standard, 10 cm$^2$ piezocones and would present technical challenges with respect to miniaturization/integration of the piezocone components. Perhaps most significantly, a miniature piezocone may suffer from increased “grain scale” environmental 'noise', especially in peat soils where the “grain scale” can be as large as several centimeters. Such noise results when the measurement scale becomes less than the grain scale, and the instrument responds to each grain as if it were a heterogeneity as significant as a soil layer. Despite this potential pitfall, miniaturization should be considered, with careful attention paid to balancing the effective measurement scale and grain scale.

3) Additional Sensors

Contrasts in soil electrical properties and levels of metal contamination suggest that additional measurements made in situ with the piezocone could be useful in detecting physical and chemical variability. Conductivity cells have already been integrated into piezocones to successfully detect changes in soil and pore fluid electrical properties (Campanella and Weemees, 1990; Strutynsky et al, 1991), and might be useful in wetland soils. Fluorescence measurements might also be useful, allowing the piezocone to identify certain chemical transitions in the peat.
References


Zeeb, P.J., 1994, “The Effects of Deposit Scale Heterogeneities on the Trapping and Transport of Toxic Metals in an Urban Riverine Peatland”, (poster), Spring Meeting, American Geophysical Unions, Baltimore, MD.

4. Interpretation of Data Gathered with a Wetland Soil Piezocone

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4.1. Introduction

4.1.1. Background

In the investigation of wetlands, soil cores typically provide the only information regarding wetland stratigraphy. Cores are often obtained to define the distribution of chemicals in wetland soils, to obtain radiocarbon ages of soil layers, to catalog and interpret pollen assemblages, and to define soil characteristics (Boelte and Verry, 1977; Chason and Siegel, 1986; Knott et al., 1987; Walton-Day, K. et al., 1990; Cole et al., 1990; Anderson et al., 1992; Orson et al., 1992; Supardi et al., 1993, Neuzil et al., 1993). Because of the relatively time-consuming process of obtaining and analyzing cores, few are obtained. Rarely can wetland stratigraphy be directly observed by other means, such as outcrops (Peters, 1988). The history of a wetland may be postulated based on the analysis of one core alone (Winkler, 1985; Anderson et al., 1992; Davis and Nickmann, 1995). Missing from this approach is the definition of horizontal soil variability and morphology of the wetland layers, from which useful interpretations regarding wetland history and the potential for present-day subsurface transport can be made.

The investigation of wetland deposits with a piezocone penetrometer for these purposes has not been attempted before this study. Because the piezocone provides high resolution profiles of soil characteristics with relatively little disturbance and efficient operation, detailed stratigraphic maps can be developed over relatively large areas.

Piezocone penetrometer data have been used to detect peat deposits buried in mineral soils, and to find thin layers of sand in softer soils such as clay. It is more difficult to distinguish between peat and other soft soils than between peat and sandy material,
because the index measurements most commonly used, tip resistance and penetration
pore pressure, are not sufficiently diagnostic of the distinctive mechanics of peat and soft
clays and silts.

Furthermore, piezocone data have not been used to discriminate among different types of
peat. However, the previous chapter has demonstrated that a highly sensitive piezocone
is capable of delineating wetland stratigraphy, including several different peat types,
organic silt, and sand.

In this chapter, the data collected with the wetland piezocone driver described in Chapter
2 and the wetland piezocone described in Chapter 3 are examined in detail (Zeeb et al.,
1996). The result is two-fold: a qualitative system for interpreting the data based on
several ‘ground-truthed’ piezocone profiles, and a set of stratigraphic cross sections
interpolated from the piezocone records.

Two aspects of these data prevent interpretation by conventional means: fine layering,
and the presence of a variety of peat soils. Together these characteristics confound both
soil models and the traditional soil type correlations that are often used to infer the
stratigraphic record from piezocone data. It is anticipated that a greatly improved
understanding of penetration of layered soils, and new and more sophisticated soil models
will be required to interpret data such as these in a quantitative manner using numerical
soil classification schemes. In the absence of these advances, however, useful
information can still be gleaned from these data using qualitative classifications.

4.1.2. Piezocone penetration

During penetration, measured pore pressure, tip resistance, and sleeve friction are
controlled by soil compressibility, permeability, and the ability of the soil to mobilize
shear strength when strained. In a relatively impermeable deposit (such as clay), penetration is essentially ‘undrained’, meaning that there is no water movement relative to the pore spaces during penetration, and large excess pore pressures develop. In a relatively permeable deposit (such as sand), penetration is fully drained, meaning that very little excess pore pressures develop, and penetration is resisted only by increases in effective stress in the soil. The greater stiffness of sand also results in increased tip resistance relative to clay. Dense sand and other overconsolidated soils also tend to dilate when sheared, causing negative excess pore pressures to develop in some cases.

Whether penetration is drained or undrained depends on the values of soil permeability, compressibility, and strength, relative to the penetration rate and the size of the cone. For a 10 cm$^2$ cross section cone pushed at 2.5 cm/sec, the typical range of permeability for undrained penetration is $< 10^{-7}$ cm/sec; for drained penetration soil permeability must exceed approximately $10^{-4}$ cm/sec, although there is also a dependence on soil compressibility (Baligh, 1986a).

Figure 1 shows how tip resistance and pore pressure typically vary through a typical profile composed of peat, sand, and clay (Baligh, 1986b). In sand layers, tip resistance is typically high, whereas penetration pore pressure is nearly hydrostatic. In the lower-permeability clay, tip resistance is lower, and penetration pore pressure is larger and variable, in some cases revealing the location of interbedded sand layers. In the peat layer at the top of the profile, tip resistance is low, and penetration pore pressure is above hydrostatic.

4.1.2.1. Soil Classification

As discussed in the previous chapters, the piezocone is unique in that it provides information over a wide range of spatial scales. A piezocone profile quantifies vertical
Figure 4-1. Typical response of a piezocone penetrometer (Baligh, 1986). In sand layers, tip resistance is typically high, whereas penetration pore pressure is nearly hydrostatic. In the lower-permeability clay, tip resistance is lower, and penetration pore pressure is larger and variable, in some cases revealing the location of interbedded sand layers. In the peat layer at the top of the profile, tip resistance is low, and penetration pore pressure is above hydrostatic.
heterogeneity at the 5 to 10 cm scale while traversing a virtually unlimited vertical distance. Because it creates only limited disturbance, profiles can be closely spaced in the horizontal dimension if necessary. A highly resolved map of the subsurface can be created in this manner, provided the data can be reliably interpreted.

Interpretation of piezocone data from stratified systems usually involves the use of classification charts that correlate piezocone measurements with soil types. These charts are often site-specific, and work only for thick soil layers (> meters) (Jones and Rust 1982; Senesset and Janbu 1984; Juran and Tumay, 1989). Application of this technique is greatly aided by previous general knowledge of site stratigraphy and the use of lithogenic models (Depret, 1982; Krajicek and De Lang, 1982). Once the expected sequence and characteristics of layers are understood, piezocone data can effectively be used to establish the continuity and extent of thick layers.

Another method for piezocone data interpretation involves the use of a quantitative soil behavior model that allows basic soil properties, and hence soil type, to be inferred from measured penetration data. Such models currently exist only for undrained and drained penetration in infinite, homogeneous soils, and are only predictive for well-studied soil types (Whittle and Germaine, 1996). Efforts to develop models of partially drained penetration, (i.e. peat, silt, and silty sand) are currently underway at MIT.

Two limitations of current piezocone data interpretation schemes that are particularly relevant to this study include the representation of organic soil behavior and the behavior of layered soils.

4.1.2.2. *Organic soils*:

Piezocones are commonly used to locate peat strata within mineral soils. Because of the high compressibility and low conductivity of peat compared to sand, contrasts in the
piezocone record are typically sufficient to differentiate the two soil types. Distinguishing peat from other soft soils using piezocone data is more difficult. The literature suggests that sleeve friction can be helpful in distinguishing peat from other soft inorganic soils (Depret, 1982; Robertson et al., 1986; Erwig, 1988; Olsen and Malone, 1988; Cheng-hou et al., 1990); however, sleeve friction data can be unreliable, especially in weak materials, and are not always collected (Jamiolkowski et al., 1985; Baligh, 1986).

We are not aware of any studies in which different peat layers have been discriminated based on piezocone data. Unlike mineral soils, different peat types are not easily categorized by consistent and substantial differences in their hydraulic and mechanical properties. The conductivity of peat spans $10^{-8} - 10^{-1}$ cm/s (see Chapter 6); penetration can be partially to fully drained. The mechanics of peat soils are complex, site specific, and highly variable. Although typically very compressible, peat can exhibit cohesion and high shear and tensile strength due to fibrous structures. Peat can also display viscous behavior (mucks).

Conventional soil models do not consider deformation mechanisms and strength characteristics that are specific to peat, such as failure of wood inclusions and tensile strength caused by interlocking of fibers and preserved roots. Even if an idealized 'granular' model of peat is assumed, the classical normalized piezocone parameters are not appropriate for data analysis, in part because the effective stress in a highly organic soil profile can be close to zero. In addition, traditional field sampling and laboratory test procedures may fail to provide adequate reference measures of strength and stiffness of peat.

Traditional normalized soil strength models are difficult to apply because the effective stress present in a soil profile of highly organic peat can is close to zero.
4.1.2.3. Layered soils

Penetration models and piezocone soil classification schemes are based on piezocone measurements that reflect steady state penetration, a condition where excess pore pressures do not change over time for any location fixed with respect to the cone geometry. Steady penetration can only occur in soil layers that are thick enough such that steady state is reached before the cone passes through the layer. Even in thick layers, as the piezocone approaches the boundary between two soil layers, measurements begin to reflect the presence of the boundary. Calibration chamber data have demonstrated that sharp boundaries often produce gradually varying piezocone response, or inflection points that do not coincide with the boundary (Kurup, 1994; Vreugdenhil et al., 1994). Van den Berg (1994) reviewed the limited literature from calibration chamber tests, finding that cone tip measurements reflect soil layer boundaries between 5 and 10 cone diameters before and after the boundary. The location of boundaries between soil types is therefore elusive even for thick layers where steady penetration is achieved within the layer.

For thin layers, the additional factors of unsteady penetration and the presence of multiple boundaries affect the ability to quantify layering. Jamiolkowski et al. (1985) have estimated that approximately 20 - 30 cm is required to reach steady penetration in clays, and 70 cm to reach steady penetration in sands. Calibration chamber results reviewed by Van den Berg (1994) indicate that steady penetration is reached in stiff layers within 36 to 72 cm and in soft layers within 7 to 15 cm. This constraint requires layers of such materials to be at least this thick if the "correct" tip resistance is to be measured within the layer. Tip resistance measured before steady state, and as the bottom of the layer is approached, will vary according to the stiffness contrast between layers. Layer thickness effects have been observed in the field by Juran and Tumay (1989), who report that thin sand layers exhibit higher pore pressures than thicker ones due to their closely confined drainage boundaries.
The sleeve load measurement is also somewhat limited in its ability to detect interface locations and thin layers because it measures skin friction over a 13 cm interval. In addition, soil that sticks to the sleeve may be carried across layer boundaries, affecting measurements.

Identification of sub-layering in the field from piezocone records is typically limited to locating transitions between soft, undrained layers, and stiffer, drained layers (i.e. clay or silt and sand). The ratio of excess pore pressure to tip resistance will approach 1 for the less permeable layers, and zero for the more permeable layers (Baligh, 1986). Identification of layering among different partially-drained soils is more difficult. The existence of sub-layers smaller than 1 m thick in partially-drained soils are sometimes inferred from ‘erratic’ or ‘scattered’ piezocone measurements; however, such layers are not typically identified.

4.1.3. Purpose and Goals

The purpose of this study is to explore the interpretability of wetland piezocone data in the absence of quantitative models of penetration in organic and highly layered soils.

The development of soil models for peat and piezocone penetration models for partially drained, layered systems will require considerable effort, and is unlikely to meet with unqualified success in the near term. On the other hand, Chapters 2 and 3 have demonstrated that the collection of piezocone data from layered wetland systems is straightforward and the data are correlated with soil layers. Although an automated, unified method of interpretation of these data will ultimately rely on successful modeling of layered soil behavior in real systems, a rational, descriptive approach has the potential to provide a useful level of interpretability.
The goal of this study is to test the hypothesis that semi-quantitative piezocone soil signatures can be used to establish the existence and continuity of (1) thin layers (2) boundaries between subtly different materials (i.e.: peat and other soils, or different types of peat).

4.2. Methods

4.2.1. Data collection

As described in previous chapters, the Wells G & H Wetland is an extensive (0.2 km²) riverine peat deposit located on the Aberjona River in Woburn, Massachusetts (Figure 2). Using the methods described in Chapters 2 and 3, a portion of this wetland was investigated in detail using a wetland piezocone and piezocone driver, as well as soil cores, revealing an often thick (8 m) section of limnic and organic soils that occupies the valley floor in this area to a distances of up to 100 m on either side of the Aberjona River (Zeeb et al., 1996). Figure 3 shows the locations of 92 piezocone profiles. Piezocone explorations are located along several cross-sections located perpendicular and parallel to the river. This distribution of measurements was chosen to better define small scale variability along these axes.

Six soil cores were obtained using a piston corer at locations identified on Figure 3. Based on these soil cores, a number of different soil layers were identified as listed in Table 1.
Figure 4-2. Map of the Aberjona Watershed showing the location of the Wells G & H Wetland.
Figure 4-3. Map of the study site showing locations of piezocone explorations and soil cores.
Table 4-1. Soil classifications in the Wells G & H Wetland

<table>
<thead>
<tr>
<th>Soil Type</th>
<th>Description</th>
<th>Organic content (by Dry Weight)</th>
</tr>
</thead>
<tbody>
<tr>
<td>P1a</td>
<td>Typha (cattail) Peat (with sandy inclusions)</td>
<td>30 - 70%</td>
</tr>
<tr>
<td>P1b</td>
<td>Sedge Peat</td>
<td>50 - 80%</td>
</tr>
<tr>
<td>P1c</td>
<td>fine-grained Sedge peat</td>
<td>80 - 100%</td>
</tr>
<tr>
<td>S/P</td>
<td>interbedded sand/peat</td>
<td>1 - 70%</td>
</tr>
<tr>
<td>P2</td>
<td>woody swamp peat</td>
<td>90 - 100%</td>
</tr>
<tr>
<td>S, S_D,</td>
<td>Sand strata within peat or Diatomaceous earth,</td>
<td>&lt;2%</td>
</tr>
<tr>
<td>pS</td>
<td>peaty Sand</td>
<td></td>
</tr>
<tr>
<td>D, sD</td>
<td>Diatomaceous silt sandy Diatomaceous silt</td>
<td>&lt;20%</td>
</tr>
<tr>
<td>S_G</td>
<td>Stratified Drift (Sand)</td>
<td>&lt;1%</td>
</tr>
</tbody>
</table>
4.2.2. Data analysis

The piezocone data were interpreted based on visual correlation of the response patterns of the three pore pressure measurements, tip resistance, and sleeve friction with each of the wetland soil layers. With the aid of the six 'ground truth' profiles, and the assumption that the layers occur in a consistent order, a number of the major soil types were readily distinguished in the piezocone records. Comparisons between the complete piezocone records for drives 94-35 and 94-50 and the corresponding soil cores are shown in Chapter 3; in this chapter we focus on piezocone response to specific soil layers in representative profiles, and present several interpreted stratigraphic cross sections that are based on the complete set of piezocone data.

Four different soil layer combinations were chosen for more extensive signature evaluation and demonstration of this technique: Typha Peat (P1a) with dense root mats, woody peat (P2) overlying sand, woody peat overlying diatomaceous silt (DE), and sand within peat. Although the wetland piezocone collects pore pressure data at three locations (see Figure 3-1), for simplicity only the tip pore pressure, tip resistance, and sleeve friction data are used to establish layer and boundary signatures.

To define these signatures, portions of a number of different piezocone drives were extracted from the complete record and plotted such that the inferred or measured boundaries between the layers of interest coincided. Table 2 summarizes the data chosen for each signature. The pore pressure and tip resistance scales were maintained within each group of plots, except for small adjustments in the tip load scale for thin layers to account (roughly) for thickness effects. The sleeve friction scale was adjusted individually by record to allow best resolution of the sleeve friction signature. This adjustment is considered reasonable due to the common measurement errors caused by intrusion of soil into the friction sleeve seals.
At this level of analysis, further data processing or rigorous correction for depth effects (stress level) and relative layer strengths is both unwarranted and unsupported by the available laboratory data. The comparisons among relatively unprocessed records serve sufficiently well to establish piezocone signatures.

Each group of sub-records includes specified record(s) for which corresponding soil core descriptions exist, such that the soil interface locations are known +/- 10 cm for this location, and marked with arrows on the appropriate plot. This error is due to vertical misalignments between the piezocone and soil core profiles caused by horizontal variability over the 30 - 50 cms that separate the two profiles, to vertical variations in wetland surface elevation, and to small errors in tracking the vertical position of the piezocone.

Based on these groups of sub-records, features were identified that are characteristic of a layer or its boundary with another layer. Characteristic feature sets were assigned as the signature of the layer or layer sequence.

The proposed signatures were then tested for each layer using records with ground truth that were not previously used in establishing the signature for that layer. Because of the limited soil core information available, some of the test records were of poor quality, but still serve to demonstrate the usefulness of this technique.

4.3. **Results and discussion**

Five major soil types were identified in the soil cores, including four different types of peat (P1a, P1b, P1c, P2), a diatomaceous silt (DE), and sand layers (S). Sandy peat (sP) layers and transitional zones (Tr) were also often identified as distinct layers in the cores.
Table 4-2. Summary of test records used in layer signature analysis. (c) indicates core data available adjacent to piezocone drive.

<table>
<thead>
<tr>
<th>Layer or sequence</th>
<th>Figure</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Typha Peat</strong></td>
<td></td>
</tr>
<tr>
<td>94-54 (c)</td>
<td>94-61</td>
</tr>
<tr>
<td></td>
<td>94-58</td>
</tr>
<tr>
<td></td>
<td>94-60</td>
</tr>
<tr>
<td>94-38 (c)</td>
<td>94-29</td>
</tr>
<tr>
<td></td>
<td>94-13</td>
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<tr>
<td></td>
<td>94-48</td>
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<tr>
<td>94-14</td>
<td>94-16</td>
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<td></td>
<td>94-49</td>
</tr>
<tr>
<td></td>
<td>94-40</td>
</tr>
<tr>
<td><strong>P2/Sand</strong></td>
<td></td>
</tr>
<tr>
<td>94-54 (c)</td>
<td>94-61</td>
</tr>
<tr>
<td></td>
<td>94-57</td>
</tr>
<tr>
<td></td>
<td>94-60</td>
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<tr>
<td>94-55</td>
<td>94-62</td>
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<tr>
<td></td>
<td>94-59</td>
</tr>
<tr>
<td></td>
<td>94-56</td>
</tr>
<tr>
<td><strong>P2/DE</strong></td>
<td></td>
</tr>
<tr>
<td>94-38 (c)</td>
<td>94-52</td>
</tr>
<tr>
<td></td>
<td>94-13</td>
</tr>
<tr>
<td></td>
<td>94-48</td>
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<tr>
<td>94-14</td>
<td>94-16</td>
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<td></td>
<td>94-49</td>
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<tr>
<td></td>
<td>94-40</td>
</tr>
<tr>
<td><strong>P/S/P</strong></td>
<td></td>
</tr>
<tr>
<td>94-38 (c)</td>
<td>94-18</td>
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<tr>
<td></td>
<td>94-51</td>
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<tr>
<td></td>
<td>94-40</td>
</tr>
<tr>
<td>94-54 (c)</td>
<td>94-52</td>
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<tr>
<td></td>
<td>94-12</td>
</tr>
<tr>
<td></td>
<td>94-60</td>
</tr>
</tbody>
</table>
Table 1 summarizes these soil types and their characteristics. The woody peat (P2) is distinguished from the fine-grained sedge peat (P1c) based on the changing assemblage of macrofossils. However, both layers are characterized by very high organic content (> 90%) and the piezocone seems to respond to them as a single layer. Therefore these layers P1c and P2 are referred to collectively in the following discussions as the woody peat (P2). Soil core profiles are included as Figures 4 - 9.

The results of additional soil testing are presented in Chapter 5 and 6. Bialon (1995) presents the full results of a number of laboratory tests of this material, including compressibility and permeability measurements. The full set of piezocone records are contained in Appendix A. Typical piezocone response to each of the major soil types is described in Chapter 3, and will not be re-iterated here. The following discussion focuses in turn on each of the signatures selected for further consideration.

4.3.1. Signatures of selected soil types/boundaries

4.3.1.1. Upper Typha peat with root mats

This soil layer is the most extensive layer found in the wetland, and similar piezocone response was observed at nearly all test locations. The currently existing vegetation in the study area is predominantly Typha (cattail), Phragmites, two wetland macrophytes that grow to substantial heights, and have rigid and substantial stems. Abundant remains of these species are found in the upper wetland soils.

Figures 10 - 12 show twelve representative sets of piezocone data from the upper 90 cm of the wetland. The sleeve friction scale has been adjusted in several cases to compensate for suspected mechanical drift in this measurement. Otherwise, the scales are the same in

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Figure 4-4. Core 17. See Table 4-1 for legend. Organic content is measured as fraction of dry weight.
Figure 4-5. Core 34. See Table 4-1 for legend. Organic content is measured as fraction of dry weight.
Figure 4-6. Core 38. See Table 4-1 for legend. Organic content is measured as fraction of dry weight.
Figure 4-7. Core 46. See Table 4-1 for legend. Organic content is measured as fraction dry weight.
Figure 4-8. Core 50. See Table 4-1 for legend. Organic content is measured as fraction dry weight.
Figure 4-9. Core 54. See Table 4-1 for legend. Organic content is measured as fraction of dry weight.
Figure 4.10. Piezocene records from Typha peat.
Figure 4.11. Piezocene records from *Typha* peat.

- Tip Pressure (MPa)
- Sleeve Load (KPa)
- Depth (cm)

[Graphs showing pressure and load variations at different depths for 94-48, 94-13, 94-29, and 94-38 samples.]
Figure 4-12. Piezocone records from *Typha* peat.
all plots. Because the depth of the features in these profiles is so consistent, the axes of these plots were not adjusted to make the features correspond any closer than they already do.

Two peaks in sleeve friction and tip resistance are apparent in all of these plots. In most cases an increase in tip pore pressure leads each tip resistance/sleeve friction peak slightly, although pore pressure is generally quite variable throughout the interval. Tip resistance peak values are consistently between 0.3 and 0.5 MPa.

The two peaks in these profiles are correlated in several locations with dense mats of preserved *Typha* roots and stems. At location 54 only one mat was detected in the soil core, but recovery of this particular portion the core is difficult. The resistance of these layers is easily detected by feel during sampling, but often not evidenced in the sample, presumably because the root mat is pushed to the side by the corer.

We believe that these piezocone data reflect the penetration resistance of the *Typha* mats, and are evidence that these mats are extensive and relatively consistent in thickness and depth. The increase in pore pressure above the leading edge of each tip resistance peak is likely a result of the increased total stress as the cone approaches the relatively stiff boundary. The lowered pore pressure commonly seen after the peak may be the result of increased soil drainage within the root mat. Similar behavior is reported by Juran and Tumay (1989) for piezocone data collected across sand inclusions in clay. The lower root mat occurs at the base of the *Typha* peat, beneath which is a well-preserved sedge peat (see Chapter 5 for a more complete description), so that the lower peak is interpreted to represent the boundary between these two peat layers.
4.3.1.2. Woody peat overlying sand

The eight records in Figures 13 and 14 illustrate the piezocone response in a layer of woody peat that is found across much of the site. Ground truth adjacent to drive 94-54 (Core 54, Figure 9) verifies the presence of this layer between the two arrows on the plot. In 94-54, the woody peat overlies a sand layer. A distinct increase in tip resistance at the peat-sand boundary is apparent in all of the eight profiles. In each case, a peak in pore pressure occurs at the peat-sand boundary before pore pressure falls to consistently low levels; this feature was used to align the base of the layer in each of the plots.

Calibration chamber data collected by Vreugdenhil et al. (1994) show tip resistance beginning to increase approximately 3 - 5 cone radii before the interface between a soft and stiff layer, and steady state tip resistance becoming re-established approximately 10 cone radii below the interface. Whereas our data do show a gradual increase in tip resistance, the likely location of the soil interface is near the start of this increase rather than 10 cm below it. It is possible that the interface is actually about 10 cm below where it is indicated in the adjacent core, although pore pressure data suggest a sharp boundary at the beginning of the tip resistance increase. The lack of an idealized layer boundary (such as that constructed in a calibration chamber) with a similar stiffness ratio may be the cause of this disagreement.

The peat layer itself is characterized by elevated and often very dynamic pore pressures. Sleeve friction is also elevated in this layer; whereas tip resistance is uniform and low throughout this layer in these examples. Although the sleeve friction measurement appears to be diagnostic of this layer, it is not useful in determining layer boundaries due to its integrated response. A rise in pore pressure and elevated sleeve friction was used to correlate the upper boundaries in these plots. One exception is drive 94-56, which shows a sharp decrease in sleeve friction and elevated tip resistance at the top of the layer. This
Figure 4-13. Piezocone records from Woody peat overlying sand. Hydrostatic pressure shown as dotted line.
Figure 4-14. Piezocone records from Woody peat overlying sand.
behavior is caused by the presence of a sand inclusion near the upper boundary of this layer with highly elevated tip resistance and sleeve friction.

4.3.1.3. Woody peat overlying diatomaceous silt

In Figures 15 and 16 eight additional piezocene records representative of the woody peat are presented. The first record in Figure 15 (94-38) has associated ground truth that verifies the layer boundaries. In this case, the woody peat has much the same characteristics as in the previous example, except that its lower boundary is a transition to diatomaceous silt. The boundary between the woody peat and the diatomaceous silt is difficult to locate even in soil cores, and consists of a gradual decrease organic content and increase in diatom frustules. Not until the organic content is known can the transition zone between the two soils be confidently located. In the piezocene data, the transition zone is correlated with a decrease in both pore pressure and sleeve friction, and continued flat tip resistance response (as opposed to the rising resistance when entering the sand layer. The pore pressure response of the underlying diatomaceous earth is also distinguished from the peat in that it lacks the large magnitude higher frequency variability typified in records 94-38, 40, 48, and 49 (in some cases this response is damped by degraded pore instrument response).

In three of these records, (94-16, 49, and 38) a sand or sandy peat layer forms the upper boundary of the woody peat. In two cases (94-16 and 49) sand inclusions are also hypothesized within the peat layer, each marked by an increase in tip resistance coupled with a large increase and decrease in pore pressure. Sand layers may also be present in 94-13. No appreciable friction sleeve response is seen for these sand inclusions, either because the scale of the friction sleeve measurement (13 cm) is too large, or there is insufficient contrast in the friction sleeve response of the two materials.
Figure 4-16. Piezocone records from Woody peat overlying Diatomaceous Earth
These records show that a signature correlated in several places with the woody peat layer is consistent regardless of the nature of the upper or lower boundary (sand, sedge peat, or diatomaceous silt), although internal layers of sandy soils can significantly alter the signal. The nature of the peat/diatomaceous silt boundary, while not exact, is consistent.

4.3.1.4. Sand layers within peat

The detection of sand layers within peat soils is further investigated using the data presented in Figures 17 and 18. Figure 17 shows sand layers 20 to 50 cm thick; Figure 18 shows sand layers from 5 to 15 cm thick. Again, the first record presented in each figure is correlated with soil core data to provide a reference for the soil layer boundaries. The thicker sand layers generate tip resistance on the order of 1 MPa. A peak in pore pressure marks the approach to the sand layer boundary, within which pore pressure is reduced and consistent. Except for the thickest of these layers (94-40), sleeve friction data do not provide any helpful information.

The pore pressure and tip resistance responses are in general agreement with data from other studies (Juran and Tumay, 1989; Kurup, 1994; Vreugdenhil et al., 1994). The brief increase in pore pressure as the sand lens is approached may be the result of increased total stress near the relatively stiff boundary. Once the sand lens is penetrated, however, the well-drained condition dissipates excess pore pressure readily. If the sand is overconsolidated, it may also dilate as it is sheared, causing further decreases (possibly negative) in pore pressure. Van den Berg (1994) measured the same pattern of pore pressure response for clay over sand, although the peak in pore pressure occurred several centimeters above the boundary.

In 94-51 and 94-40, several additional peaks in pore pressure are correlated with inflection points in tip resistance and are hypothesized to represent peat interbeds in the
Figure 4-17. Piezocone records from thin (6 - 15 cm) sand layers within peat. Note sand/peat interbeds in 94-40.
Figure 4-18 Piezocone records from thick (20 - 50cm) sand layers within peat.
sand layer. Abrupt decreases in pore pressure accompany each resumed increase in tip resistance. The hypothesized peat sub-layers are indicated at 72 to 78 and 99 to 104 cm in drive 94-40 using thin dashed lines.

In the plots of the thinner sand layers, the tip resistance scale has been substantially compressed to offset the effects of decreased layer strength. The features that distinguish these layers are much the same as the thicker layers, although the peaks in tip resistance and pore pressure are less pronounced. The one notable exception is the pore pressure peak at 132 cm in 94-52, which was plotted at a significantly expanded scale. The sand lens in this record occurs within the woody peat rather than the sedge peat, as the others do. Apparently the lower conductivity of this peat causes an enhanced pore pressure response as the sand layer is approached.

The fact that the thicker sand lenses in these examples cause hydrostatic or slightly negative pore pressures, whereas the thinner layers cause substantially negative pore pressures is not what one would expect. Juran and Tumay (1989) observed higher tip pore pressure in thin sand inclusions relative to thick layers, and point out that the close drainage boundaries for thin sand lenses should inhibit pore drainage relative to thicker layers. Several factors could explain our counter-intuitive results:

- the thinner sand layers (which are found at generally shallower depths) could be more overconsolidated than the thicker layers, and therefore more dilative

- the thicker sand layers could be more permeable than the thin layers

- the generally high variability of the shallow interbedded soils could be causing very complex drainage and stress controls that do not fit this simple model of layered penetration
It is likely that this last factor is at least somewhat applicable; however, the available data do not support assessment of *in situ* conditions at the level of detail necessary to resolve this issue.

### 4.3.2. Testing signatures against core data

The signatures described above can be applied to test records that were not used to establish the signatures. Figures 19, 20, and 21 show portions of the test records, which have been plotted in segments with scales assigned to each segment to make comparison with the signatures convenient.

In the record for drive 94-50 (Figure 19), the characteristic sleeve friction and tip resistance peaks are evident at 15 - 25 cm and 60 - 70 cm, whereas dense root mats were detected in core 50 at about 30 and 62 cm. To the extent that the lower root mat can be located in the piezocone data, the interface between *Typha* and Sedge peat (measured at 67 cm in core 50) can be identified, despite the fact that piezocone response in the sedge peat itself is neither consistent or distinctive.

Deeper in core 50, a zone of increased excess pore pressure exists between approximately 145 and 240 cm. Substantially elevated sleeve friction occurs within this zone, while tip resistance is very small except for a peak at 190 cm that is accompanied by the only negative excess pore pressures in this zone. Based on the signatures collected in Figures 13 - 16, this portion of the record is interpreted as woody peat, beginning between 140 and 150 cm where pore pressure begins to rise and shortly thereafter sleeve friction, and ending at about 218 cm where pore pressure drops sharply. Like 94-16 and 94-49, this layer of woody peat appears to contain a sand inclusion between 186 and 191 cm. The data from core 50 verify the existence of the P1c/P2 high organic peat in this location, although no sand inclusion was observed. Such thin layers can often be discontinuous however, and the core and piezocone profile are separated by as much as 0.5 meters at the
Figure 4-19. Portions of piezocone record for Drive 94-50. See previous figures for legend.
Figure 4-20. Portions of piezocone record for Drive 94-34. See previous figures for legend.
Figure 4-21. Portions of piezocone record for Drive 94-17. See previous figures for legend.
surface; each can vary slightly from vertical as well. The fact that a duplicate core later obtained for $^{14}$C analysis did not intercept the sand inclusion at approximately 120 cm is evidence for this variability.

The piezocone record for 94-34 (Figure 20) displays a series of offset peaks in tip resistance and pore pressure typical of sand inclusions in softer, less permeable material. Between 105 and 220 cm seven individual sand layers can be identified based on the characteristic response described in section 4.3.1.4. The breadth and magnitude of the peaks in tip resistance indicates a thick sand layer with peat inclusions rather than a thick peat layer with sand inclusions. The largest pore pressure peaks occur at the base of the thickest peat layers (at 108 and 165 cm— an expected pore pressure peak at 145 cm was obscured by a break to add drill rods), where drainage is most restricted prior to entering a sand layer.

Beginning at 112 cm and ending at 220 cm, the core data from this location shows thirteen interfaces between sand, peaty sand, and sandy peat, for a total of 7 transitions between layers contrasting permeability and stiffness (Figure 5).

The last peat-sand interbed (peat from 195 to 203 cm and sand from 203 to 215 cm— repeated at the top of the second plot in Figure 20) is accompanied by rising sleeve friction and followed by a broad area of elevated pore pressure and relatively low tip resistance. The start of consistent pore pressure rise (213 cm) and the return to lower pore pressures and sleeve friction (272 cm) mark the boundaries of a typical woody peat signature, matching the core data quite well. In this case the upper P1c peat, usually mapped with P2 by the piezocone, is absent in the core.

The final test record, drive 94-17 (Figure 21), shows greatly damped pore pressure response due in part to an electrical problem, but also possibly poor saturation of the porous stone. Nevertheless, the typical patterns associated with the wetland soil layers
are discernible, and adequately predict the coring data. The two Typha peaks occur at 15 - 25 and 60 - 75 cm. Although no dense root mats were mapped in this core due to poor recovery, the boundary between Typha and Sedge peat falls at 75 cm, corresponding with the bottom of the second peak (Figure 4).

A series of tip resistance peaks similar to those in drive 94-34 is present in drive 94-17 (Figure 21) from 100 - 230 cm (the pore pressure variations are present but damped). Interbedded peat and sand are mapped in the core over this same interval. The underlying woody peat signature is also present, although the degraded pore pressure data prevent its boundaries from being located accurately. The core shows high-organic peat from 220 to 300 cm.

4.3.3. Conceptualized piezocone signatures

Figure 22 shows conceptualized signatures for the soil layers and interfaces identified at this site. These generalized illustrations emphasize typical patterns of tip resistance, tip pressure, and sleeve friction that are described in greater detail in previous sections. In a), coincident peaks in tip resistance and tip pressure make the locations of two dense root mats in the Typha peat (P1). In b), elevated sleeve friction and tip pressure distinguish the woody peat (P2) from the herbaceous peat above, and an increase in tip resistance indicates the interface between peat and sand. In c) consistently low tip resistance and sleeve friction, as well as steadily rising tip pressure characterize the diatomaceous silt. The woody peat is distinguished from underlying diatomaceous earth by a drop in pore pressure and sleeve friction in d). Thicker (20 - 50 cm) sand layers (e) generate a marked increase in tip resistance, while causing pore pressure to approach the hydrostatic values. Thinner sand layers (f) cause only slight increases in tip resistance, while pore pressure drops beneath hydrostatic values.
Figure 4-22. Conceptualized piezocone signatures for soil layers and interfaces encountered in the Wells G & H Wetland. Horizontal dashed lines represent soil layer interfaces. Vertical scales are approximate, and vary from 50 to 150 cm. See Table 4-1 for soil type legend. Discussion of the specific features of these signatures is found in section 4.3.1.
Figure 4.23  Cross sections A-A' and B-B' as interpreted from piezocone data
It is important to bear in mind that the conceptualized patterns illustrated in Figure 19 do not represent actual data, and are abstractions from which there is considerable variation. These generalized signatures should be used to guide interpretation with the knowledge that a number of confounding factors (such as layer thickness, ordering, and interbedding) can cause variability in piezocone response.

4.3.4. Stratigraphic cross sections

Using most of the piezocone profiles, the cross sections shown in Figure 23 were constructed. Some of the thinner (< 20 cm) sub-layers such as those described in section 4.3.1.4 are not shown here. For example, the sand layer shown from 1 to 2 m depth in section A - A' is highly interbedded with peat close to the existing riverbed (see cores and piezocone profiles 17 and 34).

To a certain extent, data interpretation along these sections was guided by the expectation of consistent layer sequences from profile to profile, in keeping with a reasonable depositional model. However, two additional points should be made in this regard: (1) the signature data as described in section 4.3.2 did support the construction of a reasonable stratigraphic model; and (2) the data indicate the location of 'pinch outs' in several locations.

4.4. Conclusions

The correlations between well-defined transitions in the piezocone data records and the measured soil boundaries, as well as the consistent response among various records, is evidence that this piezocone reliably detects the various soil layers found in this wetland. Because of the inherent difficulty in verifying the exact position of soil boundaries penetrated during any particular piezocone drive, it is not possible to measure the response of the piezocone to thin layers in the field with complete objectivity. Nevertheless, where
data are available, a consistent response to thin layers is observed, and numerous examples of similar (although unverified) response patterns can be identified in the records included in Appendix A.

The upper Typha peat (P1a), the woody peat (P2 and the overlying P1c layer), sand layers, and the transition from woody peat to diatomaceous earth (DE) and to sand all produce recognizable signatures documented in these figures. Sand layers within the diatomaceous earth were not explicitly dealt with in the previous discussion, but review of the piezocone records in Appendix A shows that such layers are easily identified by this technique (see records such as 94-48, 16, and 51).

Piezocone response is quite variable in the sedge peat (P1b. Although this layer itself cannot be differentiated on the basis of a characteristic signature, its top is readily identified by the lower of two distinctive root Mats in the overlying Typha peat, and its lower boundary with either sandy soils or the woody peat is also distinctive. The principal of uniformitarianism and a reasonable soil formation model are important components of this interpretive framework, although rational interpretation of piezocone response and the correlation of unique piezocone signatures with soil layers are most important.

The results of this study demonstrate that various peat soils, as well as sandy units and an organic, diatomaceous silt can be identified and mapped based on their piezocone signatures. The values of pore pressure, sleeve friction, and tip resistance measured for these soils vary in a way that appears reasonable, given our basic knowledge of the structure and mechanics of these soils, and the effects of layer thickness. While some aspects of the soil signature are not immune to the effects of layering, the method of signature identification reduces the impact of these effects, and allows inclusion of interface effects as part of the signature.
Because of the difficulty of accurately locating soil interfaces coincident with piezocone profiles, the signatures as described are not certainties, but our best estimate of how the piezocone responds to these layers. These conclusions are based on adjacent soil core data, a limited amount of published calibration data, and the current level of understanding of penetration near and across boundaries.

Although the highly specific response of this piezocone to different wetland soil layers at our site provided makes it an effective profiling tool, and also makes the prediction of its performance at other sites difficult. The detection of root mats, highly organic peat layers, silts, and sand layers is expected to be straightforward once ground truth is established, but additional peat types and structures may generate responses that are not as well correlated.

The consistency of response across this site reinforces our hypothesis that these signatures are real and representative; however, laboratory chamber testing will be required to establish boundary responses with accuracy and confidence. An extensive chamber testing program is necessary to establish piezocone response near interfaces and to fully test the effects of layer stiffness ratio and layer thickness for mineral soils. Additional testing will also be required to establish the modes of penetration for interbedded peat soils. Unfortunately, replication of field conditions in test chambers is laborious and inexact even for mineral soils. Development of peat surrogates for chamber tests will present new, unanticipated challenges. Progress in this area will require considerable effort and time.
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5. Post-Glacial and Industrial History of the Wells G & H Wetland: Implications for Toxic Metal Transport

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5.1. Introduction

The Wells G & H Wetland is an extensive (0.2 km²) riverine peat deposit located along a low gradient portion of the Aberjona River in Woburn, Massachusetts (see Figure 1). The wetland overlies up to 50 m of permeable sands and gravels. This aquifer was used for municipal drinking water (pumped from Wells G & H) from 1964 - 1979.

Aberjona watershed has a long and well-documented industrial history. From the mid-1800s to the 1930s leather tanning was the dominant industry in Woburn, and the Aberjona and its tributaries the main conduits for tannery wastewater (Durant et al., 1990). During the same period, starting in the late 19th century, sulfuric acid and arsenical pesticide manufacturing took place near the headwaters of the watershed in an area now designated as a federal Superfund site known as 'Industriplex' (Aurilio, 1992; EPA, 1986). Aurilio (1992) estimates that 270 metric tons of arsenic could still be present at this site. The Wells G & H wetland itself is located at the center of a second Superfund site, added to the National Priority List in 1982 due to the widespread presence of organic solvents in groundwater (NUS, 1986).

The Wells G & H Wetland first gained notoriety in the early 1980's when contaminated drinking water from the adjacent wells was blamed for a cluster of leukemia cases (Lagakos et al., 1986; Costas et al., 1996) The long-polluted Aberjona River has since been shown to carry significant quantities (100 kg/yr) of arsenic from its headwaters to its outlet at the Mystic Lakes (Solo, 1995). The river-borne metal load is hypothesized to have been much greater in the past, depositing over 10 tons of arsenic in the Mystic Lakes during two particularly significant episodes in the 1930's and 1960's (Aurilio et al., 1994; Sliethoff and Hemond, 1996). Because the municipal wells pumped in the
Figure 5-1. Map of the Aberjona Watershed showing the Wells G & H Wetland.
underlying aquifer derived about half of their recharge from infiltrating river water, the relationship of the river to the underlying aquifer, presumably mediated by the wetland, has become a compelling topic for study (Myette et al., 1987). This question gained a further dimension when high levels of arsenic were discovered in the peat soils of the wetland in 1992 (Zeeb, 1994).

As part of an effort to understand the mechanisms responsible for distribution of arsenic in the soils of the Wells G & H Wetland, a study of its hydrogeologic history was initiated. In this study, portions of an extensive wetland that is part of the Aberjona River Watershed were investigated using a number of techniques, including soil coring, soil-profiling with a specially-designed piezocene penetrometer, radiocarbon dating, and paleoecological analysis using pollen counts. Field investigations were designed to resolve wetland features on the scale of several centimeters in the vertical and a meter in the horizontal dimension. Accordingly, data were collected along a number of profiles rather than on a broader grid. This chapter presents the results of these investigations, and an interpretation of the post-glacial history of this deposit based on our findings and the published post-glacial history of this region. The techniques used to obtain and interpret the piezocone data are fully described in previous chapters; they will not be discussed here (Zeeb et al., 1996).

5.2. Regional Post-glacial History

5.2.1. Regional models of deglaciation

Two modes of deglaciation of New England have been proposed; systematic ice retreat and regional stagnation. Koteff and Pessl (1981) demonstrate that there is justification for elements of both processes in observed landforms. Whereas a classical series of end
moraines demonstrating a rhythmic and systematic retreat of a live ice sheet is not readily observable in New England, occurrence of discontinuous live-ice deposits, southward sloping outwash plains, and meltwater-trimmed ice contact deposits does not support regional stagnation. One explanation for this apparent contradiction is a systematic retreat of the active ice margin, bordered by a narrow (several km) band of stagnant ice, the morphology of which was controlled by the irregular topography typical of New England. The extent of local topographic control increases from southeastern Massachusetts, where little physiographic variation exists, to the northeast, where ice-margin retreat features indicate that the “pattern of ice-margin retreat is related to the uncovering of successively lower and younger drainage paths, or spillways” (Stone and Peper, 1980).

The morphosequence model encapsulates this view of ice-retreat, providing mechanisms for the formation of sequences of synchronous glacial landforms formed in front of an ice margin with or without an associated end moraine. In this model, the presence of ice contact features and the sequence of lacustrine and fluvial deposition are controlled by the regional till/bedrock slope and the height of base-level control. Because these factors vary so greatly in New England, the presence of continuous features diagnostic of particular glacial stages is limited.

Morphosequence/deglaciation chronologies often rely on minimum-limiting radiocarbon dates for organic material that accumulates in glacial sedimentary depressions following ice retreat. As Davis and Davis (1980) point out, there can often be a significant time lag between ice-retreat and accumulation of datable material, due often to the persistence of stagnant ice blocks, and in some upland areas, the piping of sediments through bouldery basal material until depressions are finally sealed and begin to accumulate water and sediments. Thus basal radiocarbon dates can be many years (Davis and Davis find up to 9000 year lags) younger than the time of deglaciation. Incorporation of detrital organic matter containing ‘old’ carbon from previous inter-glacial periods can also cause radiocarbon dates that are too old. Based
on a number of dates considered reliable as minimum limiting markers, the final ice retreat from northeastern Massachusetts is believed to have been completed by 14,000 radiocarbon years before present (ybp) (Oldale et al., 1993; Davis and Jacobson, 1985). By 12,000 ybp the ice-sheet is believed to have receded as far as northern Maine.

Whether morphosequences found in eastern Massachusetts include glaciomarine landforms is determined by variations in local relative sea level, as well as the particular coastal environment (i.e. presence of abundant sea ice, which can prevent beach deposition). Relative sea level curves for coastal Maine and Massachusetts show that crustal rebound lagged eustatic sea-level rise (caused by glacial melting) sufficiently to cause inundation of a large portion of the Maine coast and parts of Eastern Massachusetts between 15,000 and 13,000 ybp. Sparse evidence of late glacial relative sea-level position at Lynn, Massachusetts and in the Merrimac River Valley suggest that this area was as much as 18 to 30 m below sea level circa 14,000 ybp, when active ice had retreated as far as the Maine coast (Kaye and Barghoorn, 1964; Oldale et al., 1993; Davis and Jacobson, 1985). By 12,000 ybp, crustal rebound caused a drop in relative sea level as much as 40 m below its present-day position. Slower, longer term sea-level rise brought relative sea level within 2 m of its present position by 3000 ybp (Oldale et al., 1993).

5.2.2. Geologic setting and history for the Fresh Pond Valley

Chute (1959) identifies a sequence of four outwash deposits located within the Fresh Pond Valley, bordered by ground moraine that covers the uplands. Chute hypothesizes that the formation of the latter three of these outwash deposits followed a re-advance of the local ice-sheet that formed the Fresh Pond End Moraine located about 10 km south of the site, at the mouth of the valley (see Figure 2).
Figure 5-2. Surficial Geology of the Fresh Pond Buried Valley (adapted from Chute, 1949).
The second and third outwash sequences were deposited while a large ice mass still existed in the lowland between the Mystic Lakes and Fresh Pond. Wedge Pond and Winter Pond, both located roughly 5 km south of the Wells G & H Wetland, are cited as examples of ice block depressions left in the third outwash. The Wells G & H Wetland is mapped as a swamp deposit within sand and gravel attributed to the third outwash.

Subsequent to the deposition of the third outwash, ice occupying the lowland between Mystic Lakes and Fresh Pond melted, and the remaining depression was filled with clay. Clay deposits up to 17 m thick are found north of the Fresh Pond Moraine, indicating the presence of an embayment in this area subsequent to the melting of local ice and the deposition of the third outwash. The maximum elevation of the top of this clay deposit is from 8 to 12 m above current sea level, indicating the possible extent of coastal inundation at this time.

Evidence of further marine incursion of the Fresh Pond Valley does not exist. Chute hypothesizes that the coastal waters were interrupted by large blocks of ice, preventing the formation of preserved and recognizable shoreline sedimentary features. Although debate exists as to the marine or non-marine origin of the Fresh Pond Valley clay, Chute believes that it was deposited in marine or brackish water of substantial depth. Regional uplift eventually drained the embayment, causing clay deposition to end.

The final outwash deposit in this sequence was deposited as a large alluvial fan entirely above the then-existing sea level. A number of ice blocks still existed during and after the deposition of the outwash and clay, leaving depressions now occupied by lakes, including Spy Pond and the Mystic Lakes. The source of the sediments making up the fourth outwash was probably the paleo-Aberjona River, flowing through the Fresh Pond Valley across the sand and gravel deposits of the third outwash that underlie the Wells G & H Wetland. The accumulation of the wetland soils began soon after this time.
A number of soil borings, advanced as part of state and federal hazardous waste site investigations starting in 1979, penetrated the wetland soils in the Wells G & H peatland. Although the wetland soils are not described in detail, their total thickness is documented, and ranges from a meter to over 8 meters (NUS, 1986). This evidence suggests the presence of ice block depressions in the Wells G & H Wetland that were later filled with sediments.

5.3. Methods

This study is focused on one portion of the Wells G & H Wetland, where it makes its closest approach to Well H. Figure 3 shows the locations of various samples and in situ tests, including 92 piezocone profiles, six descriptive soil cores, five soil cores for arsenic analysis, and seven soil cores for paleoecological analysis.

5.3.1. Soil cores

Soil cores were obtained using several techniques. The shallow (<1 m) peat was sampled using either a 10 cm diameter polycarbonate fixed piston corer with a sharpened, serrated cutting head, or a Russian peat corer. These were the only devices that could penetrate and obtain representative samples of the dense and strong root mats encountered over this interval. Below the root mat horizons, a small diameter (2.5 cm) incremental piston corer was used. Designed and built especially for this project, this device retrieves a 30 cm long core into a plastic (cellulose butyrate) sleeve which is removed from the core barrel and capped in the field. Using the same hole, progressively deeper peat horizons can be sampled by hand to depths of over five meters. Complete core recovery was achieved with this method for loose sands, silts, and moderately decomposed peat. Recovery of relatively undecomposed mats of root and stem material found in the upper peat was not always achieved.
Figure 5-3. Map of the Wells G & H Wetland showing locations of explorations.
Wetland soil cores were sub-sampled in the lab for botanical and textural classification, as well as measurement of water (drying at 85 degrees C) and organic content (loss on ignition). Each core was sub-sectioned at intervals of approximately 10 cm as well as directly above and below stratigraphic boundaries. Selected sub-samples were also analyzed for total soil arsenic. Total soil arsenic was determined by digesting soil sub-samples using a hot nitric/sulfuric acid and hydrogen peroxide process and analysis of the digestate using a Perkin Elmer 4100Z graphite furnace Atomic Adsorption Spectrophotometer. Several mineral soil samples were also analyzed using x-ray diffraction and fluorescence.

Selected soil core sub-samples were radiocarbon dated by Geochron, Inc. (Cambridge, MA). Bulk sediment was dated using conventional $^{14}$C analysis techniques; individual plant remains were dated using accelerator mass spectrometry (AMS). A total of eleven samples were dated: eight from core 94-50 and three additional samples from cores 94-34 and 94-17. Pretreatment for all $^{14}$C analyses included successive washes with 1.0 N HCl, 2% NaOH, and dilute H$_3$PO$_4$.

5.3.2. Piezocone stratigraphic cross sections

Soil type cross-sections interpreted from piezocone data were synthesized from piezocone soil profiles along several sections of the wetland. The methods used to obtain and interpret the piezocone data are described in Chapters 2 through 4 (Zeeb et al., 1996). Ground truth was provided by the soil cores described above. In several areas the presence of sandy strata in the subsurface was further inferred using a three-meter-long steel rod pushed by hand into the wetland soils until the resistance of the sand stratum was felt. A network of these probings were used to define the morphology of a particular sand stratum located between 80 and 160 cm below the surface.
5.3.3. Paleocology

Further interpretation of the botanical sequence of the peat profile, as well as pollen counting was completed by Orson Environmental (Branford, CT). A separate set of seven peat cores was obtained for this purpose using a Russian peat corer. Pollen counts of arboreal and non-arboreal species were completed to a depth of 3.3 meters in Core H2 (adjacent to Core 5u) and pollen abundances were plotted versus depth (Figure 4). Regionally synchronous events in the pollen record are rare; only several exist which can be useful in dating sediments. In most cases, pollen assemblage data can provide information about local vegetation and climate, but because they typically represent time-transgressive environments, they do not provide age information (Gaudreau and Webb, 1985).

5.4. Results

5.4.1. Stratigraphy

At least three different types of peat, a limnic sediment, and sandy strata have been identified in the wells G & H Wetland (Table 1). Stratigraphic cross-sections reveal a seven meter deep, steep-sided depression in the underlying glaciofluvial sands and gravels (Figure 5). The depression is 30 m across in the north-south dimension, and over 40 m across in the east-west dimension. At its eastern margin the wetland has been artificially filled, such that the subsurface in this area could not be fully investigated using the wetland piezocone. Hand probing data from the present eastern edge of the wetland, adjacent to the former Well H, indicate the presence of at least 3 meters of soft soils below a veneer of sand and gravel.
Figure 5-4a. Pollen abundance at Core H-2. Non-arboreal pollen.

Pollen profiles for Core H2 on Transect H. Pollen counts are expressed as total grains and reported in increments of 5,000 grains, except where noted. Dark shading denotes counts multiplied by a factor of 10. Bars show actual counts and connecting line represents extrapolations between depths. Total grains calculated based on recovered Lycopodium spores. Summary diagram based on percentage of all grains. Poaceae counts are in increments of 5,000 and separated by the size classes shown in insert. Sediment key is provided.
Table 5-1. Soil Types identified in the Well G & H Wetland

<table>
<thead>
<tr>
<th>Soil Type</th>
<th>Description</th>
<th>Organic content (by Dry Weight)</th>
</tr>
</thead>
<tbody>
<tr>
<td>P1a</td>
<td>Typha Peat (with sandy inclusions)</td>
<td>30 - 70%</td>
</tr>
<tr>
<td>P1b</td>
<td>Sedge Peat</td>
<td>50 - 80%</td>
</tr>
<tr>
<td>P1c</td>
<td>fine-grained Sedge peat</td>
<td>80 - 100%</td>
</tr>
<tr>
<td>S/P</td>
<td>interbedded sand/peat</td>
<td>1 - 70%</td>
</tr>
<tr>
<td>P2</td>
<td>woody swamp peat</td>
<td>90 %</td>
</tr>
<tr>
<td>S&lt;sub&gt;p&lt;/sub&gt;, S&lt;sub&gt;D&lt;/sub&gt;, pS</td>
<td>Sand strata within peat or Diatomaceous earth, peaty sand</td>
<td>&lt;2%</td>
</tr>
<tr>
<td>D, sD</td>
<td>Diatomaceous earth sandy Diatomaceous earth</td>
<td>&lt;20%</td>
</tr>
<tr>
<td>S&lt;sub&gt;G&lt;/sub&gt;</td>
<td>Stratified Drift (Sand)</td>
<td>&lt;1%</td>
</tr>
</tbody>
</table>
Figure 5-5. Cross sections A-A' and B-B' through the Wells G & H Wetland.
Available boring logs and other records associated with Well H are contradictory, and suggest that as little as 1 m or as much as 6 m of organic soils were originally encountered here (D.L. Maher, 1964; Dufresne - Henry, 1978; Ecology and the Environment, 1981). The high organic content, occasional bacterial contamination, and high salinity of well water led to discussion in the late 1960s of the possible excavation of organic soils located beneath the fill surrounding Well H (Whitman and Howard, 1968). The extent of these organic soils is unknown, although Dufresne and Henry (1978) report that all of the organic soil within 30 feet of each wellhead was excavated some time between 1968 and 1976, and replaced with clean “bank-run” sand and gravel. It seems certain that the wetland originally extended to the site of Well H. It is uncertain how thick wetland soils were in this locations and to what extent the fill created and additional seal around the casing.

Additional piezocone data as well as soil boring data indicate that peat strata of over a meter in thickness exist over the broader area indicated by the wetland boundaries in Figure 3 (NUS, 1986). The wetland boundaries include the low-lying land adjacent to the Aberjona River, extending approximately 100 m to the east and west of the riverbed. One portion of upland exists within this area west of the river: a small (shaded) oval about 50 m across shown in Figure 3. A sample of the top 50 cm of soils in this area reveals 30 cm of loam overlying medium to coarse sand with traces of gravel.

The depression delineated in the cross sections is filled to its edges with diatomaceous silt. This soil has a distinctive olive color and contains the remains of a number of freshwater diatom species indicative of slightly acidic or circumneutral waters (Donald Charles, Academy of Natural Sciences of Philadelphia, personal communication, 1995). Traces of sand and clay-size material are also present. X-ray diffraction/fluorescence and scanning electron microscopy analyses of this sediment reveal its primarily siliceous composition; only traces of clay minerals are present (Joseph Adario, personal communication, 1995; Bialon, 1995). The water content of this soil varies from 60 - 80 % by weight, the organic fraction of
the solids varies from 5 to 30 %, and bulk density varies from 1.1 to 1.3 g/cm³. Within this deposit are several thin sand lenses near its bottom and a thicker, more extensive sand stratum that dips into the deposit from its northern margin. The thicker lenses are composed of a medium-grained sand with organic content of less than 1 %.

Overlying the entire diatomaceous deposit as well as the sand strata is a 0.5 to 1 m thick layer of woody peat. The woody peat is distinguished by a high water and organic content (both typically above 90%) and a distinctive orange - brown color that darkens substantially within 30 minutes of exposure to air. Small, relatively undecomposed twigs are found in the woody peat, as well as numerous unidentifiable woody fragments in a very friable matrix that includes sedge stems and roots.

Above the woody peat are several herbaceous peat strata, including a sedge peat of varying degrees of humification, containing abundant and readily identifiable sedge roots and stems. Occasional well-preserved layers are found, but generally fragments are less than several centimeters in length and the material is dark brown and friable. In the deeper parts of this layer, water and organic content are similar to the underlying woody peat. In the upper part of this layer, water and organic content vary between 80 - 90 and 50 - 80%, respectively. The variation in these parameters is related to the frequent presence of sandy interbeds in the sedge peat. An extensive sand stratum up to 1.2 meters in thickness, which is itself interbedded with peat, is located in the southern end of the depression, extending from near the present eastern edge of the wetland underneath the Aberjona River. The interbedding of sand and peat within this deposit is variable, but sub-layers as thin as 5 - 10 cm are easily distinguished in both cores and piezocene records. Similar variability is found in the riverbed sediments, which consist of discontinuous layers of detrital peat, silts, and sands.

The top (from the surface to a depth of 0.5 - 0.75 m) of the wetland is composed of herbaceous peat with a highly variable structure. Composed of predominantly of Typha
stems, roots, and rhizomes, two distinct fibrous root mat horizons are distinguishable in both soil core and piezocone data. The lower root mat is located at the bottom of the section. Other sub-layers are much less structured, consisting of sapric black muck. Organic content is also highly variable, ranging from 30 - 60%. Carex (Sedge), Typha angustifolia (cattail), and other macrophytes, including Phragmites australis and Lythrum salicaria are currently flourishing in the wetland.

5.4.2. Arsenic distribution

The arsenic concentration of the wetland soils is as high as 9000 mg/kg. Representative arsenic profiles are presented in Figures 6 through 9. The highest concentrations of arsenic are found in the upper Typha peat. The average concentration of 37 samples obtained from 0 - 0.6 m depth is 2000 mg/kg. In cores 17 and 38 arsenic concentration drops sharply with depth between 0.7 and 1.0 meter. The maximum concentration below this depth in these cores is 12 mg/kg. Deeper peaks in arsenic concentration of 2900 mg/kg and 3300 mg/kg occur in Cores 6 and 11 at depths of 1.6 and 1.25 m, respectively. Average arsenic concentration in sandy layers is 2 mg/kg whereas average concentration in the woody peat is 32 mg/kg.

5.4.3. Sequence and timing of soil accumulation

The important pollen markers used to date the upper portion of this soil profile are the increase in Ambrosia (Ragweed) pollen related to agricultural development and colonization of New England in the 17th Century, and the rapid decline of Castanea (Chestnut) pollen correlated to the Chestnut Blight of 1920 (Table 2 and Figure 10). The
Figure 5-6. Arsenic, water, and organic content for core C-17. Organic content is determined as a fraction of the dry weight. Dates from radiocarbon analysis are also shown, as well as soil sub-layers. See Table 5-1 for legend.
Figure 5-7. Arsenic, water, and organic content for core C-38. Organic content is determined as a fraction of the dry weight. See Table 5-1 for legend.
Figure 5-8. Arsenic, water, and organic content for core C-11. Organic content is determined as a fraction of the dry weight.
Core C-6

Figure 5-9. Arsenic, water, and organic content for core C-6. Organic content is determined as a fraction of the dry weight.
Table 5-2. Soil horizon dates and deposition rates. Deposition/accumulation rates are calculated based on the corrected ages.

<table>
<thead>
<tr>
<th>Interval (cm - cm) or Depth (cm)</th>
<th>Bottom Marker</th>
<th>Soil Type</th>
<th>Date (AD or radiocarbon)</th>
<th>Corrected$^1$ age (yb 1995 AD)</th>
<th>Deposition/Accumulation Rate</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>mm/yr</td>
</tr>
<tr>
<td>Core 50</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0 - 40</td>
<td><em>Ambrosia</em> decline</td>
<td>P1a</td>
<td>1920 AD</td>
<td>75</td>
<td>5.33</td>
</tr>
<tr>
<td>40 - 55</td>
<td><em>Typha</em> horizon</td>
<td>P1a</td>
<td>1892 AD</td>
<td>105</td>
<td>5.33</td>
</tr>
<tr>
<td>55 - 75</td>
<td><em>Castanea</em> decline</td>
<td>P1b</td>
<td>1650 AD</td>
<td>345</td>
<td>0.83</td>
</tr>
<tr>
<td>75 - 116</td>
<td>$^{14}$C</td>
<td>P1b</td>
<td>2190 +/- 55</td>
<td>2293</td>
<td>0.23</td>
</tr>
<tr>
<td>116 - 152</td>
<td>$^{14}$C</td>
<td>P1c</td>
<td>6045 +/- 60</td>
<td>6930</td>
<td>0.08</td>
</tr>
<tr>
<td>152 - 182</td>
<td>$^{14}$C</td>
<td>P2/P1c</td>
<td>8515 +/- 75</td>
<td>9515</td>
<td>0.12</td>
</tr>
<tr>
<td>182 - 210</td>
<td>$^{14}$C</td>
<td>P2</td>
<td>9075 +/- 75</td>
<td>10030</td>
<td>0.54</td>
</tr>
<tr>
<td>210 - 233</td>
<td>$^{14}$C</td>
<td>DE</td>
<td>9650 +/- 140</td>
<td>10945</td>
<td>0.25</td>
</tr>
<tr>
<td>233 - 390</td>
<td>$^{14}$C</td>
<td>DE</td>
<td>12060 +/- 180</td>
<td>15010</td>
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<td>Core 17</td>
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<tr>
<td>121</td>
<td>$^{14}$C</td>
<td>P1b</td>
<td>795 +/- 50</td>
<td>737</td>
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<tr>
<td>222</td>
<td>$^{14}$C</td>
<td>P1c</td>
<td>4775 +/- 75</td>
<td>5568</td>
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<td>Core 34</td>
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<tr>
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<td>$^{14}$C</td>
<td>P1b</td>
<td>1035 +/- 55</td>
<td>983</td>
<td></td>
</tr>
</tbody>
</table>

$^1$ Radiocarbon years before present corrected to calendar years before 1950 using methods of Stuiver and Pearson, Pearson et al, Linick et al., Kromer and Becker, and Bard et al. (1993), then corrected to calendar years before 1995.
Figure 5-10. Soil types, radiocarbon dates, and pollen markers at core 50.
transition to cattails observed at approximately 0.55 m depth is inferred as an industrial horizon marker. A date of 1890 was estimated for this horizon by assuming that the deposition rate from 1920 to present approximates the rate established under the influence of industrial development of the watershed. This rate was used to extrapolate downward from the 1920 horizon.

Radiocarbon dates were corrected to calendar ages using the methods of Stuiver and Pearson (1993), Pearson et al (1993), Linick et al. (1993), Kromer and Becker (1993), and Bard et al. (1993). The deposition sequence and timing is discussed with reference to uncorrected radiocarbon years before present (ybp indicates radiocarbon years before 1950); calendar ages are used to calculate deposition rates.

Measured organic carbon net accumulation rates reflect several processes, including net primary productivity (annual biomass production--equivalent to litter fall to a first approximation), degradation during burial, and degradation after burial. Decay rates are in turn affected by vegetation type, water table fluctuations, temperature, and the rate of burial, which is itself a function of productivity as well as sediment influx. Therefore a particularly high measured net accumulation rate may be the result of many contributing factors, including high productivity, low rates of decay, and a high and rising water table. Clymo (1983) reviews the problem of determining production and decay rates for peat using direct and indirect measurements, showing that neglect of small post-burial decay rates can cause the original net productivity of a system to be significantly underestimated.

Clymo (1983) reports first order decay rates in shallow soils, where regular aeration might be expected (< 20 cm depth), from 0.1 to 0.7 yr\(^{-1}\). Decay rates after deeper burial may be as low as 10\(^{-5}\) yr\(^{-1}\), and are commonly reported in the range of 10\(^{-2}\) - 10\(^{-4}\) yr\(^{-1}\). Net primary productivity, as reviewed by Clymo (1983), ranges from 100 to 1500 g/m\(^2\)/yr, which, assuming 40% carbon by weight, is equivalent to 40 to 600 gC/m\(^2\)/yr.
In contrast to mass accumulation rates, accretion rates (measured in millimeters (mm/yr) are additionally affected by consolidation. Note that all of these processes cause historical wetland surface horizons to lie at lower elevations, relative to a stable mineral base, than when they were formed, and will cause rates of accretion or organic carbon production to be underestimated.

Our accretion rate calculations are approximate, and ignore the effects of consolidation and peat decay at depth. Estimates of bulk density, based on assumed specific gravities for organic (0.9) and mineral (2.55) materials, allow the reporting of estimated carbon accumulation rates, which are not affected by consolidation. Accumulation rates are shown in Table 2, calculated as mm/yr, total mass/m²/yr, mass mineral material/m²/yr, and mass of carbon/m²/yr. Whereas soil accumulation rates vary from 0.08 to 5.3 mm/yr, carbon accumulation rates vary from 2.2 to 270 g C/m²/yr.

The oldest radiocarbon dates from core 94-50 indicate that over two meters of lacustrine deposition had occurred by circa 12,000 ybp. A reasonable basal date of 14,000 ybp would require that the deposition rate for the lower portion of the diatomaceous layer was approximately 2.5 times the deposition rate for the upper portion. Woody peat began to accumulate at 9,100 ybp, and lasted until about 8,000 ybp, when sedge peat appears in the profile. In core 17, the interbedded sand layer located within the sedge peat has a basal date of 4,800 ybp. The top of the sand lens is dated in cores 17 and 34 at 800 and 1000 ybp, respectively.

The period of deposition between 8500 and 2200 ybp (7500 elapsed calendar years) is notable in that only small quantities of organic material are found in the record for this period (0.1 mm/yr, 2.5 g C/m²/yr). This observation could be the result of nearly steady wetland growth, or substantial post-burial decay. A simple first order deposition and decay model is useful for examining these possibilities:
\[
\frac{m}{t} = \frac{r}{\lambda} \left[1 - \exp(-\lambda t)\right]
\]

where:
- \(m\) = total mass accumulated
- \(t\) = elapsed time
- \(\frac{m}{t}\) = observed deposition rate in profile after time \(t\) has elapsed
- \(r\) = carbon deposition rate (net primary productivity or residual biomass)
- \(\lambda\) = decay rate (surface or post-burial)

In the first case, a steady water table and well-aerated surficial soils could have depleted a more substantial net productivity, leaving only a small residual biomass for burial beneath the oxic zone. For instance, if the “shallow” decay rate is as large as 0.7 yr\(^{-1}\) for forty years following deposition (this corresponds to 5 mm/yr accumulation for 10 cm), only 3.5 % of net productivity is preserved as “residual biomass” for deeper burial. Even if no decay occurred below 20 cm, net primary productivity could have been as high as 70 gC/m\(^2\)/yr. In the second case, “residual biomass” could have been buried at rates of 200 gC/m\(^2\)/yr if the post-burial decay rate was 0.01 yr\(^{-1}\). It is likely that both high oxic zone decay rates occurred during this period, perhaps due to water table limitations, as well as some decay after burial.

5.5. Discussion of Wetland History

5.5.1. The late glacial depositional environment in the Wells G & H Wetland area

As the fourth outwash in the sequence described by Chute was being deposited, the Valley floor in the vicinity of what is now the Wells G & H Wetland was either covered
with ice or being eroded as source material for outwash. Further erosion of the Aberjona River Valley into the third outwash occurred while an ice block remained in the depression mapped in this study. Chute believes that a number of ice blocks existed in the Aberjona Valley during its late glacial incision, as evidenced by the complex topography and variability in valley width. The presence of other thick peat sections in the Wells G & H wetland supports this hypothesis (NUS, 1986). The coarse sandy deposit that forms the upland island in the wetland was formed in late glacial time, perhaps by deposition or preservation of sediment along the margin of the same ice block that formed the kettle hole. Formation of this high energy deposit must have occurred while ice still filled the kettle hole, or it would have been filled with the same sediments.

With the further recession of the regional ice-sheet and the melting of the ice blocks, local runoff decreased, and a transition from the high-energy glaciofluvial environment to that of a moderately sized meandering river began. It is likely that the Aberjona River carried more water in the past than today, but it could not have occupied the entire valley floor after the melting of the local ice blocks, as it would have quickly filled the ice block depressions with granular sediments. A conceptual model of the sequence of riparian environments that began with this transition is illustrated in Figure 11.

5.5.2. Lacustrine deposition

Although a basal date was not obtained, the age of 12,000 ybp obtained several meters above the bottom of the deposit indicates that deposition began soon after regional ice-retreat, suggesting that the ice block that formed this depression was small and melted rapidly. The location of this kettle pond in the upper reaches of the Aberjona Valley precluded any chance of marine influence; the freshwater diatom species found in the
Figure 5-11. Conceptual model of post-glacial environments at Wells G & H Wetland.
sediments confirm this. The small pond formed in the ice block depression was fed primarily by groundwater; stream inflow to such a small water body would have contaminated the diatomaceous silt with significant quantities of granular mineral sediments.

At one time, however, a meander of the Aberjona or a tributary must have temporarily fed the pond, causing the deposition of a wedge of sandy sediments at its northern end. This sediment influx may have occurred during a cooler, wetter period known as the Younger-Dryas reversal, a post-glacial cooling trend observed in Europe, Canada, and the North Atlantic Seaboard. Peteet et al. (1990) have observed this trend in the Northeastern United States record, identified in part by the dramatic increase in Picea pollen and fossils occurring between 11,000 and 10,000 ybp. As the pollen record indicates, local vegetation during this period consisted of a conifer forest, including Picea (spruce), Pinus (pine), and Abies (fir).

5.5.3. Swamp

The exact mode of transition from lacustrine deposition to a woody swamp is not revealed by the existing data. As the pond filled with sediments, its character would have changed substantially. The shallower water could support emergent macrophytes. A transition zone in the soil record, composed of an increasingly organic-rich silt but containing no recognizable macro-fossils, may represent a community of emergent macrophytes that colonized the shoaling pond.

Alternatively, water levels may have dropped in the pond around this time, allowing shrubs and trees to begin growing here. The especially low accumulation rates (0.25 mm/yr and 5 grams carbon (gC)/m²/yr) during this time would be consistent with a drying trend associated with the end of the Younger-Dryas reversal. Further analysis of
transition zone material for pollen signatures and microfossils could help to resolve this issue. The swamp developed beyond the confines of the pond beginning after 9700 ybp, indicating an eventual increase in water levels across the developing wetland.

A very-low-mineral-content swamp peat began to accumulate approximately 9000 ybp, indicating that this portion of the wetland was isolated from stream (and surface runoff) sediment load. Although the peat formed during this stage was of very high organic content (95%), the average carbon accumulation rate is only 15 gC/m²/yr, compared to 10 gC/m²/yr for the diatomaceous silt, which had an organic carbon content of only 5 - 10%. Coincident with the change in sediment character is the appearance of deciduous pollen in the profile. Increases in Quercus (oak), Alnus (alder), and Fagus (beech) pollen indicate the presence of a mixed forest in the area. The uneven interface between the lacustrine and swamp sediments is probably the result of consolidation of the latter.

5.5.4. Marsh

The decrease in recognizable woody material from the peat beginning below 1.5 to 2.2 m, and changes in the pollen assemblages, indicate a transition to an herbaceous marsh system after 8500 ybp. Interpolating between sets of radiocarbon dates obtained above and below this boundary, the transition is estimated to have occurred between 7800 and 8300 ybp in core 50. Further changes in the local arboreal community accompany this transition with a dramatic decrease in conifer pollen, especially Pinus, and continued increase in Quercus. A decline in Tsuga (Hemlock) also occurs during this period, an event attributed to disease dated at circa 4650 ybp (Gaudreau and Webb, 1985). Few identifiable macrofossils are present in the lower portion of this layer, although Sagittaria (arrowroot) pollen and Filicales (fern) spores are numerous.
As discussed in section 5.4.3, the low rates of peat and carbon preservation in this system (0.1 mm/yr; 2.5 gC/m²/yr) indicate either high rates of degradation in a thick oxic zone, a high post-burial decay rate, and most likely both. Water table limitations could account for litter accumulation in a thick aerated zone, and a small rates of annual residual carbon preservation.

A warmer and/or drier period between 9000 and 8000 ybp has been inferred from pollen records collected across the eastern United States, including New England (Deevey and Flint, 1957; Davis et al., 1980; Winkler, 1985). This period has been referred to as the Hypsithermal interval, and its interpretation from New England pollen records has been the topic of considerable debate (Wright, 1976). As with the Younger-Dryas reversal, this climactic trend is inferred from changes in pollen abundance, which have been mapped into a series of vegetation predominance zones that change over time. Despite the significant observation that time-transgressive boundaries can also result from non-climatic tree immigration, general agreement seems to exist that this period existed. Note, however, that the influence of regional trends in temperature and rainfall can be overwhelmed by local factors such as soil type, topography, and wetland morphology that influence water levels and availability, thus controlling wetland vegetation and accumulation of its remains.

5.5.5. Alluvial deposition

The presence of well-defined sandy layers and lenses in the herbaceous peat indicates that riverine erosion and deposition occurred during this period. The largest of these sandy units appears to be a portion of a record of a meander belt. Paleochannel deposits have been identified in many settings, and may take on many different forms (Schumm and Brakenridge, 1987; Autin, 1992). The Aberjona River may have been substantially larger during this period, or may have meandered across portions of the wetland, leaving
a broad trace of its path composed of interbedded detrital peat and sandy sediments. Radiocarbon dates from the base and top of this layer (4775 and 800 - 1000 ybp, respectively) occur during a climatic trend back towards cooler, wetter conditions identified by Winkler (1985) starting at about 5000 ybp.

The varying elevation of the bottom of the sandy layers may be more consistent with a meandering streambed than a single, wider stream. Without more extensive subsurface mapping, however, it is difficult to distinguish between these possibilities with confidence. In some locations sandy strata directly overlie the woody swamp peat, in others they are underlain by younger, herbaceous peat. One such deposit directly overlies portions of the kettle hole depression. The weight of this sandy layer may have caused the consolidation of the thick section of lacustrine sediments referred to earlier.

5.5.6. Meadow

A transition to a wet meadow of grasses and sedges occurred sometime around 2200 ybp, the radiocarbon age obtained for a depth of 116 cm in core 50. This corresponds well to the transition from the warmer period described by Davis et al (1980) and Winkler (1985) to what Wright (1976) calls the Neoglacialation, a cooler, wetter period. Cyperaceae (sedge family) and Pocaceae (grasses) macrofossils and pollen become abundant at approximately the same depth in Core H2. Concurrent decline in conifer and increases in Castanea, Tsuga (Hemlock) and Carya (Hickory) pollen probably indicate forest succession and diversification during this period.

Although alluvial deposition continued through this period, the Aberjona River stabilized to its present location about 800 - 1000 ybp. This could have been related to a decrease in baseflow during through this period, although there is no reason to suspect this occurred due to climatic factors. Alternatively, the western channel of the Aberjona
River (remnants of which are observable today) may have become more active, stabilizing the channel that is observed in our section. Accumulation of organic material in this system increased over five-fold to nearly 12 gC/m²/yr, and the wetland surface began to aggrade more than twice as quickly as in the previous period.

The European colonization of the watershed began during this period, as indicated by appearance in the pollen record of cleared-land colonizers, including Ambrosia, Rumex (dock), and Plantago (plantain) at 75 cm in Core H2. Land clearing and later industrial development undoubtedly have led to increased runoff and both higher and more frequent flooding of the Aberjona River. The further increase in peat accumulation rate in the top portion of this layer, measured both as bulk accretion rate (0.23 to 0.87 mm/yr) and carbon accumulation (11.8 to 34.6 gC/m²/yr) probably reflects this anthropogenic effect on watershed hydrology.

5.5.7. Cattail marsh

The appearance of abundant Typha (cattail) remains in the peat at approximately 55 cm, including several horizons of dense, interlocked mats of Typha remains, signals the transition to a fully urbanized watershed that occurred in the mid to late 19th century. This transition likely included the increase in runoff and degradation of its quality. Higher water levels, as well as increased sediment content and salinity of runoff, could have triggered a shift from sedge meadow to Typha marsh. The influence of metal contamination in particular is an issue in this watershed, as its history clearly indicates that the Aberjona River carried industrial discharges of metals as early as 1888 (Aurilio et al., 1994). The present community of macrophytes at the wetland, including Typha as well as Phragmites and Lythrum are known to be tolerant of various pollutants including metals and high conductivity waters, and can be indicative of disturbed systems (Suzuki et al., 1989; Babcock et al., 1983; Kraus, 1987). Typha and Phragmites have been
proposed for pollutant metal treatment in constructed and urban wetlands (Zhang et al., 1990; Dunbabin and Bowmer, 1992, Chongya et al., 1990). Although not required to explain the dramatic shift from sedge meadow to cattail marsh, metal contamination existed, and may have had a significant role in this transition.

Another important change resulting from urbanization is the increase in peat accumulation rate and wetland productivity. The accretion rate increased by a factor of six with industrialization, to over 5 mm/yr; increasing a total of 25 times over undisturbed (pre-colonial) conditions. This increase is partially attributable to increased mineral deposition resulting from higher flooding frequency, but is also related to increased productivity and/or peat preservation. The mineral content of the uppermost peat is larger by weight fraction than the older peats, but the total mass of carbon preserved is still greater: 200 - 270 gC/m²/yr compared to 34.6 in colonial and early industrial times. This change may reflect increased nutrient availability and the naturally high primary productivity of these macrophytes, as well as increased preservation due to fast burial rates and inhibition of microbial degradation by toxic metals.

Increased accretion rate in wetlands is typically associated with a rising water table. Sea level rise over the past several thousand years has averaged less than 1 mm/yr. Thus, if a water table rise occurred, it was probably associated with local influences (Oldale et al., 1993). Peat growth itself can influence water retention in wetlands; such peats are classified as secondary or tertiary by Moore and Bellamy (1974). It is likely in this case, however, that anthropogenic effects influenced water levels in the Wells G & H Wetland. These effects are likely to have included: increased sedimentation/debris accumulation, culvertization, and other structures for flood control, farming, transportation, and industrial purposes.

Salem Street, which crosses the Aberjona River at the base of the wetland, appears in city records as early as 1795. A short (2 km) section of canal was dug along the western
bank of the Aberjona River from Salem Street to Montvale Avenue in 1837 (Thomas Smith, personal communication, 1996). A mill, probably a tannery, existed at the Central Street crossing of the Aberjona River about one km downstream of the wetland in the late 1800's. A mill pond was created on the Aberjona River immediately upstream of this mill, extending to within 0.5 km of the wetland. Maps indicate this impoundment was created sometime between 1831 and 1875, and then drained sometime in the 1920s (Hopkins, 1875; Sanborn, 1926; Jones, 1915; Thomas Smith, personal communication, 1996). A cranberry bog existed in the same location as the former pond for some time afterwards.

Upstream of the site, Richardson's Pond, also known as Mishawaum Lake, was created to run a saw mill in the 1750s. This mill was rebuilt in 1843, and burned down in 1895 (Thomas Smith, personal communication, 1996). Still apparent on maps drawn in 1926, Mishawaum Lake was drained in 1969, and exists as a much smaller impoundment today, known as the Halls Brook Storage area (Sanborn, 1926; Tarr, 1987).

The Boston and Maine Railroad, built in the early 1830's, was constructed on an embankment that divided the formerly much broader wetland into eastern and western halves (Tarr, 1987, Durant et al., 1990). Much of the western half has since been filled, and is now occupied by parking lots and commercial and industrial buildings.

Clearly the hydraulic impacts of documented industrial activity along the Aberjona were profound. Substantial surface impoundments were created and drained during the period of industrialization. No one singular event stands out in the late 1800s, when the transition to cattail marsh is suspected. Furthermore, the previous wetland vegetation (Tussock Sedge) is also quite flood tolerant. This observation may indicate that chemical rather than hydraulic disturbance was responsible for the rapid vegetational change. As will be discussed in the following section, there is evidence that the use (and presumably
the discharge) of arsenic-containing materials began at about the time of the observed wetland transition.

5.6. Discussion of Arsenic Distribution

To explain the observed arsenic distribution in the Wells G & H Wetland soils requires a source, a delivery mechanism, in some cases transport mechanisms within the wetland, and the existence of conditions that preserved, to a large extent, the record of sedimentary arsenic for over 50 years. These aspects of the Wells G & H arsenic story are addressed in the following sections, preceded by a brief review of the environmental chemistry of arsenic.

5.6.1. Environmental chemistry of arsenic

Inorganic arsenic species stable at typical environmental conditions include the acid oxides arsenate [As (+V); pK₁ = 2.2; pK₂ = 6.9] and arsenite [As (+III), pK₁ = 9.2]. At circumneutral pH, the dominant species are H₂AsO₄⁻, HAsO₄⁻², and H₃AsO₃. The highly surface reactive arsenate (As (+V)) is the equilibrium aqueous specie found in oxic waters; the more soluble arsenite is the expected form in reducing environments, and typically the major constituent of wetland porewaters. Laboratory and field studies have measured increases in soluble arsenic with increasingly reducing conditions (Masscheleyn et al., 1991a, 1991b; Marin et al., 1993; Moore et al., 1988; Brannon and Patrick, 1987).

It is commonly thought that equilibrium concentrations of the various arsenic species seldom occur. Although As +III/+V ratios increased with decreasing Eh, Marin found that at higher pH (7.5), As (+V) remained the most abundant specie released to solution, even at Eh = -200 mV. Much larger quantities of total arsenic were released from sediments at lower pH (5.5), with arsenate persisting, although in smaller proportion.
Masscheleyn (1991a) also observed the persistence of As (+V) under increasingly reducing conditions. Arsenate can also be biologically methylated, forming monomethylarsenate or dimethylarsenate, which are commonly reported in oxic surface waters (Aurilio, 1992).

Aurilio (1992) found that As (+III) is the dominant specie in groundwater inputs to the Halls Brook Storage in the headwaters of the Aberjona River, but that the bulk of this arsenic is oxidized to arsenate and lost from the water column before discharging to the Aberjona River. Hemond (1995) and Aurilio et al. (1994) hypothesized that newly-oxidized arsenate is scavenged by iron hydroxide precipitates. Apparent non-equilibrium distributions of arsenite and arsenate have been measured in the epilimnion and hypolimnion of the Upper Mystic Lake by Aurilio (1994) and Trowbridge (1995). Biological reduction, slow oxidation, and settling of precipitated arsenate are commonly hypothesized to explain such distributions. Biologically mediated reduction of arsenic in sediments has been investigated in the Halls Brook Storage area, where a eubacterium was isolated that reduces arsenate, gaining energy for growth (Ahmann et al., 1994).

An Eh - pH diagram in Figure 12 displays the equilibrium predominance zones for the inorganic species of arsenic. Also shown are the stability zones for various arsenic, iron, and sulfur solids when total arsenic, sulfur, and ferric iron are maintained at 10^{-5} M. An additional line shows an immobile zone below which total aqueous arsenic is reduced to less than 10^{-7} M. AsS_{2}^{-1} is the predominant As (+III) specie below about 150 mV at a pH of 6.5. It is not shown in this diagram because it does not predominate outside of the immobile zone. Although the stability zones of pyrrhotite and pyrite are shown, the effects of their precipitation on other solids or arsenic are not considered. This chart spans the pH range of 5 to 8; typical Wells G & H Wetland porewater pH is 6.3 - 6.8.
Figure 5-12. Eh - pH diagram for stable solids of arsenic. \( \text{As}_T = \text{S}_T = \text{Fe}_T = 10^{-5} \text{ M} \). Dashed lines are stability zones for iron sulfides. Fine line is boundary below which mobile arsenic is restricted to less than \( 10^{-7} \text{ M} \).
5.6.1.1. Arsenate

Iron and aluminum oxides are widely acknowledged to play a large role in the sorption and coprecipitation of arsenate (Brannon and Patrick, 1987; Moore et al., 1988; Masscheleyn et al., 1991b; Yan-Chu, 1994). The release of arsenic to solution is often correlated with the reduction of iron oxides and corresponding increases in aqueous ferrous (+II) iron. Brannon and Patrick (1987) report that the stoichiometry of concurrent iron reduction and arsenic solubilization in an anaerobic laboratory incubator suggest that arsenic in this system was co-precipitated with iron oxides. The operationally defined, 'moderately reducible' fraction of sediment extracts (acid oxalate) typically yields the overwhelming majority of sedimentary arsenic from oxic soils (Brannon and Patrick, 1987; Ahmann, 1996). This fraction is thought to remove the poorly crystallized oxides of Fe, Al, and Mn. In some studies the presence of these metals has been strongly correlated with this fraction. Ahmann (1996) used iron arsenates [Fe(+III)AsO₄ and Fe(+II)₃(AsO₄)₂] as a model for sedimentary arsenic, finding that, as in natural sediments, most of the arsenic in these compounds was released by oxalate extraction.

There is theoretical evidence that the precipitation of ferric arsenate can not limit the solubility of arsenate substantially. In fact the solubility of ferric iron is so low at circumneutral pH that arsenate can persist at molar concentrations without causing ferric arsenate precipitation. Wagemann (1978), Sadiq (1983), and Livesley and Huang (1981) all conclude that ferric arsenate is not effective at limiting arsenate solubility based on thermodynamic considerations. Wagemann proposes that Ba and Cr could be more effective at precipitating arsenate (Ba₉(AsO₄)₂ has a solubility product of 10⁻⁵⁰). He also suggests that arsenate may substitute for phosphate in minerals such as apatite [Ca₅(PO₄)₃(OH,F)], or be scavenged as ferric hydroxy arsenate [FeH₂AsO₄(OH)₂].
Both Wagemann (1978) and Livesley and Huang (1981) emphasize that sorption to oxalate extractable metals such as iron and aluminum is probably the most effective limit on arsenate mobility in most oxic environments. Livesley and Huang measure 90 to 140 µg As/g of soil on a time scale of several days. Gupta and Chen (1978) measured the sorption of 1800 µg As/g of alumina over several hours.

The pH dependency of arsenate absorption onto oxides (Fe, Al) and clays reflects the influences of surface site activation, competition by OH⁻ for sites, and the speciation of arsenate, which sorbs most strongly as the monovalent anion in the pH range of 5 - 6 (Frost and Griffin, 1977; Xu et al., 1991; Masscheleyen et al., 1991a; Yan-Chu, 1994; Marin et al., 1993). The pH and Eh dependency of sorbent precipitation/dissolution is also important in the case of more soluble materials.

The importance of organic sorbents in the immobilization of arsenate in the environment has not been comprehensively assessed (Moore et al., 1988, Vance, 1995, Bhumbla and Keefer, 1994). As a phosphate analog, biological uptake and binding with organic materials has the potential to be important (Streit and Stumm, 1993). Thanabalasingam and Pickering (1986) measured 90 mmol/kg sorption capacity for arsenate on humic acids at pH = 5.5, in the laboratory, noting that phosphate could displace arsenate in this system. Agget and O'Brian (1985), who studied the arsenic cycle in lake sediments in New Zealand, claim that despite the high organic content of these sediments, organic processes appear to be unimportant in the formation of solid arsenic phases.

5.6.1.2. Arsenite

Arsenite is the typically the mobile and more toxic form of arsenic in groundwaters (Bhumbla and Keefer, 1994). Masscheleyen et al. (1991a) found a 25 fold increase in soluble arsenic per kg of sediment with decrease of Eh to below 0 mV; As (+III) was
85% of total soluble As. Presence of sulfide in these experiments did not cause precipitation of arsenic sulfides, although several other metal sulfides precipitated.

Searches for a sorbent of arsenite for treatment of contaminated water have been unsuccessful (Vance, 1995). Arsenite is much more weakly sorbed to clays than arsenate, with peak sorption occurring at pH 7 - 8 (Frost and Griffin, 1977). Arsenite will sorb to iron hydroxide as well, but only weakly (Dzombak and Morel, 1990; Pierce and Moore, 1977), although one investigator claims that arsenite sorbs more strongly to iron hydroxides than arsenate (Bhumbla and Keefer, 1994). Ahmann (1996) suggests that the arsenite thought to be removed from solution by oxyhydroxides is in fact oxidized to arsenate in the process, but that in most studies this is not detectable by the methods employed. Several studies have indicated that manganese oxides can catalyze the adsorption of arsenic to both iron and manganese oxides by oxidizing arsenite to arsenate (Yan-Chu, 1994).

Thanabalasingam and Pickering (1986) measured 70 mmol/kg sorption capacity for arsenite on pure humic acid solids at pH = 5.5 (about 5000 mg/kg) in the laboratory. As is the case with arsenate, the importance of this interaction in natural soils is unclear. The sorption of uncharged arsenite (H$_3$AsO$_3$) by organic materials could conceivably occur due to hydrophobic partitioning. However, only minimal importance has been attributed to the interaction of arsenite with organic solids based on field studies (Aggett and O’Brian, 1985; Moore et al., 1988).

Moore et al. (1988) measured the arsenic content of reservoir sediment and porewater, and found that although arsenic concentrations in pore water increased with depth with reducing, sulfidic conditions, both the proportion and total quantity immobilized in sulfide phases of the sediment also increased to over 500 mg/kg. An abrupt transition from an arsenate-oxyhydroxide to an arsenite-sulfide system occurred as Eh dropped through 200 mV to values of about 150 mV. In the system studied, copper, zinc and
arsenic were believed to control sulfide concentrations in porewater, with \( \text{As}_2\text{S}_3 \) becoming an important sink for arsenite.

Other investigators have found that iron sulfide minerals are important in the immobilization of trace metals, and that the pyritization-depyritization of arsenic occur quickly (Huerta-Diaz and Morse, 1992; Morse, 1993). Huerta-Diaz and Morse found high levels of As association with authigenic pyrite even at low levels of pyritization in the sediments. Pyritization of As increased with increasing pyritization of Fe, indicating that at least some of the As was being incorporated into the pyrite, and not just sorbed to its surfaces. Pyritization can provide an extremely large reservoir for arsenic; in arsenopyrite (FeAsS), arsenic is 34% by weight. Although pyritization is generally favored in sulfide rich, iron poor sediments, high degrees of arsenic pyritization were observed in iron rich, sulfide poor sediments with high organic content. Under oxidizing conditions, pyritized trace metals can be released quickly.

5.6.2. Sources and riverine transport of arsenic

The Aberjona watershed became heavily contaminated by industrial discharges starting in the mid-19th century. Durant et al. (1990) describe a series of Aberjona Watershed surveys conducted by the State Board of Health beginning in 1871. Tannery wastes initially contaminated only the Horn Pond sub-basin of the Aberjona watershed, but by 1876, these and related industries had polluted the lower reaches of the Aberjona River. Although there is no direct evidence in these reports of when the upper reaches of the Aberjona River first suffered from industrial discharges, we can assume that it was not too long after this. The upper portions of the watershed, where the wetland is located, were not sewered until 1932, and even after this time overflows to the river occurred due to clogging (Durant et al., 1990). The requirement that industries pass their wastewater through settling lagoons before discharging to the sewer created massive reservoirs of
highly contaminated soil, many of which are thought to be ongoing sources of groundwater and surface water contamination today.

The largest industrial operation north and upstream of the Wells G & H wetland was located along a tributary to the Aberjona River called Halls Brook. Beginning in 1853, this site was used by a number of chemical manufacturers, including Merrimac Chemical, Monsanto, Consolidated Chemical, and later Stauffer Chemical (Durant et al., 1990). More recently, this site has become known as Industri-Plex. The use of arsenic-containing pyrites to manufacture sulfuric acid began at this site in 1888, and arsenical pesticide manufacturing began in 1899 (Aurilio, 1994). Tarr (1987) reports that pyrite cinders were stored at the site in piles in the anticipation of some then undiscovered use for this by-product. A series of photographs taken in 1921 and 1922 at the Industri-Plex site documents the discharge of waste from lagoons to surface waters (Margaret Gardner, personal communication, 1996). Aurilio estimates that 200 to 700 metric tons of arsenic were released as a result of these activities at this one site alone.

Although less well studied, two other sites, are potential sources of arsenic to the Wells G & H Wetland: the former Riley Tannery located immediately west of the Aberjona River, and the Olympia Nominee Trust Corporation property, located on the Aberjona River, at the upgradient edge of the Wells G & H Wetland. A Wells G & H Site Health Assessment issued by the Massachusetts Department of Public Health (1989) cites the Olympia Property as a known disposal site of arsenic trioxide, and reports that chromium lagoons and hide piles were present at the Riley site. According to the same report, elevated arsenic levels in groundwater and soils have been measured at both sites.

It is likely that these wastes, as well as wastes from other sites in the watershed, have been transported to Mystic Lake. A nearly 2000 mg/kg peak in arsenic concentration in the Mystic Lake sediments occurs between 1886 (70 cm) and 1943 (50 cm) (Spliethoff and Hemond, 1996). Sediment arsenic levels decrease to less than 10 mg/kg below 100
cm depth. High concentrations of arsenic exist in present-day Aberjona River sediments; the Wells G & H reach is somewhat elevated compared to other locations, ranging from 500 to 1000 mg/kg (Knox, 1991). The presence of 100 - 500 mg/kg of arsenic in the top layers of Mystic Lake and forebay sediments indicates that transport of arsenic through the river is still occurring (Seeman, 1996; Spliethoff and Hemond, 1996; Knox, 1991).

In the present day Aberjona River, the arsenic transport rate is about 100 kg/yr, with dissolved and particulate fractions accounting for roughly equal portions of the total flux (Solo, 1995). Whereas the aqueous concentration of arsenic ranged from 1 - 11 µg/L, the concentration of Arsenic on particulate matter ranged from 200 to 1400 mg/kg. In the past, perhaps during periods of high runoff from exposed waste piles, this transport rate would have been considerably higher.

As recorded in the Mystic Lake record, arsenic riverine transport rates peaked in 1930, during the height of the chemical industry, and again in 1960, when an as-yet undocumented disturbance apparently caused a large release or remobilization of riverborne arsenic. River sediments have likely been contaminated with arsenic since the late 1800s.

5.6.3. Arsenic within the Wetland

The most efficient and obvious mechanism for transport of arsenic from the riverbed to the wetland is the deposition of contaminated sediment by overbank flooding. This process would have periodically delivered arsenic-contaminated sediment to the wetland surface since the late 1800s, leaving a record beginning at the base of the *Typha* peat, at approximately 55 cm.
The occurrence of arsenic below the industrial horizon within the Wells G & H Wetland indicates that remobilization within the soils and/or additional mechanisms of delivery to the soil column were also important. Substantial quantities of arsenic are found below the inferred 1890 industrial horizon at 55 cm. In cores 17 and 38, which penetrate a substantial sandy unit at 0.7 to 1.1 cm depth, arsenic concentrations drop from 4000 - 6000 mg/kg to below 12 mg/kg between 0.7 and 1.0 m below the surface. In cores 6 and 11, which do not penetrate the sandy unit, an additional, smaller peak (3000 mg/kg) is present below 1 m.

The shallow and deep arsenic records differ not only with respect to the methods of delivery, but also with respect to the in situ chemical conditions and their variability. Whereas it is likely that the shallow peat may undergo cyclic variations in redox potential due to air entry and surface recharge, the deeper peat is unlikely to ever desaturate, and hence is expected to have a more stable and generally very low redox potential. Because of these differences, the formation and preservation of the shallow and deep arsenic records are discussed separately. The following discussions are based on the assumptions inherent in the stability and predominance diagrams shown in Figure 12.

5.6.4. Shallow Arsenic

As described above, overbank flow across the historical peat surface can explain the transport of arsenic to the peat above the industrial horizon. The arsenic found below, but close to, the hypothesized industrial horizon was probably transported there via the particulate phase by percolation through the open structure of the surface peat. It is estimated (based on the peat texture) that arsenic-contaminated sediments might be transported as deep as 10 to 20 cm into this layer during low water conditions. This observation, allowing for reasonable variability in the actual depth of the industrial horizon, supports the hypothesis that the large reservoir of arsenic in the top 80 cm of
peat is attributable to direct deposition at the surface. The preservation of this record of
deposition over more than 50 years requires that the solid chemical forms of arsenic in
this layer remain stable, or that substantial amounts of solubilized arsenic are not actively
transported out of the zone.

5.6.4.1. Solid phases of arsenic

A transition from oxic to anoxic and eventually sulfidic conditions typically occurs
within the upper meter of peat soils. The location of the redox gradient will vary with
the position of the water table and the degree of aeration of the uppermost peat. When
fully saturated, peats are typically anoxic below a depth of several centimeters. During
such conditions, the surface of an iron rod inserted into the peat at this site was found to
oxidize to ferric (Fe$^{3+}$) iron hydroxide only to a depth of 10 cm.

Arsenic initially deposited on the wetland surface would have been in the form of
arsenate, most likely sorbed to iron and other hydroxides or precipitated as ferric hydroxy
arsenate. Peat samples from above a depth of 60 cm contain from 1 to 12% iron. Upon
burial, and in saturated conditions, reduction of dissolved arsenate to arsenite occurs in
the range of 70 to 35 mV in the pH range from 6.5 to 7. Amorphous ferric hydroxide
begins to dissolve over a similar range of reduction potential (90 to 0 mV). If significant
quantities of arsenate are released, ferrous arsenate theoretically could exist in a narrow
Eh-pH zone as ferric iron is reduced to Fe$^{2+}$:

\[
3\text{Fe}^{2+} + 2\text{H}_2\text{AsO}_4^{-1} \quad \xleftrightarrow{} \quad \text{Fe}_3(\text{AsO}_4)_2(s) + 4\text{H}^+ \quad \log K = -2.47
\]

(calculated from data in Sadiq, 1990, and Morel and Hering, 1993)

As the reduction potential drops further, arsenite is expected to remain mobile until
sulfides start to precipitate in the range of -150 to -180 mV. Thermodynamically,
dissolved arsenite is reduced to less than $10^{-7}$ M below -160 to -190 mV. Orpiment (As$_2$S$_3$) formation is favored initially in a sulfidic environment:

$$\text{As}_2\text{S}_3(\text{orpiment}) + 6\text{H}_2\text{O} \rightleftharpoons 2\text{H}_3\text{AsO}_3 + 3\text{H}_2\text{S} \quad \log K^\circ = -38.1$$

(calculated from data in Sadiq, 1990, and Morel and Hering, 1993)

Realgar (AsS) has a stability zone at a lower pE than orpiment for pH < 6.4:

$$\text{AsS}(\text{realgar}) + 3\text{H}_2\text{O} \rightleftharpoons \text{H}_3\text{AsO}_3 + \text{H}_2\text{S} + \text{H}^+ + e^- \quad \log K^\circ = -13.7$$

(calculated from data in Sadiq, 1990, and Morel and Hering, 1993. Note that in realgar, As has a valence of +2, and is oxidized to As$^{+3}$ by elemental sulfur on dissolution)

Above a pH of 6.4, arsenopyrite is stable relative to orpiment at Eh values below -236 mV. Arsenopyrite is also stable relative to AsS at Eh values below -225 mV:

$$\text{FeAsS(arsenopyrite)} + \text{H}_2\text{S} \rightleftharpoons \text{As}_2\text{S}_3 + 2\text{Fe}^{+2} + 2\text{H}^+ + 6e^-$$

$$\log K^\circ = 5.35$$

$$\text{FeAsS(arsenopyrite)} \rightleftharpoons \text{AsS} + \text{Fe}^{+2} + 2e^-$$

$$\log K^\circ = 2.644$$

(calculated from data in Heinrich and Eadington, 1986; and Garrels and Christ, 1965)

The potential importance of arsenopyrite in freshwater wetlands has not been investigated, although it could clearly play a significant role in sequestering arsenic under highly reducing conditions. Huerta-Diaz and Morse (1992) have observed high degrees of so-called pyritization of arsenic in marine sediments at low redox potentials (Eh < -300 mV). Aggett and O’Brian (1985) also speculate that arsenic iron sulfides are an important sink for arsenic upon reduction of lake sediments.
Nesbitt et al. (1995) used x-ray photoelectron microscopy to determine the distribution of valence states in arsenopyrite, showing most of the sulfur to be disulfide (S$_2^{-2}$) and most of the arsenic to be As$^{-1}$. Fe$^{+2}$ was the dominant iron specie. Nesbitt et al. further monitored the surface species upon oxidation of the arsenopyrite surface with oxygen-saturated water, finding As$^{+5}$, As$^{+4}$, and As$^{+1}$, as well as ferric oxide with binding energy similar to goethite [(FeO(OH)$^{\prime}$ nH$_2$O]. Sulfur was found to oxidize less readily. They found that As migrates towards the surface of the mineral during oxidation, but does not accumulate at the surface, concluding that the potential for the mobilization of arsenic from arsenopyrite is significant.

It appears that as redox potential falls below -150 mV, soluble arsenite concentration is thermodynamically limited to less than 10$^{-5}$ M (750 $\mu$g/l) under typical freshwater conditions, and three different arsenic sulfides can be stable. It is likely that arsenic sulfides, contribute to the reservoir of solid phase arsenic in the shallow peat, although these precipitates could be partially oxidized during periods of desaturation or intense surface recharge.

It is important to note that the aqueous concentration and solid phase stability of arsenic can both increase as conditions become more reducing. Moore et al. (1988) demonstrate this fact with field data that show concentrations of arsenite and arsenate in porewater increasing dramatically (to nearly 600 $\mu$g/l) below a measured Eh of about 160 mV; at the same locations, the reservoir of solid arsenic, most of which occurred in the operationally-defined sulfide fraction, also increased. These results indicate that even in cases where appreciable dissolved arsenic exists, solid phases can account for large (Moore et al. measured 800 mg/kg) quantities of sedimentary arsenic. Observations of up to 8000 mg/kg in anoxic horizons of the Weis G & H wetland are consistent with these results. In the deepest sample from Moore’s core, porewater arsenic drops back below 100 $\mu$g/l at a measured redox potential of 140 mV, possibly due to the increased sequestering of arsenic by sulfide minerals. The fact that the mobility window in Moore
et al.'s data set is from 140 to 160 mV rather than between 0 and 150 mV as predicted may be related to differences between measured and theoretical Eh, or to the transient hydraulic and chemical conditions in their core.

5.6.4.2. Potential effects of iron: pyrite and pyrrhotite

It is important to note that the Eh - pH diagram included as Figure 12 is based on the important assumption that total aqueous sulfur and total aqueous iron exist at constant concentrations of $10^{-5}$ M. Total aqueous arsenic is also held at $10^{-5}$ M (750 μg/L), although as arsenic solids start to form its concentration becomes limited. The $10^{-7}$ M (7.5 μg/L) arsenic line is shown, delineating the area below which orpiment limits arsenic solubility to this value. A full list of the reactions and equilibrium constants used in the previous and following discussion is provided in Appendix B.

Clearly the availability of sulfide is critical in controlling the mobility of arsenic. The lack of sulfide above Eh values of -150 to -180 mV allows arsenite to remain mobile with respect to various arsenic sulfides. Precipitation of other sulfides, most notably the iron sulfides pyrrhotite (FeS) and pyrite (FeS$_2$), both have the potential to limit free ferrous iron and sulfide at lower reduction potentials:

$$FeS_{(pyrrhotite)} + 2H^+ \leftrightarrow Fe^{+2} + H_2S \quad \text{log} \ K^* = 2.82$$

such that: $[Fe^{+2}][H_2S] \leq 10^{-10.2}$ at pH = 6.5

$$FeS_2 \ (pyrite) + 4H^+ + 2e^- \leftrightarrow Fe^{+2} + 2H_2S \quad \text{log} \ K^* = -2.78$$

such that: $[Fe^{+2}][H_2S]^2/[e^-]^2 \leq 10^{-28.78}$ at pH = 6.5

(calculated from data in Morel and Hering, 1993)
Pyrite, in particular, is stable in the same range as arsenic sulfides begin to precipitate, and could limit porewater sulfide or iron. Which of these species becomes limited is dependent on how efficiently they can be delivered to the subsurface or released from solids. A fixed concentration constraint for sulfide or iron is likely to be an approximation, as well-mixed zones of constant chemical conditions are not in contact with the wetland subsurface, and transport of solutes is governed by groundwater flow and diffusion. Nevertheless, it is informative to consider the effects of sulfide or iron limitations on the idealized system shown in Figure 12.

Figure 13 depicts the stability zones for arsenic solids and iron sulfides assuming that sulfide is limited by the presence of either pyrite or pyrrhotite in the zones that these minerals are stable, and maintaining total As and Fe at $10^{-5}$ M. The first thing to notice is that the stability zones of pyrite and pyrrhotite have shifted somewhat.

Because the precipitation of pyrite occurs at nearly the same Eh - pH conditions as the precipitation of orpiment and realgar, the mobility of aqueous arsenic is not appreciably affected. However, the stability of realgar relative to orpiment increases due to the higher sulfur content of orpiment; realgar is the stable solid below an Eh of -70 mV. The arsenopyrite zone is controlled by its stability relative to realgar and pyrite, and includes an enlarged area. If iron concentrations were higher however, the stability zones of orpiment and realgar could be pushed to lower Eh overall, increasing the mobility of arsenic.

Figure 14 applies to a system in which sulfide is not limited by the precipitation of iron sulfides, and the stability of arsenopyrite relative to the other sulfides is considered. Although aqueous iron is held constant, the relative stability of the pure arsenic sulfides would be unaffected by its limitation. As long as sulfide is not limited (i.e. if iron is limited or neither is limited) arsenopyrite is unstable relative to pyrite and pyrrhotite. Thus the primary arsenic solids in this case are limited to realgar and orpiment.
Figure 5-13. Eh - pH diagram for arsenic with sulfide limited by pyrite and pyrrhotite. $\text{As}_T = \text{Fe}_T = 10^{-3} \text{ M}; \ S < 10^{-5} \text{M}$. FeAsS considered with respect to FeS and FeS$_2$. 
Figure 5-14. Eh - pH diagram for arsenic and sulfide, including pyrite and pyrrohtite, with no reactant limitations. \( \text{As}_t = \text{S}_t = \text{Fe}_t = 10^{-5} \text{ M} \)
These equilibrium calculations provide only a basic framework for the discussion of arsenic fate at this site. The way in which reactants are limited in natural systems is difficult to quantify, and the effect of such limitations potentially large. In natural systems, the rate at which precipitation-dissolution processes occur relative to each other and to the time scale of flows and changes in redox conditions can cause variations from equilibrium.

One example relative to the role of iron with respect to arsenite deserves mention. Aggett and O'Brian (1985) observed the loss of arsenite from lake sediment porewaters as the lake stratified, presumably due to sulfide precipitation ($\text{As} \sim 10^{-5} \text{ M}$; sulfide $\sim 10^{-5} \text{ M}$). Because they did not observe a decrease in iron that should have resulted from the corresponding formation of iron sulfides ($\text{Fe} \sim 10^{3.4} \text{ M}$), they hypothesized that: (1) the rate of arsenic sulfide precipitation may be much larger than that of iron sulfides and (2) that the arsenic sink could be a “mixed iron-arsenic sulfide”, which, due to arsenic limitations, does not deplete iron significantly. These observations underscore the potential importance of transient redox and hydraulic conditions, chemical kinetics, and the formation of unknown stable or meta-stable precipitates in arsenic immobilization.

5.6.4.3. Potential interactions with vegetation

As the arsenic found today above 80 cm depth is hypothesized to have been deposited on a wetland surface vegetated primarily by *Typha* and *Phragmites*, uptake and immobilization of arsenic by these macrophytes is a possible storage mechanism. Streit and Stumm (1993) generalize that plant uptake of arsenic is usually passive, although active uptake could occur as a phosphate analog. Studies of uptake by *Typha* and *Phragmites* indicate that arsenic is tolerated in high concentrations, and often translocated.
from roots to leaves, but not bioconcentrated. Wells et al. (1980) found up to 40 mg/kg in the typha leaves. Babcock et al. (1983) found less than 15 mg/kg in typha rhizomes and leaves when corresponding soil concentrations were as high as 529 mg/kg. Therefore, it is not likely that the presence of these macrophytes alone in the contaminated horizons can explain the residual sedimentary arsenic.

5.6.4.4. Mobility of shallow arsenic

A number of arsenic-containing solid phases have been identified to explain the presence of high soil concentrations of arsenic in the shallow peat soils. Given the expected migration of the redox gradient over depth in these soils, an additional question arises—how have these reservoirs of arsenic persisted over more than 50 years?

A number of investigators have noted that oxidation of sulfidic lake and river sediments could cause massive releases of arsenic to the water column (Moore et al., 1988; Mok and Wai, 1994; Morse, 1994). From the discussion above, it is apparent that a ‘window’ of mobility exists between reduction potentials of approximately 0 and -150 mV in the pH range of this wetland (6.5 - 7 in Figure 12). In this zone, greater than 750 µg/l total arsenic can exist while total iron and sulfur are maintained at 10⁻⁵ M. However, in the wetland environment, the reservoir of sedimentary arsenic, whether oxy-hydroxide or sulfide, can be depleted only if conditions are imposed that cause these arsenic solids to dissolve, and if sufficient groundwater flux exists to carry the arsenic to other locations. The requirement of groundwater flux is particularly important when transient flow conditions are common, in which case the time scales of oxidation-reduction, dissolution, and desorption must be shorter than the time scale of the flow for transport to occur, and sustained periods of chemical mobility and groundwater flux must exist for solid phases to be depleted.
In a wetland, feedback between the groundwater flow regime and redox conditions cause the issue of mobilization and transport to become related in complex ways. For instance, periods of high rainfall recharge will cause downward flow, and introduce dissolved oxygen to the upper peat. However, the fully saturated condition also caused by this condition limits further diffusion of oxygen into the soils, allowing fully reducing conditions to be established once recharge has ended. Evapotranspiration or nearby groundwater withdrawals may drain portions of the upper peat, causing oxidation and temporary mobilization of arsenic from sulfides; however, the arsenic may then be oxidized and scavenged by ferric iron. Longer-term mobilization of deeper arsenic may occur as the redox gradient and corresponding mobility window migrates downward. If this mobilized arsenic is advected downward (in the case of deep groundwater withdrawals) it may again enter a sulfidic zone where it could again be removed from solution by sulfide precipitation. On the other hand, if it is transported upward (in the case of evapotranspiration) it could again enter the oxic zone, and be immobilized by iron or other oxides. Conditions that cause flow paths to pass through oxic or sulfidic zones quickly, or remain in zones of high arsenic mobility, would contribute to the leaching of arsenic solids and the flux of arsenic out of this wetland.

The sand layer located between 0.7 and 2.0 m below the wetland surface could play a significant role in mobilizing arsenic. As a conduit for horizontal flow, it could transport dissolved arsenic away from the streambed. This process could occur due to the relatively short transit times of flow through such a permeable feature, together with a relative lack of oxygen demand in the primarily mineral soils. Such conditions might allow a moderate redox potential to be maintained that precludes the precipitation of arsenic sulfides.

Clearly, speculations as to the fate of arsenic in shallow wetland soils must be informed by the knowledge of the particular hydrology and structure of the wetland. The fact that such a large reservoir of sedimentary arsenic remains in the upper 80 cm of peat in this
wetland over 50 years since its hypothesized deposition indicates that the necessary chemical and hydraulic conditions have not coexisted over a long enough period during this time to remove the arsenic. It is possible, however, that significant quantities have been solubilized and removed, leaving what is observed today.

5.6.5. Deep Arsenic

The deeper peaks in arsenic concentration are located below the industrial horizon, and likely represent arsenic that was delivered to these locations by groundwater flow. At these depths, air entry is not expected to occur, and highly reduced conditions are expected. Arsenite would therefore be the equilibrium dissolved specie, and a variety of arsenic and iron sulfides could stable. Variability of both hydraulic and chemical conditions could have contributed to the concentration of arsenic in certain locations in the deeper wetland soils.

Spatially variable fluxes of dissolved arsenic caused by variability in the permeability of the lower peat could deliver dissolved arsenic, sulfide, and other solutes preferentially to certain zones. Precipitation or sorption of arsenic in these zones would therefore be enhanced. Alternatively, variations in peat chemistry could create subsurface zones that serve as arsenic traps, so that even a uniform flux of dissolved arsenic will leave a non-uniform soil record. Iron or sulfide enriched zones could serve as arsenic traps. The sorption of arsenite onto humic material would not explain the observed peaks in arsenic unless each exists in a preferential flow path.

It is not likely that either groundwater flux or soil properties in this wetland are uniform; the interaction between these two variables is probably responsible for the observed arsenic distribution. In Core 11, for which iron content was measured, the deeper arsenic peak corresponds to a soil layer containing over 17% iron. This quantity of iron
accounts for all of the mineral (non-ignitable) material in the duplicate sample from this depth. It is possible that the arsenic in this sample is highly pyritized.

It should also be noted that the groundwater withdrawals that occurred adjacent to this wetland sporadically from 1964 - 1979 altered hydraulic gradients and groundwater flow paths in the areas significantly (Myette et al., 1987). It is possible that the distribution of arsenic in the deeper peat reflects a period of induced infiltration of river water that carried solubilized arsenic with it for a substantial distance. The groundwater flow patterns that resumed after pumping ceased were not likely to reverse this process.

Deep arsenic peaks are absent in the cores that penetrate sand layers (C-38 and C-17) at depths that correspond to the deep peaks in other cores. The absence of deep arsenic peaks at these locations could result from a number of factors, including: (1) consistently lower oxygen demand and higher groundwater and oxygen flux, maintaining a higher degree of arsenic mobility; (2) lack of an effective sorbent; (3) preferential flushing with relatively arsenic-free water derived from ground water discharge during non-pumping conditions.

5.7. Conclusions

The post-glacial history of the Wells G & H Wetland has been interpreted from several high resolution stratigraphic cross-sections, radiocarbon dates, and paleoecological data. These data indicate that at least one ice block depression was left within the current boundaries of the wetland. After this kettle pond filled with limnic sediments circa 9600 ybp, a woody wetland developed. The wetland became more herbaceous after 8000 ybp, with a sedge meadow becoming established circa 2200 ybp. The start of the colonial period (circa 1650 AD) is apparent from the dramatic increase in Ambrosia pollen at about 75 cm depth, where sedge remains become the overwhelming constituent of the peat. An abrupt transition to Typha peat at about 55 cm is followed at 40 cm depth
by the decrease in *Castanea* pollen that identifies a soil horizon correlated with the
decline of the Chestnut Tree due to the 1920 Blight.

The evolution of the wetland from late-glacial to colonial times is consistent with the
current interpretations of local deglaciation, as well as the record of Holocene climate
change. Industrial activity in the Aberjona Watershed had substantial impacts on the
hydrology of the Aberjona River, as well as on its quality. It is the latter effect that is
suspected to have caused the most profound change in the Wells G & H Wetland system-
- namely, the transition from wet sedge meadow to cattail marsh in the late 1800s. The
horizons above this depth contain up to 8000 mg/kg arsenic, a metalloid that was released
upstream from this wetland beginning in the 1880s as a by-product of the chemical and
pesticide industry. It is quite possible that the sedge meadow that existed during colonial
times was poisoned by floodwaters that deposited toxic sediments on its surface, and that
new species with high metal tolerance, including *Typha* and *Phragmites*, subsequently
colonized the wetland.

The record of arsenic distribution in the Wells G & H Wetland is explained by several
mechanisms. Surface deposition of contaminated sediments can account for the arsenic
observed above the industrial horizon. Percolation of surface water could also have
carried sedimentary arsenic 10 - 20 cm below the surface through the open structure of
the surface peat. The sedimentary arsenic in the shallow peat has persisted for over 50
years, although it unclear to what extent the original deposits may have been depleted.
Oxy-hydroxide and sulfide phases probably dominate the immobile fraction of arsenic.
That arsenic is restricted to a narrow window of redox potential may have been important
in preserving this record of contamination. The deeper (> 80 cm) arsenic contaminated
soils are believed to reflect deeper subsurface transport of mobile arsenic, perhaps during
the pumping of Wells G & H. This arsenic eventually became immobilized in zones to
which reactive solutes (including arsenic) were preferentially delivered or within which
reactive solid or aqueous phases pre-existed.
5.8. Further Work

The potential for arsenic transport within the Wells G & H Wetland depends on both the existence of flow paths and groundwater fluxes and the mobilization of arsenic from streambed or overbank sediment deposits. The storage of arsenic within the wetland also requires stable sorption of arsenic to components of the wetland soil under highly reducing conditions. A more complete model of this system would consider the hydraulic forcing and its effects (i.e.: flow paths) in the wetland, the mobile forms of arsenic in the wetland porewater, and the sorption mechanisms for arsenic in this wetland. I begin to develop such a model in the next chapter, with particular emphasis on the hydraulic forcing (advective) aspect of solute transport.

Additional areas for further investigation of these issues include the determination of the chemical forms of arsenic in the soils of this wetland, and the relation of the deeper arsenic peaks to variability of soil hydraulic and chemical characteristics.
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6.1. Introduction

Wetland deposits are often located in groundwater discharge or recharge zones where highly three-dimensional flow regimes exist. These flow regimes, along with the chemistry of wetland sediments and pore waters, control the exchange of solutes, including pollutants, between groundwater and surface waters. Where this occurs, deposit scale heterogeneities in soil permeability and chemical characteristics may be of great significance to modeling the transport and fate of such solutes.

Deposit scale heterogeneities, especially within wetlands, are often ignored in watershed-scale modeling of groundwater flow and solute transport, most often because discrimination of such features is extremely laborious by conventional means. Networks of borings spaced closely enough to detect these features are prohibitively expensive and otherwise impractical; while typical in situ hydraulic tests necessarily average the effects of the surrounding media. Geophysical methods, while providing continuous and very broad coverage, do not appear as yet to possess the resolution or discrimination necessary to identify features that may dominate the hydraulics and chemistry of transport between surface and groundwaters. As a result, most watershed scale models use averaged bulk parameters to represent wetland deposits as single geologic and hydrologic units.

The purpose of this chapter is to measure and understand the effects of spatial and temporal heterogeneities on the flow of groundwater in wetlands and their potential to transport and store toxic metals such as arsenic. Using the Wells G & H Wetland in Woburn, Massachusetts as a study site, our primary goal is to relate observed and inferred hydrologic processes to the distribution and fate of arsenic in wetland soils. Testing the hypothesis that the arsenic carried by the Aberjona River could have been transported through this wetland to municipal wells G & H is strong motivation for this work.
Subsurface features within this wetland were mapped at a high resolution using a specially designed piezocone penetrometer described in Chapters 3 and 4. In Chapter 5 the distribution of soil arsenic was related to the post-glacial history of the wetland, and the influence of wetland chemical conditions on arsenic mobility was discussed. In this chapter we interpret additional hydraulic and chemical data that were gathered over a span of five months, including hydraulic parameter measurements, hydraulic heads, and groundwater chemical data. The monitoring period included a six week interval during which a monitoring well installed in a sand layer within the wetland was pumped continuously.

A simple numerical groundwater flow model is implemented as part of this study to accomplish the following goals:

- evaluate the accuracy, adequacy, and sensitivity of parameters measured in the field and the laboratory, and provide estimates of others that were not measured
- illustrate the qualitative and semi-quantitative aspects of flow and transport in the Wells G & H Wetland
- demonstrate the potential effects of soil heterogeneities and transient flow on arsenic flux
- reveal the possible long term effects of natural and man-induced hydraulic forcing that may help to explain the observed arsenic distribution, and evaluate the potential for arsenic transport through the wetland to Well H.

The cross-section model was not appropriate to explain observations made during the pumping interval because of the radial flow induced by pumping. For this reason, a simple analytical model was also used to explain the advection of induced stream recharge and arsenic towards the pumped monitoring well. These models are not intended to be quantitatively predictive, but are more accurately described as analytical tools useful for synthesizing and understanding a large quantity of different types of data.
As described in Chapter 5, a large reservoir of arsenic and other toxic metals in the wetland soils is believed to have been derived from contaminated riverborne sediments delivered to the wetland surface during periodic flooding. The presence of occasional peaks in arsenic concentration in deeper wetland soils (below the industrial horizon) is thought to be evidence of arsenic transport in groundwater. The data presented in this chapter demonstrate that arsenic is currently mobile within the wetland under certain conditions. Predictive model simulations provide valuable insight into the effects of natural wetland hydrologic processes, the importance of heterogeneities and transient flows, and the wetland response to municipal pumping. Arsenic sources within the wetland that are most likely to affect well water quality are identified.

6.2. Groundwater Flow in Wetlands

Numerous authors acknowledge hydrology as the most important factor in the control of vegetational and chemical dynamics in wetlands; however, only recently has groundwater flow within wetlands been studied comprehensively or quantitatively (Ingram, 1967, 1983). The following discussion summarizes general aspects of wetland groundwater hydrology as addressed in the literature, as well as the measurement of the groundwater flow parameters, hydraulic conductivity and storativity.

Peat deposits take on different hydrological roles as a function of the regional hydrology and particular features of the deposit. Strong regional or hydrologic setting controls are sometimes identified. Neuman and Dasberg (1977), postulate a regional control mechanism resulting from mountain recharge entering an Israeli peatland at depth. The resulting strong upward gradients prevent local aerial recharge from penetrating deeper than several meters. Hemond and Fifield (1982) describe a one-dimensional subsurface flow regime in salt marsh peat that is controlled primarily by evapotranspirational demand. Siegel (1988) concluded that interaction with regional groundwater was very limited in an Alaskan Wetland complex-- less than 1% of the total annual recharge, and
controlled more by the hydrologic setting than by the wetland type. Siegel and Glaser (1987) claimed that significant groundwater flow through a peatland in Minnesota is a likely source of the dissolved solids found in adjacent minerotrophic fens.

Seasonal transitions from recharge to discharge have also been observed. Motts and O'Brian (1981), found a Lincoln, Massachusetts wetland to be a groundwater discharge zone for most of the year except for six weeks in the fall, when it recharged local groundwater. Recharge-discharge relationships in the Lake Agassiz peatlands of Minnesota were modeled by Siegel (1983) to investigate the groundwater flow-bog growth feedback mechanisms that may control the evolution of peatlands. He found that most groundwater flow occurs in local flow cells driven by increased recharge over raised bogs.

Although preferred flow horizons are often identified, typically these features remain largely unexplored (Siegel, 1983; Siegel and Glaser, 1987). Several investigators record observations and hypotheses regarding high permeability “water tracks”, “windows”, or “flushes” that connect peat deposits to local aquifers, but explanations as to their origin or detailed function are scarce (Ingram, 1967; Boelter and Verry, 1977).

6.2.1. Hydraulic Conductivity

The experimental determination of hydraulic conductivity in peats is problematic. The particular difficulties associated with sample disturbance, proper application of field test methods, and widely varying properties of peat deposits are well-reviewed by Rycroft et al. (1975a). Careful analysis of seepage tube tests led Rycroft et al. (1975b) to consider the possibility of Non-Darcian flow in peat. Waine et al. (1985) found a maximum increase in hydraulic conductivity of a factor of 2.5 over a gradient range of 3 to 17.6. Porewater - matrix interactions and matrix expansion were suggested to account for these effects.
More recently the results of transient seepage tube tests have been re-interpreted with a generalized Darcy-type model which accounts for unsteady soil moisture conditions caused by pore dilation (Hemond and Goldman, 1985). This model, which observes Darcy’s linear velocity-gradient relationship, accounts for most of the non-linear behavior observed by Rycroft et al. (1975b). An effective stress model is also presented that supports Waine et al.’s hypothesis that matrix expansion causes increases in steady state hydraulic conductivity measurements under larger gradients.

It is difficult to assign a typical value of hydraulic conductivity to peat deposits: values reported in the literature range from $10^{-8}$ to $10^{-1}$ cm/sec, with most falling between $10^{-6}$ and $10^{-2}$ (Hemond and Fifield, 1982; Knott et al., 1987; Motts and O’Brian, 1981; Siegel and Glaser 1987; Siegel, 1988a; Chason and Siegel, 1986; Dasberg and Neuman 1977; Gafni and Brooks, 1990; Nuttle, 1988; Rycroft et al., 1975a).

Some of this variability is due to the natural vertical zonation of peats caused by the deposition/decomposition process. The deeper zones of progressively more humified (hemic and sapric) peat (often associated with the cutotelm in northern bogs) are generally found to be denser, fairly non-conductive, and relatively isotropic compared to the upper fibric peat of the acrotelm. (Romanov, in Ingram, 1983; Siegel, 1983). In some studies, measured peat conductivity has been successfully correlated with degree of decomposition as measured by bulk density, fiber content, or other test parameters such as the von Post Scale (Rycroft et al., 1975a; Boelter and Verry, 1976; Gafni and Brooks, 1990).

This idealized description, however, does not account for other types of heterogeneities that can affect the spatial variation in peat conductivity (e.g. concentrations of macropores, buried wood, silt or sand lenses, pipes or water tracks), and does not, in general, account for the relatively high rate of groundwater flow thought to occur in some peat deposits.
Beven and Germann (1982) report that fractures and root cavities may provide for macroporous flow in peats. Hemond and Chen (1990) used dye to trace flow in 60 to 200μm pores associated with plant roots in salt marsh peat. The largest cavities are the first to drain water, and at larger depths can easily increase hydraulic conductivity by more than an order of magnitude (Beven and Germann, 1982). Hemond also notes that these macropores could be self-perpetuating in the root zone, if roots tend to search out oxygen pockets. Siegel (1988a) found vertical flow concentration in Alaskan peat hypothesized to occur in more permeable “pipes” found near the contact between peat and mineral soils.

Siegel (1988a) found that hydraulic conductivity did not vary with depth in Alaskan peat deposits. Chason and Siegel (1986) performed field and laboratory hydraulic conductivity tests to depths of three meters in Lost River Peatland, Minnesota. They report larger hydraulic conductivity values than previously reported in the literature for similar peats (> 10^{-3} cm/sec. for moderately decomposed peat), and little correlation with depth below 50 cm.

Boelter and Verry (1977) describe vertical “windows”, or amorphous channels as the features that convey subsurface water to minerotrophic or poor fens. Ingram (1967) identifies water tracks, regions of increased percolation and permeability, that may exhibit more eutrophic vegetation due to increased nutrient concentrations or higher flow rates.

Hydraulic conductivity anisotropy calculated from permeameter tests generally follows vertical trends in conductivity. Typically the ratio $K_h:K_v$ decreases with depth as increased decomposition and consolidation destroys horizontal structures. Motts and O'Brian (1981) report 100:1 anisotropy at shallow depths (10 cm), but isotropic soils at depths >40 cm where hydraulic conductivity is two orders of magnitude lower. Some researchers report values for $K_h/K_v$ of up to 100 even for deeper samples (Dasberg and
Neuman, 1977; Chason and Siegel, 1986). Siegel (1988a) reports the presence of alternating 2.5 cm bands of sapric and fibric peat in Alaska, called “recurrence horizons” (Glaser, in Siegel, 1988a), which could create large anisotropy values at larger scales.

Field tests conducted by Boelter (1965) revealed occasional anisotropy values of as much as three or five, but generally showed the peats he tested to be isotropic. Fifield (1981) compared permeameter measurements of vertical conductivity with seepage tube measurements of horizontal conductivity and found less than order of magnitude differences. Citing Fifield, Nichols (1985) made the assumption for his seepage tube tests that peat was isotropic at the scale of the tests (5 to 30 cm.).

In summary, conductivity and anisotropy vary greatly, and not always predictably. As with any heterogeneous material, the more permeable zones will control water movement. However, the classical concept of an upper active zone and a lower inactive zone must be evaluated on a site-by-site basis for several reasons: 1) other features existing at depth may affect flow; 2) decomposition may not always be effective at destroying structures and densifying peat; and 3) hydrologic boundaries and regional flow regimes may decrease the importance of shallow horizontal flow.

6.2.2. Storage

6.2.2.1. Specific Yield \( (S_y) \)

The specific yield, or the drainable porosity, of any soil is a linear model of the relationship of storage to hydraulic head, based on the mechanism of pore drainage. Because of the strong capillarity of smaller pores, however, actual yield is seldom linear near the soil surface. Whereas larger pores (i.e. several mm radius) drain freely to within a cm of the water table, a smaller pore (i.e. 100 \( \mu \)m radius) will stay saturated to 15 cm above the water table. The water table depth at which a pore starts to drain is called the air entry pressure. Many peat soils are bi-porous, containing pore spaces at both the
millimeter and the micron scales. In addition, a certain portion of the water in peat soils is immobile due to physical or chemical binding to the peat solids. An extensive categorization of peat water content is provided by Romanov (1968).

The specific yield of peat is zero until the water table falls far enough below the surface to cause drainage of the largest pores. Thus, even in a vertically homogenous soil profile, the effective specific yield becomes larger with depth of the water table, until the air entry pressure of the smallest pores is exceeded. In the case of a rising water table, specific yield begins to fall as the distance to the ground surface decreases below the highest air entry pressure, becoming zero when the capillary fringe intercepts the ground surface.

Hemond and Chen (1990) measured the water released from samples of salt marsh peat with increasing drawdown, finding two linear portions on the curve. The steeper portion began at an average of 18 cm drawdown below the peat surface. This later, higher value of the storage coefficient corresponded to pore drainage above the air entry pressure. The linear nature of the later portion of the curve indicates a relatively small range of pore sizes. Laine (1984) measured variations of specific yield with depth in peat monoliths. She found that the properties of peat layers located 10 cm above the water table were almost as well correlated with measurements of specific yield as the properties of the water table horizon itself, revealing the importance of storage above the water table.

Nuttle (1988) measured total specific storage of salt marsh peat in a 50 cm deep lysimeter, and found a maximum at an average of 15 cm water table depth. The increase from 0 to 15 cm is attributable to increasing air entry and pore drainage, whereas the decrease below that depth is hypothesized to be related to the swelling of air-filled root cavities. As the air trapped in these cavities expands, the surrounding pore water is displaced; however, the hollow roots can swell only to a finite size, at which point the associated release of water ceases.
Boelter (1965) measured the water content of several different types of peat at a range of suction values equivalent to water table lowerings of 5 to 100 cm, finding considerable variation between surface horizons of undecomposed moss peat and deeper more decomposed herbaceous peats. While all samples had large water contents (>95% for undecomposed samples, 80 - 90% for more decomposed samples) the undecomposed peats released most of their water at lower suction values, whereas the other samples retained up to 75% of their water, resulting in low specific yields even at high suction. The strong degree of vertical zonation typical in peat deposits, particularly near the water table, thus creates another, very large source of variability in specific yield, with values typically decreasing sharply as the water table falls into more decomposed horizons. This effect could offset the predicted increase in storativity with depth due to air entry.

6.2.2.2. Elastic Storativity ($S_e$)

The elastic storage of peat can be considerable. Elastic storage coefficients for Sippewissett Salt Marsh peat have been determined by Hemon and Fifield (1982) and Nichols (1985). Their average result was 0.001 cm$^{-1}$, a value that can easily dominate the total storage of a peat deposit only several meters thick, even after some air entry occurs. Nuttle et al. (1990) measured somewhat smaller values for a different Belle Island salt marsh peat, accounting for about 20% of total specific storage.

6.2.2.3. Hysteresis

Hysteresis in soil moisture is well described in the literature, and is generally attributed to pore channel variations (Bras, 1990). The soil wetting curve lies at lower soil moisture than the drying curve, causing less water to be released upon drying than is stored during wetting. Hysteresis in $S_y$ can be large for peat in particular because of their highly variable pore structure and bi-porous nature. Macropores can fill quickly on saturation, providing a large effective porosity. If this water is subsequently carried into the connected finer pore structure by capillary forces, these same capillary forces must be overcome before drainage can occur.
Hysteresis in the compressibility (and therefore Se) of soils has long been described in the geotechnical literature, although its importance at the low effective stresses experienced during normal water table fluctuation is unclear (Lambe and Whitman, 1969).

6.2.3. Evapotranspiration (ET)

Evapotranspiration (ET) is an important part of both the water and energy balance for wetlands, due to the consistently high water table and profuse vegetation. At rates varying from several mm to several cm/day, ET can be larger than average rainfall recharge, and may the primary sink in systems without significant surface water drainage, or where soil permeability is low.

Ingram (1983) finds substantial variability in the literature on wetland ET surrounding the estimation of potential and actual ET, and faults the lack of reliable measurement techniques. He also concludes that helophyte mires (such as the Wells G & H Wetland) differ profoundly from other wetlands in their evaporative behavior. Helophyte dominated systems have a strong seasonally positive feedback, caused by the increase in winter albedo (and thus lower ET) due to coverage by dead plant remains, and potentially enhanced heating and moisture transfer in the growing season, such that ratios of actual to potential ET can exceed two. It should also be noted that ET is not necessarily a simple function of the water table elevation, especially in wetlands. Boelter and Verry (1977) note that maximum ET occurs when the water table is about 10 cm below the peat surface in some bogs, allowing for the maximum direct evaporative surface and increased root-aeration and hence water uptake.

ET rates have been measured for several of the macrophytes currently found in the Wells G & H Wetland, including Typha, Phragmites, and Carex. Although these measurements were made in different locations, the results provide a range of reasonable values for ET flux from wetlands populated by this type of vegetation. Blaney (1933)
measured up to 13.5 mm/day for *Typha latifolia* in summer months in Santa Anna, CA, falling to 6.3 mm/day in November. Rates measured at the Great Salt Lake in Utah were similar in summer, but fell to 3 mm/day by September. Preuger (1991) measured ET rates of 4 - 20 mm/day in *Typha* lysimeters located in Logan, Utah. Smid (1975) measured 5 - 6 mm/day ET from *Phragmites* during summer in Czechoslovakia, falling to 1.5 mm/day in October. Blaney and Young (1942), working at a *Carex* wetland in Fort Collins, Colorado, measured an average of 7.8 - 10.2 mm/day ET over the period of May to September. Priban and Ondak (1980) measured ET rates as high as 6.5 mm/day for sedge-grass marsh in Czechoslovakia.

ET from wetlands can be estimated by a number of techniques, including, but not limited to, water and energy balances, direct measurement by eddy correlation or weighing lysimeters, and indirect measurement using water table hydrographs.

The so-called groundwater fluctuation technique, which converts observed changes in water table elevation to ET water fluxes is used in this study (Winston, 1994; Laine, 1984; Mitsch and Gosselink, 1995). The result is obviously sensitive to the storativity of the peat. If the wetland surface is flooded, storage is equal to unity. If the peat remains fully saturated but not flooded throughout the ET cycle, then knowledge of the specific yield is not necessary; water is released by elastic storage only.

If the surface peat is not saturated, or if the water table drops far enough to cause pore drainage, the specific yield of the soil becomes important. Because specific yield is highly variable and difficult to measure, water levels are often maintained above the ground surface in lysimeters used to measure ET, although this practice can introduce errors relative to *in situ* measures (Preuger, 1991; Romanov, 1968; Priban and Ondok, 1985). In the presence of an unsaturated zone, the assumption must also be made that the ET flux is derived from the saturated zone only, and that the soil moisture profile is constant with time.
An additional problem with this method of measuring ET flux is accounting for groundwater seepage, which, if sufficient, can significantly damp the diurnal head signal attributable to ET. Hydrograph separation techniques have been applied to account for groundwater seepage. The simplest applies a correction based on the rate of water level rise between midnight and 0400 (Mitsch and Gosselink, 1995). More sophisticated methods are reviewed by Laine (1984), which attempt to account for variation in groundwater flux caused by ET drawdown.

In our case, the groundwater model itself accounts for the variable influx of groundwater that partially replenishes evapotranspirated water during the day. Winston (1994) used a distributed parameter model in assessing the use of hydrographs for ET estimation. His work shows that many factors, including temporal and spatial variability in ET and spatial variability in soil type, confound the estimation of ET from simple hydrograph data, even if the storage parameters are known with accuracy.

6.3. Study Site

6.3.1. Surficial geology

The Aberjona Watershed lies within the Fresh Pond Buried Valley, the surficial geology of which was mapped by Chute in 1949. Chapter 5 describes the late and postglacial history of the Wells G & H Wetland, which is located within the third in a series of sand and gravel outwash deposits that filled the valley as ice retreated northward.

6.3.2. Hydrogeology

The Aberjona River Valley was first investigated for purposes of water supply in 1958, and although the quality of groundwater was deemed unfit for drinking, the quantities
available were large (Tarr, 1987). As early as 1973 a report issued by the Massachusetts Division of Water Supply described the Aberjona River as “somewhat unique in that by nature it is a ‘leaky’ river,” an observation that provides an explanation for the large capacities of Wells G & H, which were installed in 1964 despite the unexplained poor quality of groundwater pumped from this aquifer.

Myette et al. (1987) describe the stratigraphy that underlies the Wells G & H Wetland, based on a number of seismic profiles and soil borings. A thin layer of glacial till is found overlying the deep bedrock in this area, in turn overlain by up to 50 m of glaciofluvial sands and gravels. The transmissivity of these deposits is reported to be 1,000 to 1,300 m²/day (11,500 to 14,000 ft²/day). The Wells G & H Wetland covers the floor of the central Aberjona River Valley, which is incised into these deposits. Peat in most places is approximately two meters thick; in several areas, much thicker (up to 8 m) deposits of lacustrine and peat soils till ice-block depressions in the valley floor. Our study site is located in one such depression which is bordered by upland, described in detail in Chapter 5.

6.3.3. Previous flow models

A number of groundwater flow models have been implemented to investigate watershed scale flow and the relationship of the Aberjona River to the former water supply aquifer. Several (but not all) of these models are briefly discussed below.

6.3.3.1. USGS (1989)

de Lima and Olimpio (1989) constructed a flow model (using MODFLOW, McDonald and Harbaugh, 1988) of a portion of the Aberjona Valley Aquifer to aid the USFPA in the evaluation of pump and treat alternatives for the widely polluted aquifer. The model domain spanned 2 km² (0.8 mi²), extending to the stratified drift/till boundaries to the east and west. The model consisted of three layers: the top 6 to 10 m of sand, silt, clay or peat, an underlying layer composed of 10 m of sandy soils, and a basal layer of 3 to 15 m
of coarse sands and gravels. The smallest grid cells, located in the immediate vicinity of Wells G & H, were 6 x 6 m (20 x 20 ft); wetland cells were approximately 6 x 30 m. The peat was included in the model only in areas where it was estimated to be at least 3 m (10 ft) thick. In these locations, limited to two small areas west of the river, adjacent to Well H and just south of Well G, the peat was assigned a uniform hydraulic conductivity of 6.4 m/day (21 ft/day) and a specific yield of 0.3 to 0.45. The river was modeled as a head-dependent flux boundary with streambed conductance of 0.6 m/day over a thickness of 0.3 m. The entire modeled portion of the wetland deposit therefore existed within one model layer with uniform properties.

Transient calibration of the model was limited to matching of a 30-day pump test at wells G & H. Temporal variability on the order of days, such as would be driven by variations in ET and rainfall, was not simulated.

6.3.3.2. Brainard (1990)

Brainard (1990) implemented a two-dimensional finite element flow model (AQUIFEM, Wilson, 1979) for the Aberjona sub-basin of the watershed. Peat soils were not included in the model, and the river was modeled as a specified-head boundary with no bed resistance. This model was intended to be conceptual in nature, and was used to illustrate the need to consider vertical flow, especially near discharge points like the Aberjona River. An approximate analytical solution was also developed for a hypothetical, homogenous aquifer to evaluate the effects of vertical flow on transport. This solution combines Dupuit flow to a fully penetrating stream in two dimensions perpendicular to the valley, and radial flow to the river as a line sink with a finite width and depth. The results of this analysis show that even in homogenous aquifers the effects of vertical flow are substantial, causing transport paths to intersect the streambed some distance down-valley from where two-dimensional analyses would predict.
6.3.3.3. Reynolds (1992)

Reynolds (1992) implemented another watershed scale model using the Dynflow code (CDM, 1984), a three dimensional finite-element computer code. The modeled area in this case spanned 11 km² (4.25 mi²). The model grid provided the finest discretization along the river, although the smallest elements were larger than 50 m. Ten layers were used, the bottom of which provided for some flow in fractured bedrock. The thickest peat deposits intersected the top four layers of the model, and were assigned a hydraulic conductivity of 15m/day (50 ft/day). The river was modeled as a specified-head boundary. Effective riverbed conductivity is then the vertical conductivity of the peat, 0.3 m/day (1 ft/day). Although vertical flow was more adequately considered in this model, flow in the wetland area was only resolved at the scale of the entire deposit, internal heterogeneities were not modeled.

Calibration was to steady state conditions only, although transient simulations of seasonal and storm scale behavior were performed. Seasonal simulations indicated that extended periods of river infiltration could occur, especially along breaks in riverbed slope. Storm simulations duplicated observed periods of river infiltration with varying degrees of success. Differences between observed and simulated heads during storms was attributed to the use of long (1 day) time steps.

6.3.4. Industrial influence

Industrialization of the Aberjona Watershed began in the mid-19th century, and seriously degraded Aberjona River water quality by the late 1800's. Flow records for the Aberjona River clearly show a trend towards faster storm response and higher hydrograph peaks over the last century (Solo, 1995). As described in chapter 5, profound changes in the Wells G & H Wetland vegetation and hydrology probably resulted from man's impacts.
6.4. Methods

6.4.1. Mapping of site conditions

Previous chapters have described the methods used to map the stratigraphy at the site, including soil cores and piezocone profiling. Arsenic distribution was measured in six cores; methods and results are described in Chapter 5.

6.4.2. Measurement of hydraulic parameters

6.4.2.1. Hydraulic conductivity

Hydraulic conductivity of wetland strata were measured in both the field and the laboratory. Bialon (1995) conducted a total of 44 laboratory tests with samples of peat and diatomaceous earth from the wetland. Techniques included Constant Rate of Strain (CRS), and flexible wall permeameter tests, and are fully described in Bialon (1995).

Several vertical profiles of hydraulic conductivity were also generated using in situ methods, including rising head, constant drawdown, and constant flow rate tests. The somewhat elaborate and time-intensive method of Nichols (1985) was not employed due to resource limitations. Instead a 4.2 cm (1.25” NPT) PVC wellpoint screened over 20 cm (8”) was driven to successively deeper depths for testing of the various wetland soil layers. In the sand layer previously described, several short and long term pumping tests were also conducted. Data from these tests were analyzed using the ADEPT electronic notebook for Mathcad (Levy, 1994), applying the methods of Hantush (1956), Hvorslev (1951), Jacob and Lohman (1952), Theis (1935), and Theim (1906). The data collected, as well as the analytical results are included for each of these tests in Appendix D.

6.4.2.2. Storativity

Although the void ratio was determined for a number of peat samples, specific yield was not measured due to the difficulties involved in relating such measurements to field scale, in situ values. Elastic storativity ($S_e$) was measured in all laboratory tests using the
methods described in Bialon (1995). Effective stress - strain curves generated during the CRS tests are typically plotted to strains of 50% or more to emphasize the break in slope that occurs at the maximum past pressure. Estimated maximum past pressure for these soils was typically in the range of 0.2 to 0.4 ksc (kg/cm$^2$), or the equivalent of 200 to 400 cm soil suction or drawdown.

The value of elastic storativity appropriate for use under average field conditions required close inspection of the data over the stress range of 0 to 0.1 kg/cm$^2$ (0 - 100 cm H$_2$O), and revealed that the stress-strain curve is often linear in this range. $S_e$ was estimated as the initial slope of this portion of the curve, representing the strain response to a range of soil effective stresses caused by drawdowns on the order of 50 cm. Plots of stress vs. strain for these samples are included in Appendix D.

Estimates of $S_e$ were also derived from constant drawdown and leaky aquifer analysis of some of the in situ test data from the sand layer (Jacob and Lohman, 1952; Hantush, 1956).

6.4.3. Groundwater monitoring

6.4.3.1. General considerations

6.4.3.1.1. Time scales

The short (diurnal) time scale of head fluctuations caused by evapotranspiration (ET) as well as the potential importance of short rainfall events necessitated automated groundwater head monitoring at short (< 1 hr) intervals. The added advantages of automated monitoring include the lack of measurement effects (total stress increases due to the presence of the observer walking on the wetland surface), and high accuracy of measurements. The less conductive soils and desired response times also required the use of fast-response piezometers.
Because of the large seasonal variation anticipated in wetland hydraulic regimes, measurements were also needed over a long time scale, spanning a range of conditions.

### 6.4.3.1.2. Spatial scales

A large range in the necessary scale of measurements exists over space as well as time. The stratigraphy of the wetland suggested that vertical scales on the order of 0.5 m may be important. Soil units can pinch out over several meters; river location and geometry varies horizontally over similar distances. In the study area, the river is located 20 m from the wetland edge, and 50 m from the nearest upland monitoring well.

### 6.4.3.2. Piezometer construction

Piezometer screens were constructed of 30 cm long, 3.34 cm (1” NPT) diameter polyvinyl chloride (pvc) pipe with 0.0254 cm (0.010”) slots. The bottom of the screen was plugged with a conical piece of pvc. Each screen was fitted with a compression fitting for passage of a pressure transducer cable and a pvc riser or standpipe with a 0.77 cm inside diameter and 1.37 cm outside diameter (1/4” NPT schedule 80). O-rings were fitted to grooves machined into the top and bottom of each screen such that it would slide snugly into a section of 4.21 cm (1.25” NPT) pvc pipe. A pressure transducer was then installed in the screen, and the outer pipe (casing) driven to the desired depth with the screen retracted into it. While holding the riser pipe at a constant height relative to the wetland surface, the casing was then retracted 30 cm to expose the screen (Figure 1 shows a cutaway diagram of the piezometer).
Figure 6-1 Cutaway view of the cased wetland piezometer.
After installation, the transducer cable and riser are both accessible at the top of the casing. The riser is used to make manual measurements to periodically calibrate and thus correct for pressure transducer drift.

This design proved quite easy to install, and has several other distinct advantages over simpler designs. First, the large contrast between intake size and riser volume ensured fast response time. Second the retractable outer pipe allowed installation without “smearing” soil into the screen slots, providing long life and also ensuring fast response. Finally, the constant diameter of the outer pipe from the surface to the top of the screen minimized the chance of vertical leakage along the casing.

6.4.3.3. Piezometer distribution

A total of seven piezometers were installed. Two clusters of three piezometers each were located at 4 meters from the river and 14 m from the river, screened in the upper peat, the sandy strata, and the lower peat. Another piezometer was located in the sandy layer directly beneath the streambed. Pressure transducers were also installed on a staff gauge in the river and in a monitoring well screened in the aquifer directly below the wetland (see Table 1 and Figures 2 and 3).

6.4.3.4. Monitoring schedule

A shortage of pressure transducers prevented all of the above locations to be monitored at the same time. From five to eight locations were monitored simultaneously throughout the monitoring period, using two Campbell Scientific Data Loggers (Logan, UT).

The monitoring period began on August 10, and ended December 18, 1995. Recording intervals varied from 5 to 30 minutes. Each recorded value was computed as an average of approximately 10 to 20 measurements made during the recording interval.
Table 6-1. Wetland soil layers and piezometer locations

<table>
<thead>
<tr>
<th>Layer (depth range)</th>
<th>Symbol</th>
<th>Piezometers</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surface</td>
<td>SW</td>
<td>streamstage</td>
</tr>
<tr>
<td>Surface Typha Peat (top 20 cm)</td>
<td>P1aS</td>
<td>none</td>
</tr>
<tr>
<td>Typha Peat (20 - 50 cm)</td>
<td>P1a</td>
<td>none</td>
</tr>
<tr>
<td>Sedge Peat (50 - 100 cm)</td>
<td>P1b/c</td>
<td>17s, 38s</td>
</tr>
<tr>
<td>Sand (100 - 200 cm)</td>
<td>Sand</td>
<td>17m, 38m, sb</td>
</tr>
<tr>
<td>Woody Peat (200 - 260 cm)</td>
<td>P2</td>
<td>17d, 38d</td>
</tr>
<tr>
<td>Diatomaceous Silt (260 - 500 cm)</td>
<td>DE</td>
<td>none</td>
</tr>
<tr>
<td>streambed (40 cm - 100 cm)</td>
<td>sb</td>
<td>none</td>
</tr>
<tr>
<td>Aquifer Sand (&gt; 500 cm)</td>
<td>SG</td>
<td>s89s</td>
</tr>
</tbody>
</table>
Figure 6-2. Plan of the study site showing locations of monitoring well, piezometer clusters, and miniature pumping well. Cross-section A-A' is also indicated.
Figure 6-3 Cross-section A - A’, showing piezometer screen locations. Vertical exaggeration is 2:1.
Each pressure transducer was calibrated before deployment. At intervals during the monitoring experiment, depth to water measurements were made at each of the pressure transducers, and the calibration offset corrected.

6.4.3.5. **Pumping test**

The long term pumping test was conducted using a 4.21 cm (1.25” NPT) diameter well screened within the sand strata, located 2 meters from the river (see Figure 2). Pumping at approximately 0.6 L/min was supported without dewatering of the sand layer in step drawdown tests. This rate was eventually reduced to approximately 0.52 L/min for the long term pump test which took place between August 21 and October 3, 1995.

Rainfall records for the monitoring period were obtained from a privately operated weather station in Reading Massachusetts located several km from the site, within the Aberjona Watershed.

6.4.3.6. **Chemical measurements**

Throughout the long term pump test and the remainder of the monitoring period, a number of chemical measurements were made. Both pumped water and river water were sampled regularly. Water samples were obtained less frequently from several piezometers and two temporary porewater sampling points. Measurement parameters included pH, specific conductance, temperature, dissolved oxygen, and total arsenic. On several occasions redox potential and total sulfide were also measured. All parameters except arsenic and sulfide were measured in the field in a flow-through cell using the appropriate electrode.

Arsenic was measured by atomic absorption spectrometry in samples returned to the lab. A 1.5 μm glass fiber filter was used to field-filter samples pumped from the sand layer. Peat porewater samples were field-filtered through a 0.45 μm high capacity glass fiber cartridge filter. Measurements were obtained an average of twice daily during the first week of the pump test, every two days for the second week, and between once and twice
per week for the remainder of the test. An effort was also made to make measurements directly before and after rainfall events.

Total sulfide was measured in the lab for one set of samples (December 18) with a silver sulfide electrode. Sulfide samples were preserved using ascorbic acid pH-adjusted with sodium hydroxide. Field platinum electrode redox potential measurements were made on two occasions (December 5 and 18) using a calomel reference electrode.

6.4.4. Modeling

6.4.4.1. Software

The MODFLOW code (McDonald and Harbaugh, 1988) for simulating confined and unconfined saturated flow was implemented for this study. Our version of MODFLOW was obtained from Waterloo Hydrogeologic (Waterloo, Ontario), along with the Visual Modflow user interface. This version of MODFLOW supports several more recent capabilities, including soil re-wetting and variable head boundaries. The variable head feature interpolates between boundary heads specified for each stress period, changing them every time step. This feature allows a transient boundary to be accurately simulated without an excessive number of stress periods.

The MODFLOW code does not simulate unsaturated flow, and as such does not allow for any hysteresis in soil moisture storage during wetting and re-wetting, or any variability in specific yield due to air entry processes. Soil water storage is simulated by linear elastic compressibility and linear reversible specific yield mechanisms.

6.4.4.2. Boundaries

Our model is constructed along a hypothesized flow line from the upland to the Aberjona River, and oriented as a vertical, two-dimensional slice of the wetland and adjacent aquifer soils (see Figure 2 and 3). The assumption inherent in this geometry is that there
is no net flow of groundwater into or out of the model cross section perpendicular to its length. The base of the cross section, located 10 m below the wetland surface and 3 m below the deeper parts of the wetland is simulated as a flow-line (no flux) boundary. In reality, there is a vertical component of flow at this depth, but the boundary is sufficiently far from the area of interest that effects are small. The western and downgradient edge of the cross-section is also taken as a flow-line (no flux) boundary due to the presumed symmetry of flow to the Aberjona River at the valley axis.

The Aberjona River, located at the (typically) downgradient end of the cross section, is simulated as a specified-head. For calibration and verification simulations, this head was measured. The upgradient vertical boundary of the model, located within the adjacent aquifer, is also a specified-head. The hydraulic head in the aquifer was measured directly below the wetland, and extrapolated to the upgradient boundary using typical gradients calculated from previously measured heads at a number of monitoring wells in the area. Several transition zones were used between the vertical constant head boundary and the modeled area of interest to match the observed head loss between the adjacent aquifer and heads at cluster 38 (see Figure 4).

The ground surface of the model was simulated as a specified flux boundary during rainfall events or daylight ET, a no-flux boundary during non-daylight hours, and a surface water boundary during flooding conditions. The surface water boundary was constructed using an additional row of model cells located at ground surface with a specific yield of 1 and very high hydraulic conductivity. These cells are typically dry, becoming wet under river flooding conditions. Surface water flow per se was not simulated for the high conductivity surface nodes; it was assumed that the surface wets from and drains to the river on a much shorter time scale than that for groundwater flow.
Figure 6-4  Soil type zones as represented in the model. Top layer without stippling is a surface water flow layer. Eastern edge is a specified head boundary. Bottom and western edges are flow line (no flow) boundaries. The Aberjona River is a specified head cell in the upper western corner.
6.4.4.3. Soil types and parameter zones

The model is composed of six different soil types: surface peat (P1aS), upper peat (P1a, P1b, P1c), wetland sand (S_p), lower peat (P2), diatomaceous silt (DE), and aquifer sand (S_o) (see Figure 4 and Table 1). As mentioned in the previous section, two transition zones were also delineated and assigned parameters such that the head loss between the upgradient boundary and the wetland was matched. The boundaries of these soil layers were taken from the piezocone/soil core cross section. The data in most cases were not sufficient to define horizontal variability in soil properties; therefore the intent was to minimize horizontal zoning of model parameters. The results of the tests described in section 6.4.2 were used as initial parameter estimates. Results of these tests and calibration adjustments to these estimates are described in sections 6.5.1 and 6.5.4.

6.4.4.4. Grid

Because the model domain is small and computational limitations were not an issue, a fine discretization was possible (see Figure 5). Horizontal grid spacing was on the order of one meter, with half-meter spacing near the river. Vertical layering was on the order of 20 cm to several m, using a total of 20 layers; finer layer discretization was used in less permeable layers. The surface water boundary constituted the top model layer. The surface peat was represented with a single 20 cm layer directly below the surface water layer. The upper peat, sand layer, and the lower peat are each composed of two layers. The diatomaceous sediments are represented by seven model layers, and the underlying aquifer has five layers. The cells in each layer are offset vertically as necessary to best follow the soil layers, thus optimizing the benefits of fine discretization.
Figure 6-5 Model grid. Specified head cells are designated with a hatched pattern.
Figure 6-6. Calibration data for October 4, 1995 rainfall. Numbers refer to hydrograph features described in section 6.4.4.5.
6.4.4.5. Calibration

Model calibration was achieved by varying a subset of the model parameters and comparing the model output, a time series of simulated heads at the piezometer locations, to measured data. The model was calibrated to rainfall response, because a relatively well known flux of water is added to the system during rainfall, whereas ET, which dominates the system during dry weather, withdraws an unknown flux.

The October 4 rainfall was chosen for calibration, for several reasons: (1) it occurred after the pump was shut down, (2) a maximum number of transducers deployed, providing the best possible data set; (3) the wetland was not flooded, eliminating the need to approximate surface water flow in the simulation; and (4) the rainfall ended in the late afternoon, so the effects of evapotranspiration following the rainfall could be ignored.

The wetland water table fell below the average wetland surface elevation of 12.95 m on September 29, approximately five days prior to the rainfall. As a result, the calibrated specific yield for the upper peat reflected the influence of a degree of pore drainage and re-saturation.

Field and laboratory measurements provided initial parameter estimates. During the calibration process, parameters with the most uncertainty were varied, including the specific yield and conductivity of the peat, conductivity and storativity of the sand layer, and conductivity of the diatomaceous silt. The following important features of the hydrographs recorded for this storm were identified to guide the calibration process (numbers refer to Figure 6).

1. shortly after the rainfall, both P-17s and P-17m peak at about 12.85 m
2. the head increase at P-38m and P-38d was much smaller, and no peak occurred
3. the gradient between cluster 38 and 17 reverses for several hours while heads at cluster 17 peak
4. heads at P-17m P-17d do not increase until the river stage increases abruptly at hour six
5. a vertical gradient reversal occurs between P-17m and P-17d at hour six and again at hour 14
6. no vertical gradient reversal occurs between P-38m and P-38d
7. the head measured in the upper peat at P-17s increases more steeply and earlier than in deeper layers, creating a downward gradient that is reversed only as rainfall tapers off at about hour eight
8. heads measured at cluster 17 fall at the same rate as river stage after peaking

The best fit to the observed hydrographs (judged against the preceding criteria) was further tested by varying a number of parameters within reasonable limits and noting changes in model performance. It was not possible to test every possible combination of parameters, but in some cases parameters were varied in groups in such a way that a change in one parameter that might be expected to worsen the fit was offset by a corresponding change in another. A listing of these tests, discussed later in section 6.5.5.1, is included as Appendix F.

A pre-storm simulation was necessary to provide initial conditions for rainfall simulation, and also to calibrate the model to evapotranspirative (ET) flux. Pre-storm simulations were started high on the receding limb of the previous storm hydrograph, with starting heads set to reasonable values, constant by layer. ET flux was withdrawn from the third model layer, corresponding to the middle of the upper peat. This arrangement was necessary in order to maintain ET flux during periods when the surface peat model cells dried out. ET was withdrawn from 0800 to 1600 hrs daily. As no direct measurements of the ET flux rate were available, it was essentially a calibration parameter, and was adjusted within a realistic range until measured daily head excursions were matched. Once the heads measured just prior to the rainfall event were matched roughly, the model
output heads were used as initial conditions to the calibration rainfall simulation. After calibration to the known rainfall, the pre-storm simulation was repeated using the calibrated model parameters, to derive improved estimates of ET and verify model performance.

The assumption of constant ET flux between 0800 and 1600 introduces an unquantified error, as does the discounting of unsaturated flow. Other errors introduced by unmodeled spatial heterogeneities are considered acceptable for the purposes of these simulations. The model parameters were measured in the field or lab (see Table 2 and Appendix D), and measured ranges used to constrain model values. Although specific yield was not measured directly in this study, it was estimated using the water table response to known rainfall, as described in the previous section.

6.5. Results and Discussion

6.5.1. Parameter Measurements

Table 2 lists the results of laboratory and in situ measurements by soil type. Refined estimates derived from the numerical model calibration are also listed, and are discussed in section 6.5.4. All laboratory measurements were conducted by Bialon (1994), which contains detailed results.

Values of elastic storativity were remarkably similar for the range of peat types tested, spanning only one order of magnitude. No values were as high as the $10^{-1}$ m$^{-1}$ value for salt marsh peat measured by Knott (1987). The sand layer had an elastic storativity of the order of magnitude expected for a “leaky” aquifer (0.05 m$^{-1}$). Interpretation of this value is difficult with respect to assigning parameters to the numerical model. If the
There is no text material missing here. Pages have been incorrectly numbered.

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Table 6-2. Parameter measurements and model values

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Soil Layer</th>
<th>Lab Range</th>
<th>Field Range</th>
<th>Model Parameters Run 23R8</th>
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</thead>
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<tr>
<td></td>
<td></td>
<td>$K_v$</td>
<td>$K_h$</td>
<td>$K_{eff}$</td>
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<tr>
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<td>P1a (surface)</td>
<td>0.00006</td>
<td>0.00001</td>
<td>---</td>
</tr>
<tr>
<td>[cm/s]</td>
<td>P1a (Typha)</td>
<td>0.00002</td>
<td>0.00002</td>
<td>0.000006-0.00001</td>
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<tr>
<td></td>
<td>P1b/c (sedge)</td>
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<td>0.00004</td>
<td>0.00006</td>
</tr>
<tr>
<td></td>
<td>Sand</td>
<td>---</td>
<td>---</td>
<td>0.001 - 0.01</td>
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<td>0.00003</td>
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<td>0.000001-0.000015</td>
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</tr>
<tr>
<td></td>
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<td>---</td>
<td>---</td>
<td>---</td>
</tr>
<tr>
<td></td>
<td>Aquifer</td>
<td>---</td>
<td>---</td>
<td>0.047$^1$</td>
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<table>
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<th>$S_y$</th>
<th>$S_e$</th>
<th>$S_y$</th>
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<td>0.004</td>
<td>0.05</td>
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<tr>
<td>P1b/c (sedge)</td>
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<td>0.02</td>
</tr>
<tr>
<td>Sand</td>
<td>---</td>
<td>---</td>
<td>0.002 - 0.005$^2$</td>
<td>0.008</td>
</tr>
<tr>
<td>P2</td>
<td>0.014</td>
<td>---</td>
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</tr>
<tr>
<td>Aquifer</td>
<td>---</td>
<td>---</td>
<td>0.0005$^1$</td>
<td>0.0005</td>
</tr>
</tbody>
</table>

---

1 From de Lima and Olimpio (1989), USGS Aquifer Model
2 From Leaky Aquifer Analysis, includes leakage from aquitard
“leakiness” of this layer is caused by interbedded organic material of a spatial scale smaller than that represented in the model, than the measured elastic storage is a valid effective parameter for the model sand layer. If however, the measured $S_e$ reflects flow from the adjacent peat layers, at a spatial scale larger than that represented in the model, this effect should be accounted for by the model layering, and an intrinsic value of $S_e$ should be estimated for the sand layer. The results of model calibration described in 6.5.4 help to resolve this issue by providing additional estimates of the effective storage coefficient of the semi-confined sand layer.

Hydraulic conductivity estimates from laboratory and field data were in general agreement, although, as expected, the range of values for any one soil type was larger for the lab measurements than for the field measurements. Significant disagreement does exist for tests in the *Typha* peat. The lower values derived from laboratory tests are believed to be the result of the small test sample size ($<6$ cm diameter). Because the most permeable structures in the peat (such as large *Typha* roots and stems) are larger than the tested sample, they are either destroyed by sample trimming or simply not tested.

The relative values of horizontal and vertical conductivity determined from the laboratory measurements also require some explanation. Although horizontal conductivities are generally lower than vertical conductivities, in most cases the two measurements are not made on the same sample, and thus are not comparable. In addition, few horizontal conductivity measurements were made, lowering the degree of confidence in the results. Based on the texture and small scale layering observed in these soils, the ratio of horizontal to vertical conductivity is judged to be larger than one everywhere, although probably close to one in the woody peat and diatomaceous earth.

Disturbance of the soil structure cannot be completely avoided with either laboratory or field methodologies, although its effects can perhaps be minimized in situ, especially if a larger volume of soil, lying outside of the disturbed zone, is tested. In this case the field
measurements provide better estimates of the bulk hydraulic conductivity of the layers of interest; consequently these measurements were used as initial estimates for the model.

6.5.2. Groundwater monitoring data: hydraulic head

Figures 7 - 10 present representative portions of the records of hydraulic head. Complete records of the monitoring data are contained in the Appendix C. Gaps in the data record indicate that data were not recorded or samples not obtained over the corresponding interval.

6.5.2.1. Pre-pumping period

Three different flow regimes were observed during the monitoring period, and are represented schematically in Figure 11. From the beginning of the monitoring period (August 10) till the start of pumping (August 22), flow in the sand layer occurred in the direction away from the stream, and an upward gradient existed from the sand layer to the upper peat. During this period, head data were collected only at P-38 s, m, and d, P-17m, and P-SB. Heads measured at the deep piezometer at cluster 38 show downward flow from the sand layer, a condition that is not maintained later in this period, and is considered either a transient effect or the result of an inaccurate transducer offset for this recording period. A strong diurnal variation with an amplitude of up to 6 cm is recorded for all of the wetland piezometers. A damped response (2 cm) is recorded for piezometer P-SB beneath the streambed. Barometric pressure measured at the Reading station was also plotted for this period to assure that this aquifer is not stiff enough to be subject to barometric pumping effects.

Consistent flow away from the streambed and upward gradients in shallow soils indicate that ET in the wetland was a local sink for groundwater during this period. Prior to August 6, relatively frequent recharge occurred (weekly rainfall, averaging 6 - 9
Figure 6.7: Groundwater head measurements. Pre-pumping, pre-rainfall period. Barometric pressure recorded at the Reading, MA station is also shown (Pam).
Figure 6.8. Groundwater head measurements. Pumping, pre-rainfall.
Figure 6-9. Groundwater head measurements. Pumping, rainfall period.
Figure 6.10. Groundwater head measurements. Post-pumping, rainfall period.
Figure 6-11. Flow regimes observed during monitoring period. (a) pre-pumping: August 10 - August 21; (b) pumping, pre-rainfall: August 22 - September 15; (c) rainfall period: September 16 on. Arrows in 6-11c indicate general flow directions for non-pumping condition (after October 3). Dashed line shows pumping water table, solid lines shows non-pumping water table.
cm/month), whereas no rainfall occurred between August 6 and September 15. It is therefore likely that this flow regime developed after August 6.

6.5.2.2. Pre-rainfall period

From the start of pumping until mid-September, the pre-pumping flow regime was not substantially changed aside from the local influence of the pumping well. Additional head measurements were made during this period in the stream itself, as well as S89S and P-17s and P-17d. In the near-stream area, gradients away from the stream were enhanced by pumping. The measured horizontal gradient between P-17 and P-38 remained in the direction away from the river, although it was greatly decreased by the drawdown at 17m caused by pumping. An upward gradient from the aquifer beneath the wetland (S89S) continued to exist. Upward gradients were also consistently measured between P-17d and P-17m. Upward gradients are believed to have existed between P-38d and P-38m as well. Poor performance of the transducer at P-38d is believed to have resulted in unreliable measurements at this location. Heads measured in the upper peat (P-17s and P-38s) were the lowest heads measured in the wetland other than at the pumping well itself. The stream stage is the highest head (other than S89S), with a consistent downward gradient between the surface and the sand layer directly below the streambed (P-SB).

Because pumping-induced drawdown at the piezometers closest to the pumping well was on the order of the diurnal head excursions caused by ET, drawdown is not readily discernible over most of the record. Observations of rapid head changes at the start and end of pumping indicate that pumping drawdown was on the order of 3 cm at 17m and 17d, and 1 cm at 17s and P-SB. Drawdown in the pumping well was approximately 55 cm. Head changes in the sand layer during the three days prior and 16 days after the start of pumping are shown in Figure 12.

Strong diurnal fluctuations were also measured during this period. The largest (5 cm) are measured in the shallow peat, with smaller fluctuations (3 - 4 cm) measured in the sand
layer and the lower peat. A significant (2 - 3 cm) diurnal fluctuation was also measured in the monitoring well beneath the wetland (S89S). Within the sand layer (P-38m, P-17m, and P-SB), the amplitude of fluctuation increases with distance from the river. P-SB fluctuates only about 1 cm.

6.5.2.3. Rainfall period

Starting in mid-September, a series of rainfall events spaced an average of 4 - 7 days apart altered this flow regime significantly. Ambient horizontal flow in the sand layer reversed and was consistently toward the stream within a week of the initial rainfall, with the exception of the zone between the pumping well and the stream. The gradient towards the stream varied from 0.002 to 0.01, depending on the time elapsed since the most recent rainfall. Heads in the upper peat were also typically higher than those in the sand layer after mid-September; however, upward gradients existed everywhere else in the wetland and aquifer below. Only during brief intervals during rainfalls did the streamstage exceed heads in the wetland. A temporary flow reversal (downward gradient) also occasionally occurred between the sand layer and the lower peat during rainfall events when stream stage increased substantially.

The magnitude of diurnal fluctuations decreases throughout the period of frequent rainfall (late September - December), although daily head excursions are detectable in the record until early November. During certain periods (most notably 9/27 through 10/4), the diurnal variations in groundwater and streamstage are out of phase. This observation may be evidence of vertical movements in the wetland surface. The daily groundwater head decrease is expected to cause consolidation and movement of the piezometers and staff gauge relative to a stable datum. Such movement partially offsets the head changes in the piezometers, and completely offsets movements of the more stable river surface, which appears to rise as the entire wetland consolidates throughout the day. Based on the measured compressibility of the peat, such errors are expected to be less than 0.5 cm.
Figure 6-12. Groundwater heads measured near pumping well: first two weeks of pumping. Plots with symbols are from hand measurements, others are from data logger files.
6.5.3. *Groundwater monitoring data: chemistry*

The variation of arsenic and conductivity with time is depicted in two plots (Figures 13 and 14). Hourly rainfall data are also included for each plot. Other measured water quality parameters varied less; average values and ranges are presented in Table 3.

6.5.3.1. *Hypothesis*

Observations of chemical changes in the pumped water and samples obtained from various piezometers and the stream are discussed in the following section. These data lead to the hypothesis that the streambed sediments are currently the major source of mobile arsenic. The historical evidence and recent measurements of streambed sedimentary arsenic are discussed in Chapter 5. Specifically, it appears that during periods when the ET exceeds rainfall and the stream recharges the wetland, this arsenic is mobilized and transported downward from the streambed sediments into the sand layer beneath the stream (P-SB). It is then advected away from the stream at a rate enhanced by pumping. Conversely, during periods of increased rainfall recharge, gradients reverse and wetland porewater discharges through the streambed. Any dissolved arsenic transported as far as the oxic horizon or the stream itself would tend to be oxidized to arsenate and scavenged by iron oxides.

The chemistry of the arsenic release was not directly measured, although possible explanations are discussed in Chapter 5. In particular, oxidizing conditions are known to mobilize arsenic from sulfidic sediments such as those found in the streambed (Moore et al., 1988). The mobility of arsenic in the streambed sediments is also apparently related to the direction of flow across the streambed. It is possible that the oxic streamwater that infiltrates during periods of low recharge and high ET (or due to groundwater withdrawals) influences arsenic mobility. This hypothesis has not been investigated directly in this study.
Figure 6-13. Total arsenic in water samples, August - December, 1995. Open symbols represent filtered samples (f). Filled symbols represent unfiltered samples. Note that streambed porewater results are plotted on a separate scale.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>wetland groundwater at pumping well (P-3)</th>
<th>streambed porewater (PW-SB)</th>
<th>stream</th>
<th>aquifer (S89S)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Temperature (°C)</td>
<td>15-18</td>
<td>--</td>
<td>10 - 27</td>
<td>--</td>
</tr>
<tr>
<td>Conductivity (μmhos/cm)</td>
<td>600 - 800</td>
<td>750</td>
<td>350 - 1100</td>
<td>300</td>
</tr>
<tr>
<td>pH</td>
<td>6.1 - 6.4</td>
<td>6.8</td>
<td>6.9 - 7.3</td>
<td>6.1</td>
</tr>
<tr>
<td>Eh (mV)</td>
<td>-224 to -167</td>
<td>-182 to -95</td>
<td>--</td>
<td>-25</td>
</tr>
<tr>
<td>arsenic (μg/l)</td>
<td>30 - 120</td>
<td>150 - 600</td>
<td>1 - 12</td>
<td>1</td>
</tr>
<tr>
<td>dissolved O₂ (mg/l)</td>
<td>0</td>
<td>0</td>
<td>3 - 6</td>
<td>0</td>
</tr>
<tr>
<td>sulfide (M)</td>
<td>$10^{-6}$</td>
<td>$10^{-7}$</td>
<td>--</td>
<td>--</td>
</tr>
</tbody>
</table>
Measurements of sulfide and arsenic throughout the wetland indicate supersaturation of pore water with respect to orpiment ($\text{As}_2\text{S}_3$; $K_{sp} = 10^{-38}$). The presence of $10^{-7}$ M sulfide at $\text{pH} = 6.8$ should limit arsenite solubility to approximately $10^{-8}$ M. The presence of up to 700 $\mu$g/L ($10^{-5}$ M) total arsenic in pore water may be explained by the presence of stable or meta-stable arsenite and arsenate colloids or complexes, such as iron oxides, clay particles, organic molecules, or polysulfides. Even the lowest arsenic concentrations measured in wetland porewater in this study (30 $\mu$g/L or $4 \times 10^{-7}$ M) indicate supersaturation.

Groundwater conductivity data show a similar trend to arsenic, with the exception that the conductivity of the stream water varies considerably. The extremely high conductivity of the streamwater in late summer is probably the result of relatively undiluted urban runoff contributions. The sharp decrease after mid-September is attributable to rainfall recharge dilution, whereas the increases in late November and December may be due to road salting. Weather records from the Reading, MA station indicate a 13 cm of snow and 1 cm of rain on November 29, followed by sub-freezing temperatures. Conductivity may also be generated in the streambed sediments much as arsenic is, although the varying quality of the surface water itself could mask this process. The use of conductivity as a tracer in this data set is precluded due to the large variability in the source concentration.

6.5.3.2. Arsenic in streambed porewater

The stream water itself has consistent concentrations of arsenic less than 10 $\mu$g/l. However, the porewater in the streambed sediments and from the sand layer beneath the streambed had much higher concentrations during the monitoring period. Arsenic in streambed porewater (“streambed” in Figure 13) was 600 $\mu$g/l on November 16, decreasing to 150 $\mu$g/l between November 16 and December 17. It is likely that
concentrations in the streambed sediments were even higher in mid-September, based on the observed decrease in concentration below the streambed ("P-SB") starting in mid-September.

If increases in mobile arsenic beneath the stream are causally related to the infiltration of stream water, then it is expected that mobile arsenic increased from soon after the most recent rainfall (August 6) until mid-September. Based on the same assumption, the frequent rainfall events occurring until August 6 likely limited streambed arsenic during late spring and early summer to the levels observed in December under similar conditions. As the mechanism for release to porewater is uncertain, the time scale on which porewater concentrations increased between early August and late September is difficult to estimate.

6.5.3.3. Arsenic beneath the streambed

During the pumping test, the arsenic concentration in the sand layer below the streambed peaked at nearly 400 µg/l, but diminished to 100 µg/l within a few weeks of the gradient reversal that occurred in mid-September ("P-SB" in Figure 13). Samples from P-SB were not collected prior to September 15. If the source of arsenic is indeed the overlying streambed sediments, arsenic concentrations probably increased at P-SB throughout the weeks prior to September 15. Under pre-pumping conditions, the vertical gradient across the streambed was approximately 0.02. Based on a vertical conductivity estimate of 0.002 cm/sec and an effective porosity of 0.2, the travel time through the streambed is on the order of several weeks under non-pumping conditions. It is therefore likely that the arsenic concentration at P-SB was still substantially lower than 400 µg/l at the start of pumping (only two weeks after the last rainfall) and then increased rapidly under the roughly doubled vertical gradient caused by pumping.
6.5.3.4 Arsenic transport towards the pumping well

Lower concentrations of arsenic were measured in the piezometers further from the river, although a distinct increase (from 30 to 120 ppb in the pumping well) occurred during the three weeks of pump test monitoring preceding the mid-September rainfalls.

The pumping well probably enhanced the mobilization of arsenic relative to that caused by ambient flow alone, resulting in a substantial increase in arsenic concentration at the wellhead shortly after pumping started. A gradient of 0.001 to 0.002 away from the river existed between piezometers P-17m and P-38m several weeks before pumping began on August 21. Seepage velocity under these conditions would be approximately 2 - 4 cm/day, resulting in a travel time from the area beneath the streambed to well P-3 of 50 to 100 days. This explains why arsenic concentrations in the pumped water were initially close to background despite the fact that stream discharge to the wetland began in early August.

To support this interpretation of the data, several modeling experiments were conducted using a simple two dimensional steady state analytical solution for flow to a well adjacent to a stream with ambient flow. This solution is simply the superposition of radial flow, including an image well to maintain a constant head at the stream, and uniform (ambient) flow in the prescribed direction. The solution provides a two-dimensional head distribution, as well as the stream function, which can be used to define the capture zone of the well, the contribution of induced recharge from the stream to well pumpage, and the travel time between sources and sinks. Although this solution assumes homogenous soil properties, a confined aquifer, a fully penetrating stream, and provides only a steady state hydraulic solution, it is still useful to constrain the expected advective transport of conservative solutes between the river and pumping well. Figure 15 shows a schematic representation of this model space, including the source of arsenic at the streambed.
Because the sand layer was not dewatered by pumping, and is at least semi-confined by peat layers above and below, the approximation is reasonable in this respect. The assumption of steady state conditions can be evaluated using the solution of Jenkins (1968) which gives the ratio of stream depletion to well pumpage as a function of time:

\[
\frac{q}{Q} = \text{erfc} \left( \frac{a^2S}{4tT} \right)
\]

where:
\[a = \text{distance to stream}\]
\[S = \text{storativity}\]
\[T = \text{transmissivity}\]
\[q = \text{rate of stream depletion}\]
\[Q = \text{pumping rate; and}\]
\[t = \text{time}\]

Using the parameters estimated from field measurements and development of the numerical model (Table 2), \(q/Q\) reaches 90% within 0.5 days, a time sufficiently short to allow us to disregard the transient hydraulic response in the evaluation of flow to the well over several weeks.

Figure 16 shows the two-dimensional flow net solution that includes uniform ambient flow away from the stream. All recharge is derived from the stream under these conditions. The flow net shows a substantial capture zone, 12 m wide at the river and close to 11 m wide at the well. In the two-dimensional, steady state solution, the rate at which the chemistry of the pumped water approaches that of the river depends only on the distribution of travel times for the flow paths from the river to the well. The shortest travel times in this simulation are slightly more than two days. The seventh streamline (darkened), with a travel time of seven days, encloses 48% of the total well recharge spans the time lag before the well discharges 48% stream water.
Figure 6-15. Schematic representation of flow to a well from the streambed, showing the hypothesized arsenic source and the predominant flow paths to the well through the streambed and the (stippled) sand layer. Conceptualized flow paths in a two-dimensional plan view are projected onto the ground surface.
Figure 6-16. Flownet for a well near a stream with uniform flow away from the stream. Ambient gradient = 0.001; Q = 8.68 cm$^3$/sec; T = 0.3cm$^2$/sec.
The longest paths are on the order of hundreds of days. However, about 2/3 of the flow occurs within stream tubes that are less than 23 days in length. Thus, the concentration of arsenic at the well could attain 2/3 of the source concentration within the three weeks between the start of pumping and the flow reversal that occurred in mid-September. Whereas the predicted concentration at the well is 220 μg/l after three weeks of pumping, the measured concentration was 120 μg/l (see Figure 13). This level of agreement is reasonable given the simplicity of this model.

A number of confounding effects can cause the predications of such a simple model to differ from observations. Among these are:

- heterogeneities in the two-dimensional domain, including variations in aquifer properties, pinching out of layers, and smaller scale features that cause dispersion

- three dimensional flow, including leakage between the sand layer and the upper and lower peat, and the effects of partial penetration of the streambed

- vertical variability in the source-- this model assumes a constant source concentration over the entire thickness of the sand layer below the streambed, as measured in samples from P-SB

- non-conservative solute behavior, including sorption and precipitation

The influence of several of these effects can be evaluated in an approximate manner by further modification of the model. Pinch-outs (such as depicted in Figure 4) can be simulated as no-flux boundaries by adding the appropriate image wells to the solution. Other types of heterogeneities in aquifer properties cannot easily be addressed with this
type of model. Large scale features will cause partial boundary type effects, whereas small scale variability will cause a spreading of the breakthrough curve.

The effect of vertical flow can be predicted using a leakance term that is first order in time along any particular flow path. The first order dependence on leakance occurs if water added to the two-dimensional flow tube from above or below contains no appreciable amount of the solute in question, and serves only to dilute the flow. A mass-balance along a streamtube yields:

\[
\frac{dC}{ds} = (C_l - C) \frac{V_l}{B \cdot V_s}
\]

where:

- \( C = \) aquifer concentration
- \( s = \) streamtube coordinate
- \( C_l = \) leakance concentration
- \( V_l = \) leakage velocity = \( \frac{dh}{dz} K_z \)
- \( V_s = \) seepage velocity along streamtube = \( \frac{dh}{ds} K_h \)
- \( B = \) aquifer thickness

if \( C_l = 0 \), and the velocity along the streamtube is unaffected by leakance, then using \( S = V_s \cdot t \), this equation can be written as:

\[
\frac{dC}{dt} = -\frac{V_l}{B} C;
\]

such that: \( C = C_{source} \exp\left(-t \cdot V_l / B\right) \) along any streamtube

In this way, dilution along each contributing streamtube can be estimated and used to modify the predicted breakthrough curve.
The effects of partial stream penetration can be accounted for by formulating a three-dimensional problem. Chow (1981) developed a three-dimensional solution for groundwater flow from a partially penetrating stream to a well with ambient flow, but his solution requires the aquifer to be thick relative to stream penetration and the stream to be narrow relative to the distance between well and stream. Because this is not the case for this example, an effective distance approximation is used. Kazmann (1948) developed an empirical method for determining the effective distance between a partially penetrating stream and a well, but it requires a number of drawdown measurements between the river and the well. Hantush (1965) describes an alternate method that minimizes the spurious storage effects that are introduced when the well is positioned further from the stream in model space. Because insufficient data are available to use Kazmann's technique, and we are concerned primarily with steady state conditions, an effective distance was estimated simply as:

\[ d_{\text{eff}} = d \left(1 + \frac{B_{sb}}{K_{sb}} \frac{K}{d}\right) \]

where:

- \( B_{sb} = \text{streambed thickness} \)
- \( K_{sb} = \text{streambed conductivity} \)
- \( K = \text{aquifer conductivity} \)

This formulation considers purely horizontal and purely vertical flow only, and is therefore an overestimate.

Source function variability undoubtedly exists, and could be important. Without any measurements of this variability, its effects cannot be integrated with the model. However, it is likely to be large enough that more detailed chemical modeling, such as simulation of chemical interactions, is not warranted.
For example, linear sorption, leading to retardation, could easily be accommodated within the uncertainty of the source function. It is also highly unlikely, however, that interactions of arsenic with aquifer solids can be described adequately by a simple extension of the model, as by including linear sorption. Immobilization by precipitation of arsenic solids would lead to much more complex and profound effects on transport which are not quantified in this model.

Based on the techniques outlined above, several additional flow nets were generated, as shown in Figures 17 and 18. The predominance of shorter flow tubes between the river and well increases as the effects of boundaries are imposed on the sand layer. Figure 17 shows the results of a no-flow boundary imposed 12 m east of the pumping well, in accordance with the stratigraphic information in Figure 4. The effects of partial penetration using an effective well-stream spacing of 300 cm offset this effect to a degree (Figure 18).

The additional effect of leakage dilution was evaluated using estimates of leakance velocity from field head measurements and calibrated parameters from the numerical model described in section 6.5.4. Leakage between the upper peat and the sand lens was ignored based on the assumption that most of the ET flux in the area immediately adjacent to the river is replenished directly through the peat. The leakage velocity through the lower peat is difficult to evaluate with confidence due to the variation of pumping drawdown between the stream and well. An average head difference of 17 cm over the distance of 360 cm that separates the aquifer from the sand layer was used (see Figure 4). The weighted harmonic mean of the vertical conductivities of the diatomaceous earth and the woody peat is $8.9 \times 10^{-8}$ cm/sec. The dilution decay term is therefore:

$$K_I = \frac{1}{B} \frac{\Delta h}{\Delta z} K_z = \frac{1}{100 cm} \frac{17 cm}{360 cm} \left( \frac{360 cm}{0.00003 cm/s + 60 cm/0.00005 cm/s} \right) = 4.2 \times 10^{-8} \text{ sec}^{-1}$$
Figure 6-17. Flownet for a well near a stream with uniform flow away from the stream. Impermeable boundary located at x = 1400 cm. Ambient gradient = 0.001; Q = 8.68 cm$^3$/sec; T = 0.3 cm$^2$/sec.
Figure 6-18. Flownet for a well near a stream with uniform flow away from the stream. Impermeable boundary located at x = 1500 cm. Effective distance to stream = 300 cm. Ambient gradient = 0.001; Q = 8.68 cm$^3$/sec; $T = 0.3$ cm$^2$/sec.
Therefore, over a travel time of 20 days, a dilution of only about 10% occurs.

The calculated breakthrough curve for a system with a no-flow boundary 12 m from the pumping well and a partially penetrating stream is shown in Figure 19. This breakthrough curve peaks at 220 μg/l after three weeks, when the flow reversal occurs. This model illustrates in a useful way the importance of several features of flow between the stream and well, including flowpath dispersion, "aquifer" boundaries, and partial penetration. The model also highlights the probable importance of unmodeled processes, such as potential heterogeneities along the river axis, source function variability, and chemical interactions. Vertical flow, at least during periods of low rainfall, appears to be of lesser importance. We believe that this model provides the appropriate level of analysis relative to the available data.

6.5.3.5. Well capture and arsenic transport during rainfall period

Concentrations fell from 120 μg/l at the pumping well (P-3) and 55 μg/l at P-17m to a final steady concentration of about 30 μg/l in the three weeks following the onset of rainfall. The arsenic concentration at P-SB fell from nearly 400 μg/l to 70 μg/l, and was continuing to decrease at the end of the monitoring period. The conductivity of samples pumped from these piezometers converged in a similar way during the rainfall period to an average value of about 650 μmhos/cm. Measured values of conductivity and arsenic concentration for the adjacent aquifer (from S89S) are 400 μmhos/cm and 1 μg/l, respectively. The apparently increased wetland "background" levels of conductivity and mobile arsenic are probably related to the chemistry of the wetland soils. As described in Chapter 5, the mobility of arsenic in these soils is believed to be limited under equilibrium conditions by the precipitation of arsenic sulfides. Depending on the specific redox and pH conditions of the soil, arsenic sulfides could maintain arsenic in porewater...
Figure 6-19. Total arsenic at P-3 predicted using the steady-state flow regime depicted in Figure 6-18. Measured concentrations at P-3 and P-SB included for comparison. Open symbols indicate filtered samples. Closed symbols indicate unfiltered samples.
at a concentration in the range of 5 to 50 μg/l, provided equilibrium is attained (reactions 2 and 5 in Appendix B).

The quality of water pumped from the well under the conditions of discharging groundwater is simply a function of the mixing of the induced stream recharge and captured ambient flow. The rainfall events that occurred regularly after mid-September caused the ambient gradient to increase as rainfall recharge infiltrated the wetland from above and the aquifer discharged through the wetland from below. Although hydraulic steady state was never observed, an average hydraulic gradient towards the stream can be used to estimate the relative contribution of stream and groundwater to the well.

Under the same assumptions used in the previous section, applying a gradient of 0.006 towards the stream, the two dimensional model predicts well capture of stream water along about 8 m of its length, deriving 55% of its recharge from the stream (Figure 20). This prediction is not consistent with the observation that arsenic concentration fell below 50 μg/l very quickly at the pumping well in mid-September, while arsenic concentration at P-SB remained above 300 μg/l until the end of pumping.

The rapid response to rainfall at the pumping well is suggests that the capture zone of the well is dramatically altered under these flow conditions, with direct vertical recharge contributing substantially to the dilution of arsenic in the pumped water. The wetland surface was intermittently flooded for much of the time after mid-September, providing a recharge boundary only a meter above the wellscreen. The apparently strong influence of vertical flow during these conditions makes the two-dimensional stream-well model inappropriate for interpretation of the pumping well data after mid-September.
6.5.4. Numerical model calibration

Calibration of the numerical model to observed conditions served to verify laboratory and field measurement of hydraulic parameters, and provide estimates of both evapotranspiration rate and effective specific yield.

6.5.4.1. Non-flooding storm--hydrograph match and parameter estimates

The measured (symbols) and simulated (lines) hydrographs for the October 4 storm are presented in Figure 21. Early rainfall response is evident in the simulated head at the shallow piezometer (P-17s), although the timing is slightly shifted, possibly due to differences between the rainfall records from the Reading station and the actual rainfall at the wetland. As inferred from the measured data, simulated heads at P-17m and P-17d are somewhat isolated from early recharge, and do not increase significantly until the river stage begins to rise. A vertical gradient reversal occurs between the sand layer (P-17m) and the underlying peat (P-17d) as the river stage increases heads within the more permeable sand.

A generally more damped response at cluster 38 is evident in the simulated hydrographs, as it is in the recorded ones. The gradient reversal between cluster 17 and 38 is simulated during the correct time period. The match to P-38m is not as quantitative, especially with respect to the vertical gradient reversal at cluster 38. A small reversal is simulated, whereas none is measured. As stated in section 6.5.2, inconsistencies in the recorded heads for these piezometers suggests that these data are not as accurate as other head data. The parameter estimates generated in the calibration process are listed in Table 2. Model parameters are typically within the field or laboratory measured ranges. As discussed in section 6.5.1, the field measurements of hydraulic conductivity are Figure 6-20. Flow-net for flow to a well near a stream: uniform flow towards the stream.
Figure 6-20. Flownet for a well near a stream with uniform flow towards the stream. Ambient gradient = 0.006; \( Q = 8.68 \text{ cm}^3/\text{sec} \); \( T = 0.3 \text{ cm}^2/\text{sec} \)
Figure 6-21. Measured and simulated hydrographs for October Rainfall. Lines are simulated heads; symbols are measured heads.
considered more reliable than the laboratory measurements; the choice of model parameters reflects this. The only model parameters that lie outside their measured ranges are the elastic storativity of the sand layer and the conductivity of the aquifer. The increased value of conductivity for the aquifer was used because of the way the boundary condition was implemented, as described in section 6.4.4.2. Because the aquifer head was measured directly beneath the wetland and the boundary was imposed at the edge of the wetland, the conductivity values for the aquifer and the transition zones near the boundary are effective values that are used to replicate the hydraulic forcing imposed on the wetland as accurately as possible. As revealed in section 6.5.5.2, the simulated heads at cluster 17 are insensitive to changes in the characteristics of the transition zone, which is used to ensure that the required head loss occurs between the specified head aquifer boundary and the location of cluster 38.

The increased value of elastic storativity (0.008 m\(^{-1}\)) used for the wetland sand layer provided a significantly improved fit to the measured data relative to the measured value (0.05 m\(^{-1}\)) or other, seemingly more reasonable estimates for this soil type (< 0.001 m\(^{-1}\) is more common). The use of 0.008 m\(^{-1}\) in this layer may be thought of as an effective parameter that accounts for the prominent interbeds of detrital peat found in the sand unit described in Chapters 4 and 5.

The two values of specific yield listed for two layers in the upper peat (surface *Typha*: P1aS; and *Typha*: P1a) layers were estimated as fitting parameters for the non-flooding rainfall simulation. Because rainfall rate and hydraulic conductivity were relatively well constrained, the storage values necessary to match the observed hydrograph apply under the antecedent conditions specific to this simulation. The values of 0.05 and 0.08 are reasonable based on measured values for other peat soils, which typically have large air entry pressures. The smaller value used for the surface peat layer reflects the fact that capillary rise cannot occur above wetland surface, which reduces the effective specific yield when the water table is high, as discussed in section 6.2.2.1. Specific yield of the
deeper layers was not estimated because the water table never dropped further than 50 cm below the wetland surface during this period.

The only notable mismatch to the measured data is the lag in the receding limbs of the simulated hydrographs for P-17s and P-17m. In the monitoring data, the head in the sedge peat layer (P1bc) decreases quickly, following river stage, a result that could not be matched exactly by the model. Unmodeled hysteresis in storativity could account for this difference. In the simulation, the volume of leakage from the slowly draining peat layer is equivalent to the previous increase in storage resulting from the rise in water levels. This drainage helps to maintain the head in the sand layer, which also decreases slowly. Section 6.2.2.3 describes a possible bi-porous model for peat saturation and drainage that would allow head to drop in such layers with very little pore drainage.

Head in the upper peat could also decline earlier and more quickly than can be simulated due to the decrease in total stress that occurs as the surface layers of peat drain. In the extreme, the screened layer can be idealized as follows:

- no capacity to mobilize effective stress on the time scale of the storm
- small enough permeability that insignificant porewater flow occurs during the storm

In this case, increased saturation of the more permeable surface peat (P1a) would cause pore pressure increase in the underlying sedge peat layer (P1bc) equal to the weight of the added water. A corresponding decrease in pore pressure would occur as the upper peat drains, without requiring any water flow in the layer below. More moderate assumptions regarding peat characteristics would lead to a model in which finite but small effective stresses are mobilized and some porewater flow occurs in the sedge peat, but head changes are still more rapid than if the layer was considered rigid.
6.5.4.2. Pre-storm: water level recession and ET

Measured and simulated heads for the five day period preceding the storm are shown in Figure 22. Approximate initial heads (constant by layer) were used, so that simulated heads at early times are not expected to match the data. Drawdowns due to pumping at P-3 decrease the measured heads at cluster 17 (and to a lesser extent at P-38m) until hour 158 (0700, October 3), when the pump is shut down; the recovery of water levels at P-17m and d at hour 158 is apparent in these data. The model is expected to over-estimate heads by approximately 1 - 3 cm before this time.

The parameters from the October 4 rainfall calibration were used in this simulation; no additional calibration was performed to achieve these results, except for assigning an ET rate. A daily rate of 3.7 mm, withdrawn entirely between 0800 and 1600 hrs, matched both the daily head excursion and the total water table recession well. This ET rate falls within the published ranges listed in section 6.2.3.

6.5.5. Model performance assessment

Whereas a two-dimensional horizontal flow model (section 6.5.3) fails to represent important vertical flow components, similar limitations exist for the numerical cross section model with respect to horizontal flow. Other stratigraphic cross sections (see Chapter 4) across the model domain indicate that variability exists in three dimensions. Despite these limitations, a good calibration was achieved. The following performance assessment is made within the context of the two dimensional cross section geometry and the relatively sparse data outside of the modeled domain.
6.5.5.1 Parameter sensitivity

The sensitivity of the calibration results (Table 2) to a number of different model parameters was tested using the criteria outlined in 6.4.4.5. As expected, model results are relatively insensitive to the horizontal conductivity of the less permeable layers and the vertical conductivity of the more permeable layers. A table listing these tests is included in Appendix F.

The horizontal hydraulic conductivity of the *Typha* peat (P1a) was varied from 0.00025 to 0.004 cm/sec. Model results were insensitive to decreases below the calibration value of 0.001 cm/sec. However, higher values caused the head at 17s to peak at a later time and recede more slowly, aggravating the mismatch at late times. The vertical conductivity of the *Typha* peat was also varied from 0.00025 to 0.004. While decreases caused slight damping of the response at cluster 17, increases had little or no effect on the model results. The relative lack of sensitivity to changes in the conductivity of this layer is due partly to the fact that it is unsaturated during much of the simulation. In addition, whereas the sand layer tends to control the rate of horizontal drainage through the cross section, the less permeable sedge peat (P1bc) (which separates the *Typha* peat and the sand layer) controls the rate of vertical leakage. Thus the uppermost peat layers function mostly to store water, and are most sensitive to the values of specific yield used to calibrate the modeled response to rainfall.

Horizontal conductivity of the sedge peat (P1bc) was varied from 0.00006 to 0.001. As was the case with the *Typha* peat above, model results were essentially insensitive to this parameter. The vertical hydraulic conductivity of the upper peat layers was varied from 0.000025 to 0.0002. Lower values of vertical conductivity severely limited the simulated early response to rainfall at P-17s, and shifted the simulated hydrograph peak to later times. Higher values (> 0.0003) caused excessive rainfall response at P-17s and also
allowed recharge to the sand layer below, resulting in a poor hydrograph match at P-17m, especially on the rising limb. The highest values of vertical conductivity, equal to that of the sand layer itself, cause the peat and sand to respond to rising river stage as one layer with an average horizontal conductivity and storativity, displaying an over-damped response.

The combined sensitivity to vertical conductivity and specific yield of the sedge peat was also tested by setting these parameters to 0.00025 and 0.06, respectively. The decreased specific yield increases the early response at P-17s close to measured values despite the lower conductance, but causes an excessive peak height and slower drainage of the layer at later time.

Streambed conductivity ($K_v(sb)$) was tested over the range of 0.001 to 0.008 cm/sec. Higher values cause slight decreases in simulated head at all the piezometer locations; model results are otherwise insensitive to increases in streambed conductivity. The highest value in the range is greater than the horizontal conductivity of the sand layer itself, and is therefore considered an unreasonable number. Lower values caused the simulated hydrographs for cluster 17 to become over-damped relative to measurements. Although no data are available for the P-SB piezometer during the calibration period, earlier data indicate that very little head loss occurs across the streambed, supporting the relatively high calibrated value of vertical conductivity.

The horizontal conductivity of the sand layer was varied from 0.0025 to 0.01. Lower values caused over-damping of the hydrographs simulated for P-17m and P-17d. Higher values caused increased peak heights as well as a decrease in the heads at cluster 38 over the entire time period of the simulation. Based on the fact that the confined, one-dimensional flow of groundwater depends only on the ratio of conductivity to storativity, several attempts were made to match measured heads by changing both the conductivity and storativity of the sand layer while maintaining their ratio. Increases in these parameters lowered simulated heads at cluster 38, and required an unrealistic value of
storativity. Decreases in these parameters could maintain a good match at P-17m, but caused poor matches at other locations. The simulated hydrograph at P-17d became over-damped, and heads at cluster 38 increased. As a result, the observed gradient reversal between cluster 17 and 38 was not simulated. These observations reinforce the importance of vertical flow in this system that was noted earlier during the development of the analytical model (section 6.5.3).

Finally, the vertical conductivity of the diatomaceous silt was varied from 0.0000075 to 0.00012. Simulated hydrograph response was less sensitive to decreases in vertical conductivity, although a four-fold decrease caused the rising limbs of the hydrographs at P-17d and P-17m to lag, and the heads at cluster 38 to decrease. Increases in vertical conductivity caused the hydrograph peak at P-17m to increase and delayed the peak at P-17d. In addition increases in vertical conductivity increased heads at cluster 38.

6.5.5.2. Upland boundary and sand layer connectivity

The connection to the upland aquifer boundary was also tested by changing the shape and parameter assignments of the transition zones. The simulated hydrographs at cluster 17 were insensitive to these changes. The simulated hydrographs at cluster 38 were somewhat sensitive to changes in the boundary transition zones. As described in section 6.5.4.1, the particular shape and character of these transition zones is based only partly on the limited subsurface data for this area; the primary criterion for designing the transition zone was to simulate reasonable hydraulic forcing on the wetland but matching the observed head losses between the aquifer and the wetland. The model is not intended to simulate the details of flow in this area.

It is apparent that the sand layer in this cross section is hydraulically well-connected to the stream; however, where probing data (as described in Chapter 5) were obtained, it appears that the sand layer does not extend past the eastern wetland boundary, a fact that limits significantly its ability to provide an efficient flow path between the river and the aquifer. The model was used to test this assumption. A simulation in which the sand
layer was connected to the aquifer material at the eastern boundary results in the simulated hydrographs for the October 4 rainfall as shown in Figure 23. The simulated hydrographs at cluster 17 appear to be insensitive to the change in model geometry. Simulated hydrographs for cluster 38, however, do not match the measured hydrographs. In addition, the large head adjustment that occurs in the first hour indicates that simulation of the previous pre-rainfall measurements would also be inadequate. These results support the interpretation of the sand lens as mapped.

6.5.5.3. Verification simulation

To take advantage of the larger number of head measurements made earlier in the monitoring period, a rainfall event recorded on September 22 was chosen to test the model. Because the pump was operating at P-3 during this time, approximate drawdown corrections were made to the measured heads for purposes of comparison to the simulated heads. The measured and simulated heads show good qualitative agreement (Figure 24). The hydrographs for cluster 17 show relatively rapid response and large peaks, whereas the response at cluster 38 is more damped. A gradient reversal occurs between P-17m and P-17d, as the head in the sand layer responds most quickly to the rising river stage. An earlier response is evident in the shallow peat (P-17s) due to rainfall infiltration. The record of head at P-SB is of poor quality, and only plotted until hour 10, when excessive noise interferes. During this time period, however, the model output matches the small head loss between the riverbed and P-SB.

A small inflection point can be seen in the measured hydrograph for P-17s, which peaks above the lower piezometers. The inflection point occurs when the river stage is approximately 12.92 m (at approximately hour 4 in the simulation), and appears to represent the effects of surface infiltration as the wetland surface becomes flooded. This behavior is mirrored by the simulated hydrograph; further inspection of the model results verifies that wetting of the surface nodes causes increased recharge to the upper peat. Because the model wetland surface is located at 12.95 m, the inflection point occurs somewhat later. The rapid decrease in measured heads is not well-matched, probably
Figure 6-23. Simulated heads for October 4 rainfall: sand layer connected to aquifer.
due to the failure of the model surface water cells to drain rapidly after flooding. As described in section 6.4.4.3, the simulation of surface water flow across the wetland surface is approximate. Because the surface water cells are removed from the model domain when they become dry, numerical instabilities and inaccuracies can result when attempting to simulate small changes in head.

The lack of any modeled hysteresis in peat storage may also contribute to this discrepancy, which would be aggravated by the flooding of the wetland surface. Whereas surface water would be expected to infiltrate peat macropores quickly, re-distribution of pore water to smaller pores due to capillary forces could allow negative pore pressures to develop as the peat drains.

6.5.6. **Modeling experiments**

Several questions are raised in Chapter 5 regarding the distribution of arsenic in wetland soils: (1) How was arsenic delivered to and immobilized in the deeper, pre-industrial peat layers?, and (2) How have high concentrations of arsenic been preserved in the shallow peat for more than 50 years? Chapter 5 describes the chemical mechanisms that are likely to control arsenic mobility, and concludes that significant and sustained groundwater flux, combined with maintenance of chemical conditions within a “window of mobility” for arsenic, is necessary to transport significant quantities of arsenic or deplete the reservoir of sedimentary arsenic. An additional related question is raised regarding the potential for arsenic transport through and out of the wetland. The data presented in the current chapter have provided direct evidence of arsenic mobility in this wetland under a specific set of natural and imposed conditions.
Figure 6-24. Measured and simulated hydrographs: September 19 rainfall. Lines are simulated heads; symbols are measured heads. Dashed lines are deep (d) piezometers.
The numerical model described in previous sections helps to verify hydraulic parameter measurements, infer additional parameter values, and visualize the flow field during portions of the groundwater monitoring experiment. The model is also useful to explore the potential consequences of longer term natural and imposed hydraulic forcing that cannot easily be observed directly. The results of several simulations aimed at explaining the observed arsenic distribution and the potential for transport are discussed in the following section.

Transient hydraulic conditions are common in this system due to the varying stream stage typical of high run-off urban watersheds and daytime evapotranspiration flux. Average advective transport is approximated in the following examples as the result of several steady state flow regimes. For each simulation, constant head boundaries and either rainfall or ET flux values were chosen based on sets of conditions observed during the five month monitoring period. The following flow regimes are considered: 1) “average” groundwater discharge under flooded and non-flooded conditions; 2) summer low flow; and 3) municipal pumping under high and low water conditions.

Where specific flow paths and travel times are referenced, note that they apply within the wetland only. As explained in sections 6.4.4.3 and 6.5.5.2, the model domain includes several transition zones which are constructed to apply representative hydraulic forcing on the area of interest. Simulation of flow in these areas is not quantitative.

6.5.6.1. Average discharge

During much of the year, the underlying aquifer discharges to the wetland, and the wetland discharges to the stream (Figure 25). Under these conditions flow patterns can be simulated in this cross section model as quasi-steady state. The specified-heads chosen for this simulation are an intermediate stream stage (between low water and bank-full) and the corresponding aquifer head observed during the monitoring period. Rainfall
recharge is applied at a rate of 75 cm/yr. Equipotentials and path lines for aerial and aquifer recharge resulting from these conditions are depicted in Figure 25.

These results show that the flux through the upper peat is due to aerial recharge only, which is conveyed, along with aquifer recharge, along the sand layer to the streambed. Thus the flux of arsenic within the contaminated upper peat depends on the redox conditions during infiltration of rainwater. Advective transport of any mobilized arsenic is limited in downward extent by the horizontal flow in the sand layer. The travel times through the upper peat are on the order of 50 days, as are the travel times along the sand layer.

The total recharge to the stream under these conditions is only 0.003 m$^3$/hour/m of river reach, indicating that the modeled cross section provides a relatively unimportant flow path for stream recharge compared to average discharge to the river. The only available measurements for comparison were made in October and November of 1995 by the USGS in preparation for their pump test. The reach of the Aberjona River within the Wells G & H wetland (~ 1000 m) gained an average of 0.15 m$^3$/hour/m (total = 1.5 ft$^3$/sec) during this time (Myette et al, 1987).

According to the monitoring data (see Figures 7 - 10 and Appendix C), the wetland is flooded for several days after each rainfall of more than approximately 2 cm, and for longer periods when rainfall is substantial and frequent. Under flooded conditions, the flow patterns in the shallow wetland are much different (Figure 26). Direct rainfall recharge is conveyed to the stream as runoff, and groundwater recharge flows upward through the peat, discharging at the wetland surface.
Figure 6-25. Simulated equipotentials and pathlines. Average discharge to wetland. Light stippling indicates dry model cells; bold line below dark stippling indicates position of water table. Model coordinates are in meters.
Potential arsenic transport in the upper peat during groundwater discharge conditions depends on both the chemistry of the inflowing water and the flow rate and direction through the peat. Groundwater inflow from the underlying aquifer is expected to be chemically reduced due to its isolation from the atmosphere and the availability of organic material in the diatomaceous silt and the deep peat, whereas rainfall recharge is saturated with oxygen. During unflooded conditions, the flow of rainfall recharge downward through the upper peat could, in theory, mobilize arsenic by oxidizing arsenic sulfides. To the extent that redox conditions are maintained in the “mobile window” described in Chapter 5, arsenic would be transported towards the stream, primarily within the sand layer. The potential of rainfall to alter redox conditions in the upper peat is uncertain. Oxygen may be rapidly depleted even at high recharge rates. No appreciable dissolved oxygen was measured in any of the wetland porewater samples, even during rainfall events (Table 3).

Flooding of the wetland is expected to occur at shorter intervals than the travel time for transport through the wetland under unflooded conditions (50 to 100s of days). During flooded conditions, relatively reduced groundwater flows upward through the upper peat, reversing the advective transport due to surface recharge. Because arsenic is immobilized at both high and low reduction potentials, a redox balance must be maintained within this zone during sustained period of net advective transport for the leaching of arsenic to occur. The dual constraints of flow reversals and changing redox conditions offer one explanation of why a large reservoir of arsenic remains in the upper peat.

The average rate of steady flow through the upper peat is an additional constraint on arsenic leaching. Based on the non-flooding model simulation, the average areal flux through this layer is only slightly greater than rainfall, about 0.86 m/yr. If arsenic were soluble at 700 µg/l, the flux of arsenic through this layer would be approximately 600 mg/yr/m². This aqueous concentration is slightly higher that the maximum measured
Figure 6-26. Simulated equipotentials and pathlines. Discharge to wetland during flooded conditions. Light stippling indicates dry model cells; bold line below dark stippling indicates position of water table. Model coordinates are in meters.

MIT Parson’s Lab. – Cambridge, MA
Project: Wells G & H Wetland Model
Description: Flood stage discharge
rain = 0.75 m/y, dt = 50d
15 Aug 96

Visual MODFLOW v1.50, (c) 1995
Waterloo Hydrogeologic Software
NC: 52 NR: 3 NL: 20
Current Row: 2
during the monitoring period, and also corresponds with the maximum measured by
Moore et al. (600 µg/l; 1988). Based on an average soil arsenic concentration of 1000
mg/kg throughout the top 0.5 m of peat, and a soil bulk density of 300 kg/m³, the arsenic
reservoir is 150,000 mg/m². Thus, even employing unrealistically favorable groundwater
flux conditions and high arsenic mobility, over 250 years would be required to deplete the
arsenic reservoir that remains.

It is clear that under “average” conditions, flow from neither the streambed nor the upper
peat (the two potential sources of arsenic) is expected to advect arsenic to the deeper
peat. However, such transport might be expected under the conditions caused by
municipal pumping, a scenario that is explored in section 6.5.6.3.

6.5.6.2. Summer flow

During summer months, a wetland flow pattern develops in which evapotranspirative flux
is replenished partly by local infiltration of stream water. This flow system can exist
only in the absence of significant rainfall recharge, and probably occurs for a total of four
to six weeks in an average summer. The steady state equipotentials resulting from ET of
6 mm/day, no rainfall recharge, and average stream stage and aquifer head measured in
early September, 1995, are depicted in Figure 27. The heads measured at P-17 and P-38
are approximately matched in this steady state simulation. The pathlines plotted in this
simulation show that river water discharged into the wetland under these conditions
travels as far as 10 meters from the river, and is carried predominantly by the sand layer.
The longest pathlines from the river are about 100 days long. The top two peat layers
(about 50 cm) are unsaturated or partly unsaturated during these conditions. ET flux is
withdrawn from the third model layer (20 to 50 cm below the surface). Flow above this
layer is through plant tissues or as unsaturated flow that evaporates near the wetland
surface.
Figure 6-27. Simulated equipotentials and pathlines. Summer conditions, including 6 mm/day ET. Light stippling indicates dry model cells; bold line below dark stippling indicates position of water table. Model coordinates are in meters.
Arsenic transport was observed under these conditions during the monitoring period. However, due to the limited period over which these conditions persist, arsenic from the streambed can only travel the length of the shortest pathlines (30 to 60 days). Within the upper peat itself, aqueous arsenic can be transported towards the surface, but would eventually be immobilized after oxidation in the unsaturated zone due to the presence of effective sorbents such as iron. Such a process may have occurred since the last hypothesized episode of arsenic deposition in the 1960s, concentrating sedimentary arsenic near the wetland surface. Leaching of arsenic from the upper peat would not be expected during the summer flow regime. In the long term, burial of arsenic-contaminated surface sediments could lead to the mobilization of arsenic as stable oxide forms are reduced. Such mobilized arsenic would of course be subject to the same limitations on further transport as described in this and previous sections.

6.5.6.3. Municipal pumping

To simulate flow within the wetland during the pumping of Well H, drawdowns beneath the wetland were estimated from the results of the USGS pump test (Myette et al., 1987), and used to apply a specified-head boundary at the edge of the wetland. Flow to the well itself can not be simulated in a cross section, but flow within the wetland, along a radial flowline to the well, can be approximated.

Simulated flow patterns for bank-full stream flow are shown in Figure 28. Total flow through the wetland cross-section is 0.011 m³/hr/m. For comparison, total measured stream losses during the USGS pump test were 148 m³/hr (1.45 cfs) at the end of the test. Based on the relative pumping rates at Well G & H and the measured groundwater heads, 54 m³/hr of this can be assigned to Well H, induced from a 300 m reach of the river. Thus the flow through the modeled cross section is substantially less than the average flow of 0.18 m³/hour/m. Under these assumptions, most of the stream discharge flows
through the sand layer, with travel times through the wetland of sixty days or more. The sand layer remains saturated, but much of the peat above it does not.

Figure 29 shows simulated flow during summer conditions; stream stage is maintained at the lowest head observed during the monitoring period. Total flow through the wetland is about 0.0095 m$^3$/hr/m. Less of the upper peat is saturated during summer conditions, and flow is still concentrated in the underlying sand layer.

An additional pumping simulation is shown in Figure 30. In this simulation, the sand layer was connected directly to the aquifer specified-head nodes to estimate the potential importance of such a layer. A bank-full stream level was used. Although travel times from the river to the aquifer are reduced to as little as 20 days, total flow is still small: 0.016 m$^3$/hr/m. Flow through the upper peat is still limited by its unsaturated condition.

In summary, simulated flow through the modeled cross section subjected to municipal pumping drawdowns does not explain the measured loss of streamflow during pumping or suggest a substantial flux of water or solutes to the well was likely to have occurred in this area. Greatly increased river leakance must occur at other locations, where a better hydraulic connection exists between the river and the aquifer.

Pumping-induced desaturation of the upper peat also precludes the leaching of arsenic from this layer under high flux conditions. The highly-contaminated streambed sediments, which remain saturated even during dry periods, are thus the most important potential source of arsenic under pumping conditions. During occasional flood conditions, infiltration of flood water near the margins of the wetland could also contribute to municipal well recharge along short flow paths, depending on the nature of the fill material near the well. Such conditions could also temporarily increase the importance of the arsenic source in the shallow peat in a small area close to the well.
Figure 6-28. Simulated equipotentials and pathlines. Municipal pumping and bank-full streamflow. Light stippling indicates dry model cells; bold line below dark stippling indicates position of water table. Model coordinates are in meters.
Figure 6-29. Simulated equipotentials and pathlines. Municipal pumping and low (summer) streamflow. Light stippling indicates dry model cells; bold line below dark stippling indicates position of water table. Model coordinates are in meters.
Figure 6.30. Simulated equipotentials and pathlines. Municipal pumping and bank-full streamflow: sand layer connected to aquifer. Light stippling indicates dry model cells; bold line below dark stippling indicates position of water table. Model coordinates are in meters.
Figure 6-31. Simulated equipotentials and pathlines. Municipal pumping and bank-full streamflow, no sand layer. Light stippling indicates dry model cells; bold line below dark stippling indicates position of water table. Model coordinates are in meters.
Modeling results do suggest that the hydraulic gradients induced by pumping may be the only identifiable mechanism for transport of arsenic to the lower peat. Wetland flow paths illustrated in Figures 28 - 30 indicate potential transport from the upper contaminated peat and the contaminated streambed to the lower peat. For flow through a cross-section without a sand layer (such as in the areas where arsenic is found in the lower peat-- see Figures 5-6 through 5-10), we would expect an increased number of vertical flow paths. A simulation that reflects these conditions is illustrated in Figure 31.

6.6. Conclusions

Groundwater elevations measured in wetland piezometers indicate that in periods of low rainfall the Aberjona River recharges the wetland locally, replacing water evapotranspired at 3 - 6 mm/day. The measured period of river discharge corresponded with relatively high concentrations of arsenic in streambed porewater (600 \( \mu \text{g/l} \)) and groundwater beneath the stream (120 \( \mu \text{g/l} \)). With the onset of frequent rainfall recharge, local flow directions reversed, with wetland groundwater discharging to the stream. Arsenic concentrations decreased rapidly at all measured locations during the first week of this flow reversal, returning to an apparent "steady state" concentration of approximately 30 \( \mu \text{g/l} \), although porewater from the streambed sediments was still above 100 \( \mu \text{g/l} \) when last measured in mid-December.

These observations are direct evidence that arsenic is mobile in the Wells G & H Wetland groundwater under certain conditions. The local source of arsenic appears to be the streambed sediments, and mobilization was observed under conditions of stream discharge to the wetland. Such conditions are expected to occur during periods of low rainfall and high evapotranspiration, and for short periods of time after rainfall events. A sustained period of induced infiltration also occurred when the municipal Wells G & H were pumped for 30 days in 1985, approximating the conditions that existed for most of
the period between 1964 and 1979, when one or both of these wells were in operation (Myette et al., 1987; Murphy, 1986).

A depth-averaged two-dimensional analytical model of flow to a well near a stream can explain qualitatively the effects of travel time dispersion, impermeable boundaries, and partial penetration of the streambed. This simple model over-predicts arsenic concentration at the test well, which is likely the combined result of horizontal heterogeneities, source variability, chemical interactions, and vertical flow.

Calibration of a numerical cross-section model of the studied portion of the wetland supports parameter estimates made from laboratory and field tests. The results of flow simulations using this model indicate that this portion of the wetland carries a small share of the total leakage between the river and aquifer. Although the sand layer represents a preferred flow path within the defined cross section, and is of great consequence to the wetland itself, it is not an important flow path between the river and the aquifer. Substantially better flow paths must exist elsewhere in the wetland to explain measured baseflow increases through the reach as well as the degree of induced infiltration observed during the USGS pump test. It is likely that such flow paths consist of large areas where the riverbed is relatively well-connected to aquifer sands, as that the peat itself is not sufficiently permeable to account for this quantity of leakage.

During typical groundwater discharging conditions, infiltration of rainfall recharge causes downward vertical flow in the upper peat, which is discharged along the underlying sand layer to the river. Frequent flooding conditions reverse this flow, causing the discharge of reducing groundwater at the wetland surface, and decreasing the net flow (and hence transport) through the upper peat. Even under favorable chemical conditions and consistent downward flow, only a small advective flux of arsenic is expected in the upper peat, insufficient to deplete the existing sedimentary reservoir of arsenic over the time scale of its deposition there. Therefore it is likely that the
originally-deposited arsenic concentrations in the upper peat were not substantially larger than they are today, although the chemical form may have changed many times.

The sand layer and others like it may be an important factor in the mobilization of arsenic within wetland soils. Under pumping conditions especially, arsenic may have been mobile in this sand layer due to localized differences in chemical conditions caused by increased groundwater flux and contrasting soil type. The relatively short travel times within this layer could also impose kinetic limitations on the chemical transformations that immobilize arsenic. Such arsenic may have then gradually precipitated or sorbed as flow lines diverged at the boundaries of the layer and local chemical conditions changed. Similarly, the sand layer provides a preferential pathway for the flushing of arsenic during typical groundwater discharge conditions, which may in part account for the lack of arsenic currently stored in the sand layer, compared to similar depths in the peat.

Naturally-occurring flow patterns can not account for the delivery of arsenic to the deeper (> 1 m) wetland soil layers by groundwater flow. However, groundwater flow induced by municipal pumping could have advected arsenic into the deeper peat, which, once immobilized there, would be poorly flushed under the lower gradients typical of average discharge conditions. Limited mobilization of arsenic from this zone may contribute to measured “background” porewater concentration of arsenic.

Municipal pumping is not likely to influence the mobility of arsenic in the upper peat, because much of the upper peat is de-watered by pumping. However, the streambed sediments still remain a potentially important source of arsenic under pumping gradients. Along other adjacent reaches of the river where streambed-aquifer flux is inferred to be much larger, significant arsenic transport may have occurred towards the wells. The release of arsenic from the streambed under such conditions parallel the process observed during pumping of the monitoring well in this study; the continued mobility of arsenic along flow paths to the well would depend on the local redox conditions and the relative time scales of groundwater flow and arsenic transformation.
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7. Summary and Conclusions

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Collection of hydrogeologic data in wetlands is hindered by limited access, difficulties with soil and groundwater sampling, and by the fact that the scales of variability are small whereas the magnitude of variability can be large. Although the important hydrologic role of wetlands is widely acknowledged, the nature of flow within them is typically not investigated, in part due to the difficulties of access and sampling, but also because the assumption is made that not much will be learned by doing so. This study demonstrates the utility of a new piezocone penetrometer for wetland exploration, and shows that in fact much can be learned from the internal structure of wetland deposits. In particular, insights are gained regarding the genesis of such deposits as well as their role in sequestering, mobilizing, and transporting pollutants.

7.1. Wetland Exploration with the Piezocone Penetrometer

The mapping of wetland stratigraphy can be successfully accomplished with a piezocone penetrometer. Standard measurements, if made with increased sensitivity and resolution, are sufficient to discriminate sand from soft sediments, peat from other soft sediments, and different peat layers from each other. Friction sleeve measurements are integral to the identification of wetland soil layers, and any mechanical innovations that enhance their reliability would further improve the ability to identify various peat strata.

The highly layered nature of wetland deposits and the sometimes subtle differences in soil properties among these layers prevents the interpretation of piezocone data based on correlations between soil type and 'steady state' values of measurements. In most cases, steady state penetration is not achieved. A 'fingerprint' method has been successfully employed to interpret the data gathered in this study, allowing the construction of several detailed stratigraphic cross-sections through the study site. Automation of this method
could probably be achieved based on pattern recognition and change-detection signal processing techniques.

Although the response of the various piezocone measurements that define these signatures can be interpreted within a rational soil mechanics framework, the effects of small interbeds, layer ordering, and complex peat mechanics may prevent indefinitely the development of a model that quantitatively predicts soil penetration in these types of soils.

For purposes of portability and efficiency of operation, a miniature piezocone should be considered for wetland use, although the relationship of measurement scale to 'grain' scale needs to be considered carefully in interpretation of the data.

As defined by our piezocone and verified in soil cores, significant variability exists on the 10 cm scale in the vertical, and on the one meter scale in the horizontal dimension. Sand layers thinner than 10 cm are identified, and thin (probably less than 10 cm) root mats within the peat are also consistently distinguished in the piezocone record. Thicker (1-1.5 m) sand layers are observed to pinch out over distances of several meters.

7.2. The Post-Glacial and Industrial Evolution of the Wells G & H Wetland

The stratigraphy of the Wells G & H Wetland is interpreted within a post-glacial depositional model by integrating piezocone and soil core data with both radiocarbon dates and pollen abundance information. In the area of exploration, an ice block depression was left in the plain of the paleo-Aberjona River, which probably existed as a much larger braided stream until much of the local ice had melted. The diatom assemblage in limnic sediments that fill this depression indicate a circumneutral fresh water lake, predominantly groundwater fed. A broader wetland system covered the infilled depression by about 9000 ybp, accumulating up to three meters of woody and
herbaceous peat. A sand layer located within the peat is believed to represent the depositional record of the Aberjona River as it meandered across the wetland surface from 5000 to 1000 ybp.

Post-colonial vegetational changes and the presence of extremely high soil arsenic concentrations in the shallow (<80 cm) peat are consistent with the documented industrial impacts on the Aberjona River, culminating with the transport of large quantities of arsenic from its headwaters to its outlet from the late 1800s to the 1960s. The deposition of arsenic-contaminated soils on the wetland surface is hypothesized to have contributed to a shift in vegetation from sedges to *Typha* and *Phragmites*. The widespread use of arsenic in pesticides and herbicides, as well as its more recent use in products such as rot-resistant lumber, testifies to the powerful toxic effects of arsenic.

Occurrence of soil arsenic below the colonial horizon is believed to be the record of subsurface transport and trapping in specific zones within the wetland. It is possible that the transport of arsenic to these locations occurred under the influence of large groundwater withdrawals at Wells G & H. Gawel (1996) has observed indicators of metal stress in trees near the wetland border that also suggest that metal-contaminated groundwater may have been transported through this area, presumably from the Aberjona River.

Although the chemical form of the sedimentary arsenic in this wetland was not determined, equilibrium calculations indicate it could be immobilized in several different sulfide forms, including AsS, As$_2$S$_3$, and FeAsS. Sorption to humic material is also possible. In the highly reducing wetland environment the mobility of arsenic is, at equilibrium, probably controlled by the solubility of these minerals, with significant transport of arsenic only occurring within a narrow window of redox conditions between oxic and sulfidic conditions.
7.3. Groundwater Flow Within the Wells G & H Wetland: Implications for Past and Future Transport of Arsenic

7.3.1. Mobile arsenic

The chemical and hydraulic measurements presented in Chapter 6 are evidence that arsenic can become mobile (600 µg/l measured in porewater) in the Wells G & H Wetland. Increased mobility corresponded with a period of stream discharge, during which oxic stream water flows through the contaminated streambed sediments into the wetland soils. It is possible that such flow causes a change in subsurface chemical conditions that contributes to the release of arsenic. The mobility of arsenic under other conditions and in other locations within the wetland is difficult to predict quantitatively due to the large dependency of arsenic solubility on redox conditions and the effect of groundwater flow on redox potential. Reduction potential in wetland soils is likely to at least intermittently overlap the “mobility window” for arsenic defined in Chapter 5.

7.3.2. Sand layer

Within the investigated area of the site, a sand layer existing at about one meter below the surface provides a preferred flow path for river infiltration, acting as a leaky “aquifer” with a good connection to the streambed. During river discharge conditions, the sand layer enhances the transport of arsenic away from the stream, recharging the upper peat when evapotranspiration flux is large. The influence of vertical flow and horizontal heterogeneities on the advection of arsenic between the stream and the sand layer, as well as probable source variability, contribute to the inability of a depth-averaged two dimensional model to predict arsenic concentration at a well pumped in the layer. During conditions of average rainfall and groundwater discharge to the wetland, vertical flow from above and below the sand converge within the layer and are carried horizontally towards the stream.
The sand layer modeled in this study is not a particularly important flow path between the river and the aquifer; however, it has a strong influence on local internal wetland water and solute fluxes.

The effect of this sand layer, and others like it, have interesting parallels to construction techniques that are used to control groundwater levels and flow rates. This work highlights these parallels, and suggests that similar techniques might be used to create certain desirable chemical and hydrologic regimes in constructed wetlands used for stormwater control or water treatment. Redox conditions, root zone fluxes, and response times are among many characteristics that could be “tuned” by controlling wetland subsurface as well as surface features.

7.3.3. Mobility of arsenic in upper peat

The large reservoir of arsenic in the highly contaminated upper peat is expected to have limited mobility. This conclusion helps to explain the persistence of this arsenic, which was deposited beginning at the turn of this century. Its mobility after burial is limited for several reasons. First, for arsenic to remain soluble, redox conditions must remain in a mobile window between oxic and sulfidic zones. Substantial advection of arsenic upward or downward within the peat is likely to carry arsenic into zones of increased or decreased redox potential, precipitating soluble arsenic and limiting mobility.

Second, during typical groundwater discharge conditions, infiltrating rainfall recharge penetrates the wetland only to the depth of the sand layer, which conveys this flow to the stream. Rainfall does not infiltrate during frequently flooded conditions, in which case discharging groundwater flows upward through the peat to the flooded surface. Such flow reversals limit the net advective transport through the contaminated upper peat. Third, even if the estimated rainfall recharge rate through the upper peat is applied as an average, and redox conditions are assumed that allow 700 µg/l mobile arsenic, the
resulting arsenic flux is insufficient to deplete the present stock of arsenic over a time scale of less than 100 years.

Finally, under the influence of municipal pumping, when groundwater flow rates through the wetland are locally increased, the upper peat becomes de-saturated (even during bank-full streamflow), and thus is poorly flushed under these conditions as well. This last observation indicates that arsenic contained in the streambed sediments, which remain saturated during pumping, is a more important potential source of arsenic with respect to the pumping wells.

7.3.4. Arsenic in deeper peat

Naturally occurring flow paths are not expected to advect arsenic from the streambed or the upper peat towards the lower peat. It is therefore likely that the arsenic present at certain locations in the deep peat (120 - 160 cm) was advected there under the influence of municipal pumping. The lack of arsenic at similar depths in the sand layer is due either to contrasting local chemical conditions during municipal pumping, or enhanced flushing with 'clean' water subsequent to the period of pumping.

Approximately 1 kg of arsenic per meter of river reach is estimated to be stored in the lower peat. Based on groundwater discharge through the lower peat of 4.5 m$^3$/yr per meter of river reach, local groundwater arsenic concentrations of greater than 4500 μg/l would have been required to deliver this quantity of arsenic from outside the wetland. As the highest current concentrations in the aquifer are on the order of 100 μg/l, this requirement further supports the assertion that efficient flow paths between the river and the lower peat are responsible for the delivery of arsenic to these horizons from a streambed source (Geotrans and Retec, 1994).
7.3.5. Groundwater flow and arsenic transport during municipal pumping

Results of a numerical two-dimensional cross-section model indicate that the studied wetland section, located at the river’s closest approach to Well H, carries a disproportionately small quantity of flow between the adjacent aquifer and the stream. Contrary to previous conclusions, as well as assumptions made as part of several watershed-scale models, the peat soils in this wetland are not “loose and permeable”, although they are in places interbedded with sand. Subsurface features outside of the study area must provide better connections between the stream and the underlying aquifer in order to explain the quantity of induced infiltration that was observed during the USGS pump test, as well as typical (non-pumping) baseflow pick-up. It is likely that such features consist of places where peat is very thin or absent below the streambed, as the wetland soils are not permeable enough to convey the volume of leakage observed.

Clearly, arsenic transport through the studied section during municipal pumping is somewhat limited by the small advective flux. In addition, the contribution of the arsenic contained in the upper peat is limited due to the de-wetting of this layer. The streambed represents the most likely source of arsenic to the wetland and hence the adjacent aquifer. Within the zones of high leakance, arsenic may be mobilized and advected towards the wells if the necessary balance of redox conditions is maintained.

References


Appendix A

Piezocone Data Records

The piezocone records contained in this appendix include several corrections. Each measurement has been corrected to the tip location. The sleeve friction and point resistance measurements have been pore pressure corrected based on measured pore pressures at the tip and along the shaft, if applicable.

The records from 1993 (93- ) were obtained with the piezocone described in Chapter 2. The records from 1994 (94- ) were obtained with the piezocone described in Chapter 3.

Missing records in the numerical sequence were not included due to poor data quality.
Drive 94-4
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)  Tip Load (MPa)

Depth (cm.)
0.00  0.03  0.06  0.09

Breaks
◊ tip
▼ lower Shaft

0.00  0.03  0.06  0.09

Sleeve Friction (MPa)

0.000  0.015  0.030
Drive 94-13
(All Depths Corrected to Tip Position)

Pore Pressure (MPa) vs. Depth (cm.)

- Shaft
- Tip
- Breaks
- Tip
- Lower Shaft

Tip Load (MPa)

- Sleeve
- Tip
- Breaks
- Tip
- Sleeve

Pore Pressure (MPa) vs. Depth (cm.)

Sleeve Friction (MPa)
Drive 94-9
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)  Tip Load (MPa)

Depth (cm.)

Pore Pressure (MPa)  Sleeve Friction (MPa)
Drive 94-12
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)

Tip Load (MPa)

Depth (cm.)

Pore Pressure (MPa)

Sleeve Friction (MPa)

Breaks

Upper Shaft

Shaft

Tip

Diamond

Square

tip

lower Shaft

upper Shaft

Sleeve

Tip

Breaks

Diamond

Square

tip

sleeve
Drive 94-14

(All Depths Corrected to Tip Position)

Pore Pressure (MPa)

Tip Load (MPa)

Depth (cm.)

0.00 0.03 0.06 0.09

0.0 1.5 3.0

Upper Shaft
Shaft
Tip

Breaks

| tip |
| lower Shaft |
| upper Shaft |

Sleeve
Tip

Breaks

| tip |
| sleeve |

Sleeve Friction (MPa)
Drive 94-16
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)

Tip Load (MPa)

Depth (cm.)

0.00 0.03 0.06 0.09

0.0 1.5 3.0

Upper Shaft
Shaft
Tip

Breaks

tip
lower Shaft
upper Shaft

Sleeve
Tip

Breaks

tip
sleeve

Pore Pressure (MPa)
Sleeve Friction (MPa)
Drive 94-17
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)

Tip Load (MPa)

Depth (cm.)

Pore Pressure (MPa)

Sleeve Friction (MPa)
Drive 94-18
(All Depths Corrected to Tip Position)

- Pore Pressure (MPa)
- Tip Load (MPa)

Depth (cm.):
- Upper Shaft
- Shaft
- Tip

Breaks:
- Tip
- Lower Shaft
- Upper Shaft

Pore Pressure (MPa): [Graph]
Tip Load (MPa): [Graph]

Sleeve Friction (MPa): [Graph]
Drive 94-22

(All Depths Corrected to Tip Position)

Pore Pressure (MPa)

Tip Load (MPa)

Depth (cm.)

Breaks
- tip
- lower Shaft
- upper Shaft

Upper Shaft
- Shaft
- Tip

Sleeve
- Tip

Breaks
- tip
- sleeve

Pore Pressure (MPa)

Sleeve Friction (MPa)
Drive 94-23
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)

0.00 0.03 0.06 0.09

Tip Load (MPa)

0.0 1.5 3.0

Depth (cm.)

0

Upper Shaft
Shaft
Tip

Breaks

○ tip
△ lower Shaft
▽ upper Shaft

Pore Pressure (MPa)

0.00 0.03 0.06 0.09

Sleeve Friction (MPa)

0.000 0.015 0.030

Sleeve
Tip

Breaks

○ tip
□ sleeve
Drive 94-24
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)

Tip Load (MPa)

Depth (cm.)

Pore Pressure (MPa)

Sleeve Friction (MPa)
Drive 94-25
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)

Depth (cm.)

0.00 0.03 0.06 0.09
0 50 100 150 200 250 300 350 400 450 500 550 600 650 700

Upper Shaft
Shaft
Tip

Breaks
tip
lower Shaft
upper Shaft

Tip Load (MPa)

0.0 1.5 3.0

Sleeve
Tip

Breaks
tip
sleeve

Sleeve Friction (MPa)
Drive 94-26
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)

0.00 0.03 0.06 0.09

0 50 100 150 200 250 300 350 400 450 500 550 600

Depth (cm.)

Upper Shaft
Shaft
Tip

Breaks:

• tip
• lower Shaft
• upper Shaft

Tip Load (MPa)

0.0 1.5 3.0

0.000 0.015 0.030

Sleeve Friction (MPa)

Sleeve
Tip

Breaks:

• tip
• sleeve
Drive 94-27

(All Depths Corrected to Tip Position)

Pore Pressure (MPa)

Tip Load (MPa)

Depth (cm.)

Upper Shaft
Shaft
Tip

Breaks:
- tip
- lower Shaft
- upper Shaft

Sleeve
Tip

Breaks:
- tip
- sleeve

Sleeve Friction (MPa)
Drive 94-28

(All Depths Corrected to Tip Position)

Pore Pressure (MPa) vs Depth (cm.)
- Upper Shaft
- Shaft
- Tip

Breaks:
- ▲ tip
- ▼ lower Shaft
- △ upper Shaft

Tip Load (MPa) vs Depth (cm.)
- Sleeve
- Tip

Breaks:
- ◇ tip
- □ sleeve

Sleeve Friction (MPa) vs Depth (cm.)
Drive 94-29

(All Depths Corrected to Tip Position)

Pore Pressure (MPa)

Tip Load (MPa)

Depth (cm.)

Upper Shaft
Shaft
Tip

Breaks

tip
lower Shaft
upper Shaft

Sleeve
Tip

Breaks

tip
sleeve

Sleeve Friction (MPa)
Drive 94-30
(All Depths Corrected to Tip Position)

Pore Pressure (MPa) vs. Depth (cm.)
- Upper Shaft
- Shaft
- Tip

Tip Load (MPa) vs. Depth (cm.)
- Sleeve
- Tip

Breaks
- tip
- lower Shaft
- upper Shaft

Pore Pressure (MPa) vs. Depth (cm.)
- Sleeve Friction (MPa)
Drive 94-31
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)

Tip Load (MPa)

Depth (cm.)

Pore Pressure (MPa)

Sleeve Friction (MPa)
Drive 94-33
(All Depths Corrected to Tip Position)

Pore Pressure (MPa) and Tip Load (MPa) graphs are shown, with depth in cm. on the y-axis and pore pressure (MPa) and sleeve friction (MPa) on the x-axis. The graphs display different lines and symbols indicating upper shaft, lower shaft, sleeve, and tip breaks.
Drive 94-34 (Sleeve Corrected)
(All Depths Corrected to Tip Position)

Pore Pressure (MPa) vs. Depth (cm.)
- Upper Shaft
- Shaft
- Tip

Breaks:
- Tip
- Lower Shaft
- Upper Shaft

Tip Load (MPa)

Sleeve Friction (MPa)
Drive 94-34 (Sleeve Corrected)
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)  Tip Load (MPa)

Depth (cm.)

Upper Shaft
Shaft
Tip

Breaks

- tip
- lower Shaft
- upper Shaft

Sleeve
Tip

Breaks

- tip
- sleeve

Pore Pressure (MPa)  Sleeve Friction (MPa)
Drive 94-35
(All Depths Corrected to Tip Position)

Pore Pressure (MPa) vs. Depth (cm.)

- Upper Shaft
- Shaft
- Tip

Breaks:
- **tip**
- **lower Shaft**
- **upper Shaft**

Tip Load (MPa)

Sleeve Friction (MPa)
Drive 94-36
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)

Tip Load (MPa)

Depth (cm.)

Upper Shaft
Shaft
Tip

Breaks

tip
lower Shaft
upper Shaft

Sleeve
Tip

Breaks

tip
sleeve

Sleeve Friction (MPa)
Drive 94-37
(All Depths Corrected to Tip Position)

Pore Pressure (MPa) vs. Depth (cm.)
- Upper Shaft
- Shaft
- Tip

Breaks:
- tip
- lower Shaft
- upper Shaft

Tip Load (MPa) vs. Depth (cm.)
- Sleeve
- Tip

Breaks:
- tip
- sleeve

Pore Pressure (MPa) vs. Sleeve Friction (MPa)

332
Drive 94-38
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)

Tip Load (MPa)

Depth (cm.)

Upper Shaft
Shaft
Tip

Breaks

tip
lower Shaft
upper Shaft

Sleeve
Tip

Breaks

tip
sleeve

Pore Pressure (MPa)
Sleeve Friction (MPa)
Drive 94-39
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)

Tip Load (MPa)

Depth (cm.)

Breaks

- Upper Shaft
- Shaft
- Tip

- Sleeve
- Tip

Breaks

- tip
- lower Shaft
- upper Shaft

- sleeve

Pore Pressure (MPa)

Sleeve Friction (MPa)
Drive 94-40
(All Depths Corrected to Tip Position)

Pore Pressure (MPa) vs. Depth (cm.)

Tip Load (MPa) vs. Depth (cm.)

Sleeve Friction (MPa) vs. Depth (cm.)
Drive 94-41

(All Depths Corrected to Tip Position)

Pore Pressure (MPa)

Tip Load (MPa)

Depth (cm.)

Upper Shaft

Shaft

Tip

Breaks

Diamond tip

Inverted triangle lower Shaft

Triangle upper Shaft

Sleeve

Tip

Breaks

Diamond tip

Box sleeve

Pore Pressure (MPa)

Sleeve Friction (MPa)
Drive 94-42
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)

Tip Load (MPa)

Depth (cm.)

Upper Shaft
Shaft
Tip

Breaks
tip
lower Shaft
der upper Shaft

Sleeve
Tip

Breaks

Sleeve Friction (MPa)
Drive 94-43

(All Depths Corrected to Tip Position)

Pore Pressure (MPa) vs. Depth (cm.)

Tip Load (MPa) vs. Depth (cm.)

Breaks:
- tip
- lower Shaft
- upper Shaft

Sleeve Friction (MPa) vs. Depth (cm.)
Drive 94-45
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)  Tip Load (MPa)

0.00 0.03 0.06 0.09

Depth (cm.)

Upper Shaft
Shaft
Tip

Breaks
tip
lower Shaft
upper Shaft

0.0 1.5 3.0

Sleeve Friction (MPa)

0.00 0.015 0.030
Drive 94-46

(All Depths Corrected to Tip Position)

Pore Pressure (MPa) vs. Depth (cm.)
- Upper Shaft
- Shaft
- Tip

Breaks
- tip
- lower Shaft
- upper Shaft

Tip Load (MPa) vs. Depth (cm.)
- Sleeve
- Tip

Breaks
- tip
- sleeve

Pore Pressure (MPa) vs. Depth (cm.)
Sleeve Friction (MPa) vs. Depth (cm.)
Drive 94-47
(All Depths Corrected to Tip Position)

Pore Pressure (MPa) vs. Depth (cm.)

Tip Load (MPa)

Sleeve Friction (MPa)

Breaks:
- Tip
- Lower Shaft
- Upper Shaft

Upper Shaft
- Shaft
- Tip

Sleeve
- Tip

341
Drive 94-48
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)  Tip Load (MPa)

Depth (cm.)

Upper Shaft  Shaft  Tip

Breaks

tip  lower Shaft  upper Shaft

Sleeve  Tip

Breaks

tip  sleeve

Pore Pressure (MPa)  Sleeve Friction (MPa)
Drive 94-50

(All Depths Corrected to Tip Position)

Pore Pressure (MPa)

Tip Load (MPa)

Depth (cm.)

Pore Pressure (MPa)

Sleeve Friction (MPa)
Drive 94-51
(All Depths Corrected to Tip Position)

Pore Pressure (MPa) vs. Depth (cm.)
- Upper Shaft
- Shaft
- Tip

Tip Load (MPa)

Pore Pressure (MPa) vs. Tip Load (MPa)

Sleeve Friction (MPa)
- Sleeve
- Tip

Breaks:
- tip
- lower Shaft
- upper Shaft
- sleeve
Drive 94-52

(All Depths Corrected to Tip Position)

Pore Pressure (MPa)  Tip Load (MPa)

Pore Pressure (MPa)  Sleeve Friction (MPa)

Depth (cm.)

Upper Shaft
Shaft
Tip

Breaks

tip
lower Shaft
upper Shaft

Breaks

tip
sleeve
Drive 94-53
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)  Tip Load (MPa)

Depth (cm.)

Upper Shaft
Shaft
Tip

Breaks
- tip
- lower Shaft
- upper Shaft

Sleeve
Tip

Breaks
- tip
- sleeve

Pore Pressure (MPa)  Sleeve Friction (MPa)
Drive 94-54
(All Depths Corrected to Tip Position)

Pore Pressure (MPa) vs. Depth (cm.)

Tip Load (MPa)

Breaks:
- □ tip
- ▼ lower Shaft
- ▲ upper Shaft

Upper Shaft
Shaft
Tip

Sleeve
Tip

Breaks:
- □ tip
- △ sleeve

Pore Pressure (MPa) vs. Depth (cm.)

Sleeve Friction (MPa)
Drive 94-58
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)  Tip Load (MPa)

Depth (cm.)

0.00 0.03 0.06 0.09

Upper Shaft
Shaft
Tip

Breaks
tip
lower Shaft
upper Shaft

Sleeve
Tip

Breaks
tip
sleeve

Pore Pressure (MPa)  Sleeve Friction (MPa)
Drive 94-59

(All Depths Corrected to Tip Position)

Pore Pressure (MPa)  Tip Load (MPa)

Depth (cm.)

Upper Shaft
Shaft
Tip

Breaks
□ tip
△ lower Shaft
△ upper Shaft

Sleeve
Tip

Breaks
□ tip
□ sleeve

Pore Pressure (MPa)  Sleeve Friction (MPa)

0.00  0.03  0.06  0.09

0.00  0.030

0.00  0.015  0.030

354
Drive 94-60
(All Depths Corrected to Tip Position)

Pore Pressure (MPa) vs. Depth (cm.)

Tip Load (MPa) vs. Depth (cm.)

Breaks:
- Tip
- Upper Shaft
- Lower Shaft

Sleeve Friction (MPa)
Drive 94-61
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)  Tip Load (MPa)

0.00  0.03  0.06  0.09

0.0  1.5  3.0

Depth (cm.)

Upper Shaft
Shaft
Tip

Breaks

Diamond tip
Triangle lower Shaft
Triangle upper Shaft

Sleeve
Tip

Breaks

Diamond tip
Rectangle sleeve
Drive 94-62
(All Depths Corrected to Tip Position)

Pore Pressure (MPa) vs. Depth (cm.)

Tip Load (MPa)

Sleeve Friction (MPa)

Depth (cm.)

Pore Pressure (MPa)

Breaks:
- Upper Shaft
- Shaft
- Tip

Breaks:
- Sleeve
- Tip

Breaks:
- Tip
- Sleeve
Drive 93-11
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)

Depth (cm.)

0.00 0.03 0.06 0.09

50
100
150
200
250
300
350
400
450
500
550
600
650
700

0.0 1.5 3.0

Tip Load (MPa)

Tip Breaks

Tip Breaks
Drive 93-20
(All Depths Corrected to Tip Position)

Depth (cm.)

Pore Pressure (MPa)  Tip Load (MPa)

0.00  0.03  0.06  0.09

0.0  1.5  3.0

Pore Pressure (MPa)  Tip Load (MPa)

Tip

Breaks

Tip

Breaks

360
Drive 93-21
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)

Tip Load (MPa)

Depth (cm.)
Drive 93-22
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)

Tip Load (MPa)
Drive 93-23
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)  Tip Load (MPa)

Depth (cm.)

Pore Pressure (MPa)  Tip Load (MPa)
Drive 93-24

(All Depths Corrected to Tip Position)
Drive 93-25
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)  Tip Load (MPa)

Depth (cm.)

0.00 0.03 0.06 0.09
0.0 1.5 3.0

Tip
Breaks

Tip
Breaks

Pore Pressure (MPa)
Tip Load (MPa)
Drive 93-26
(All Depths Corrected to Tip Position)

Pore Pressure (MPa) vs. Depth (cm.)

Tip Load (MPa) vs. Depth (cm.)

- Tip
- Breaks

366
Drive 93-27
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)  Tip Load (MPa)

Depth (cm.)

Pore Pressure (MPa)  Tip Load (MPa)

Tip
Breaks

Tip
Breaks
Drive 93-29
(All Depths Corrected to Tip Position)

Pore Pressure (MPa)

Tip Load (MPa)

Depth (cm.)

Pore Pressure (MPa)

Tip

Breaks

Tip

Breaks
Drive 93-30

(All Depths Corrected to Tip Position)

Pore Pressure (MPa) vs. Depth (cm.)

Tip Load (MPa) vs. Depth (cm.)

---

Tip

Breaks
Appendix B

Reactions and Equilibrium Constants Used in Creating the Predominance/Stability Diagrams

The equations and equilibrium constants listed in this appendix have been calculated based on the following sources:


1. \[ \text{AsS}_2^- + 3\text{H}_2\text{O} + \text{H}^+ \leftrightarrow \text{H}_3\text{AsO}_3 + 2\text{H}_2\text{S} \]
   \[ \log K^\circ = -6.707 \]

2. \[ \text{AsS(realgar)} + 3\text{H}_2\text{O} \leftrightarrow \text{H}_3\text{AsO}_3 + \text{H}_2\text{S} + \text{e}^- + \text{H}^+ \]
   \[ \log K^\circ = -19.01 \]

3. \[ \text{AsS(realgar)} + \text{H}_2\text{S} \leftrightarrow \text{AsS}_2^{-1} + 2\text{H}^+ + \text{e}^- \]
   \[ \log K^\circ = -12.3 \]

4. \[ \text{As}_2\text{S}_3(orpiment) + 2\text{e}^- + 2\text{H}^+ \leftrightarrow 2\text{AsS(realgar)} + \text{H}_2\text{S} \]
   \[ \log K = -0.066 \]

5. \[ \text{As}_2\text{S}_3(orpiment) + 6\text{H}_2\text{O} \leftrightarrow 2\text{H}_3\text{AsO}_3 + 3\text{H}_2\text{S} \]
   \[ \log K^\circ = -38.1 \]

6. \[ \text{As}_2\text{S}_3(orpiment) + \text{H}_2\text{S} \leftrightarrow 2\text{AsS}_2^{-1} + 2\text{H}^+ \]
   \[ \log K = -24.68 \]

7. \[ \text{FeAsS(arsenopyrite)} + 3\text{H}_2\text{O} \rightarrow \text{H}_3\text{AsO}_3 + \text{Fe}^{2+} + \text{H}_2\text{S} + 3\text{e}^- + \text{H}^+ \]
   \[ \log K^\circ = -11.24 \]

8. \[ \text{FeAsS(arsenopyrite)} + \text{H}_2\text{S} \leftrightarrow \text{As}_2\text{S}_3(orpiment) + 2\text{Fe}^{2+} + 2\text{H}^+ + 6\text{e}^- \]
   \[ \log K^\circ = 5.35 \]

9. \[ \text{FeAsS(arsenopyrite)} \leftrightarrow \text{AsS(realgar)} + \text{Fe}^{2+} + 2\text{e}^- \]
   \[ \log K^\circ = 2.644 \]

10. \[ \text{FeAsO}_4(s) + 2\text{H}_2\text{O} + \text{H}_2\text{O} \leftrightarrow \text{Fe(OH)}_3(\text{am}) + 2\text{H}^+ + \text{HAsO}_4^{-2} \]
    \[ \log K^\circ = -13.2 \]

11. \[ \text{Fe}_3(\text{AsO}_4)_2(s) + 4\text{H} \leftrightarrow 3\text{Fe}^{2+} + 2\text{H}_2\text{AsO}_4^- \]
    \[ \log K = 2.47 \]

12. \[ \text{Fe(OH)}_3(\text{am}) + 3\text{H}^+ + \text{e}^- \leftrightarrow \text{Fe}^{2+} + 3\text{H}_2\text{O} \]
    \[ \log K^\circ = 16.0 \]
\( FeS(\text{pyrrhotite}) + 2H^+ \leftrightarrow Fe^{+2} + H_2S \)

13. \( \log K^o = 2.82 \)

\( FeS_2(\text{pyrite}) + 4H^+ + 2e^- \leftrightarrow Fe^{+2} + 2H_2S \)

14. \( \log K = -2.78 \)

\( FeS(\text{pyrrhotite}) + H_2S \rightarrow FeS_2(\text{pyrite}) + 2H^+ + 2e^- \)

15. \( \log K = 4.5 \)

\( FeAsS(\text{arsenopyrite}) + H_2S + 3H_2O \leftrightarrow FeS_2(\text{pyrite}) + H_3AsO_3 + 5H^+ + 5e^- \)

16. \( \log K = -14.45 \)

\( FeAsS(\text{arsenopyrite}) + 3H_2O \rightarrow FeS(\text{pyrrohtite}) + H_3AsO_3 + 3H^+ + 3e^- \)

17. \( \log K = -18.96 \)
Appendix C

Records of Groundwater Elevation
August - December, 1995
Appendix D

Records of Field Hydraulic Conductivity Tests
and
Laboratory Compressibility Tests

Wellpoint test data for various depths at site 17 and 38 described in the text are reproduced here. Unless otherwise noted, conductivity estimates were derived from flow rate vs. time data using the method of Lohman (1972) as described in Chapter 6. Estimates are also provided based on flow rate and drawdown using Theim (1906). In some cases, head vs. time data were analyzed using Hvorslev (1951).

Head recovery data from piezometer 17m, located two meters from P-3 was analyzed using residual drawdown analysis as implemented in Levy (1994).

All laboratory testing was completed by Jason Bialon, and is fully described in Bialon (1995). Certain portions of his data are reproduced here for the purpose of re-interpretation.
Site 38
<table>
<thead>
<tr>
<th>min</th>
<th>sec</th>
<th>time (sec)</th>
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<th>Sw/Q</th>
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slope= 6

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\[ T = 0.001525 \text{ cm}^2/\text{sec} \quad \text{slope}=120 \]

0.004344 by Theim equation
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"steady state" Q = 0.054746

slope = 400

T = 0.000459 cm²/sec

0.001657 by Theim equation

![Test 3 Site 38 Graph](attachment:image.jpg)
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0.199462 by Theim equation

continued next page
Test 5
110-130 cm Site 38
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"steady state" Q= 4.005

T = 0.065367 cm²/sec slope=2.8

0.121192 by Theim equation

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steady state Q = 2.193

T = 0.036606 cm²/sec
slope = 5

0.066363 by Theim equation
Test 7  150-170 cm  Site 38
Sw=30

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"steady state" Q = 1.007268

T = 0.036606 cm^2/sec  slope=5

0.030479 by Theim equation
Test 1  70-90 cm  Site 17
Sw= 30 cm

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"steady state" Q = 0.266

T = 0.006101 cm²/sec  slope=30

0.008043 by Theim equation

![Test 1 Graph](image-url)
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"steady state" Q = 2.911496
T = 0.052294 cm²/sec, slope = 4.5

0.0881 by Theim equation

![Graph](image-url)
Test 4  140-160 cm  Site 17  
Sw=30

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"steady state" Q = 2.439769  
slope=3.6

T = 0.050841 cm^2/sec

0.073826 by Theim equation
Test 5  161-181  Site 17
Sw=60

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"steady state" Q= 0.111949

T= 0.000508 cm²/sec  slope=360

0.003388 by Theim equation
Test 6  180-200 cm.  Site 17
Sw=30

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"steady state" Q= 0.30882

T= 0.009151 cm³/sec  slope=20

0.009345 by Theim equation

![Graph of Test 6 Site 17](image)
Test 7 200-220 cm Site 17
Sw=30

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<td>63</td>
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<tr>
<td>0</td>
<td>46</td>
<td>46</td>
<td>0.4</td>
<td>75</td>
</tr>
<tr>
<td>1</td>
<td>11</td>
<td>71</td>
<td>0.4</td>
<td>75</td>
</tr>
<tr>
<td>1</td>
<td>36</td>
<td>96</td>
<td>0.4</td>
<td>75</td>
</tr>
<tr>
<td>2</td>
<td>1</td>
<td>121</td>
<td>0.4</td>
<td>75</td>
</tr>
<tr>
<td>2</td>
<td>26</td>
<td>146</td>
<td>0.4</td>
<td>75</td>
</tr>
<tr>
<td>2</td>
<td>51</td>
<td>171</td>
<td>0.4</td>
<td>75</td>
</tr>
<tr>
<td>3</td>
<td>16</td>
<td>196</td>
<td>0.4</td>
<td>75</td>
</tr>
</tbody>
</table>

"steady state" Q= 0.4

T= ? cm²/sec

0.012104 by Theim equation
<table>
<thead>
<tr>
<th>Site 17</th>
<th>Test 9</th>
<th>240 - 260 cm</th>
<th>Zeab drawdown</th>
</tr>
</thead>
<tbody>
<tr>
<td>hour min</td>
<td>time</td>
<td>dtw</td>
<td></td>
</tr>
<tr>
<td>8</td>
<td>55</td>
<td>0</td>
<td>117</td>
</tr>
<tr>
<td>9</td>
<td>4</td>
<td>9</td>
<td>113.5</td>
</tr>
<tr>
<td>9</td>
<td>20</td>
<td>25</td>
<td>109.5</td>
</tr>
<tr>
<td>9</td>
<td>29.5</td>
<td>34.5</td>
<td>108</td>
</tr>
<tr>
<td>9</td>
<td>42</td>
<td>47</td>
<td>106</td>
</tr>
</tbody>
</table>

$r = 1$
$L = 20$

$K = 8.48E-05 \text{ cm/sec}$
<table>
<thead>
<tr>
<th>Test 10</th>
<th>260 - 280 c</th>
<th>7/7/95</th>
<th>Zeeb</th>
</tr>
</thead>
<tbody>
<tr>
<td>hour</td>
<td>min</td>
<td>time</td>
<td>dtw</td>
</tr>
<tr>
<td>9</td>
<td>49.5</td>
<td>54.5</td>
<td>115</td>
</tr>
<tr>
<td>9</td>
<td>54</td>
<td>59</td>
<td>110</td>
</tr>
<tr>
<td>9</td>
<td>58</td>
<td>63</td>
<td>109</td>
</tr>
<tr>
<td>10</td>
<td>0</td>
<td>65</td>
<td>107</td>
</tr>
<tr>
<td>10</td>
<td>4</td>
<td>69</td>
<td>105.5</td>
</tr>
</tbody>
</table>

$r$ = 1
$L$ = 20
$K$ = 1.99E-04
Site 17
Test 11 280 - 300 cm 7/7/95 Zeeb drawdown

<table>
<thead>
<tr>
<th>hour</th>
<th>min</th>
<th>time</th>
<th>dtw</th>
<th>drawdown</th>
</tr>
</thead>
<tbody>
<tr>
<td>10</td>
<td>8</td>
<td>73</td>
<td>118.5</td>
<td>84.5</td>
</tr>
<tr>
<td>10</td>
<td>10</td>
<td>75</td>
<td>128.5</td>
<td>94.5</td>
</tr>
<tr>
<td>10</td>
<td>15</td>
<td>80</td>
<td>122</td>
<td>88</td>
</tr>
<tr>
<td>10</td>
<td>20</td>
<td>85</td>
<td>119.5</td>
<td>85.5</td>
</tr>
<tr>
<td>10</td>
<td>24</td>
<td>89</td>
<td>117.5</td>
<td>83.5</td>
</tr>
</tbody>
</table>

r = 1
L = 20

K = 1.99E-04

Test 11

![Test 11 diagram]
<table>
<thead>
<tr>
<th>Site 17</th>
</tr>
</thead>
<tbody>
<tr>
<td>Test 12</td>
</tr>
<tr>
<td>hour</td>
</tr>
<tr>
<td>------</td>
</tr>
<tr>
<td>10</td>
</tr>
<tr>
<td>10</td>
</tr>
<tr>
<td>10</td>
</tr>
<tr>
<td>11</td>
</tr>
</tbody>
</table>

\[ r = 1 \]
\[ L = 20 \]
\[ K = 8.53E-05 \]
Residual drawdown analysis at P-17fm.
SUMMARY SHEET Leaky Aquifer Residual Drawdown Analysis.

Name of data file

Number of data points in file

Units of drawdown

Units of time

Pumping rate

Radial distance from pumping well to observation well

Aquifer thickness

Aquitard thickness

Transmissivity estimate

Storage coefficient estimate

Radial hydraulic conductivity

Time match-point

Leakage match-point

Drawdown match-point

Estimate of r/L

Estimate of leakage factor

Estimate of hydraulic resistance of aquitard

Vertical hydraulic conductivity of aquitard

Squared sum of errors

RECOVER.prm

n = 94

L1 = 100 · cm

T1 = 1 · min

Q = 500 · cm³

min

r = 2 · m

b = 1 · m

bp = 1 · m

T = 6.03 · 10⁻¹ · cm²

sec

S = 4.70 · 10⁻³

K_r = 0.006 · cm

sec

t_S = 1.30

β = 1.30 · 10⁻¹

s_D = 0.01

rL = 0.26

L = 7.69 · m

c = 11.36 · day

K_p = 1.02 · 10⁻⁴ · cm

sec

SS_e = 8.32 · 10⁻⁶
Leaky Aquifer Residual Drawdown Analysis

Residual Drawdown

\[ s_{g_i}, F_{g_k}, s_{g_k} \]

\[ t \frac{t_i}{t'_i}, t_gk, t_gk' \]
SUMMARY SHEET for Theis Residual Drawdown Analysis.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Name of data file</td>
<td>RECOVER.pm</td>
</tr>
<tr>
<td>Number of data points in file</td>
<td>n = 94</td>
</tr>
<tr>
<td>Units of drawdown</td>
<td>L1 = 1 · m</td>
</tr>
<tr>
<td>Units of time</td>
<td>T1 = 1 · min</td>
</tr>
<tr>
<td>Pumping rate</td>
<td>Q = 500 · cm³/min</td>
</tr>
<tr>
<td>Time at which pumping ceased</td>
<td>t₀ = 1 · 10⁴ · min</td>
</tr>
<tr>
<td>Transmissivity estimate</td>
<td>T = 6.82 · 10⁻¹ · cm²/sec</td>
</tr>
<tr>
<td>Piezometer distance to pumping well</td>
<td>r = 2 · m</td>
</tr>
<tr>
<td>Change in drawdown over one log-cycle of distance</td>
<td>ds = 0.07 · ft</td>
</tr>
<tr>
<td>Sum of squared errors</td>
<td>SSₑ = 1.40 · 10⁻⁶</td>
</tr>
<tr>
<td>Correlation coefficient</td>
<td>R² = 0.9786</td>
</tr>
<tr>
<td>Lower regression limit</td>
<td>n₁ = 2</td>
</tr>
<tr>
<td>Upper regression limit</td>
<td>n₂ = 20</td>
</tr>
<tr>
<td>Number of points in regression</td>
<td>nn = 19</td>
</tr>
</tbody>
</table>

![Theis Residual Recovery Analysis](image URL)

405
Elastic Storage Estimates From CRS Data
Wells G & H Peatland, Woburn

The following comments are in reference to a series of stress-strain plots, where low stress is defined as < 0.2 ksc, and high stress is defined as < 2 ksc. Elastic storage is calculated as follows from the slope of % strain vs. stress. Data were generated by Jason Bialon (1995).

The plots provided show data by soil type and core location. The upper plot shows low stress data, whereas the lower plot shows high stress data. Slopes, or elastic storage estimates are noted next to the upper plot.

\[
\frac{\% \text{ strain}}{kg \cdot cm^{-2}} \left[ \frac{cm \cdot m^{-1}}{\%} \cdot \frac{kg}{1000 g} \cdot \frac{g H_2 O}{cm^3} \right] \Rightarrow S_e [m^{-1}]
\]

I. Surface Typha Peat

A. three samples, 10 - 27 cm
B. horizontal sample is more non-linear at small stress, but similar at large stresses
C. both vertical samples are very noisy up to 0.6 ksc
D. best estimate of \( S_e \) is 0.004

II. Typha Peat

A. 12 samples, 39 - 70 cm
B. only one horizontal sample, which is more non-linear at low stress, but similar at larger stresses
C. samples from core 2 are noisy around 0.5 ksc, others are smooth
D. samples form core 3, which are actually taken from the same depth at a number of closely (1 m) spaced locations, are very close at low and higher stresses
E. samples form core 1 are more variable, especially at low stress, but also at higher stresses; low stress data from CRS 112 is used for estimate
F. after disregarding results of CRS 100 and 103, best estimate of \( S_e \) is 0.02 at Core 3, 0.016 at core 2, and 0.014 at core 1
III. Sedge Peat

A. four samples, 84 - 100 cm, only at core 2
B. CRS 110 (100 cm) is smooth at small and large stress, others are non-linear at small stress and noisy until 0.5 ksc
C. best estimate of $S_e$ is 0.0026 - 0.016 (CRS 110 gives larger estimate)

IV. Woody Peat

A. eight samples, 103 - 184 cm
B. core 2 samples are noisy at small and large stresses and inconsistent
C. core 1 samples are smooth and very consistent
D. best estimate of $S_e$ is 0.014 for core 1; no estimate possible for core 2

V. Diatomaceous Earth

A. five samples, 184 - 243 cm, only at core 1
B. except for CRS 122, slopes are similar. Only CRS 121 has data at stresses < 0.05 ksc, but all are fairly linear to 0.15 ksc
C. best estimate of $S_e$ is 0.015
Surface Typha Peat in Core 2
Axial strain vs Vertical effective stress

Se = 0.004
Typha Peat in Core 2
Axial strain vs Vertical effective stress

\[ Se = 0.016 \]

Axial strain vs Vertical effective stress

\[ \Delta \ 133 \\
\bullet \ 135 \\
136H \]

Axial strain vs Vertical effective stress

\[ \Delta \ 133 \ (39 \text{ cm}) \\
\bullet \ 135 \ (52 \text{ cm}) \\
136H \ (61 \text{ cm}) \]
Typha Peat in Core 3
Axial strain vs Vertical effective stress

Se = 0.02

Axial strain vs Vertical effective stress

Δ 157 (42 cm)
□ 161 (41 cm)
□ 160 (39 cm)
× 159 (40 cm)
× 158 (44 cm)
Sedge Peat in Core 2
Axial strain vs Vertical effective stress

Se = 0.016 - 0.02

Axial strain vs Vertical effective stress

Se = 0.016 - 0.02
Woody Peat in Core 1
Axial strain vs Vertical effective stress

Vertical Effective Stress (ksc)

Axial Strain (%)

Se = 0.014
Appendix E

Formulation of Depth-Averaged Model of Flow to a Well Near a Stream

The notation used in this appendix follows that of O. D. Strack, (1989, *Groundwater Mechanics*, Prentice-Hall). Once a flow-net was established, the average travel time for each flow tube was calculated using:

\[ V_s = \frac{Ki}{n_e} \]

where:
- \( V_s \) = seepage velocity
- \( K \) = hydraulic conductivity
- \( i \) = hydraulic gradient
- \( n_e \) = effective porosity = 0.3
Notation, as depicted in following figure:

\( \phi = \text{head; [L]} \)
\( b = \text{aquifer thickness; [L]} \)
\( K = \text{hydraulic conductivity; [L} / \text{T]} \)
\( \Phi = K b \phi; \text{flow potential; [L}^3 / \text{T]} \)
\( \Psi = \text{Stream function; [L}^3 / \text{T]} \)
\( Q_{xo} = K b \frac{\partial \phi}{\partial x}; \text{ambient flow; [L}^2 / \text{T]} \)
\( Q = \text{pump rate; [L}^3 / \text{T]} \)
\( r_n = \text{radial coordinate relative to well n; [L]} \)
\( \theta_n = \text{angular coordinate relative to well n; [rad]} \)
\( D = \text{distance from well to river; [L]} \)
\( D' = \text{distance from well to impermeable boundary; [L]} \)

Flow to a well near a stream with ambient flow perpendicular to the stream:

\[
\Phi = Q_{xo}x + \frac{Q}{2\pi} \ln \left( \frac{r_1}{r_2} \right) + \Phi_0; \quad \Psi = -Q_{xo}y + \frac{Q}{2\pi} \ln(\theta_1 - \theta_2) + \Psi_0
\]

\[
r_1 = \sqrt{y^2 + (x - D)^2}; \quad r_2 = \sqrt{y^2 + (x + D)^2}
\]

\[
\theta_1 = \tan^{-1}\left( \frac{y}{x - D} \right); \quad \theta_2 = \tan^{-1}\left( \frac{y}{x + D} \right)
\]

Add an impermeable boundary:

\[
\Phi = Q_{xo}x + \frac{Q}{2\pi} \ln \left( \frac{r_1}{r_2} \right) + \frac{Q}{2\pi} \ln \left( \frac{r_3}{r_4} \right) + \Phi_0; \quad \Psi = -Q_{xo}y + \frac{Q}{2\pi} \ln(\theta_1 - \theta_2) + \frac{Q}{2\pi} \ln(\theta_3 - \theta_4) + \Psi_0
\]

\[
r_3 = \sqrt{y^2 + (x - D')^2}; \quad r_4 = \sqrt{y^2 + (x + D')^2}
\]

\[
\theta_3 = \tan^{-1}\left( \frac{y}{x - D'} \right); \quad \theta_4 = \tan^{-1}\left( \frac{y}{x + D'} \right)
\]
Appendix F

Sensitivity of Model Performance to Changes in Model Parameters
<table>
<thead>
<tr>
<th>Change implemented</th>
<th>Performance Relative to Run 23R7</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>P-17s upper peat</td>
</tr>
<tr>
<td><strong>Sedge Peat (P1bc)</strong></td>
<td></td>
</tr>
</tbody>
</table>
| 8A:  
$K_v(P1bc)$ $\downarrow$ 0.000025 | Very little rainfall response, peak shifted later | NSD | NSD | NSD | Rainfall infiltration shunted by surface layer |
| 8B:  
$K_v(P1bc)$ $\uparrow$ 0.0001 | Moves peak to correct time, causes excess response to early rainfall | Head increases too early | NSD | NSD | Surface recharge to sand increases |
| 8C:  
$K_v(P1bc)$ $\uparrow$ 0.0002 | Decreases slope of receding limb | Increases heads at m and d at early time, flattens peak. m/d gradient reversal too early. | Head increases too early | Increases heads on rising limb | Further increases in recharge to sand. Sand and peat acting as single layer with more damped response |
| 8D:  
$K_v(P1bc)$ $\downarrow$ 0.000025  
$S_x(P1bc)$ $\downarrow$ 0.06 | Early response matched, but peaks higher above 17M, and requires much longer to drain | NSD | NSD | NSD | Matching the rising limb w/o the receding limb may indicate storage hysteresis |
| 8E:  
$K_v(P1bc)$ $\downarrow$ 0.00025 | NSD | NSD | NSD | NSD | Not much horizontal flow in Peat |
| 8F:  
$K_v(P1bc)$ $\downarrow$ 0.00012 | NSD | |
| 8G:  
$K_v(P1bc)$ $\downarrow$ 0.00006 | NSD | |
<table>
<thead>
<tr>
<th>Change implemented</th>
<th>Performance Relative to Run 23R7</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>P-17s upper peat</td>
<td>P-17m Sand</td>
</tr>
<tr>
<td>Sand (S)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>8H: $K_h (P1bc) \uparrow 0.001$</td>
<td>NSD</td>
<td>NSD</td>
</tr>
<tr>
<td>8I: $K_h \ (sand) \uparrow 0.01$</td>
<td>NSD</td>
<td>Peaks too high</td>
</tr>
<tr>
<td>8J: $K_h \ (sand) \downarrow 0.0025$</td>
<td>NSD</td>
<td>Peak too low, slower recession</td>
</tr>
<tr>
<td>8K: $K_h \ (sand) \uparrow 0.01$ $S_e \uparrow 0.016$ (maintains $K_h/S_e = 0.625$)</td>
<td>NSD</td>
<td>NSD</td>
</tr>
<tr>
<td>8L: $K_h (sand) \downarrow 0.0025$ $S_e \downarrow 0.004$ (maintains $K_h/S_e = 0.625$)</td>
<td>Peak slightly lower</td>
<td>better match</td>
</tr>
<tr>
<td>Change implemented</td>
<td>Performance Relative to Run 23R7</td>
<td></td>
</tr>
<tr>
<td>--------------------</td>
<td>---------------------------------</td>
<td>--</td>
</tr>
<tr>
<td></td>
<td>P-17s upper peat</td>
<td>P-17m Sand</td>
</tr>
<tr>
<td>Sand (S) with zoning east-west</td>
<td></td>
<td></td>
</tr>
<tr>
<td>8M: same as 8L with Zoned $K_h$ $K_h$ (sand)east = 0.005</td>
<td>NSD</td>
<td>Vertical gradient reversal late. Heads rise too quickly early time</td>
</tr>
<tr>
<td>8N: same as 8L with $K_h$ (sand)east = 0.007 Tr1 = 0.001 Tr2 = 0.003</td>
<td>NSD</td>
<td>OK, but rises too early</td>
</tr>
<tr>
<td>Streambed (SB)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>8P: $K_s$(SB) ↓ 0.002</td>
<td>Peak damped slightly</td>
<td>Peak damped slightly</td>
</tr>
<tr>
<td>8O: $K_s$(SB) ↓ 0.001</td>
<td>Peak damped</td>
<td>Peak damped</td>
</tr>
<tr>
<td>8Q: $K_s$(SB) ↑ 0.008</td>
<td>NSD</td>
<td>NSD</td>
</tr>
<tr>
<td>Change</td>
<td>Performance Relative to Run 23R7</td>
<td></td>
</tr>
<tr>
<td>-------------</td>
<td>----------------------------------</td>
<td></td>
</tr>
<tr>
<td>implemented</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Typha Peat (P1a)</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
| **8R:**  
Kₙ(P1a) ↓ 0.0005 | NSD          | NSD          | NSD          | NSD          | P1a not saturated much of the time |
| **8S:**  
Kₙ(P1a) ↓ 0.0025 | Rising limb shifted          | Peak and rising limb shifted slightly forward | Peak slightly damped and shifted forward | Starts to restrict recharge | Lowered conductivity starts to restrict recharge at early times |
| **8T:**  
Kₙ(P1a) ↑ 0.002  | NSD          | NSD          | NSD          | NSD          | Vertical flow is controlled by P1bc when P1a is highly conductive |
| **8U:**  
Kₙ(P1a) ↑ 0.004  | NSD          | NSD          | NSD          | NSD          | Vertical flow is controlled by P1bc when P1a is highly conductive |
| **8V:**  
Kₙ(P1a) ↓ 0.00025 | NSD          | NSD          | NSD          | NSD          | No: much horizontal flow in P1a |
| **8W:**  
Kₙ(P1a) ↑ 0.004  | Causes more lag in recession of 17S, shifts peak later | NSD          | NSD          | NSD          | Not much horizontal flow in P1a |
<table>
<thead>
<tr>
<th>Change implemented</th>
<th>Performance Relative to Run 23R7</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>P-17s upper peat</td>
<td>P-17m Sané</td>
</tr>
<tr>
<td>Diatomaceous Silt (DE)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>8X:</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$K_v$(DE) ↓ 0.000015</td>
<td>NSD</td>
<td>NSD</td>
</tr>
<tr>
<td>8Y:</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$K_v$(DE) ↓ 0.0000075</td>
<td>NSD</td>
<td>delays head rise slightly</td>
</tr>
<tr>
<td>8Z:</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$K_v$(DE) ↑ 0.00006</td>
<td>NSD</td>
<td>Peaks to high</td>
</tr>
<tr>
<td>8ZA:</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$K_v$(DE) ↑ 0.00012</td>
<td>All heads too high recession too low</td>
<td></td>
</tr>
</tbody>
</table>

↑ indicates parameter increased to following value  
↓ indicates parameter decreased to following value  
**bold type** indicates that change is an improvement relative to calibration, otherwise the recorded effect causes a worse fit to measured data  
NSD = no significant difference