ELEMENTS OF SALT MARSH HYDROLOGY

by

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Submitted to the Department of Civil Engineering on May 20, 1986, in partial fulfillment of the requirements for the degree of Doctor of Philosophy in Civil Engineering

ABSTRACT

Movement of the pore water controls the solute balance and air entry in salt marsh sediments, and through these affects primary productivity of the ecosystem. The objective of this thesis is to describe the water balance in the sediments of the Belle Isle marsh, a New England salt marsh, and in doing so describe the general pattern of water movement in the sediment. This study is based on the relationship that exists between hydraulic head and the storage and conductance of water in the sediment. Observed fluctuations in the distribution of head are compared to models of head response to the component fluxes of the water balance to assess the importance of each flux in the sediment water balance. Physical properties of the sediment determined by fitting the models to head data are compared to independent estimates of the sediment properties.

Four mechanisms link changes in the water content of the sediment to changes in head. These are changes in the degree of saturation, changes in the bulk volume of the sediment, compression of air-filled roots and rhizomes, and compression of trapped pockets of gas. A general expression is derived for specific storativity, the storage property of the sediment, that accounts for all four storage mechanisms. The contribution of each of the storage mechanisms is estimated from water balance observations in a lysimeter, tension plate tests and direct measurements of changes in the bulk volume of the sediment in the field. Methods that have been used to estimate the hydraulic conductivity in salt marshes are summarized and results are presented for two salt marshes.

This study finds that the water balance in salt marsh sediments is dominated by vertical fluxes. The addition of water to the sediment by precipitation and tidal inundation is balanced by a loss of water by evapotranspiration. Water loss by drainage through the sediment and out across the creek bank is significant only within 10-15 m of the creek. The specific storativity of the sediment is determined by three storage mechanisms, changes in the degree of saturation, changes in bulk volume, and compression of air-filled roots. Changes in water content occur throughout the sediment, not just above the water table as is sometimes thought. Marsh grasses have a significant effect on the hydraulic properties of the sediment, and the possibility exists that ecological processes affect hydrology of the sediment in salt marshes.

Thesis Supervisor: Dr. Harold F. Hemond
Title: Associate Professor of Civil Engineering
Work, Finish, Publish.

- Michael Faraday
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PREFACE

Ever since it became apparent to me that there would be a thesis to write I have been looking forward to writing its preface. In many ways this is the best part. These are my thoughts at the finish of a project that has occupied me for six years, a sort of formalized scribbling in the margins of a graduate school yearbook. I have known some of you who are reading this; you have contributed to this work and witnessed its progress. You are in my thoughts even though not all of you can be mentioned here. Others of you I do not know; you may have retreated here for a respite or in search of the story behind the thesis, as I have done with others’ theses. Let me begin then with a story that has been on my mind lately.

A young couple was dismayed by the extreme and curiously divergent attitudes that had developed in their twin sons. One was a dismal pessimist while the other was an undaunted optimist. The parents became concerned that each boy would have a hard time of it if these attitudes were not somehow moderated and set themselves to the task. On Christmas morning, when the family assembled to see what Santa had brought, the pessimist had been treated generously and he found all of the toys that he had ever wished for. His brother on the other hand had received a single gift, a heap of stable sweepings. The ploy did not have the intended effect as the pessimist was just as despondent as ever. When asked what was the matter he replied that this was the wrong color and that the wrong model and in any case all of the toys would soon be broken. Surely the stable sweepings had moderated the other son’s optimism the parents thought, but
no, he was found in the middle of his heap gleefully digging away. The parents, puzzled, asked how he could be so happy. "I just know that there's a pony in here somewhere!" he replied.

I have not always been nearly so good natured mucking around in the marshes, nor have I been alone. The scope of this work is broad, and I am indebted to those whose work I have drawn on and to those who have assisted me directly. Many thanks to Harry Hemond and Keith Stolzenbach who advised me; in particular to Harry for encouraging me to follow my interests no matter where they seemed to be leading and to Keith for making sure that that did not get too far out of hand.

I have relied heavily on Diane Chen and the results of her thesis. Diane has been responsible for what I known about the peat of Belle Isle marsh. Eric Nichols helped me figure out how piezometers work, and parameter estimates based on his analysis of piezometer response are featured here. Roger Burke and Jayne Fifield Knott preceded me and were the first man and first woman of salt marsh hydrology. They received the allegorical apple from John Teal of Woods Hole and thus began an eight-year flight from ignorance in which this is only the latest step.

Mark Schaefer seems to have been around nearly as long as I have and has been a valued companion in exploring the uncharted world of instrumentation electronics. Mark Madson wandered in one day looking for something to do and ended up working on various aspects of the lysimeter, particularly the data logger, which, I am proud to say, was conceived and built from scratch over a period of nine months by two guys who had absolutely no idea what they were doing.
The Undergraduate Research Opportunities Program office has been generous in its support of this research. Ray Schmitt spent a summer figuring out how to watch the marsh surface rise and fall with changes in water content. Martin Anderson, Ruth Fricker and Alison Cullen helped in the early stages of the field work in Belle Isle marsh. Helen Han contributed to the work done over the summer of 1985.

I am fortunate to have studied within the Parsons Lab at MIT. The people I have worked with and lived with are an extraordinary bunch and have left a deep impression on me. Of this nameless herd I would like to mention Lenny Richardson in particular who made the early part of writing this thesis interesting, my colleagues in 48-420, and the two classes of Civil Engineering undergraduates who were supposed to learn something from me and who taught me a lot in return.

Finally, I would like to express my appreciation to the Friends of the Belle Isle Marsh, the Metropolitan District Commission, and the Boston Conservation Commission for letting me work in their marsh. Belle Isle marsh was uniquely suited for this work and I have no doubt that less would have been accomplished had I been working elsewhere.

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SYNOPSIS

Salt marshes are among the most productive ecosystems of the world. In terms of the annual production of organic matter, salt marshes are more productive than agriculture in the United States (1.8 kg/m² yr in a Georgia marsh versus 1.3 kg/m² yr for corn). The material produced in the salt marshes is thought to be eventually exported to adjacent estuaries, where it is an important component of the food chain. Salt marshes are all the more remarkable for attaining their high productivity in an environment that otherwise would be considered to be hostile to plant growth. The marshes are inundated with highly saline water and the soil is continually wet. The near saturation of marsh sediments restricts the availability of oxygen to the roots where it is required for respiration.

It is important to understand how the salt marsh ecosystem works, and sediment hydrology is an important part of the system. The abundance of water in the sediment is one of the principal stresses on the marsh grasses, and the movement of water in the sediment plays a role in the major geochemical cycles in the marshes. There are zones of active pore water movement and zones of little or no water movement. This characteristic has been correlated with the productivity of the marsh grasses and the chemical characteristics of the pore water. The problem is that virtually nothing is known about the general pattern of water circulation in the sediments, the principal mechanisms driving water movement, and typical values of water fluxes. This is the problem to be addressed in this study.
A Brief History

Teal (1962) was the first to describe the mechanics of a salt marsh ecosystem by detailing the flow of energy through the food web. His was one of the first studies to apply this technique to any ecosystem. Teal concluded with the following paragraph which set an agenda that is still followed in salt marsh research.

The tides are of supreme importance in controlling the environment of the salt marshes. They limit the number of species that can occupy the system and so make it simple enough to be studied in detail reported here. They are responsible for the high production of Spartina, as witnessed by the luxuriant growth along the tidal creeks as compared with that on the smooth Spartina areas. At the same time the tides remove 45% of the production before the marsh consumers have a chance to use it and in so doing permit the estuaries to support an abundance of animals.

It is clear from comments made earlier in his paper that Teal had in mind the specific marsh/estuary system at his study site in Georgia. Nonetheless, the hypothesis that highly productive salt marshes export the fundamental components of the food chain to support estuarine fisheries was a popular one at the time; one that probably saved many marshes from being filled and stimulated interest in further research in salt marshes.

Valiela et al. (1976) published the first, and to date the only, estimate of the rate of root production in a salt marsh, which showed that primary production below-ground is as large as the above ground production used by Teal (1962) in his energy flow calculations. The estimated rate of root production is sufficient to raise the surface of the marsh 1 cm per year if it were allowed to accumulate. Attempts to account for the fate of all of this organic material have led to the formulation of two different
views of the below-ground metabolism (Howarth and Teal 1979, 1980; Howes et al. 1984, 1985). Two points of contention are the rate of oxygen exchange across the marsh surface and the nature of the solute balance in the sediment, i.e., the rate of export of reduced sulfur compounds. Others have established that these issues bear directly on the productivity of the marsh grasses (Linthurst 1979, Mendelssohn et al. 1981 King et al. 1982).

Oxygen exchange and the solute balance are both controlled by the movement of water in the sediment. It is known that generally pore water in the creek bank regions is more dynamic than the pore water in the interior regions of a marsh (Gardner 1973). However, very little has been done to quantify water movement in salt marsh sediments, and there has been no recognition of the fact that the mechanisms for water storage in salt marsh sediments are significantly different than those in upland soils.

**This Work**

This study can be divided into three stages. The first stage is the formulation of an expression for the conservation of mass for the pore water in the sediment. This has the form of a differential equation describing the distribution of hydraulic head in the marsh, in which the hydraulic properties of the sediment are parameters. The components of the water balance are evapotranspiration, precipitation, infiltration of tidal water, and drainage out of the sediments through the creek banks, Figure 1. Vertical water fluxes across the bottom of the sediments may be a factor in some marshes, but they are thought to be negligible in Belle Isle because the sediment is underlain by a layer of impermeable clay.
Figure 1
COMPONENTS OF THE SEDIMENT WATER BALANCE IN A SALT MARSH

PRECIPITATION
EVAPOTRANSPIRATION
TIDAL INUNDATION
DRAINAGE
EXCHANGE WITH UNDERLYING AQUIFER (?)
The second stage establishes the values of the hydraulic properties of the sediment, the mean specific storativity and hydraulic conductivity. Particular attention is paid to describing the physical mechanisms that contribute to the storage of water in the sediment. Recent efforts at modeling head response in salt marsh sediments have assumed that the principal mechanism for water storage is compression and expansion of the sediment itself (Hemond and Fifield 1982, Hemond et al. 1984). The importance of the degree of saturation of the pores in determining the rate of oxygen transport into the sediment and evidence that changes in water content by desaturation may be important hydrologically led to the decision that a full accounting of all of the mechanisms of water storage in the sediment should be made. A lysimeter study and observations of volumetric strain of the sediment in the field were undertaken as part of this study for this purpose. The results of tension plate tests and dye studies by Chen (1986) are also used.

In the third stage, models of the head response to the component fluxes of the water balance are formulated and compared to observations of the distribution of hydraulic head observed in Belle Isle marsh. The aim of the study is a broad, general description of the patterns of water movement in the sediment. This is accomplished through the use of simple models that aim to describe the head response to specific fluxes of the water balance rather than the formulation of a single comprehensive numerical code that is able to describe all aspects of the water balance. The process of "dissecting" the record of hydraulic head using a suite of simple
models leads to a description of the mechanics of the water balance in the sediment. The model parameters determined by fitting models to the head data are compared with the independent estimates of the sediment properties obtained in the second stage of the study as a check.

Organization of the Thesis

This thesis has been written as five papers. Together the first three papers describe the investigation of the water balance. The fourth and fifth papers describe the lysimeter study and the observations of volumetric strain (surface displacement) in the marsh. A summary of each paper follows.

"Elements of Salt Marsh Hydrology: I. The Storage and Conductance of Water by Peat" presents the equation of mass conservation for water in peat that is the basis for the analysis of head observations. A general expression is developed for specific storativity in which four mechanisms of water storage are considered; changes in the degree of saturation, changes in bulk volume, compression of gas pockets and compression of the air-filled roots and rhizomes. Methods for estimating the specific storativity and hydraulic conductivity of peat are reviewed, and the results from two salt marshes, Belle Isle marsh in Boston, Massachusetts, and Sippewissett marsh in Falmouth, Massachusetts, are summarized.

"Elements of Salt Marsh Hydrology: II. The Water Balance in Belle Isle Marsh" presents the observations of hydraulic head in Belle Isle marsh and their analysis to describe the water balance in the sediment. Models of head response to the component fluxes in the water balance are derived from
the mass conservation equation developed in the previous paper and compared
to the observed head response. Model parameters determined by fitting are
compared to independent estimates of specific storativity and hydraulic
conductivity.

"Elements of Salt Marsh Hydrology: III. How Much Seeps on the Neap?" focuses on the horizontal drainage in the sediment from interior regions of
the marsh towards the creek banks. The significance of this water flux is
a point of contention between two views of the below-ground metabolism in
the marsh. A model of head response to evaporation and drainage is formu-
lated and calibrated using observations of head. The model is used to
discuss the factors controlling horizontal drainage and to estimate the
magnitude and distribution of horizontal water fluxes in the sediment.

"Hydraulic Properties of Salt Marsh Sediment Determined from a Lysi-
meter" presents the results of a water balance study on a sample of Belle
Isle sediment. The specific storativity is estimated directly from the net
change in head observed in response to a change in sediment water content.
Hydraulic conductivity is estimated from the transient head response to the
addition of water to the lysimeter. Both hydraulic conductivity and speci-
fic storativity vary with depth and pressure due to the effects of desatu-
ration. In spite of this, the linearized mass balance equation, in which
the properties of the peat are assumed to be constant and uniform, is
adequate to describe the general features of the head response.

"Peat Rose!" describes the displacements of the marsh surface observed
in response to changes in water content and surface loading by tidal inun-
dation. These observations are the basis for estimates of the components
of specific storativity due to bulk volume change and the compression of trapped gas pockets.

Conclusions:

There are three mechanisms for water storage active in the marshes studied; changes in the degree of saturation of the pores, changes in the bulk volume of the sediment, and compression of air-filled roots and rhizomes. In Belle Isle marsh, changes in bulk volume account for 20% of the change in water content in the 170 cm-deep sediments. The remaining 80% is divided more or less evenly between root compression and changes in saturation in the upper 20 cm of the sediment. Observations of surface displacements in Sippewissett marsh during tidal inundation of the surface suggest that the sediments may contain as much as 1-2% gas by volume. Using this figure, the specific storativity due to compression of the gas trapped in the sediment is on the order of $0.2 \times 10^{-4}$ cm$^{-1}$, compared to $2-4 \times 10^{-4}$ for bulk volume change, $4 \times 10^{-4}$ for root compression, and $10-20 \times 10^{-4}$ for desaturation. The storativity due to desaturation is smaller in salt marsh sediments than it is typically in sand or undisturbed fresh water peats, but it is larger than for clays in the same range of pressures. The compressibility of the sediment is one to two orders of magnitude higher than for consolidated sediments and is comparable to the compressibility of unconsolidated clays. The contribution of root compression to water storage has not been reported prior to this study. The magnitude of the storativity due to root compression was obtained indirectly so more study of this phenomenon is needed to confirm the result.
The mean hydraulic conductivity of salt marsh sediments is comparable to that of a fine sand or silt, $5 \times 10^{-4}$ cm/s in the root zone. The saturated hydraulic conductivity of the sediment in the root zones of both of the marshes studied here are virtually identical, in spite of significant differences in the composition of the sediments. Chen (1986) has found that water flows preferentially along channels associated with the roots of the marsh grasses, and this may account for the similarity in the hydraulic conductivity in the two marshes.

The water balance in the sediment is dominated by vertical water fluxes. Water supplied to the sediment by precipitation and infiltration during tidal inundation is balanced by the loss of water to evapotranspiration. Water loss by drainage horizontally through the sediment and out through the creek banks is significant only within 10-15 m of the creek bank. The results of the model study indicate that the area over which horizontal drainage is significant is controlled by the topography of the marsh surface. The cumulative effect of evapotranspiration on mean head reduces the amount of drainage, and during extended periods when the water content of the sediment is not replenished by precipitation or tidal inundation, the drainage flux at the creek bank may reverse so that there is a net flow of water into the sediment.

The infiltration of saline tidal water and its loss principally by evapotranspiration presents a problem for the control of solute concentrations in the sediment. It is estimated that the evaporation of tidal water from the sediments of Belle Isle marsh results in a net flux of 10 kg m$^{-2}$ yr$^{-1}$ of salt into the sediment. This flux must be balanced by
some process that is yet to be described. The results of this study suggest that salt removal by advection to the creeks is unlikely to be sufficient to balance its input by infiltration.

A close coupling between the marsh grass and the hydraulic properties of the sediment is indicated. Root compression is an important mechanism of water storage in the sediment and it seems likely that the water yield by desaturation of the pores is also affected by the roots through their inevitable effect on the distribution of pore sizes. The hydraulic conductivities of sediments from the root zones of three different salt marshes are virtually identical. A comparison of the sediment response in the lysimeter and in tension plate tests (Chen 1986) suggests that a network of relatively large pores exists in the sediment through which the bulk of the water flow occurs. Dye studies by Chen (1986) have demonstrated this to be the case, with flow occurring predominantly along root channels.

The magnitude of fluctuations in the water content of the sediment is imposed by the rate of evapotranspiration and the frequency of tidal inundation. The range of pressure variation in the pores in response to changes in the water content of the sediment is governed by the storage properties of the sediment. The larger the range of pressure change, the greater the probability that air will enter the sediments to a given depth. Anoxic conditions in the sediments, which are caused by their high degree of saturation, are a significant stress to the marsh grasses.
REFERENCES


ELEMENTS OF SALT MARSH HYDROLOGY:

I. STORAGE AND CONDUCTANCE OF WATER IN PEAT

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May 1986
ABSTRACT

The mechanisms of dynamic water storage in salt marsh sediments and their distribution are important to understanding the hydrology of salt marshes. There are four storage mechanisms to be considered: changes in saturation, changes in the bulk volume of the sediment, compression of gas pockets isolated from the atmosphere in the sediment, and compression of air-filled roots of the marsh grasses. The water storage characteristics of the sediment are described by the mean specific storativity, the change in water content per unit change in hydraulic head per unit volume of sediment. A general expression for the specific storativity of peat is developed, and estimates of specific storativity and hydraulic conductivity of the sediment in two New England salt marshes are presented. Interaction between the processes of water storage and water conductance results in a lag in the response of pore water to meteorological and tidal forcing. The roots of the marsh grasses have a large effect on the hydraulic properties of the sediment that control the pore water response.
I INTRODUCTION

Sediment hydrology exerts an important influence on the productivity of the marsh grasses through its role in the balance of solutes in the peat and the coupling between water movement and the advection and diffusion of atmospheric gases into the sediment (Mendelssohn et al. 1981, King et al. 1982). The primary productivity of salt marsh grasses is high which makes salt marshes a valuable component of the coastal environment. However, in spite of the recognized importance of hydrology in the last 10 years, few studies of sediment hydrology have been attempted in salt marshes, and "an overall picture of salt marsh hydrology is not now available" (Correll 1985).

Pore water fluxes have been estimated from the distribution and variation of pore pressures in salt marsh sediment. Jordan and Correll (1985) estimated the velocity of drainage toward a creek from the slope in the water table by using Darcy's law and estimates of hydraulic conductivity determined in situ. Several investigators (Howarth et al. 1983, Dacey and Howes 1984, Agosta 1985) have used observed fluctuations of the water table and estimates of the difference in water content of the sediment above and below the water table, presumably due to desaturation, to estimate water fluxes. Hemond and Fifield (1982) and Hemond et al. (1984) have constructed models of pore water movement in salt marsh sediment based on the assumption that the compressibility of the sediment, not changes in saturation, is the principal mechanism of water storage. Questions related to the relative importance of mechanisms of
dynamic water storage and their location in the sediment are timely and are fundamental to understanding the patterns of water movement.

This paper reviews the current knowledge about the hydraulic properties of salt marsh sediment. Much of this information derives from an eight-year study of salt marsh hydrology at MIT, and is being published for the first time. Four storage mechanisms are identified as potentially significant in marsh sediments: changes in the degree of saturation, changes in the bulk volume of the sediment, compression of gas pockets, and compression of air-filled roots and rhizomes. A general parameterization for water storage in sediment is presented that is valid for both fully saturated and partially saturated conditions. The contribution of each mechanism is estimated from laboratory and field observations of related phenomena.

II STUDY SITES

Sediment hydrology has been studied in two Massachusetts salt marshes: Belle Isle marsh in Boston and Sippewissett marsh in Falmouth. An extensive program of core sampling was conducted in each marsh as part of investigations into the depth of the sediment deposits and the physical properties of the sediment (Fifield 1981, Knott et al. 1984, Chen 1986). In addition, Belle Isle has been the site of a study of the water balance in the sediment (Nuttle, 1986b,c), and Sippewissett has been the site of studies of major geochemical cycles (Valiela et al. 1978, Howarth et al. 1983, Howes et al. 1984, Howes et al. 1985).
The Belle Isle marsh is the last remnant (100 hectares) of the once-extensive marshes in Boston, Massachusetts. The marsh sediments consist of a layer of clayey peat (90% ash free) 70 cm deep which grades into a layer of gray silty clay. A layer of relatively impermeable clay underlies the study area at 170 cm. The vegetation over most of the study area consists of a mix of Distichlis spicata and Spartina patens typical of drier, high marsh conditions. The average range of tides is 2 m. The hydrologic regime is characterized by alternating periods of daily flooding of the marsh surface and periods in which the surface does not flood for several days. The periods of surface flooding are related to the spring-neap cycle but do not follow a regular 28-day cycle.

Sippewissett marsh is a Spartina alterniflora-dominated salt marsh on the eastern shore of Buzzards Bay, on Cape Cod. The range of tides is 1.2 m, and the marsh surface is flooded more frequently than in Belle Isle. The sediment is 1 m to 4.5 m deep and is underlain by a thick (30 m) deposit of glacial outwash sand. The sediment is much "peatier" in appearance than that in Belle Isle marsh and contains less mineral material, 70% ash-free dry weight.

III THE WATER BALANCE IN PEAT

The term "peat" will be used here to refer to any salt marsh sediment that has built up over a period of time as the result of the presence of salt marsh vegetation. It therefore includes both the conventionally peaty soil found in some salt marshes, which is obviously organic in origin, and marine sediments that have been co-deposited with
organic matter as the result of the filtering action of the vegetation. The latter may be predominantly clay or silt, with roots, rhizomes and other plant remains in small proportion by weight. The important role played by salt marsh vegetation in the structure of these sediments distinguishes them from other sediments deposited by purely mechanical means. The hydrologic concepts and techniques for estimating the hydraulic properties discussed in this and the following sections have been applied to both the clayey peat from Belle Isle marsh and the "peaty" peat from Sippewissett marsh.

Consider a volume of peat $V_B$ that encloses a constant mass of solid material. The condition of mass conservation for pore water in the peat provides that the rate of change in the water content of $V_B$ is equal to the net water flux across the boundary of the control volume. The density of water in marsh sediments can be taken as uniform and constant, and the conservation of mass can be expressed as

$$ \frac{dV_W}{dt} = -\int \bar{q} \cdot ds \quad [3.1] $$

in which $V_W$ is the volume of water contained in the peat volume $V_B$, $\bar{q}$ is the specific discharge of water in the peat (cm$^3$/cm$^2$/s), and $ds$ is a differential element of the boundary of $V_B$ with an outward oriented normal. By Darcy's law specific discharge is proportional to the gradient in the hydraulic head $h$.

$$ \bar{q} = -K \nabla h \quad [3.2] $$
Hydraulic conductivity is the constant of proportionality K. The hydraulic head at a point in the peat is the sum of the pressure head and elevation (gravity head):

\[ h = \psi + z ; \quad \psi = P/\gamma_w \]  \hspace{1cm} [3.3]

in which P is the pore pressure and \( \gamma_w \) is the unit weight of water

The water content of soils can be related to pore pressure theoretically, either by assuming that the soil components are in thermodynamic equilibrium (Sposito 1972, 1973) or by assuming that a state of mechanical equilibrium exists (Narasimhan and Kanehiro 1980, Brutsaert and El-Kadi 1984). The mechanical approach will be used in the following sections to describe the mechanisms of water storage in salt marsh peat. Briefly, changes in pore pressure affect the distribution of stress in peat. Decreases in pore pressure cause the bulk volume to shrink and expands the air-filled roots and gas pockets isolated from the atmosphere in the peat. The pressure difference between gases in the peat that are in equilibrium with the atmosphere and the pore water, which is usually at pressures less than atmospheric near the peat surface, is balanced by the surface tension of the air-water interface acting through the curvature of myriad menisci in the pores. As pore pressures decrease the menisci retreat into smaller and smaller pores to increase the interfacial curvature. The net effect of decreasing pore pressures is to decrease the proportion of the bulk volume of the peat that is occupied by water.
The relationship between pore pressure changes and changes in the water content of peat is characterized by a material property, the specific storativity. For the volume of peat considered above, the mean specific storativity is defined as

\[
\langle S \rangle_s = \frac{1}{V_B} \frac{dV_w}{d\langle \psi \rangle}
\]  

in which \(\langle \psi \rangle\) is the mean pressure head in \(V_B\). The mass conservation equation, [3.1], can be written as

\[
V_B \langle S \rangle_s \frac{d\langle \psi \rangle}{dt} = -\iint \vec{q} \cdot d\vec{s}
\]  

Volumetric strain in salt marsh peat is small \((\Delta V_B/V_B < 0.01)\) so that the effects of changes in bulk volume that are not directly related to changes in water content can be neglected in [3.5].

In general, \(d\langle \psi \rangle\) in [3.4] and [3.5] is defined only for changes between equilibrium distributions of \(\psi\) in \(V_B\). A similar condition is imposed by assuming equilibrium conditions (thermodynamic or mechanical) in the theories relating pore pressure to water content. These conditions can be relaxed by conceptually shrinking \(V_B\) so that equilibrium conditions are assumed to occur only locally at each point in the peat. However, often it will be useful to apply [3.5] to finite volumes of peat when interpreting changes in head observed in the peat following the addition or loss of an unknown volume of water, as by evapotranspiration or infiltration. It follows from [3.5] that the net change in
head due to a changes in water storage is determined by the specific
storativity of the peat. The transient response of head, the length of
time that it takes for the head to respond completely to a new water
flux, depends on both the hydraulic conductivity and the specific stor-
ativity of the peat.

IV SPECIFIC STORATIVITY OF PEAT

Mechanisms of Water Storage

An arbitrary volume of peat $V_B$ can contain as many as five separate
components, Figure 1. The solid matrix of the peat is made up of a
volume of living roots and rhizomes $V_R$, and a volume of lifeless solid
material $V_S$, which is composed of organic debris, silt, sand, and clay.
The lifeless material is incompressible and by definition the amount of
this material in $V_B$ does not change. The roots of salt marsh grasses
are hollow and air-filled (60% air by volume, Dacey 1981) as an adapta-
tion for survival in saturated soils. Therefore root volume can be
expected to change with pore pressure. The pores in the solid matrix
are filled by a volume of water $V_W$, a volume of air or gas that is
pressure equilibrium with the atmosphere, $V_A$, and a volume of gas that
is isolated from the atmosphere, $V_G$. The air contained in the roots is
included in $V_R$. There are two reasons for the occurrence of isolated
gas pockets in the peat. Atmospheric gases can be trapped when the
pores near the surface are saturated during infiltration, and methane
generated in the peat by anaerobic decomposition is likely to form
bubbles due to its low solubility.
A relation for the change in water content of $V_B$ follows directly from the schematic representation of the peat shown in Figure 1.

\[ dV_W = dV_B - dV_R - dV_G - dV_A \]  \[ [4.1] \]

Each term on the right corresponds to a particular mechanism of water storage. They are, respectively, changes in bulk volume, compression of roots, compression of gas pockets, and changes in the air content of the peat. The volume of the lifeless solid phase in $V_B$ is constant, as noted above, and does not contribute to changes in water content. All that remains in deriving a general expression for specific storativity is to relate the volume changes on the right side of [4.1] to changes in pore pressure.

Degree of Saturation

The air content of a volume of soil is parameterized by its degree of saturation, $S$, the proportion of the pore volume $V_V$ occupied by gases in equilibrium with the atmosphere.

\[ S = 1 - \frac{V_A}{V_V}; \quad V_V = V_B - V_S - V_R \]  \[ [4.2] \]

The degree of saturation of a soil varies with pore pressure. The position of the interface between the pore fluid and atmospheric gases depends on the pressure difference across the interface and the geometry of the pores in the soil. The pressure difference is balanced by the
interfacial tension acting through the curved interfacial surface, as in a capillary tube.

\[ P_{ATM} - P = \frac{2\tau \cos \theta}{r} \]  \hspace{1cm} [4.3]

Here \( \tau \) is the interfacial tension, \( \theta \) is the wetting angle (\( \theta = 0 \) for hydrophilic surfaces) and \( r \) is the radius of curvature of the interface. The radius of curvature is roughly equal to the mean radius of the pore bridged by the interface.

Decreases in fluid pressure require the air-water interface to retreat to smaller pores and deeper into the peat where the fluid pressure is higher. The volume of air in the peat increases and the degree of saturation decreases. Increasing pore pressure has the opposite effect. In general, the relation between saturation and pore pressure depends on whether the soil is wetting or drying, a situation known as hysteresis (Hillel 1980). However, for the purposes of this discussion, \( S(P) \) is assumed to be unique.

The rate of change of the air content with pressure

\[ \frac{dV_A}{dP} = -V_V \frac{dS}{dP} \]  \hspace{1cm} [4.4]

depends on the magnitude of the fluid pressure through the distribution of pore sizes in the peat. In particular, if pressure is decreased starting with peat in a fully saturated state, \( P = 0 \) at the surface, there will be no significant change in the degree of saturation until
the radius of curvature of the air-water interface is less than the largest significant pore size. The pressure difference $P_{\text{ATM}} - P$ at which this occurs is known as the air entry value, or the bubbling pressure. In 4.4 it has been assumed that changes in void volume due to changes in bulk volume or changes in root volume have an insignificant effect on the air content of the peat.

**Gas Compressibility**

The volume of a gas pocket trapped in the peat is a function only of the pressure of the surrounding fluid, assuming that the rate of gas production by biological processes is negligible. If the gas follows Boyles law, then the rate of change of gas volume with pressure is

$$\frac{dV_G}{dP} = \frac{-V_G}{P}$$ \hspace{1cm} [4.5]

If a Henry’s law equilibrium exists between the gas mixture in the gas pocket and substances in solution in the pore fluid then, the rate of volume change is increased by multiplication by a factor.

$$\frac{dV_G}{dP} = \frac{-V_G}{P} \left[ 1 - \chi (1) \left( 1 + \frac{M_{(1)}}{M_1} \right)^{-1} - \chi (2) \left( 1 + \frac{M_{(2)}}{M_1} \right)^{-1} \cdots \right]^{-1} \hspace{1cm} [4.6]$$

For a particular volume of peat, $M_{g(n)}$ is the mass of the $n^{\text{th}}$ gas in the gaseous phase, $M_1(n)$ in the mass of the $n^{\text{th}}$ gas contained in the solution, and $\chi (n)$ is its mole fraction in the gas mixture. Henry’s law
controls the partitioning of the total mass of a gas between its gaseous and dissolved states.

\[
\frac{M_g^{(n)}}{M_L^{(n)}} = H^{(n)} \frac{M_w V_G}{\rho_w RT V_w} V_B
\]  \[4.7\]

\(H^{(n)}\) is the Henry's law constant for the \(n^{th}\) gas, \(M_w\) and \(\rho_w\) are the molar mass and density of water, \(R\) is the universal gas constant, and \(T\) is the absolute temperature. Generally, soluble gases will give rise to large Henry's law effects in the compressibility of gas pockets.

The Mechanical Behavior of Peat: Bulk Volume and Root Compressibility

The total stress and its distribution among a soil's constituents determines the mechanical behavior of the soil, of which changes in bulk volume and root volume are of interest here. The total stresses can be calculated at any point in a soil body from the forces carried by the soil body by considering the soil to act as a continuous solid. The stresses are distributed between the fluid in the pores, the pore pressure, and the "effective stress" acting on the solid matrix. Changes in the bulk volume of the soil are related to changes in the effective stress component.

A simple stress distribution can be assumed through much of a deposit of salt marsh peat: the vertical normal stresses at a point in the peat balance the weight of the overlying material. This holds except in the region of the creek banks where lateral stresses may also be impor-
tant. The total vertical stress at a point can be considered to be constant, except during tidal inundation when the weight of the water ponded on the marsh surface adds to the stresses already present in the peat. The small changes in stress related to changes in the water content of the peat above a point are neglected. Hemond et al. (1984) have considered the effects of tidal loading on the movement of water in the peat during infiltration.

The distribution of the total vertical stress \( \sigma_T \) at a point in a partially saturated soil is given by the relation

\[
\sigma_T = \sigma' + xP - (1 - x)P_{ATM}
\]  \[4.8\]

in which \( \sigma' \) is the vertical effective stress in the soil matrix, and \( x \) is known as Bishop's parameter and is a function of the degree of saturation of the soil (Bishop and Blight 1963). For a soil at or near saturation \( S = 1 \), \( x \) is unity. \( x(S) \) decreases with the degree of saturation so that pore pressure \( P \) plays a small role in the stress distribution in a dry soil. Fredlund and Morgenstern (1976) argue that the use of Bishop's parameter alone is insufficient to describe the behavior of a partially saturated soil with variable total stress. However, salt marsh peat is always at or near saturation and the total stress is constant except as noted above. The formulation for the distribution of stress in \[4.8\] is sufficient for the purposes of the present discussion.
Consider the classical problem of the vertical consolidation of a soil composed of an incompressible solid phase (Lambe and Whitman 1969). The distribution of total stress is defined by 4.8 and the total stress in the soil is assumed to be constant. If the soil is saturated, then changes in pore pressure induce equal and opposite changes in effective stress. The soil matrix, the porous structure composed of the solid phase of the soil, is assumed to be an elastic structure in which bulk volume is a function only of the applied effective stress;

\[ V_B = \frac{dV_B}{\sigma'} + C_o \]  

[4.9]

in which \( C_o \) is a constant and \( \frac{dV_B}{d\sigma'} \) is a measure of the elasticity of the soil matrix. Inelastic deformations of a soil decay with time, so where stresses vary cyclically within a limited range, as is the case in salt marshes sediments, it is reasonable to assume elastic soil response.

The theory of consolidation is a useful model for relating observed changes in bulk volume of marsh sediments to changes in pore pressure. Day-to-day changes in the surface elevation of the sediments are linearly related to changes in the mean hydraulic head in the sediments, of which pressure is the only variable component (Nuttle 1986e). However, the unique structure of salt marsh peat limits the application of the classical theory to the level of a first-order approximation.
Peat is made up of both an incompressible solid phase (organic debris and mineral material) and the roots of salt marsh grass which are hollow, air-filled, and compressible. The actual behavior of the root volume is unknown, but because the roots are subject both to changes in effective stress, as a consequence of being a component of the peat matrix, and to changes in the pressure of the surrounding pore fluid, it must be assumed that root volume varies with both components of stress. Therefore, for conditions of constant stress,

\[
\frac{dV_R}{d\sigma} \bigg|_{T} = \frac{\partial V_R}{\partial P} + \frac{\partial V_R}{\partial \sigma'} \frac{d\sigma'}{dP} \bigg|_{T} \tag{4.10}
\]

in which \(\frac{d\sigma'}{dP}\bigg|_{T}\) is equal to the modified Bishop's parameter \(\chi^e = \chi^e(P)\) as a consequence of [4.8] and the condition of constant total stress, Narasimhan and Kanehiro (1980). Also, the relationship between root volume and pressure must be affected by the degree of saturation, so that root volume is less sensitive to changes in pore pressure as saturation decreases and an increasing proportion of the adjacent pores contain air at constant pressure.

The potential for root volume change with pore pressure has implications for the behavior of the bulk volume of the peat. The bulk elasticity of the peat matrix depends on the elasticity, strength, and the arrangement of the components of the matrix structure. The size and geometry of the roots, and thus their contribution to the bulk elastici-
ty of the peat, depend on the surrounding fluid pressure. Therefore, the bulk volume of peat may vary significantly with changes in pore pressure, independent of the effect of pore pressure on effective stress, through the dependence of the elasticity of the peat matrix on pore pressure. Therefore, the change in bulk volume of peat with pressure under the condition of constant total stress is

\[
\frac{dV_B}{dP}\bigg|_{\sigma_T} = \frac{d}{dP}\left[\frac{dV_B}{d\sigma'}\right]_{\sigma_T} + \frac{dV_B}{d\sigma'} \frac{d\sigma'}{dP}\bigg|_{\sigma_T} \tag{[4.11]}
\]

The relation between the elasticity of the peat and pore pressure is expected to depend on saturation, as in the case of effective stress.

**General Expression for Specific Storativity**

A general expression for the mean specific storativity in a volume of salt marsh peat follows directly from [3.4], [4.1], [4.4], [4.5], [4.10], and [4.11].

\[
\langle S_s \rangle = \gamma \frac{d}{dP}\left[\frac{dV_B}{d\sigma'}\right] + \frac{dV_B}{d\sigma'} \chi^* - \frac{\partial V_R}{\partial P} + \frac{\partial V_R}{\partial \sigma'} \chi^* + \frac{V_G}{P} + V_Y \frac{dS}{dP} \tag{[4.12]}
\]

This result is based on the assumption that the total stress in the peat is constant, water is incompressible, gas pockets trapped in the peat follow Boyle's law, and changes in void volume due to the compressibility of the peat matrix and roots do not contribute significantly to changes in the volume of the air in the peat. The bracket notation for volume average, which applies to the stress variables, and the Henry's law
correction to gas compressibility have been omitted from [4.12] for the sake of simplicity. Each term in [4.12] is evaluated for the entire volume \( V_B \), and it is assumed the meaningful average values of the stress state variables \( \langle P \rangle \) and \( \langle \sigma' \rangle \) can be defined for \( V_B \). This is the case if \( \langle P \rangle \) and \( \langle \sigma' \rangle \) are considered to be defined only when \( P(\bar{x}) \) and \( \sigma(\bar{x}) \) are statically distributed within \( V_B \), and the derivatives with respect to the stress variables are defined for changes from one equilibrium state to the next. These conditions can be relaxed by assuming that equilibrium conditions occur locally at each point in the peat. The point average specific storativity is defined as

\[
S_s(\bar{x}) = \lim_{V_B \to 0} \frac{\langle S_s \rangle}{S}\tag{4.13}
\]

The point value of specific storativity defined by [4.12] and [4.13] is equivalent to expressions for specific storativity discussed by Narasimhan and Kanehiro (1980) and Brutsaert and El-Kadi (1984) except that in peat there is a pore fluid compressibility that arises from trapped gas pockets and the effects of root volume change are considered. The present treatment of the compressibility of gas pockets follows the development of Verruijt (1969) with the addition of the Henry's law correction term. Although specific storativity can be defined at a point, estimates of specific storativity, such as will be discussed in the next section, are always based on the mean storativity of a finite volume of soil.
The average specific storativity reflects the distribution of the degree of saturation as well as spatial variations of the mechanical properties within the volume of peat. Consider the case where \( V_B \) is taken to be a column of peat extending through the entire depth of a marsh deposit. In the zone of partial saturation, above the water table, the gas in the pores is largely continuous with the atmosphere, \( V_G = 0 \). At low values of saturation bulk volume and root volume become insensitive to changes in pore pressure, and the local average specific storativity simplifies to

\[
\langle S_s \rangle = \frac{\gamma W}{V_B} \left[ V_V \frac{dS}{dP} \right]
\]  \hspace{1cm} [4.14]

Below the water table the peat is fully saturated and [4.12] simplifies to

\[
\langle S_s \rangle = \frac{\gamma W}{V_B} \left[ \frac{d}{dP} \left( \frac{\partial V_B}{\partial \sigma} \right) \sigma' - \frac{dV_B}{d\sigma'} \frac{\partial V_R}{\partial \sigma'} + \frac{V_G}{P} \right]
\]  \hspace{1cm} [4.15]

Below the zone of active roots the expression for mean specific storativity simplifies further.

\[
\langle S_s \rangle = \frac{\gamma W}{V_B} \left[ -\frac{dV_B}{d\sigma'} + \frac{V_G}{P} \right]
\]  \hspace{1cm} [4.16]
Specific Yield versus Specific Storativity

The water storage capacity of the peat column can be described by its specific yield, the rate of change of water content with changes in head per unit area of marsh surface.

\[ S_y = \frac{\gamma_W \frac{dV_W}{dP}}{A} \]  \hspace{1cm} [4.17]

The specific yield of a vertical column of peat, with depth \( D \), is related to its mean specific storativity by

\[ S_y = D \langle S_s \rangle \]  \hspace{1cm} [4.18]

and is also referred to as the storage coefficient of the peat. Both specific storativity and specific yield are valid measures of water storage capacity. The choice of which to use can be made on the basis of convenience or convention. However, it is important to remember that the storage capacity of a peat deposit is not solely determined by the area of the marsh surface. This would be the case if, for instance, the only mechanism for water storage was the change in the degree of saturation of the peat at the water table. This simple model of water storage seems to be the basis of some of the past estimates of the changes in water storage in salt marsh peat (Howarth and Teal 1979, Howarth et al. 1983, Dacey and Howes 1984, Agosta 1985). The result is to underestimate the storage capacity of the marsh sediments by ignoring the storage mechanisms that are active below the water table: root compression, compression of gas pockets, and bulk compressibility of the peat.
V ESTIMATES OF SPECIFIC STORATIVITY

Estimates of the specific storativity of salt marsh peat and of the importance of each of the four storage mechanisms have been obtained from observations of the mechanical behavior of peat, water balance studies, and model fitting. Observations of the mechanical response of the peat to changes in the state of stress are used to estimate the components of specific storativity due to gas compressibility and bulk volume change. To obtain estimates of the total specific storativity based on the water balance, changes in either the water content or pore pressure are imposed in a known volume of peat while observing the corresponding change in the uncontrolled variable. The observed response of the water level in a piezometer following the sudden addition of a volume of water is compared to a model of the response to estimate the hydraulic properties of the peat.

Observations of the Mechanical Behavior of Peat

An accurate estimate of the component of specific storativity due to changes in bulk volume can be obtained for a deposit of peat simply by observing the movement of the marsh surface. The peat expands and contracts in response naturally occurring changes in water content. The resulting vertical displacement of the marsh surface and the corresponding changes in head have been measured in Belle Isle marsh and Sippe-wissett marsh (Nuttle 1986e), Figure 2. Estimate of the specific storativity due to changes in bulk volume are obtained by dividing the slope of data in Figure 2 by the depth of the peat deposits. The specific
storativity is $1.4 \times 10^{-4}$ cm$^{-1}$ for the data shown. The storativity due to bulk volume change at a second site in Sippewissett, at which 2 m of freshwater wetland sediments underlies 2.5 m of salt marsh peat, was found to be $4 \times 10^{-4}$ cm$^{-1}$.

Consolidation Theory

If the classical theory of one dimensional soil consolidation is assumed to describe the behavior of saturated peat then observations of volume changes in response to changes in the total stress on the peat, in the marsh during tidal inundation and in core samples in the laboratory, can be used to estimate the elasticity of the peat matrix and of the pore fluid. The elasticity of the pore fluid is caused by the gas pockets trapped in the peat. This information can then be used to estimate the components of specific storativity due to bulk volume change and gas compressibility.

Consider the case of a soil composed of an incompressible solid phase (no roots) that is saturated except for the presence of trapped gas pockets. Changes in water content of a volume $V_B$ of this soil are given by

$$dV_W = dV_B - dV_G$$  \[5.1\]

and the distribution of total vertical stress is governed by Equation [4.8] with $\chi = 1$. Further, let the soil be confined against horizontal deformation and let the pore pressures be in equilibrium with pressure
at the boundaries so that initially there is no flow of water into or out of the soil. If the soil is suddenly loaded vertically, while holding pore pressures at the boundaries constant, then pore water will be forced out of the soil and the bulk volume will decrease until, after a period of time, the entire increment in stress has been taken up by an increase in the effective stress in the solid matrix. This describes the procedure of the consolidation test, a standard test in the field of soil mechanics.

The transient response of pore pressure, effective stress, and bulk volume to a sudden change in total stress occurs in two stages, Figure 3. In the first stage, which is coincident with the imposition of the load on the soil, the increase in total stress is distributed between the pore pressure and effective stress components. This occurs before any drainage of water can take place, so the constraint on the distribution of the stress increment is, from [5.1],

$$0 = \frac{dV_B}{d\sigma'} \Delta\sigma' - \frac{dV_G}{dP} \Delta P$$

[5.2]

from which it follows that

$$\frac{\Delta P}{\Delta\sigma'} = \frac{dV_B}{d\sigma'} \frac{dP}{dV_G}$$

[5.3]

This constitutes the so-called "undrained" response of a compressible soil.
The change in bulk volume of the ideal soil during the undrained response is equal to the compression of the internal gas spaces, Figure 3. Nuttle (1986e) has observed the deflection of the marsh surface due to loading by flooding tides. By assuming that the surface deflection is entirely due to the compression of gas pockets (i.e., not root compression) and that the entire increment of stress was taken up by pore pressure, the component of specific storativity due to gas compression \( \frac{\gamma_w}{V_B} \frac{dV_G}{dP} \) was estimated to be \( 2.2 \times 10^{-5} \text{ cm}^{-1} \). This corresponds to a gas content of the peat of 1 to 2 percent by volume.

The increase in pore pressures during the undrained response causes a pressure imbalance at the soil boundaries and water drains out of the soil. Water will drain from the soil, causing further decrease in volume beyond the initial compression, until the pore pressures have returned to their original value. At this point the entire increment in total stress has been transferred to the effective stress component, and \( V_G \) has returned to its pre-loaded value. The elasticity of the soil matrix can be calculated from the net change in bulk volume. The consolidation test has been applied to salt marsh peat to estimate compressive storativity, \( \frac{\gamma_w}{V_B} \frac{dV_B}{d\sigma} \) (Nuttle 1982; Knott et al 1984), with the results summarized in Table 1.

Salt marsh peat is highly compressible, two orders of magnitude more compressible than unconsolidated sands, and the component of specific storativity due to this mechanism is on the order of \( 8 \times 10^{-4} \text{ cm}^{-1} \). Table 1. The consolidation test results are similar to results reported
for freshwater peats (MacFarlane 1969, Dasberg and Neuman 1977, Hanrahan 1954). The significance of this comparison is unclear; with the exception of Dasberg and Neuman, the consolidation tests on freshwater peat were done for the purpose of assessing the strength of the peat for engineering purposes. Consequently, changes in stress imposed during the tests were not matched to natural variations in stress that accompany changes in water content.

The storativity derived from consolidation tests are significantly larger than the storativities derived from actual surface displacements, Table 1. Consolidation theory assumes that the components of the soil matrix are incompressible so the application of this theory to calculate specific storativity from the consolidation test results ignores the possible contribution of root compressibility in the observed mechanical behavior of peat. Comparison of the field observations with laboratory test results is further complicated by the fact that the laboratory tests did not duplicate the natural variation of pore pressure and effective stress (total stress was increased while holding pressure constant). Also tests were conducted on core samples that were disturbed to some degree by sampling and storage.

**Lysimeter Water Balance**

The storage properties of salt marsh peat have been investigated through manipulation of the water balance in lysimeters (Dacey and Howes 1984, Nuttle 1986d). A lysimeter is basically a tank containing an undisturbed sample of soil maintained with surface vegetation intact.
This is an important consideration for salt marsh peat given the role of the compressibility of living roots as a mechanism for water storage. The water content of peat in the lysimeters is varied by adding or withdrawing water, and the resulting change in pore pressure is monitored. The total specific storativity can then be calculated.

\[ S_s = \frac{\gamma_w}{V_B} \frac{AV_W}{AP} \]

in which \( V_B \) is the total volume of peat in the lysimeter.

The lysimeter of Dacey and Howes was 40 cm deep and 26.3 cm in diameter (\( V_B = 22 \text{ l} \)), and contained peat from Sippewissett marsh. Nuttle's lysimeter was larger, 52 cm deep and 58 cm in diameter (\( V_B = 137 \text{ l} \)), and contained peat from Belle Isle marsh. The estimates of total specific storativity are similar; \( 8 \times 10^{-4} \text{ cm}^{-1} \) for Dacey and Howes and \( 3-10 \times 10^{-4} \text{ cm}^{-1} \) for Nuttle. Nuttle observed a variation in specific storativity with mean head in the lysimeter, which is to be expected from the influence of the pore size distribution in the relation between pore pressure and the degree of saturation.

**Tension Plate Test**

The tension plate procedure for determining the storage characteristics of a soil is described in general texts on soil physics (e.g., Hillel 1980). Chen (1986) has adapted the procedure for use on salt marsh peat and applied it to peat from Belle Isle marsh. Briefly, the pore pressures in a small sample of peat (5-10 cm deep) are controlled
externally, and the volume of water that drains from the peat following a decrease in pressure is measured.

The result of a typical tension plate experiment on a sample of peat from the surface of Belle Isle marsh is shown in Figure 4. Characteristically, the specific storativity changes at a critical value of pressure head $\psi_{\text{crit}}$ in the range $-10$ to $-20$ cm. The mean values of total specific storativity from five tension plate tests were $8 \times 10^{-4}$ cm$^{-1}$ for $\psi > \psi_{\text{crit}}$ and $20 \times 10^{-4}$ cm$^{-1}$ for $\psi < \psi_{\text{crit}}$. This behavior is related to the pore size distribution, as discussed earlier, and may correspond to the first entry of air into the peat. The difference in specific storativity, $12 \times 10^{-4}$ cm$^{-1}$, can be taken as the magnitude of specific storativity due to changes in saturation. This figure is similar to that found by Dasberg and Neuman (1977) for peat in a drained freshwater wetland, but it is significantly less than the storativity of undisturbed sphagnum moss, found in freshwater bogs (Boelter 1964).

**Contribution of Root Compressibility**

The total specific storativity of Belle Isle peat estimated for the depth interval 0-10 cm by tension plate tests (Chen 1986) is significantly larger than the storativity estimated for depth interval 0-52 cm in lysimeter experiments (Nuttle 1986d), when the comparison is made at similar pore pressures in the peat. The principal reason for this is that specific storativity is not uniform throughout a peat deposit. The storativity measured in the tension plate tests is affected by the pore structure right at the marsh surface and includes the effect of desatu-
ration of the sample. In the lysimeter, desaturation occurs only near the surface of peat. The specific storativity in the surface layer is averaged in with the contributions of other mechanisms acting throughout the 52 cm sample, which results in a lower mean value of storativity.

This situation can be used to estimate the specific storativity of the peat in the lysimeter by mechanisms other than the desaturation of the surface layer. The peat in the lysimeter is conceptually divided into two layers at the water table, the depth at which \( \psi = 0 \). The peat in the lower layer is saturated \( (\psi > 0) \), and the specific storativity of the upper layer is taken to be the specific storativity determined by the tension plate tests. The specific storativity estimated by the lysimeter tests is the depth-weighted mean specific storativity of the two layers together, so the specific storativity of the saturated layer can be calculated by difference. Nuttle (1986d) finds this value to be \( 4 \times 10^{-4} \text{ cm}^{-1} \).

The specific storativity of saturated peat peat layer in the lysimeter is the sum of storativities due to the mechanisms of bulk compressibility, gas compression, and root compression. Nuttle (1986d) has determined that the storativity in the lysimeter sample due to bulk volume change is on the order of \( 3 \times 10^{-6} \text{ cm}^{-1} \), by monitoring movements of the surface of the peat. The storativity due to gas compression in Sippewissett marsh was also determined to be small (Nuttle 1986e), as reported above. All of this suggests that the compression of roots is the mechanism responsible for most of the specific storativity in the lysimeter not associated with desaturation \( (4 \times 10^{-4} \text{ cm}^{-1}) \). Given the nature of this result, the actual value of storativity due to root
compression is not as significant as finding that it is comparable to the other mechanisms of water storage.

**Slug Test**

The third general approach to estimating specific storativity of a soil is based on the recovery of the level of the water surface in a piezometer when perturbed by the addition or removal of a volume of water. The hydraulic parameters of the soil are estimated by comparing a record of water level versus time during the recovery to the theoretical response predicted by a model of water flow in the region of the piezometer. This is a well-established technique for estimating the hydraulic conductivity of conventional soils that Nichols (1985) has adapted for use in salt marshes. The large compressive storativity of salt marsh peat makes it possible to estimate both the hydraulic conductivity and the specific storativity of the peat from the piezometer response. The technique is limited to application in regions deep in the peat, away from the marsh surface where desaturation is important to the total storativity.

The slug tests were carried out at depths below the root zone, where only the specific storativity due to bulk volume change and gas compression would be active. However, the slug test estimates of specific storativity, Table 1, are similar to the estimates based on the consolidation tests and are higher than the estimates of storativity based on vertical displacements of the marsh surface, which are believed to be more accurate. The reason for this is unknown.
Distribution of Water Storage

Changes in the water content of peat occur throughout the depth of a peat deposit. In recognition of this fact, the discussion of water storage has been presented here in terms of specific storativity, which describes the storage capacity per unit volume of the peat. The specific storativity is not uniform through the depth of a peat deposit. Figure 5. Changes in the degree of saturation is the principal mechanism of water storage in the upper 10-20 cm of peat, when pore pressures drop below the air entry threshold. Root compressibility determines the amount of water storage in the saturated region of the root zone and bulk volume change is important at depth. The specific storativity due to changes in saturation may be five to ten times greater than specific storativity due to root compression and bulk volume change, but changes in the volume of water stored below the water table by the latter two mechanisms can be as large as that above the water table because of the larger volume of peat below the water table.

VI ESTIMATES OF HYDRAULIC CONDUCTIVITY

Methods of estimating hydraulic conductivity can be divided into two general classes, those based on steady or uniform flow conditions and those based on the interpretation of transient head response to perturbations in water content. Field and laboratory methods that have been applied to salt marsh peat are included in both classes.

In the first class of methods, a steady flow of water is established in the region of a well, if in the field, or in a core sample, if
in the lab, and the corresponding head drop is observed. The falling head permeameter method is a variation on the laboratory method in which both the head drop and the flow rate vary continuously during the experiment but flow conditions in the sample are uniform at any given instant. The hydraulic conductivity is determined from the geometry of the flow field, and the observed relation of flow rate to head drop.

In the second class of methods, the hydraulic head in the peat is perturbed and the observed recovery to equilibrium conditions is matched to a theoretical head response based on a model of the flow in the soil. Both the storativity and the hydraulic conductivity of the soil can be determined by the slug test if the soil is highly compressible (see above). In another method, Nettle (1986d) has used head transients observed in a lysimeter following the addition and infiltration of a volume of water to estimate specific storativity and hydraulic conductivity under partially saturated conditions.

Estimates from Core Samples

Extensive coring and testing programs have been carried out to characterize the saturated hydraulic conductivity of peat in Belle Isle marsh and in Sippewissett marsh. The falling head permeameter method was used to determine the hydraulic conductivity of 65 samples of Belle Isle peat (Chen 1986) and 52 samples of Sippewissett peat (Fifield 1981, Knott et al 1984). The hydraulic conductivity of the root zone is on the order of $5 \times 10^{-4}$ cm/s, similar to the hydraulic conductivity of very fine sand or silt. Values of hydraulic conductivity are highly
variable. Estimates of hydraulic conductivity of the core samples were found to be lognormally distributed in each of the marshes.

There is a systematic variation in hydraulic conductivity with depth in both marshes, but the variation is different for each, Table 2. In Belle Isle marsh, the hydraulic conductivity is highest in the upper 70 cm of peat and decreases toward the hydraulic conductivity of the underlying clay ($10^{-7}$ cm s$^{-1}$) with increasing depth. In Sippewissett marsh, the hydraulic conductivity is generally higher at mid-depth than in either the root zone or the highly humified peat and silt layer just above the underlying sand. By comparison, the hydraulic conductivity of undisturbed peat in freshwater bogs ranges from values near 1 cm/s at the surface to very low values ($< 10^{-7}$ cm/s) at depth (Ingram 1983). The composition of the peat in the two marshes studied is different. The peat in Belle Isle marsh is largely silt and clay, 90% ash-free dry weight, and the peat in Sippewissett marsh is "peatier", 70% ash-free dry weight. It is remarkable that the hydraulic conductivities of the root zone are virtually identical in the two marshes.

**In situ Estimates of Hydraulic Conductivity**

The permeameter test is susceptible to errors caused by disturbance of the peat during coring and storage and to errors caused by the flow of water between the core tube and the relatively impermeable peat. There are also some regions in the peat, particularly at depth, that cannot be sampled by coring due to the consistency of the peat. Techniques for estimating hydraulic conductivity in the field are easy to
apply and serve as a check on the more extensive laboratory results. The disturbance of the peat is minimized by testing in situ; however head gradients in excess of natural conditions are often required, which can result in disturbance of the peat structure (Rycroft et al. 1975).

In situ estimates of hydraulic conductivity have been made in both Belle Isle and Sippewissett marshes, Table 2. Two field methods have been employed. A steady state pumped well method was used by Fifield (1981) in Sippewissett marsh, and the slug test procedure developed by Nichols (1985) was used in both Belle Isle and Sippewissett marsh. The in situ estimates are in good agreement with the permeameter results in Sippewissett marsh, but the slug test results are less than the permeameter results at comparable depths in Belle Isle marsh. This is thought to be the result of wall flow in the permeameters.

**Anisotropy**

Water in salt marsh peat flows preferentially along pathways left by dead and decaying roots (Chen 1986). If roots run predominantly horizontally or vertically then anisotropy in hydraulic conductivity will result. Permeameter tests measure the hydraulic conductivity of vertical flow in the peat and the in situ methods measure a directionally averaged hydraulic conductivity, weighted to the horizontal conductivity by the vertical orientation of the well screen. A comparison between the hydraulic conductivities determined by the two methods would reveal extreme anisotropic behavior if it is present, but it is not sensitive to small anisotropy ratios because of the variability of the
hydraulic conductivity estimates obtained by any one method. Fifield (1981) and Knott et al. (1984) have compared permeameter data with in situ measurements of hydraulic conductivity made nearby in Sippewissett marsh. A plot of laboratory versus in situ estimates of hydraulic conductivity is shown in Figure 6. There are no systematic differences in the two estimates except at lower hydraulic conductivities. If the in situ technique averages conductivity by weighting proportionally to the flow rate of the material then it would be relatively insensitive to small lenses of low conductivity material, which is one explanation for the results in Figure 6. The results of this comparison, taken along with the general agreement between the permeameter and in situ estimates in Table 2, are evidence that salt marsh peat is not highly anisotropic. Similar results have been reported for freshwater peats (Rycroft et al. 1975a).

Estimates from a Lysimeter

In situ techniques based on pumped or transient well response are unable to estimate the hydraulic conductivity of the near-surface peat which may control the process of infiltration. The hydrologic properties of peat in the root zone are influenced by the flora and fauna that reside there and are therefore more susceptible to disturbance by conventional handling techniques. A procedure for determining hydraulic conductivity from transient head response in a lysimeter has been developed by Nuttle (1986d) using on peat from Belle Isle marsh. Use of a lysimeter maintains natural conditions with respect to the peat structure and head gradients, and it minimizes the effects of sample disturb-
ance because a large volume of peat is used. The method results in estimates of the bulk average hydraulic conductivity of the peat in the lysimeter for saturated and partially saturated conditions.

Results from the lysimeter tests show that the average saturated hydraulic conductivity of Belle Isle peat in the depth interval 0-52 cm is $5.0 \times 10^{-4}$ cm/s, exactly the result obtained from the permeameter tests on the root zone of Belle Isle. Values of the mean hydraulic conductivity decrease exponentially with increasing depth to the water table, down to a value of $3.0 \times 10^{-6}$ cm/s for a depth to the water table of 36 cm. The mean conductivity decreases due to desaturation of the peat near the surface because desaturation reduces the proportion of the pore volume active in the conduction of water.

VII DISCUSSION

Delayed Head Response

Delayed response in pore pressures has been observed in response to evapotranspiration (Dacey and Howes 1984) and in response to tidal flooding (Hemond et al. 1984, Agosta 1985, Carr and Blackley 1986). The effect of changes on a boundary must "diffuse" into the peat. For instance, if an evaporative flux is suddenly imposed, then the head at the boundary must decrease through the loss of water from the peat near the surface first, before water will be removed from depth. Thus the movement of water and the decline in head in the interior of the peat deposit are lagged relative to the start of evaporation.
This behavior is intrinsic to the relation of mass conservation in the peat. If the specific storativity and hydraulic conductivity are uniform and the hydraulic conductivity is isotropic, then [3.5] can be written in differential form

\[ \frac{\partial h}{\partial t} = \frac{K}{S_s} v^2 h \]  

in which the principle of conservation of mass is applied at a point and hydraulic head is a field variable, \( h(x,t) \). The ratio of hydraulic conductivity to specific storativity plays an important role in determining the temporal and spatial scales of the response of head to transient conditions in the peat and often will be considered as a separate parameter, \( \alpha = K/S_s \). The value of \( \alpha \) is in the range 0.1-1.0 cm²/s for peat in the two salt marshes studied here.

The general form of [7.1] is one that is frequently encountered in other contexts (e.g., heat diffusion). Solutions presented in Carslaw and Jaeger (1959) can be used to construct models of head response for some simple boundary conditions that are useful analogs to hydrologic conditions in a marsh. Refinements to account for variation in hydraulic properties of the peat and complicated boundary conditions can be implemented in numerical solutions, e.g., Hemond and Fifield (1982), Nuttle (1986d).

For example, the initial and boundary conditions for the sudden imposition of a constant evaporative flux, \( E \), from the surface of a peat
deposit underlain by an impermeable material are shown in Figure 7 (inset).

\[ t < 0 \quad -D < z < 0 \quad h(z) = h_i \]

\[ t > 0 \quad z = 0 \quad \frac{\partial h}{\partial z} = -\frac{E}{K} \]

\[ z = -D \quad \frac{\partial h}{\partial z} = 0 \] [7.2]

The solution for the rate of change of head in the peat is

\[ \frac{\partial}{\partial t} h(z,t) = \frac{-E}{S_D} \left[ 1 + 2 \sum_{n=1}^{\infty} e^{-n^2 \pi^2 t / t^*} \cos \frac{n\pi z}{D} \right] \] [7.3]

in which \( t^* = D^2 / \pi \alpha \).

The solution is the sum of two parts, a steady decline in head

\[ -\frac{E}{S_D} \] in which the rate of change of mean head is simply the ratio of the rate of evaporation and the specific yield of the peat deposit. The transient term, which is responsible for the lagged head response at depth, has a time scale that is determined by \( \alpha \) and the depth of peat. Typical values of \( t^* \) are 0.28–2.8 hours for a peat deposit that is 1 m deep and the range of \( \alpha \) discussed above.

Dacey and Howes (1984) observe a lag in the response of pore pressure to evapotranspiration in a lysimeter, which they attribute entirely to delayed uptake by the plants. Plants are known to respond to transpiration water demand in part by reducing the water content of their
tissues, so the same type of interaction between water storage and water movement that is responsible for delay in the head response in peat will cause a delay in water uptake by roots in response to transpiration at the leaf surface (Molz 1981).

Another situation of interest is the head response in a semi-infinite deposit of peat \(0 \leq x \leq \infty\) to a step change of head on the boundary. This constitutes a first-order model for infiltration during tidal inundation (Hemond et al. 1984), \(x\) positive downward into the peat, and for drainage horizontally into the creeks during a period of neap tides (Nuttle 1986c), in which \(x\) is the distance away from the creek bank. The solution of [7.1] for this case is

\[
\frac{h(x,t) - h_0}{h_0 - h_1} = \text{ERFC}\left(\frac{x}{2\sqrt{at}}\right) \tag{7.4}
\]

in which \(h_1\) is the initial, uniform value of hydraulic head and \(h_0\) is the head at the boundary \(x = 0\). The complementary error function \(\text{ERFC}(\xi)\) has the value 1.0 for \(\xi = 0\) and decreases to zero as \(\xi\) increases; \(\text{ERFC}(1.0) = 0.157, \text{ERFC}(2.0) = 0.004\). The water flux across the peat surface can be calculated from [7.4] by applying Darcy's law

\[
F(t)\bigg|_{x=0} = -K \frac{h_1 - h_0}{\sqrt{\pi at}} \tag{7.5}
\]

The dimensionless group \(x/(2\sqrt{at})\) governs the head response. By imposing the condition
\[
\frac{x}{2\sqrt{\alpha t}} \sim 1 \quad \text{[7.6]}
\]

one can estimate the time lag of the head response at a particular location in the peat, \( x \) known, or one can estimate the length over which head varies from \( h_1 \) to \( h_0 \) in the sediment at a particular point in time.

A third solution that has application in salt marshes is the response of head in a semi-infinite peat deposit to a sinusoidal variation in head on the boundary, i.e., the effect of tides. The solution for this case has been discussed by Werner and Noren (1951) and by Tison (1965) and can be found in Carslaw and Jaeger (1959);

\[
h(x,t) = h_o e^{-x/x^*} \cos(\omega t - x/x^*) \quad \text{[7.7]}
\]

Here \( h_o \) is the amplitude of the depth averaged head at the creek bank, \( x^* = \frac{2\alpha}{\sqrt{\omega}} \) is the wavelength of the head fluctuation in the peat, and \( x \) is the distance from the creek bank. The water fluxes in the peat due to the effect of tides are

\[
F(x,t) = \frac{-\sqrt{2}}{x^*} K_h h_o e^{-x/x^*} \cos(\omega t - x/x^* + \pi/4) \quad \text{[7.8]}
\]

Two features of this response are notable. First, peak water fluxes across the creek bank occur before the highest and lowest excursions of the water level in the creek. Second, both the amplitude of the head fluctuations and of the water fluxes decay exponentially with distance from the creek. For a tidal period of 12.4 hours, typical wavelengths
in the peat are 0.4–1.2 m. This implies that virtually all of the
effect of tidal fluctuations in the creeks is limited to a narrow region
(1–3 m) of peat near the creek bank. The effects of vertical water
movement induced by flooding and infiltration or by water fluxes across
the bottom boundary of a peat deposit are not included in this analysis
and may be responsible for periodic fluctuations in hydraulic head
farther away from the creek bank.

Interaction between Ecology and Hydrology

The structure of salt marsh peat is controlled by a dynamic equi-
librium among physical and biological processes, e.g., sedimentation,
autocompaction, growth of roots and rhizomes, and their death and decay.
Chen (1986) notes that water flows preferentially along channels formed
by roots. The hydraulic conductivity of the root zone peat from Belle
Isle and Sippewissett marshes have been found to be virtually identical,
and the peat in the Ebbon Creek marsh in Gloucester, Massachusetts, has
similar properties (Fifield 1981, Knott et al. 1984), which suggests
that the hydraulic properties are more dependent on the presence of the
marsh grass than on other factors in the composition of the peat. The
compressibility of the roots contributes to the dynamic storage of water
in the peat, and channels left by dead roots are likely sites of air
entry, another factor in water storage.

The hydraulic properties of the peat govern the movement of pore
water in response to hydrologic forcing through weather and tides. The
movement of pore water controls air entry and the solute balance in the
peat. This affects the productivity of the marsh grasses and other biological processes (Howes et al. 1981, King et al. 1982) which, in turn, affect the hydraulic properties of the peat. Therefore, a set of interrelationships exists through which a perturbation in the biological component of the peat can affect pore water hydrology. The high productivity of the marsh grasses, especially the high turnover rates in the root mass (Valiela et al. 1976) provides for rapid feedback between biology and hydrology, so a close coupling of the two is possible. Several authors (Howes et al. 1981, Hemond and Fifield 1982, Dacey and Howes 1984) have speculated that productivity, air entry, and the oxidation state in the sediment are linked in this manner.

It has been suggested by Nuttle (1986e) that the formation of unvegetated pannes in older sections of many marshes is another manifestation of the interaction of hydrology and ecology. As the gradual accumulation of sediment increases the depth of peat at a site, the storage capacity of the deposit due to bulk compressibility and gas compression increases, increasing the total storage capacity. Changes in water content are imposed externally by the balance between evaporation, precipitation, and the infiltration of tidal water (Nuttle 1986b), and the result of the increase in the storage capacity is to decrease the range of variation of pore pressure and the extent of air entry into the peat. This imposes a chronic stress on the vegetation, eventually leading to a loss in vegetation. With no root channels in the peat, aeration of the sediment becomes even more difficult and less frequent, and grasses are discouraged from recolonizing areas that have lost vegetation.
REFERENCES


NUTTLE, W. K. 1986e. Peat rose!


Table 1

Estimates of Specific Storativity
$(10^{-4} \text{ cm}^{-1})$

<table>
<thead>
<tr>
<th>Method</th>
<th>Belle Isle</th>
<th>Sippewissett</th>
</tr>
</thead>
<tbody>
<tr>
<td>Consolidation Test; Chen 1986, Knot et al. 1985</td>
<td>6.5 (1.7)</td>
<td>9.1 (2)</td>
</tr>
<tr>
<td>Gas Compression; Nuttle (1986a)</td>
<td>--</td>
<td>0.22</td>
</tr>
<tr>
<td>Compressive Storativity; Nuttle 1986e</td>
<td>1.1-1.6</td>
<td>1.4, 4.1</td>
</tr>
<tr>
<td>Total Storativity from Lysimeter; Nuttle 1986d</td>
<td>3.2-10.0</td>
<td>8.0</td>
</tr>
<tr>
<td>Total Storativity from Tension Plate Tests; Chen 1986</td>
<td>7.9 (1.2) $\psi &gt; \psi_{\text{crit}}$</td>
<td>--</td>
</tr>
<tr>
<td>Well Test; Nichols 1985</td>
<td>3.4</td>
<td>9.2</td>
</tr>
</tbody>
</table>

`1` Values of storativity are the geometric mean where more than one estimate was obtained. The numbers in parentheses are multiplicative factors that, when applied to the mean, define the range of storativities corresponding to ±1 standard deviation of the log transformed $S_s$, which is normally distributed. Example: mean = 6.5; range $= \left[ \frac{6.5}{1.7} \right]$, $1.7 \times 6.5$. 

- 70 -
Table 2

Estimates of Saturated Hydraulic Conductivity ($10^{-4}$ cm/s)

<table>
<thead>
<tr>
<th>Method</th>
<th>Belle Isle</th>
<th>Sippewissett</th>
</tr>
</thead>
<tbody>
<tr>
<td>All samples</td>
<td>1.7 (6.6)$^1$</td>
<td>12 (10)</td>
</tr>
<tr>
<td>Root zone (0–40 cm)</td>
<td>5.7 (2.3)</td>
<td>3.1 (6.7)</td>
</tr>
<tr>
<td>Non-root zone</td>
<td>1.2 (5.0)</td>
<td>18 (10)</td>
</tr>
<tr>
<td>Slug test; Nichols (1985)</td>
<td>0.092 (2.0)</td>
<td>26</td>
</tr>
<tr>
<td>100 cm depth</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pumped well; Fifield (1981)</td>
<td></td>
<td>1.7 (6.3)</td>
</tr>
<tr>
<td>Lysimeter; Nuttle (1986d)</td>
<td>5.0</td>
<td></td>
</tr>
<tr>
<td>0–52 cm</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

$^1$ Values of conductivity are the geometric mean where more than one estimate was obtained. The numbers in parentheses are multiplicative factors that, when applied to the mean, define the range of conductivities corresponding to $1$ standard deviation of the log transformed $K$, which is normally distributed. Example: mean $K = 1.7$; range = $\left[ \frac{1.7}{6.6}, 6.6 \times 1.7 \right]$. 
The components of peat are shown schematically. Air, water and gas are contained in a porous structure composed of roots and rhizomes and a solid phase of lifeless organic and mineral material. Air in the peat is in equilibrium with the atmosphere. Gas pockets in the peat are isolated from the atmosphere and are in pressure equilibrium with the pore water. The bulk volume of peat and the volumes of the roots, gas pockets, and air vary independently. The volume of the solid phase, excluding the volume of the roots, is constant.
The elevation of the marsh surface varies linearly with depth-averaged head in two marshes: Sippewissett marsh, August 1984 (■); and Belle Isle marsh, August (●) and September (▲) 1984. The peat depths are 1 m, Sippewissett, and 1.7 m, Belle Isle. The compressive storativities are $1.4 \times 10^{-4}$ cm$^{-1}$, Sippewissett, and $1.6 \times 10^{-4}$ cm$^{-1}$, Belle Isle.
Figure 3

The response of bulk volume, pore pressure, and effective stress in a simple saturated soil to a step change in uniform vertical stress occurs in two stages. In the first stage, which occurs immediately, the water content remains constant and the stress increment is distributed between pore pressure and effective stress in inverse proportion to the compressibilities of the solid matrix and the gas pockets. The initial change in bulk volume $\Delta V_B^{(1)}$ is equal to the change in volume of the internal gas spaces $\Delta V_G$. If pore pressures are held constant at the soil boundaries then the initial "undrained" response is followed by a gradual loss of water from the soil and a redistribution of stress until the entire change in stress has been added to the effective stress component. The ultimate change in bulk volume is equal to the change in water content, $\Delta V_W$. 
UNDRAINED RESPONSE

TOTAL STRESS

EFFECTIVE STRESS

PORE PRESSURE

\[ \Delta \sigma_T \]

\[ \Delta V_B^{(i)} \frac{d\sigma'}{dV_B} \]

\[ \Delta V_B^{(i)} \frac{dP}{dV_G} \]

\[ \Delta V_B^{(i)} = \Delta V_G \]

BULK VOLUME

\[ \Delta V_W \]

TIME
Cumulative water loss during a tension plate experiment on an 8 cm deep surface sample of Belle Isle peat. Pressure head is measured at the bottom of the sample. The rate of change of water content with pressure head in the specific storativity. The change in storativity at \( \psi_{\text{crit}} \) is characteristic of all tests.
The mechanisms of water storage are distributed throughout the depth of a peat deposit. Specific storativity is dominated by three storage mechanisms, each over a different interval of depth. Changes in the degree of saturation dominate in the upper 10-20 cm. Root compression is the principal storage mechanism in the saturated region of the root zone, and bulk compressibility is the principal storage mechanism below the root zone. A fourth mechanism, compression of gas pockets, contributes slightly to the total water storage in peat.
AIR ENTRY

P < ATM
P = ATM
P > ATM

LIMIT OF SATURATION

DEPTH OF WATER TABLE

30-40 cm

DEPTH BELOW SURFACE

SPECIFIC STORATIVITY
$10^{-4} \times \text{cm}^{-1}$

$\Delta$ SATURATION
10-20

ROOT ZONE

ROOT COMPressibiliTY
~4.0

$\Delta$ BULK VOLUME
1-4

GAS COMPressibiliTY
0.1-0.2
Figure 6

Horizontal (seepage tube) and vertical (permeameter) hydraulic conductivities of peat in the same location in Sippewissett marsh are similar in magnitude.
The transient head response to a suddenly imposed evaporative flux at the surface depends on the ratio of the hydraulic properties $\alpha = K/S_s$. The head distribution is initially uniform and there is no flow of water across the bottom boundary. The quantity $-E/S_s D$ is the steady state rate of change of head in the peat. The rate of change of head is uniform through the depth of peat only after a characteristic time $4t^\star$ following the start of evaporation, $t^\star = D^2/\pi^2 \alpha$. 
ELEMENTS OF SALT MARSH HYDROLOGY:

II. THE WATER BALANCE IN BELLE ISLE MARSH

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May 1986
ABSTRACT

Fluctuations in hydraulic head observed in a salt marsh over a two-month period are correlated with hydrologic boundary conditions to describe the water budget of the sediment. Horizontal drainage of pore water into a creek does not have a detectable effect on head at distances greater than 10 m from the creek. Beyond this distance the water content of the sediment varies cyclically in response to water loss by evapotranspiration and replenishment by precipitation and infiltration during flooding of the marsh surface on the highest high tides. The lack of significant drainage from the interior regions of a salt marsh, other than by evapotranspiration, has implications for the concentration of conservative solutes in the sediment.
I. INTRODUCTION

The study of the ecology of salt marshes and the biogeochemistry of their sediments has directed interest to two aspects of the water balance of the sediments, lateral drainage of water through the sediments to creeks and advection of air into sediments, which is coupled to changes in water content. The former is of interest with respect to the export of reduced sulfur compounds (Howarth and Teal 1980, Howarth et al. 1983, Jordan and Correll 1985) and to myriad factors bearing on the differentiation of the tall and short growth forms of Spartina alterniflora (King et al. 1982, Mendelssohn and Seneca 1980). The latter is a factor in the productivity of marsh vegetation (Howes et al. 1981, Mendelssohn et al., 1981, Dacey and Howes 1984).

Horizontal drainage controls the concentration of solutes, and also can have an effect on the oxidation state in the sediment. Drainage increases fluctuations in sediment water content, and therefore the probability of air entry, relative to that due to evapotranspiration alone, and the loss of sulfides by drainage improves the oxidation state of the sediment by removing below-ground oxygen demand. There is ample evidence to suggest that pore water in interior regions of marsh sediments, away from creeks, is stagnant while pore water in creek bank regions is relatively active (Gardner 1973, Nestler 1977). Jordan and Correll (1985) have estimated water fluxes in the sediment draining towards a creek. However, until now, there has not been a study of the complete water budget in the sediment.
the rates of the principal water fluxes, and the scales over which they vary.

This paper addresses these questions by analyzing the temporal and spatial variation of hydraulic head in a New England salt marsh observed over a two-month period. Nuttle (1986a) has shown that storage and conductance of water by salt marsh sediments are directly related to the hydraulic head in the sediment. Fluctuations in hydraulic head are proportional to fluctuations in the water content of the sediments. The effect of each component of the water balance, evapotranspiration, precipitation, infiltration of tidal water, and drainage, on the water content of the sediment adds to the effects of the other components to produce the observed time course of hydraulic head.

Here, the water balance is considered in three parts; the net atmospheric water flux (precipitation less evapotranspiration), infiltration of tidal water, and horizontal drainage through the sediment into a creek. Observed head is correlated first with the combined effects of net atmospheric flux and infiltration. This is sufficient to account for 96% of the variation in head observed at distances greater than 10 m from the creek and 75% of the head variation within 10 m of the creek. The residual head variation within 10 m of the creek is related, in part, to the net horizontal water flux out of this region across the creek bank. The hydraulic properties of the marsh sediments are estimated from the correlation of head to the various hydrologic boundary conditions. Estimates of pore water velocities are obtained from observed head gradients.
II. MATERIALS AND METHODS

Study Site

The Belle Isle marsh is the last remnant (100 hectares) of the once-extensive marshes in Boston, Massachusetts. The study site is located 3 km northeast of Logan International airport, Figure 1. The marsh sediments consist of a layer of clayey peat (90% ash free, Chen 1986) 70 cm deep which grades into a layer of gray silty clay. A layer of relatively impermeable clay underlies the study area at 170 cm. The vegetation over most of the study area is a mix of Distichlis spicata and Spartina patens typical of drier high marsh conditions. The hydrologic regime is characterized by alternating periods of daily flooding of the marsh surface and periods in which the surface does not flood for several days. The average tidal range in the creeks is 2 m. The periods of surface flooding are related to the spring-neap cycle but do not follow a regular 18-day cycle.

Pressure Measurement

An array of 18 piezometers was installed for the purpose of long-term observation of the distribution of hydraulic head in the sediments with respect to a creek. All of the piezometers shown in Figure 1 were installed to a depth of 50 cm. Additional piezometers were installed to depths of 100 and 150 cm, adjacent to piezometers 1 and 4, and at 150 cm, adjacent to piezometer 6. The staggered arrangement of the piezometer array allows for the estimation of head gradients both parallel and perpendicular to the creek.
The piezometers were constructed of 0.5 inch (nominal) PVC pipe slotted over 15 cm at the lower end. These were installed in cavities augered with a 2.54 cm soil auger and backfilled with gravel over the slotted interval, then bentonite to the surface. The piezometers were protected above ground by a casing of 1.25 inch (nominal) PVC pipe, capped and vented in such a way that the piezometer tubes would not be filled by flooding tides. The outer casing also protected the piezometers from disturbance by ice and debris rafted on the flooding tides.

Water was pumped out of the piezometers to clear the tube and the gravel pack of fines and thus "develop" the well. The time constant for recovery of the water levels following pumping was on the order of two hours. This is too long to be accounted for by the hydraulic conductivity of the sediments at the 50 cm depth (Chen 1986) so it seems likely that disturbance of the sediment during installation and development resulted in partial clogging of the piezometer screen. This suspicion is corroborated by the observation during core sampling of a highly fluid layer of silt at the 50 cm depth.

The 2-hour lag in piezometer response is long with respect to the semi-diurnal tides but it is small with respect to day-to-day variations in head, which are of interest here. The piezometers were not read directly following a flooding tide to avoid errors introduced by the piezometer recovery from the static pore pressure response to the weight of water ponded on the surface during flooding (Hemond et al. 1984).
SOME TEXT ON THE FOLLOWING PAGE(S) IS ILLEGIBLE ON THE ORIGINAL MATERIAL.
Elevations of the piezometers and the surrounding marsh were determined relative to a common datum, approximately mean low water in Boston Harbor, with an accuracy of 1 cm. Elevations of the piezometer tubes were checked from time to time as the expansion and compression of the sediment tends to work the piezometer tubes out of the sediment. This was not a problem during the period of head observation reported here. Water levels in the piezometers were read to 0.5 cm by lowering an electrode-tipped dowel into the tube. Contact with the water in the piezometer causes a change in the resistance between the electrode and the sediment.

Observations of hydraulic head were made approximately every other day during the two-month period August-September 1984. More frequent observations were made to capture the head response to precipitation events and the resumption of daily flooding following a dry period. A total of 40 head measurements were made at each piezometer in the 60-day period.

Evaporation and Precipitation

Evaporation and precipitation during the period of head observation were estimated from meteorologic data from the National Weather Service observatory at Logan Airport, 3 km southwest of the study site. Estimates of evaporation are based on an estimated heat budget at the marsh surface. Heat budget terms were estimated from observations of air temperature, humidity and pressure and observed cloud cover by the methods described below. Measurements of local temperature and relative humidity were made at the study site during visits to read the piezometers. Air temperatures were always within 2.5°C of the airport record and were generally closer
than 1°C. Differences in humidity were on the order of 5% relative humidity.

Evapotranspiration was estimated by the method of Priestly and Taylor (1972). The rate of evapotranspiration, in units of langleyes per day, is given by the relation

\[ q_E = \alpha_E \frac{A}{A+\gamma} (R_N - G) \]  \[2.1\]

in which \( A \) is the rate of change of saturation vapor pressure with air temperature at a temperature typical of the air near the ground and \( \gamma \) is the ratio of the sensible heat content of air to the latent heat content of water vapor in vapor saturated air, also known as the psychrometric constant. \( R_N \) is the net radiation flux to the surface, \( G \) is the soil heat flux and \( \alpha_E \) is an empirical constant established to be 1.26 by Priestly and Taylor. This method has been reviewed by Brutsaert (1982) and found to perform well for well-watered grassed surfaces. A similar expression has been found to apply to freshwater bogs (Romanov 1968).

Net incident radiation at the marsh surface is the sum of incoming shortwave solar radiation and longwave radiation from the atmosphere, less reflected shortwave and black body radiation from the surface. Incident shortwave radiation is a function of latitude, season, time of the day and cloud cover. Clear sky radiation was estimated by the procedure outlined by Curtis and Eagleson (1982) with the cloud cover correction from TVA (1972). Reflected shortwave radiation was calculated by assuming an albedo of 0.20, which is typical for grassed surfaces (Brutsaert 1982).
Longwave radiation follows the Stephen-Boltzman relation

\[ R_{\text{LW}} = \varepsilon\sigma T^4 \]  

where \( \sigma \) is a constant with the value \( 0.826 \times 10^{-10} \text{ ly min}^{-1} \text{ °K}^{-1} \) and \( \varepsilon \) and \( T \) are the emissivity and the absolute temperature of the radiation source, respectively. Swinbank's (1963) relation for the emissivity of the atmosphere and the cloudy sky correction from TVA (1972) have been used here to calculate incident longwave radiation. An emissivity of 1.0 was assumed for calculation of black body radiation from the marsh surface. The air temperature was used for the calculation of both incoming and outgoing longwave radiation fluxes.

III RESULTS

Pressure Measurements

The effects of three factors bearing on the variation of hydraulic head are immediately evident in Figure 2, in which depth-averaged head at piezometers 1, 4, and 6 is shown for the entire period of observation. First is the high degree of correlation of head fluctuations with hydrologic conditions at the marsh surface. Hydraulic head declines during non-flooding periods when there is a net loss of water from the sediment due to evapotranspiration. Recovery occurs in response to precipitation and to the resumption of daily flooding. Long periods of daily flooding maintain the hydraulic head at levels near the marsh surface. The second effect is the result of the difference in surface elevation at the piezometers, which
is responsible for the offset between the head at piezometers 1 and 4. Finally, hydraulic head at piezometer 6 shows the influence of vertical forcing from the surface, surface elevation effects, and and additional influence introduced by its proximity to the creek.

Components of the Heat Budget

The estimated heat budgets for August and September 1984 are summarized in Table 1. The soil heat flux was calculated from changes in the heat content of the sediment estimated from four temperature profiles taken over the depth 0–260 cm during the period 217 to 263, Julian date. The energy consumed by photosynthesis was not included in the heat budget estimates. Teal (1962) estimates this component to be 6% of the incident solar radiation during the height of the growing season. While this figure is larger than the soil heat flux, it is negligible compared to errors in the radiation flux estimates. The radiation fluxes and rates of evapotranspiration were estimated every three hours, corresponding to the frequency of data available from the airport observatory record. The quantity \( \frac{A}{A+\gamma} \) is a function of temperature and pressure and was calculated from empirical formulae.

Evapotranspiration was arbitrarily set to zero for periods in which either precipitation was recorded or there was a net loss of heat by radiation. The Priestly–Taylor evaporation formula implicitly assumes conditions of net influx of heat by radiation balanced by an efflux of heat by latent and sensible heat fluxes and is not applicable to periods of radia-
tion cooling. These conditions typically occur in the evening and early morning when evaporative fluxes are small.

The largest sources of error in the estimate of evapotranspiration are related to in the estimates of the radiation fluxes. Estimates of daily incident longwave and shortwave radiation from the airport weather record were compared to values measured at MIT in Cambridge, Massachusetts (6 km from the airport), during November 1984 (E. Adams, unpublished data). The estimated values showed both random error and bias; daily shortwave bias = 34 ly, error = 29 ly; daily longwave bias = -31.3 ly, error = 24 ly. The mean daily totals of recorded incident radiation were 165 ly, shortwave, and 551 ly, longwave. The use of air temperature in the computation of longwave radiation from the marsh surface under-predicts this flux during the daytime when surface temperatures are higher than air temperatures. Teal and Kanwisher (1970) report that the temperature of Spartina leaves can be 3°C above the temperature of the air. If this temperature difference is taken as typical for the marsh surface then the error in estimated outgoing longwave radiation is about 4% of the total. If the errors in estimates of incident radiation are assumed to vary with the total radiation flux then the bias in the estimation of evapotranspiration is on the order of +30% of the estimates in Table 1. If the errors in incident radiation fluxes are assumed to be independent of the total flux then the bias in estimated evapotranspiration is on the order of +20% of the estimated values. Random errors in daily values are similar in magnitude to the biases.
IV DISCUSSION

Nuttle (1986a) has discussed the relation between hydraulic head and the water balance in the sediment. Hydraulic head observed at each piezometer can be taken as a measure of the mean head in an arbitrary volume of peat in the region of the piezometer. Water fluxes across the surface of this volume result in changes in the water content of the sediment and thus in the mean hydraulic head. This can be written as

$$V_B \langle S_s \rangle \frac{dh}{dt} = \iint (\bar{q}_I + \bar{q}_p + \bar{q}_E + \bar{q}_H) \cdot ds$$  \hspace{1cm} [4.1]

in which $\langle S_s \rangle$ is the mean specific storativity of the volume $V_B$ of sediment in the region of the piezometer. $q_E$, $q_p$, and $q_I$ are the water fluxes at the marsh surface due to evapotranspiration, precipitation, and infiltration of tidal water respectively, and $q_H$ is the horizontal water flux within the sediment. A layer of clay prevents water movement across the bottom boundary of the salt marsh sediments in Belle Isle marsh. Vertical fluxes at depth may be important to the water balance in marshes underlain by permeable materials (Hemond and Fifield 1982, Carr and Blackley 1986). An integrated form of [4.1] is used in the interpretation of the head data in order to reduce the effects of random errors.

$$h(t) = h(t_0) + \frac{1}{V_B \langle S_s \rangle} \int_{t_0}^{t} \iint (\bar{q}_I + \bar{q}_p + \bar{q}_E + \bar{q}_H) \cdot ds \, dt$$  \hspace{1cm} [4.2]

in which $V_B$ and $\langle S_s \rangle$ are assumed to be constant.
The analysis of the record of hydraulic head will proceed in two stages. Horizontal fluxes are neglected in the first stage of the analysis. Models of head response to vertical water fluxes are constructed and used to remove the head variation due to surface fluxes from the data. In the second stage, the residual head variation from the vertical flux correlation is examined for effects due to horizontal fluxes in the sediment. The region of creek bank influence is, by definition, the region in which both vertical and horizontal fluxes in the sediment towards the creek are necessary to account for the total variation in head.

Correlation with Evapotranspiration and Precipitation

During periods of no surface flooding [4.2] can be written

$$h(t) = h(t_0) + \frac{1}{D_s} \int_{t_0}^{t} (q_p - q_E) dt$$  \[4.3\]

in which $h$ now represents the hydraulic head averaged over the depth $D$ of the sediment. $S_s$ is the bulk average specific storativity, as above, but with the $<>$ notation dropped. A simple linear relationship exists between depth-averaged head and net water flux across the marsh surface.

Observations of depth-averaged head, the average of head measurements at 50, 100 and 150 cm at piezometer 1, and estimates of net surface flux are plotted in Figure 3a for the non-flooding periods 227 to 236 and 249 to 266, Julian dates. The arrows connect successive points in which the mean head increased due to precipitation on the dates noted next to the arrows;
otherwise mean head decreased through these periods in response to a net loss by evapotranspiration, Figure 1. The precipitation flux is estimated from the precipitation recorded at Logan Airport, assuming that the entire depth of precipitation infiltrates into the sediment. The colinearity of the precipitation and the evapotranspiration responses in this figure is a rough check on the accuracy of the estimates of evapotranspiration.

A linear regression of head on the net surface flux results in estimates of $h(t_0)$, 332.6 cm, and the storage coefficient $D_s$, 0.11-0.16. The correlation coefficient for this relation is 0.976 and the standard error of head prediction is 1.4 cm. The range in the estimate of the storage coefficient results from the estimated 30% bias in evaporation estimates. The storage coefficient is the product of the depth of sediment, 170 cm, and the average specific storativity, $6.5 \times 10^{-4} - 9.4 \times 10^{-4}$ cm$^{-1}$, which is determined by various storage mechanisms reviewed by Nuttle (1986a).

The phenomenon of sediment bulk volume change coupled to water content was studied concurrently with the piezometer observations (Nuttle 1986e). Surface movement in response to changes in water content is evident in the elevation data shown in Figure 3b. Bulk volume change of the sediment accounts for 0.03 of the total storage coefficient. If the remainder, 0.08-0.13, is assumed to be the result of processes such as desaturation and root expansion, active in the upper 50 cm of the sediment, then the mean specific storativity for this layer would be $1.6 \times 10^{-3} - 2.6 \times 10^{-3}$ cm$^{-1}$. This range is in reasonable agreement with the results of a lysimeter study of Belle Isle sediment (Nuttle 1986d) considering the errors that arise
here from the estimation of evaporation and the possible experimental errors in the lysimeter study.

**Head Response to Tidal Inundation**

Hemond et al. (1984) have shown that the depth of water that infiltrates into salt marsh sediments during a period of surface inundation, $T$, depends on the hydraulic properties of the sediment $K$ and $S_s$:

$$I(z_s - h_0) = 2(z_s - h_0) \left( \frac{KS_s T}{\pi} \right)^{1/2}$$ \[4.4\]

where $K$ is the hydraulic conductivity, and $h_0$ is the initial value of mean head in the sediment relative to the elevation of the sediment surface $z_s$. An equivalent expression is presented by Smiles (1974) for infiltration into a saturated swelling clay. Equation [4.1] can be written as

$$\frac{dh}{dt} = \frac{1}{DS_s} (q_I - q_E)$$ \[4.5\]

during periods of no precipitation and when there is no horizontal water movement in the sediment. Interest here is on day-to-day variation in head. The daily average infiltration flux can be obtained from equation [4.4] by multiplying by the frequency of tidal inundation $\eta$, dy$^{-1}$. Late in an interval of daily flooding, infiltration balances evaporative losses

$$q_I = \eta I(z_s - h_{EQ}) = q_E$$ \[4.6\]
SOME TEXT ON THE FOLLOWING PAGE(S) IS ILLEGIBLE ON THE ORIGINAL MATERIAL.
The daily average head is held constant at $h_{EQ}$, which is observed to be 1-2 cm below the sediment surface in Beile Isle marsh.

During the first few days of a period of daily flooding, infiltration rates are large enough to balance evaporation and to replenish the net loss of water from the sediment during the preceding non-flooding period. The rate of change of daily mean head during this period of recovery can be obtained from [4.5] and [4.6].

\[
\frac{dh}{dt} = \frac{1}{S_s} \eta_{1}(h_{EQ} - h) \tag{4.7}
\]

This is first order with respect to $(h_{eq} - h)$ so an exponential recovery of the form

\[
(h_{EQ} - h) = (h_{EQ} - h_1)e^{-\Gamma t} \tag{4.8}
\]

\[
\Gamma = 2\eta \left[ \frac{KT}{S_s \pi D^2} \right]^{1/2}
\]

can be expected, in which $h_1$ is the mean head in the sediment at the beginning of the period of daily surface flooding.

Mean head data observed at piezometer 1 during the two periods of head recovery (Figure 1) are plotted versus time since the onset of surface flooding on a semi-log plot, Figure 4. The plot is linear indicating that the recovery of head is exponential. Scatter in the data for time greater than $30 \times 10^4$ s is caused by limits on the precision to which hydraulic head can be measured in the field. The value of $\Gamma$ in Equation [4.8] is
estimated by the slope of the line fitted arbitrarily to the data, \( \Gamma = 7.5 \times 10^{-6} \text{ s}^{-1} \).

The correlation of mean head to the cumulative net atmospheric flux during periods of no surface flooding and the exponential recovery at the resumption of daily flooding at piezometer 1 forms the basis for a model for head variation due to surface fluxes, Figure 5a. If it is assumed that the only source of variation in head between piezometers in the array is surface elevation, which controls the value of \( h_{\text{EQ}} \), then the model shown in 5a can be used to predict the head response at each piezometer, \( h_{\text{VERT}} \), making the appropriate adjustment for \( h_{\text{EQ}} \). If all of the systematic variation in head is accounted for in this model, then the variance of the residuals \( h_{\text{OBS}} - h_{\text{VERT}} \) (Table 2) should then be comparable to the mean square error of observation (0.25 cm²) and model error (1.96 cm²).

**Effect of Vertical Fluxes within the Sediment**

The variance of the residuals is too large to be accounted for by model and observational errors alone. The residuals of the three piezometers closest to the creek are significantly higher than the others and will be analyzed for the effect of horizontal fluxes below. Residuals at the remaining piezometers include the effect of systematic differences between head at the 50 cm depth and the mean head in the sediment profile that are related to vertical water movement in the sediment. This can be seen in the mean residual \( r_{M} \) averaged over all of the piezometers, Figure 5b, except 5, 6 and 15 where the residuals are much larger. The observed heads at the 50 cm depth are generally lower than the predicted mean head during
periods of evaporative loss and upward movement of water, and higher than mean head during periods of infiltration and downward water movement in the sediment. The mean residual shown in Figure 5b is taken to represent the systematic variation between mean head and head observed at 50 cm, which cannot be accounted for by a depth-averaged model, and is used to correct for this effect in the residuals. The variances of the corrected residuals, $h_{OBS} - h_{VERT} - r_M$, are shown in the third column of Table 2.

The corrected residuals for piezometers at the 50 cm depth for 1, 3, and 4 are shown in Figure 6. Large values occur during precipitation events and at the onset of surface flooding, and positive residuals at piezometer 1 correlate with negative residuals at piezometers 3 and 4. This behavior results from a lag in the head response at 3 and 4 relative to 1 that is related to differences in surface elevation. Low areas, lower by only a few centimeters, can flood a day before the rest of the marsh surface and be flooded for a longer period of time during each flooding tide. Low areas can also collect runoff from precipitation and hold it until it has a chance to completely infiltrate. Therefore the head response to infiltration and precipitation in low areas can be more rapid and begin earlier than in other areas of the marsh.

Effect of Horizontal Fluxes

Large positive values for the uncorrected head residual at piezometers 5, 6 and 15 (Figure 6) can be explained by early flooding of the low areas of the marsh near the creek by rising spring tides, which causes a recovery in head at those piezometers a day or two in advance of the flooding and
recovery of head in the rest of the marsh. The large negative residuals
during the non-flooding periods require a mechanism for water loss in addi-
tion to evapotranspiration, the effects of which have already been account-
ed for in $h_{\text{VERT}}$. The logical mechanism to consider is water loss by drain-
age to the creek.

Piezometers 5 and 15 are close enough together that their average
residual head will be taken as the mean for the sediment volume shown in
Figure 7. If only water fluxes perpendicular to the creek are considered
then the variation in mean residual of these piezometers follows

$$\frac{d}{dt}(h - h_{\text{VERT}})\bigg|_{5,15} = \frac{Z}{S_D(\ell_1 + \ell_2)} (Q_1 - Q_2) \tag{4.9}$$

The total influx of water from the direction of piezometer 4, $Q_1$, and the
efflux in the direction of piezometer 6, $Q_2$, can be estimated from Darcy's
law.

$$Q_1 = KD \frac{h_4 - h_{5,15}}{\ell_1}$$

$$Q_2 = KD \frac{h_{5,15} - h_6}{\ell_2} \tag{4.10}$$

Combining [4.9] and [4.10] and integrating with respect to time yields, for
the mean residual,
\[ (h - h_{\text{VERT}})^{5,15} = \frac{K}{S_s} \int_{t_0}^{t} \left\langle \frac{d^2h}{dx^2} \right\rangle dt \]

\[
\left\langle \frac{d^2h}{dx^2} \right\rangle = \frac{2[\ell_2h_4 - (\ell_1+1)\ell_5,15 + \ell_1h_6]}{[\ell_1+\ell_2]\ell_1\ell_2} \tag{4.11}
\]

The quantity in the integral is an estimate of the second derivative of head calculated from discrete values. The integral is estimated from the data by averaging and summing over discrete time intervals. Values of the time integral of the second derivative of head are plotted against the mean head residual \((h_{\text{OBS}} - h_{\text{VERT}})^{5,15}\) in Figure 8 for the two non-flooding periods in the record excluding the periods following the precipitation events. The residual head at piezometers 5 and 15 is seen to be largely the result of the drainage of water through the sediment into the creek during these periods. The slope in Figure 8 is the ratio of the hydraulic properties of the sediment, \(K/S_s = 0.18 \text{ cm}^2/\text{s}\).

**Estimates of Pore Water Fluxes**

The mean flux velocity of pore water drainage across the creek bank can be estimated from the results above. In the period 227-232, hydraulic head 6.5 m from the creek declined 16 cm due to evapotranspiration and 12 cm due to drainage to the creek. Therefore, 43% of the water loss in the sediment at this point was caused by drainage to the creek. There was no detectable influence of drainage greater than 10 m from the creek. If the 12 cm drop is taken as the average for the interval 0-10 m from the creek then the mean water flux out of the sediments was 30 \( \ell \text{ dy}^{-1} \) for each meter along the creek. This rate is comparable to those discussed by Jordan and
Correll (1985). Assuming a depth of sediments of 170 cm, the average flux velocity of the pore water at the creek bank was $2 \times 10^{-5}$ cm s$^{-1}$.

Estimates of horizontal gradients in hydraulic head were obtained by linear interpolation among groups of adjacent piezometers and are summarized in Table 3. Significant gradients in head occur near the creek and mean head gradients follow the slope of the marsh surface. Positive gradients in the x direction correspond to increases in head away from the creek and are proportional to pore water fluxes towards the creek. Gradients in the y direction are proportional to pore water fluxes parallel to the creek. Errors in the estimates of head gradients arise chiefly from the error of determining the relative elevation of the piezometers, estimated to be $\pm 1$ cm. Typical spacings in the piezometer array are 5 m perpendicular to the creek and 2.5 m parallel to the creek. Therefore, the uncertainty in mean head gradients is 0.002 and 0.004 cm/cm, respectively. Changes in hydraulic head are known to 0.5 cm, so errors in the variation of the head gradient are 0.001 and 0.002 cm/cm.

The means and standard deviations of the components of pore water discharge are easily computed using Darcy's law (Figure 9). The hydraulic conductivity of the sediment is $1.4 \times 10^{-4}$ cm s$^{-1}$, estimated by taking the mean specific storativity to be $7.9 \times 10^{-4}$ cm$^{-1}$ and multiplying by the value of $K/S_s$ determined in Figure 8. This value is in agreement with estimates obtained by Chen (1986) from core samples and by Nuttle (1986d) from head response in a lysimeter. Actual water velocities can be much higher than the mean flux velocities (Figure 9) if the bulk of the flow is
constricted to a few high conductance channels, as suggested by dye studies (Chen 1986).

The response of hydraulic head to periodic variation in the level of an adjacent body of water has been considered by several authors (Tison 1965, Jacob 1949, Werner and Noren 1951). If the head imposed at the creek bank by the action of tides is roughly sinusoidal then the variation of head in the sediment is

\[
h = h_0 e^{-x \frac{\omega}{2\alpha}} \sin\left(\omega t - x \frac{\omega}{2\alpha}\right)
\]

[4.12]

in which \(\alpha\) is the ratio \(K/S_s\) and \(\omega\) is the radian frequency of the head variation. By application of Darcy's law, the horizontal flux due to tidal forcing is

\[
q = K h_0 \frac{\omega}{\alpha} e^{-x \frac{\omega}{2\alpha}} \sin\left(\omega t - x \frac{\omega}{2\alpha} + \frac{\pi}{4}\right)
\]

[4.13]

The amplitude of the tidally driven pore water flux at the creek bank is \(2.3 \times 10^{-4}\) cm/s for the 2-meter range of water levels in Belle Isle marsh. The amplitude of tidal fluctuations in pore water fluxes decays exponentially with distance away from the creek, Figure 9.

Mean pore water flux velocities are closely related to the slope of the sediment surface and decrease rapidly with increasing distance from the creek. Estimated flux velocities farther than 15 m from the creek are small relative to the error associated with estimates of the head gradient.
Significant pore water fluxes associated with the semi-diurnal tides are restricted to within 5 m of the creek, given the hydraulic properties of the Belle Isle sediments determined here. Fluctuations in horizontal pore water velocities farther from the creek must be the result of localized topographic features of the surface. Locally low areas of the surface may flood sooner in the spring-neap cycle and remain flooded longer during each inundation. The recovery of head in these areas early in a period of daily flooding will be faster than in the surrounding sediment, causing pore water to flow from the low areas to the higher areas. Later in the period of daily flooding this pattern will be reversed. When the hydraulic head is uniformly high throughout the marsh, the horizontal distribution of head is determined by surface topography immediately following the recession of the tides from the marsh. The value of head everywhere will be the elevation of the surface and pore water will drain from areas of higher surface elevation to lower areas. It can be shown that the effects of these topographically induced pore water flows are local and short lived. They have no effect on the overall water balance of the sediment, but they may be important to controlling the concentration of solutes in salt marsh sediments.

Solute Balance

The evaporative loss of sea water from the sediments of Belle Isle marsh presumably concentrates salts in the sediments, which in turn poses a long-term threat to the viability of the salt marsh vegetation. Evaporation from moist environments in the Boston area is on the order of 66 cm per year (Kohler et al. 1959). If this is balanced by equal parts precipi-
tation and infiltration of tidal water at a salinity of 32 ppt then an annual influx of 10 kg m\(^{-2}\) of salt must be balanced by some mechanism of salt export. If 10% of the water loss attributed to evaporation is actually lost through pore water drainage to the creek from interior regions of the marsh then salinities on the order of 160 ppt and greater would be needed for the annual influx of salt to be balanced by advection in the sediment. However the salinity of water pumped from the piezometers is about 40 ppt and is nearly the same for all of the piezometers.

Clearly some other mechanism is responsible for exporting salt from the sediment since burial seems unlikely. Gardner (1975) notes that water draining from the marsh surface after inundation is enriched in silica, phosphate and bicarbonate, but it is not clear whether this is the result of export from the sediment or of processes occurring at the surface. The question of an unaccounted for mechanism of solute transport from interior regions of the marsh is particularly important to the fate of sulfur that enters the sediment as sulfate and is reduced to sulfide as a result of the below-ground metabolism (Howarth and Teal 1980, Howes et al. 1984). The rate of export of the reduced sulfur compounds is a factor in estimates of the rate of anaerobic metabolism in salt marsh sediments.
REFERENCES


NUTTLE, W. K. 1986e. Peat rose!


Table 1
Summary of Daily Heat Budget

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<th>August</th>
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Ly = 1 \( \text{cal} \) / \( \text{cm}^2 \)
Table 2
Head Variance for Piezometers at 50 cm Depth

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<thead>
<tr>
<th>Piezometer</th>
<th>Distance from creek bank (m)</th>
<th>VAR[h_{OBS}]{\textsuperscript{1}}</th>
<th>VAR[h_{OBS} - h_{VERT}]{\textsuperscript{1}}</th>
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1. (in cm\textsuperscript{2})
Table 3

Gradients of Head and Surface Elevation

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<th>Distance from creek (m)</th>
<th>Piezometers</th>
<th>$\frac{dh}{dx}$</th>
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<td>(mean) 4.36</td>
<td>(S.D.) 1.02</td>
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1. (in cm/m)
An array of 18 piezometers was installed perpendicular to a creek in Belle Isle marsh. The 13 piezometers shown are screened over the interval 43-58 cm below the marsh surface. The remaining 5 piezometers were installed deeper in the sediments, one each at 100 and 150 cm near piezometers 1 and 4, and one at 150 cm near piezometer 6. Surface contours are in centimeters relative to mean low water in Boston Harbor. The location of the creek bank is approximately at the 300 cm contour.
Changes in the depth-averaged hydraulic head observed during August and September are highly correlated with hydrologic conditions at the marsh surface. Daily flooding of the marsh by the highest spring tides maintains the hydraulic head at levels near the marsh surface. Hydraulic head declines as a result of a net loss of water from the sediment by evapotranspiration during the intervening dry periods. Periods of rainfall cause a temporary recovery in head. Surface elevation and proximity to a tidal creek are factors in the head differences between piezometers.
A) Depth averaged hydraulic head at piezometer 1 is highly correlated with the net atmospheric water flux (evaporation less precipitation) during periods when the marsh is not flooded daily by tides. Arrows show short-term recovery in mean head following precipitation on the Julian dates shown. Otherwise head generally declined in response to evaporative water loss. Data are for the periods Julian dates 227-236 (■) and 249-266 (□).

B) Changes in the elevation of the marsh surface follow changes in depth averaged head throughout the same two month period. Change in the bulk volume of the sediment accounts for 20% of the changes in water content that occurred in this period.
Figure 4

Depth-averaged head tends exponentially towards an equilibrium value during daily flooding of the marsh surface by tides. Data are for the periods Julian dates 237-240 (●) and 266-274 (○).
A) The correlation of mean head with evapotranspiration, precipitation, and infiltration of tidal water (——) accounts for 96% of the variation in depth averaged head at piezometer 1.

B) The difference between head values observed at 50 cm depth and the predicted depth averaged head depends on the direction of the water flux in the sediment. The head at 50 cm is less than the mean during periods of net water loss and upward water movement and greater than the mean during periods of replenishment and downward water movement. Data are the mean residuals of all piezometers at 50 cm except piezometers 5, 15, and 6.
The nature of the head residuals changes with distance to the creek. Far from the creek, the residuals at piezometers 1, 3, and 4, corrected for the vertical flux in the sediment, show that practically all of the observed variation in head has been accounted for by the correlation with surface fluxes. Nearer the creek, the large positive values of the uncorrected residuals for piezometers 5, 6, and 15 result from the flooding of the low areas of the marsh earlier in the spring tide cycle. Large negative residuals during periods of drying indicate additional water loss from the sediment due to drainage to the creek.
Figure 7

Scheme for calculating the effect of net horizontal water flux on the average of the hydraulic heads at piezometers 5 and 15.
Figure 8

The average of the residuals of piezometers 5 and 15 is correlated with the second derivative of hydraulic head with respect to distance from the creek, estimated by differences between observed values of head. The slope is the ratio of hydraulic conductivity to specific storativity of the sediment.
SOME TEXT ON THE FOLLOWING PAGE(S) IS ILLEGIBLE ON THE ORIGINAL MATERIAL.
\(- \sum \left( \frac{d^2h}{dx^2} \right) \Delta t \left( \frac{cm \times day}{m^2} \right)\)
Tidal forcing dominates the movement of pore water in sediments within 2-3 m of the creek. Gravity drainage is important in the interval 3-10 m. Evaporation is the dominant cause of water loss from the sediments beyond 10 meters. Horizontal fluxes are small but may be important for controlling the concentration of solutes. Data are mean (○) and fluctuating (○, standard deviation) horizontal flux velocities estimated from observed gradients in hydraulic head.
ELEMENTS OF SALT MARSH HYDROLOGY:

III. HOW MUCH SEEPS ON THE NEAP?

William Nuttle
Massachusetts Institute of Technology
May 1986
ABSTRACT

A model of water drainage through the sediments of a salt marsh into a tidal creek is constructed and compared with observations of hydraulic head. Significant horizontal water movement is limited to a region within 15 m of the creek bank. The factors that determine the size of this region are surface topography, which influences the distribution of hydraulic head, and the frequency of inundation of the marsh surface. Frequent inundation of the marsh increases the rate of water flow at the creek bank, but it prevents the drainage of water from regions more than a few meters from the creek bank. The interior regions of the marsh require long periods of no surface inundation to drain.
I INTRODUCTION

Beneficial effects of tidally induced circulation of pore water through the sediments of salt marshes have been claimed as one explanation for the high productivity of the marsh grasses. It is becoming clear that a simple pattern of infiltration and drainage through the sediment is not adequate to describe the range of hydrologic and biogeochemical regimes that are observed in the sediments. There is a growing body of geochemical (Gardner 1973, Nestler 1977, Howes et al. 1984) and hydrologic (Hemond and Fifield 1982, Dacey and Howes 1984, Jordan and Correll 1985, and Nuttle 1986b) evidence that the movement of pore water in interior regions of the marsh is slow. Rapid movement of pore water in response to semi-diurnal tides, which has been cited as contributing to the differentiation of tall and short growth forms of Spartina (King et al. 1982) is limited to a region within 3-4 meters of the creek bank (Gardner 1973, Nuttle 1986b). Agosta (1985) has suggested that water draining from the interior regions of a marsh is an important source of nitrogen for the vegetation in the hydrologically active creek bank region.

This paper presents a model for horizontal water movement that spans the transition from stagnant conditions in the interior regions to active drainage near the creek bank. Nuttle (1986b) has determined that significant horizontal fluxes are limited to a region within 10 m of the creek bank at a site in Belle Isle marsh. This result is confirmed by the model, which predicts the distribution of horizontal water movement from the factors that drive the flow of water in the sediment. Horizontal drainage
is controlled by the surface topography of the marsh, the frequency of tidal inundation, and the rate of evapotranspiration. The slope of the marsh surface towards the creek bank plays a large role in determining the extent of horizontal water movement, through its effect on the slope of the water table. The model reveals that, although frequent inundation of the marsh by tides maximizes water flow across the creek bank, it reduces drainage from interior regions of the marsh. And, the effect of evapotranspiration on the water table throughout the marsh may cause a reversal of the outflow of water across the creek bank.

II MODEL FORMULATION

The hydrology of Belle Isle marsh is typified by alternating periods of daily flooding by tides and periods of a week or more in which the surface does not flood (Nuttle 1986b). The water content of the sediment increases through the flooding periods until it reaches full capacity and declines through the non-flooding periods. Evapotranspiration contributes to water loss from the sediment throughout the marsh. Horizontal drainage in the sediments has been observed to contribute to the decrease in water content up to 10 m from the creek. The objective of this section is to formulate a model of the head distribution during periods in which there is a net loss of water from the sediment due to evapotranspiration and drainage across the creek bank.

Consider the idealized water balance for the column of sediment shown in Figure 1. The water content of the sediment is related to the mean hydraulic head. Horizontal water fluxes in the sediment are parallel to
the x axis, which is perpendicular to the creek, and there is a constant, uniform loss of water at the sediment surface due to evapotranspiration, \( q_E \). The equation for mass conservation for water in the sediment can be written as (Nuttle 1986b)

\[
X' D S \frac{d\langle h \rangle}{dt} = \int_0^D q_H \, dz - q_E X'
\]  

[2.1]

which has the differential form

\[
S_s \frac{d\langle h \rangle}{dt} = \frac{d}{dx} \langle q_H \rangle - \frac{q_E}{D}
\]  

[2.2]

Application of Darcy's law leads to a differential equation that describes the distribution of depth-averaged hydraulic head perpendicular to the creek bank.

\[
\frac{\partial h}{\partial t} = \frac{K}{S_s} \frac{\partial^2 h}{\partial x^2} - \frac{q}{DS_s}
\]  

[2.3]

The hydraulic conductivity \( K \) and the specific storativity \( S_s \) of the sediment are assumed to be constant and uniform. The \( \langle \rangle \) notation has been dropped for the sake of simplicity.

**Initial and Boundary Conditions**

The solution of [2.3] for the appropriate initial and boundary conditions constitutes a model of the head response to water loss by evapotranspiration and drainage. Daily flooding and infiltration of tidal water
increases the depth-averaged head everywhere in the sediment to an equilib-
rium value at which the water table is typically only 2–3 cm below the
marsh surface (Nuttle 1986b). The distribution of head at the start of a
non-flooding period $h_1(x)$ is closely related to the surface profile of the
sediment, Figure 2.

The depth-averaged head at the creek bank ($x = 0$) can be considered to
be the sum of two components, a constant value imposed by the time-averaged
creek level and a fluctuating component that reflects the influence of the
tides. It has been argued that tidal fluctuations in creek level have
little effect on water movement in the sediment beyond 3 m from the creek
bank (Nuttle 1986b). The interest here is on drainage towards the creek
that occurs over 10s of meters and periods of several days. Therefore, the
effects of the tides can be ignored, and the boundary condition at the
creek bank will be represented by a constant value of head, $h_0$, to be
determined by fitting.

The initial and boundary conditions on [2.3] are formally stated as

\[ t = 0 ; \quad h = h_1(x) \]

\[ t > 0 ; \quad h = h_0 \quad x = 0 \]  \hspace{1cm} [2.4]

In addition, it is necessary to specify that the solution for head remains
finite over the interval $0 < x < \infty$ to obtain a solution for the case of a
semi-infinite deposit.
Solution for Head

In order to obtain a solution for \( h(x,t) \) it is necessary to describe the initial distribution of head, \( h_1(x) \). The principal of superposition of solutions for \( h(x,t) \) can be used (because \([2.3]\) is linear) to separate the effects of a non-uniform initial distribution of head from the effects of evapotranspiration and creek bank boundary condition. Consider \( h(x,t) \) to be the sum

\[
h(x,t) = h_1(x,t) + h_2(x,t) \tag{2.5}
\]

such that the components \( h_1 \) and \( h_2 \) are the solutions to

\[
\frac{\partial h_1}{\partial t} = \frac{K}{S} \frac{\partial^2 h_1}{\partial x^2} - \frac{q_E}{D_S} \frac{\partial S}{\partial x}
\]

\[ t = 0 ; \quad h_1 = h^* = \text{constant} \]

\[ t > 0 ; \quad h_1 = 0 ; \quad x = 0 \tag{2.6} \]

\[
\frac{\partial h_2}{\partial t} = \frac{K}{S} \frac{\partial^2 h_2}{\partial x^2}
\]

\[ t = 0 \quad h_2 = h_1(x) - h^* \]

\[ t > 0 \quad h_2 = 0 ; \quad x = 0 \tag{2.7} \]

It is easily confirmed that the sum \( h_1 + h_2 \) satisfies \([2.3]\) and \([2.4]\).
The solution for $h_1(x,t)$ is

$$h_1(x,t) = h^* + (h_0 - h^*) \text{ERFC}\left[\frac{x}{2\sqrt{at}}\right] +$$

$$\frac{q_E^t}{D_S^*} \cdot \left[ -1 + \left[ 1 + \frac{x^2}{2at} \right] \text{ERFC}\left[\frac{x}{2\sqrt{at}}\right] - \frac{x}{\sqrt{4at}} \exp\left[\frac{-x^2}{4at}\right] \right]$$

[2.8]

in which $\alpha = K/S^*_S$. This result was obtained by first making the transformation of variables $h' = h_1 + q_E^t/D_S^*$ in [2.6] and then applying techniques presented in Carslaw and Jaeger (1959).

It remains to find a function $h_2(x,t)$ that satisfies the conditions in [2.7]. The initial distribution of head is characterized by a nearly uniform value in the interior region of the marsh and a marked gradient in head near the creek bank, following the general features of the surface topography, Figure 2. If $h^*$ is set equal to the uniform value of head in the interior then $h_2|_{t=0}$ is negative and decreases in magnitude as the distance from the creek increases. A solution to [2.7] that has these general features is

$$h_2(x,t) = \phi^* \left\{ \text{ERFC}\left[\frac{x}{2\sqrt{at}} (t^{*\#}+t)^{-\frac{1}{2}}\right] - \text{ERFC}\left[\frac{x}{2\sqrt{at}}\right]\right\}$$

[2.9]

in which $\phi^*$ and $t^{*\#}$ are fitting parameters. $\phi^*$ is nominally equal to the difference $h_1 - h^*$ at the creek bank, $x = 0$, and $t^{*\#}$ is chosen to match the length scale over which $h_2$ increases from $\phi^*$ to zero.
III MODEL CALIBRATION

The sum of [2.8] and [2.9] is a model of the response of depth-averaged hydraulic head to water loss from the sediment by a constant rate of evapotranspiration and drainage perpendicular to a creek bank. This model is calibrated using observations of head and estimates of $\alpha$ and of the storage coefficient $D_s$ obtained in the water balance study by Nuttle (1986b). An array of piezometers was installed in Belle Isle marsh in Boston, Massachusetts, Figure 3. Observations of hydraulic head were made every other day during August and September 1984. A full description of sediment characteristics and pore water hydrology is given by Nuttle (1986b) and Chen (1986).

Initial Conditions

The distribution of head preceding a period of net water loss from the sediment is shown in Figure 2. Also shown is the fitted head distribution $h_2 + h^*\|$ for the model. A reasonable fit is obtained with the values $h^*\| = 334.5 \text{ cm}$, $\phi^*\| = -45 \text{ cm}$, and $t^*\| = 9.06 \times 10^5 \text{ seconds}$.

Evapotranspiration

Evapotranspiration was estimated by the method of Priestly and Taylor (1972). The rate of evapotranspiration, in units of langleys per day, is given by the relation

$$q_E = \alpha_E \frac{A}{\Lambda + \gamma} (R_N - G)$$  [3.1]
SOME TEXT ON THE FOLLOWING PAGE(S) IS ILLEGIBLE ON THE ORIGINAL MATERIAL.
in which $A$ is the rate of change of saturation vapor pressure with air
temperature at a temperature typical of the air near the ground and $\gamma$ is
the ratio of the sensible heat content of air to the latent heat content of
water vapor in vapor saturated air, also known as the psychrometric con-
stant. $R_N$ is the net radiant heat flux to the surface, $G$ is the soil heat
flux and $a_E$ is an empirical constant established to be 1.26 by Priestly and
Taylor. This method has been reviewed by Brutsaert (1982) and found to
perform well for well-watered grassed surfaces. The errors of estimation
for evaporation from Belle Isle marsh have been described by Nuttle
(1986b). The net radiation flux $R_N$ was estimated from air temperature,
humidity, pressure and cloud cover. The meteorologic data was taken from
the National Weather Service record for Logan Airport, 3 km southeast of
the study site. Humidity and air temperature measurements were made in the
marsh as a check on the applicability of the airport record. The soil heat
flux was estimated from changes in the temperature distribution in the
sediment.

The cumulative atmospheric flux (evapotranspiration minus precipita-
tion) for the beginning of a non-flooding period is shown in Figure 4. The
cumulative net flux declines due to 14 mm of precipitation on September 4,
1984, Julian date 248. There was no substantial precipitation for the
following 10 days during which the mean rate of evapotranspiration was $a_E =
3.2 \text{ mm/day}$. 

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Sediment Properties

The hydraulic properties of the sediment in Belle Isle marsh have been determined by correlating the observed variation of head with the hydrologic boundary conditions (Nuttle 1986b). An estimate of the storage coefficient, $D_S = 0.16 \text{ cm/cm}$, was obtained by correlating the depth averaged head 24 m from the creek with the cumulative net atmospheric flux. An estimate of the sediment parameter, $\alpha = 0.18 \text{ cm}^2/\text{s}$, was obtained by analyzing the response of head near the creek bank for the effects of horizontal drainage. These estimates of the hydraulic properties of the peat agree well with independent estimates summarized by Nuttle (1986a).

Creek Bank Boundary

The last remaining model parameter, the mean hydraulic head at the creek bank, is estimated by fitting the model to the observed distribution of head during the period September 4 to September 13, 1984 (Julian dates 248-257), Figure 5. The fitted value is $h_0 = 287 \text{ cm}$.

IV DISCUSSION

The model reproduces the general distribution of head throughout the period of net water loss. The uppermost curve in Figure 5 represents the fitted initial conditions, $t = 0$, while the other curves are model "predictions." The method of determining the storage coefficient $D_S$ and the rate of evapotranspiration in effect constrains the modeled head distribution to pass through the average of the three data points at 24 m from the creek. The discrepancy between the model and the data at distances greater than
20 m is believed to be caused by the daily course of evapotranspiration, Figure 4, and by the vertical distribution of head within the sediment. The model predicts the head averaged over the 170 cm depth of salt marsh sediment in Belle Isle marsh, while the observed head data is for the 50 cm depth only, except for the three data points at 24 m which are distributed through the depth of the sediment. Upward flow of water in the sediment in response to evapotranspiration at the surface causes the hydraulic head at 50 cm to be lower than the depth averaged head.

Infiltration, Evapotranspiration, and Drainage

The water content of the sediment in Belle Isle marsh follows a cycle driven by the spring/neap tide cycle. When high tides exceed a threshold (~10 ft relative to low water in Boston Harbor) the marsh floods and water infiltrates into the sediment. A period of daily flooding drives the water table to an equilibrium position near the marsh surface. As long as tidal flooding continues then the water table is held at this position, and the water content of the sediment remains constant. During periods of neap tides, in which the marsh surface is not flooded, there is a net loss of water from the sediment by evapotranspiration and drainage, near the creek bank, and the water table declines. On the return of the spring tides and flooding of the marsh surface, infiltration replenishes the water content of the sediment, and the cycle begins again. The amount of infiltration that occurs during a period of daily flooding is determined by the net loss of water from the sediment in the preceding non-flooding period. Precipitation occurs at random with respect to this cycle and affects the net water loss from the sediment during the non-flooding periods. However,
fluctuations in water content are dominated by the cycle driven by the spring tides.

Evapotranspiration is the dominant mechanism for water loss from the sediment throughout most of the marsh. In the interior regions, away from the creeks, the water balance is characterized by infiltration of tidal water and precipitation and water loss by evapotranspiration. The magnitude of fluctuations in water content, which is an important factor in the degree of desaturation and oxygen availability to the sediment, depends on the rate of evapotranspiration and the frequency of surface flooding. The cycle of infiltration and evapotranspiration also tends to concentrate solutes in the sediment.

Additional water loss by horizontal drainage through the sediments near the creek has two effects that are beneficial to the marsh grasses and primary productivity. The additional water loss translates into larger fluctuations in the pore pressures in the sediment which increases the probability of air entry into the sediment and the extent to which it occurs. The additional water loss near the creek bank results in an increased volume of infiltration relative to the interior regions of the marsh. This additional infiltration returns to the creek through the sediment, "flushing" the pore water and providing a means of transporting nitrogen from the interior of the marsh to the sediment at the creek bank where active tidal flushing may otherwise limit its concentration (Agosta 1985).

The one-dimensional model of head response to water loss by evapotranspiration and drainage can be used to estimate the magnitude of the
fluxes that occur as a result of drainage and the scales over which drainage occurs. First consider the magnitude of water loss (and subsequent infiltration) in excess of evapotranspiration. Cumulative water loss in excess of evapo-transpiration can be calculated from the net change in depth averaged head during a non-flooding period.

\[ D_w(x,t) = DS_s [h_1(x) - h(x,t)] - q_E t \]  \hspace{1cm} [4.1]

Here \( D_w \) is the "depth" of the additional water loss by drainage (volume per unit surface area of the marsh) cumulative since \( t = 0 \) when the distribution of head was given by \( h_1(x) \). An expression for \( D_w(x,t) \) can be obtained by substituting the model of head 2.8 and 2.9 for \( h(x,t) \).

\[ D_w(x,t) = DS_s \left\{ k \left[ \text{ERFC} \left( \frac{x}{2\sqrt{\alpha t}} \right) - \text{ERFC} \left( \frac{x}{2\sqrt{\alpha}} (t^* + t)^{-1/2} \right) \right] + \text{ERFC} \left( \frac{x}{2\sqrt{\alpha t}} \right) \right\} \]

\[- DS_s (h_0 - h^*) \text{ERFC} \left( \frac{x}{2\sqrt{\alpha t}} \right) - q_E t \left\{ \left[ 1 + \frac{x^2}{2\alpha t} \right] \text{ERFC} \left( \frac{x}{2\sqrt{\alpha t}} \right) \right\} \]

\[- \frac{x}{\sqrt{\pi \alpha t}} \exp \left( \frac{x^2}{4\alpha t} \right) \]  \hspace{1cm} [4.2]

The excess water loss for Belle Isle marsh is shown in Figure 6 for several periods of drainage. The hydraulic head near the creek bank, and thus the water content of the sediment there, is held constant, so the cumulative water "loss" in excess of evapotranspiration becomes increasingly negative after the initial change in head in the first day of drainage. One can interpret the curves in Figure 6 as the distribution of the depth of infiltration in excess of that needed to make up for evaporative losses.
The curves show how the distribution of infiltration depends on the length of the preceding non-flooding period. It is instructive to compare the magnitude of the excess infiltration with cumulative volume of evapotranspiration, which occurs at the rate of 3.2 mm per day. For short non-flooding periods, the excess water loss is comparable to the evaporative loss. As the length of the non-flooding period increases, the maximum excess water loss does not change significantly, so it becomes increasingly less important relative to the cumulative evapotranspiration loss.

**Horizontal Water Fluxes**

The cumulative flux towards the creek at a point can be computed from the total excess water loss that occurred by drainage landward of the point.

\[ q(X,t) = \frac{1}{D} \int_X^\infty D_w(x,t)dx \]

[4.3]

in which the computed flux is the average over the entire depth of the sediment. The distribution of cumulative flux corresponding to the results in Figure 6 are shown in Figure 7. Mean horizontal fluxes in the region 0-8 m are on the order of $6 \times 10^{-6}$ cm/s for the long 10-15 day non-flooding periods. Higher fluxes, $5 \times 10^{-5}$ cm/s, occur at the creek bank in the first day of drainage. Generally, mean water fluxes due to drainage are maximized by short periods of drainage; however this limits the extent of significant drainage to regions close to the creek bank. Long periods of drainage result in smaller fluxes acting over longer distances.
The horizontal extent of water loss due to drainage is determined by the interaction of two length scales: the length scale imposed by the initial distribution of head $\sqrt{\alpha t}$, which is ultimately related to the surface topography, and the length scale associated with the transient head response to horizontal drainage $\sqrt{\alpha t}$. The length scale associated with the initial distribution of head and its propagation away from the creek bank $\sqrt{\alpha(t+t^*)}$ determines the scale of the drainage response. The effects of drainage predicted by the model extend to 15-20 m from the creek, somewhat farther than determined by the analysis of the head data (Nuttle 1986b). There are two reasons for the difference between the model predictions and the data analysis. First, the model idealizes the conditions in the sediment, particularly the distribution of the hydraulic properties of the sediment which were assumed to be uniform and constant. Second, the analysis of the head data was not sensitive enough to observe the drainage effects predicted by the model. For instance, a difference in water content of the sediment of 3 mm corresponds to a 2 cm difference in mean head. This is comparable to the unexplained variation in the head data.

In the early stages of drainage, the flow of water from farther back in the sediment is sufficient to balance the loss due to evapotranspiration from the sediment near the creek, with the excess draining out of the sediment into the creek. The rate of drainage decreases with time, so at some time, between 5 and 10 days in Belle Isle marsh, the rate of drainage is insufficient to match evaporative losses. In order to maintain a constant value of head at the creek bank, the flow of water reverses at the creek bank and creek water and pore water converge in the sediment. The signifi-
cance of this finding is unknown because tide-forced fluxes, which have been neglected so far, have an amplitude on the order of $5 \times 10^{-6}$ cm/s in the same region of the sediment (Nuttle 1986b), so a more detailed study of water fluxes in the creek bank region is called for.

**Long-Term Effects on Drainage**

This analysis identifies the length of non-flooding periods (i.e., the frequency of the underlying flooding/drying cycle) as an important factor in the movement of water far from the creeks. The length of periods of flooding and non-flooding are determined in part by the random influence of weather and storm effects on water levels on the coast, but cyclic variations related to the relative positions of the earth, moon, and sun are also present. In particular, an 18.6-year cycle in the frequency of extreme tides is thought to have ecological significance on intertidal organisms (Bleakney, 1972).

The number of non-flooding periods in Belle Isle marsh in the months April-September follows an 18-year cycle, Figure 8. These data have been determined by applying the flooding threshold (10 ft predicted tide for Boston harbor) to published tide predictions (U.S. Commerce Dept.). Horizontal water movement in the interior of salt marshes is never large. Solute concentrations may vary from week to week in the marsh interior due to constant concentration by evaporation and periodic removal by episodes of horizontal water movement. Changes in the number of non-flooding periods that are sufficiently long enough to remove water from the interior regions may also cause year to year variations in pore water chemistry.
REFERENCES


Figure 1

Control volume for mass conservation of pore water in the model of horizontal drainage in Belle Isle marsh.
Figure 2

The actual distribution of head at the start of drainage is determined largely by the surface topography. The smooth curve is the complementary error function fitted to the head data to describe initial conditions in the model.
Figure 3

An array of 18 piezometers was installed perpendicular to a creek in Belle Isle marsh. The 13 piezometers shown are screened over the interval 43-58 cm below the marsh surface. The remaining 5 piezometers were installed deeper in the sediments, one each at 100 and 150 cm near piezometers 1 and 4, and one at 150 cm near piezometer 6. Surface contours are in centimeters relative to mean low water. The location of the creek bank is approximately at the 300 cm contour.
Figure 4

Mean surface flux was estimated from the trend in the cumulative estimated evaporation for the 10-day period 248-258, 1984. A rain storm just before this period resulted in the initial head distribution shown in Figure 2.
Figure 5

The model of head response is fitted to piezometer observations during a period of net water loss from the sediment due to evapotranspiration and drainage. The uppermost curve is the fitted initial head distribution. The other curves are model results for the Julian dates shown.
The depth of water loss in excess of evapotranspiration due to hori-
zontal water movement in Belle Isle sediments is estimated by the model of
head response to drainage and evapotranspiration.
The cumulative horizontal water flux at a point is calculated by integrating the total change in storage landward of the point as indicated by the curves in Figure 6 and dividing by the depth of sediment to obtain a depth average flux. Fluxes at the creek bank decrease with time and have reversed by day 10. Peaks occur in regions in which drainage from the interior of the marsh and seepage of water from the creek converge to balance water losses due to evapotranspiration.
Figure 8

The number of periods of 10 days or longer during the months April-September in which Belle Isle marsh is not inundated at high tide (O) changes from year to year and follows an 18-year cycle. The total number of non-flooding periods (●) does not show the same cyclic behavior.
HYDRAULIC PROPERTIES OF SALT MARSH SEDIMENT

DETERMINED FROM A LYSIMETER

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ABSTRACT

A lysimeter is used to investigate the response of pore pressures to changes in water content in the sediment of a New England salt marsh and to estimate the hydraulic properties of the sediment. Desaturation causes the hydraulic properties to vary with mean hydraulic head in the lysimeter. The mean specific storativity varied from $4 \times 10^{-4}$ cm$^{-1}$ to $1 \times 10^{-3}$ cm$^{-1}$, and the saturated hydraulic conductivity was $5 \times 10^{-4}$ cm/s. Saturation-desaturation and changes in root volume were the dominant mechanisms of water storage.
I. INTRODUCTION

The hydraulic properties of salt marsh sediments, hydraulic conductivity and specific storativity, are difficult to determine by techniques usually applied to soils. The use of core samples for laboratory tests introduces errors related to sample disturbance and wall flow in permeameters. The roots of marsh grasses influence the physical properties of the sediment and it is difficult to maintain their natural state when conventionally sized cores are used. Techniques for estimating hydraulic properties in situ, such as those that rely on pumped or transient well response, are limited by theoretical considerations to use in saturated conditions, away from the sediment surface where the contribution of desaturation to changes in water storage is of interest. Large pressure gradients imposed by laboratory and field tests can also change the structure of the peaty sediments (Rycroft et al 1975), affecting their hydraulic properties.

Good estimates of the hydraulic properties of marsh sediments can be obtained by using a lysimeter, a very large, relatively undisturbed sample of sediment maintained under natural conditions in the laboratory. Dacey and Howes (1984) have used this approach in an investigation of the response of pore pressures to evapotranspiration. They were able to determine the storage coefficient of the sediment by manipulating the water content of the sediment in the lysimeter and observing the resulting changes in pore pressure. This paper presents a method for estimating the mean hydraulic conductivity of marsh sediments in a lysimeter by analyzing the transient response of pore pressure to an addition of water to the sedi-
ment. The hydraulic properties vary with pore pressure; however, numerical simulations show that a linearized form of the differential equation for mass conservation of pore water, in which the hydraulic properties are assumed to be constant and uniform, is an adequate basis for the estimation of the mean hydraulic conductivity from the observed response of pore pressure.

II. MATERIALS AND METHODS

Study Site

Belle Isle marsh is the last remnant (100 hectares) of the once extensive marshes in Boston, Massachusetts. The marsh sediments consist of a layer of clayey peat (90% ash free, Chen 1986), 70 cm deep, which grades into a layer of gray silty clay. Relatively impermeable clay underlies the study area at 170 cm. The vegetation over most of the marsh consists of a mix of Distichlis spicata and Spartina patens typical of drier high marsh conditions. The hydrologic regime is characterized by alternating periods of daily flooding of the marsh surface and periods in which the surface does not flood for several days. The average tidal range in the creeks is 2 m. The marsh surface is flooded at high tide about 20 days per month during the summer months. The periods of surface flooding are related to the spring-neap cycle but do not follow a regular 28-day cycle.

Lysimeter Construction and Instrumentation

An undisturbed sample of Belle Isle marsh sediment (52 cm deep by 58 cm in diameter) was obtained with vegetation intact by the following
procedure. A steel cylinder 60 cm deep was made by cutting the bottom off of a standard 55-gallon drum. The cylinder was carefully worked into the sediment until the desired sample depth (~50 cm) was reached by simultaneously excavating a trench around the outside of the cylinder, a little in advance of the leading edge. A pallet was then constructed under the cylinder and the enclosed sediment so that both could be lifted to the marsh surface. A bottom constructed from 1 in (nom.) plywood and a neoprene gasket was installed in the field, immediately after retrieval, to minimized water loss and the disturbance of the sediment. Final waterproofing of the lysimeter was done in the lab with silicone caulk. A small gap between the sediment and the cylinder wall was filled by rodding in bentonite through the full depth of the lysimeter around its circumference, Figure 1.

Four tensiometers were installed horizontally through the side of the lysimeter to the center of the sediment at 7, 20, 33, and 47 cm below the sediment surface, as shown in Figure 1. The tensiometers were connected to pressure transducers (MicroSwitch model 143PC) that were connected in turn to a digital data logger through signal conditioning circuitry. The use of pressure transducers assured nearly instantaneous (< 1 second) response in the tensiometer readings to changes in pore pressures in the sediment. The pressure transducer outputs were conditioned to provide a proportional 0-4.5 V output in response to a 0-55 cm variation in hydraulic head in the lysimeter.

The digital data logger provided 8-bit precision over the input range of 0-5 V and averaged each signal over a period of 2 seconds to provide a
measure of noise immunity. Tensiometer data were recorded during the experiments at the rate of one reading every 64 seconds for each tensiometer. The resolution of the overall tensiometer data system obtained in practice was ±1.5 mm of head. The sampling frequency and data resolution were sufficiently good to characterize transient hydraulic head responses of typically 7 cm occurring over a period of 1-2 hours.

Lighting for the lysimeter was provided by direct sunlight, diffuse sunlight, and artificial light provided by a bank of four 250-watt incandescent lamps and a fluorescent "grow light" suspended approximately 1 meter above the sediment surface. The water table was maintained within 20 cm of the sediment surface except during experiments. Fresh water was added to offset evaporative losses in order to avoid the build up of dissolved salts in the sediment. The natural vegetation and a community of amphipods, isopods, spiders and mites remained active in the sediment from the time of collection in September 1984 to the end of testing in July 1985.

III PROCEDURE AND RESULTS

The experimental procedure was simple; a quantity of water is poured on the surface of the sediment and the response of each tensiometer is recorded and analyzed to determine the hydraulic properties of the sediment. The response of hydraulic head generally follows the sequence sketched in Figure 2. Initially, the pressure head

$$\psi = \frac{P}{\gamma_w} ; \gamma_w = \text{unit weight of water}$$  \hspace{1cm} [3.1]
is hydrostatically distributed, $\frac{d\psi}{dz} = -1$, and the hydraulic head, $h = \psi + z$ is uniform with a value $h_i$. When water is poured on the sediment surface it infiltrates, causing the hydraulic head near the surface to increase, and pore water begins to move farther down into the sediment due to the head gradient. This is the infiltration phase of the response. In most cases infiltration stops before equilibrium head conditions are reestablished because the entire volume of the water addition has been incorporated into the sediment near the surface. A period of vertical redistribution of pore water follows in which water continues to flow until a new equilibrium head, $h_f$, is reached.

Two types of water additions were used. The majority were limited to 500 ml so that the associated changes in pressure head were small, in the range 3-10 cm. These water additions were made in series of four or five pulses, beginning when the pore pressures in the sediment were low and continuing until the equilibrium pore pressure at the sediment surface was at or near atmospheric pressure when no more water would infiltrate into the sediment. The lysimeter was covered by a plastic bag at least 2 hours in advance of the first water addition and remained covered throughout each series of additions to prevent evaporation and to impose a condition of zero vertical flux in the sediment at equilibrium. Subsequent water additions were made only after the tensiometer output indicated that static pressure conditions had been reestablished. Four series of 500 ml pulse additions were made. The record of hydraulic head in Figure 3 is a typical result.
In five additional cases, the amount of water added was made large enough to exceed the amount that could be infiltrated into the sediment. This was done to simulate the conditions in the marsh during inundation and infiltration by tides. Changes in pressure head in response to these water additions were large, 15-40 cm.

Changes in the bulk volume of the sediment were monitored during some of the water additions by observing the movement of the surface of the sediment with a dial gauge mounted on the rim of the lysimeter. The sediment was confined by rigid walls except at the surface, so volume changes were entirely manifested as vertical movement of the surface. The dial gauge was mounted so that its plunger rested on a metal target in the middle of the sediment surface. Surface movements of 10 \( \mu \text{m} \) could be detected. Actual displacements were on the order of 0.2 mm for a 500 ml water addition. The displacement of the sediment surface was roughly linear with the square root of time, as would be expected for infiltration into a saturated soil (Hemond et al 1984, Smiles 1974), Figure 4, but the change in sediment volume was too small to account for most of the change in water content.

IV DISCUSSION

The response of head to changes in water content of a variably saturated soil is governed by the mass conservation equation (Bear 1979)

\[
S_s(\psi) \frac{\partial \psi}{\partial t} = \frac{\partial}{\partial z} \left[ k(\psi) \left( \frac{\partial \psi}{\partial z} + 1 \right) \right]
\]  

[4.1]
in which $S_s$ and $K$ are the specific storativity and the hydraulic conductivity respectively. Hillel et al. (1972) present a means of estimating local averages of $S_s(\psi)$ and $K(\psi)$ at a number of points in a soil profile from changes in the distribution of pressure head following infiltration of water at the surface. However, the method assumes that simultaneous point estimates of the volumetric water content $\theta(\psi)$ are also made. Salt marsh sediments are characterized by high porosity and near-saturated conditions. Typical values of the volumetric water content are on the order of 80%, whereas changes in the mean water content of the sediment in the lysimeter due to the addition of 500 ml of water are less than 0.5%, too small to be detected by non-destructive means. Consequently, only the bulk averages of $S(\psi)$ and $K(\psi)$ can be determined from the analysis of the pressure response in the lysimeter.

**Specific Storativity**

An estimate of the mean specific storativity of the sediment in the lysimeter is obtained from the net change in head that results from each water addition.

$$<S_s> = \frac{1}{V_B} \frac{V_W}{(h_f-h_i)}$$

[4.2]  

in which $V_w$ is the volume of water added (500 ml) and $V_B$ is the bulk volume of the sediment in the lysimeter (137 l). The estimates derived from 21 experiments are shown in Figure 5. The precision of the digital head measurements is the major source of error in the determination of specific
storativity using [4.2]. The effect is more pronounced for smaller changes in head, corresponding to larger values of storativity, but the error is limited to ±25 per cent of the storativity estimates in all cases.

There are four mechanisms of water storage in salt marsh sediment; saturation-desaturation of the pores, compression of air-filled roots and rhizomes, compression of trapped gas pockets, and changes in the bulk volume of the sediment (Nuttle 1986a). Water storage by compression of gas pockets is thought to be much less than that due to bulk volume change. The net surface displacement shown in Figure 4 (Δz = 0.46 mm) can be used to estimate the specific storativity by bulk volume change:

\[ S_e = \frac{1}{D} \frac{Δz}{Δh} \]  [4.3]

in which D is the depth of the sediment (52 cm) and Δh is the net change in head, 24 cm. The resulting storativity is 3.7 × 10⁻⁶ cm⁻¹, less than 10% of the total specific storativity. Therefore, changes in water storage in the lysimeter are largely the result of the mechanisms of saturation-desaturation of the pores and changes in the volume of hollow roots. The relative contribution of these two mechanisms is of interest.

Chen (1986) has studied the storage properties of small cores from Belle Isle marsh in detail using the tension plate procedure. Hillel (1980). Basically in this procedure the pore pressures in the sediment are controlled externally, and the storativity of the sediment is determined by incrementally decreasing the pressure and observing the volume of water that drains. The results of a typical experiment are shown in Table 1.
Characteristically, the storativity changes at a critical pressure head in the range -10 to -20 cm.

The tension plate data can be used to abstract the storativity due to root compressibility from the estimated average lysimeter storativity, Figure 5. The lysimeter can be divided into two layers at the water table, the depth at which $\psi = 0$. The mean lysimeter storativity is a depth-weighted average of the storativities of each layer. The storativity of the upper layer can be taken as the mean value from the tension plate tests for water table depths less than 10-15 cm, $10 \times 10^{-4}$ cm$^{-1}$. The mean storativity in the lysimeter is $5 \times 10^{-4}$ cm$^{-1}$ for water table depths less than 12 cm, so the storativity below the water table can be calculated as $4 \times 10^{-4}$ cm$^{-1}$. As the contribution of changes in gas volume and bulk volume is small, change in the volume of hollow roots is assumed to account for most of the specific storativity below the water table in the lysimeter.

The mean storativity increases sharply as the mean hydraulic head decreases below 38 cm (-14 cm pressure head at the surface). This would be expected if the storativity of the sediment near the surface in the lysimeter increases at some critical pressure as in the case of the tension plate data, Figure 6. However if the sediment in the lysimeter continues to behave in the manner of the tension plate results, one would expect the lysimeter storativities to continue to increase as more sediment desaturates and yields water at the higher rate, but the lysimeter storativities inexplicably decline with further decrease in head.

One reason for the inconsistency between the lysimeter and tension plate results may be related to differences in the direction of the change
in water content. The lysimeter method determines the storativity of the sediment on resaturation while the tension plate method measures the storativity on draining. In general, draining and drying measures of storativity are not comparable if there is significant hysteresis in the water content versus $\psi$ relation, but it is difficult to say what effect hysteresis has here because the surface sediment in the lysimeter saturates and drains with every water addition.

Another reason for the aberrant behavior of the mean lysimeter storativity is suggested by the different ways in which equilibrium conditions are defined in the lysimeter and in the tension plate procedure. The lysimeter is assumed to be at equilibrium with respect to pore water movement when the hydraulic head is uniformly distributed, within the precision of observation. Theoretically there is no water movement and no change in water content anywhere in the sediment under uniform head conditions. Equilibrium conditions are determined in the tension plate tests by observation of zero drainage. That these two conditions may not be the same is indicated by the difference between time it takes to reach equilibrium in the lysimeter and in the tension plate tests. In the lysimeter this is about 2–3 hours in all cases, about the same as the time required for equilibrium in the tension plate test for pressure heads greater than $\psi_{\text{crit}}$ (Chen 1986). However, as the pressure head drops below $\psi_{\text{crit}}$ the time to equilibrium in the tension plate tests increases to 6–9 hours or more. This implies that changes in water content persist in the sediment under "equilibrium" pressure conditions. This may explain a drift in the lysimeter head observations that was observed from time to time following the recovery from a water addition and attributed to evapotranspiration.
Failure to recognize the persistence of transient conditions in the lysimeter would lead to an underestimation of the storativity in the lysimeter. The occurrence of disequilibrium at uniform head conditions is also an indication that the representation of the water balance in the sediment in terms of continuum properties averaged over small scale variations in the sediment, which is the basis for Equation [4.1], is incorrect. The details of the sediment structure may have importance on the large scale.

There is reason to believe that voids and channels associated with roots of the marsh grasses play a large role in the movement of water in the sediment (e.g. dye studies by Chen 1986). The consequences of water flow in a soil with macro-pores have been reviewed by Beven and Germann (1982). If a significant fraction of the water storage is in fine-grained material or dead-end pores, and if the conductance of water is by a network of a relatively few larger pores, then the approach to equilibrium following the addition of a water pulse would occur in two stages. First, water flows in the conductance network until static pressure equilibrium is established in the sediment. At the end of this stage the distribution of head is uniform in the large pores. Second a redistribution of water occurs between the large pores that make up the conductance network and the small pores that make up the storage network. The end of this stage corresponds to the true equilibrium conditions, indicated by cessation of water drainage in the case of the tension plate experiments. This view of water movement in sediment has implications for mass transfer and the maintenance of anoxic zones in partially saturated sediment.
Analysis of Transient Head Response

The analysis of the transient response of head in the sediment for the purpose of estimating the mean hydraulic conductivity is based on a linear approximation to the equation of mass conservation [4.1];

\[
\frac{\partial h}{\partial t} = \alpha \frac{\partial^2 h}{\partial z^2} : \langle \alpha \rangle = \frac{\langle K \rangle}{\langle S_z \rangle}
\]  

[4.4]

The hydraulic properties of the sediment are assumed to be constant, uniform, and equal to the mean value of the actual properties of the sediment in the lysimeter. The boundary conditions on head are shown in Figure 2 and discussed in detail in the appendix. The theoretical head response obtained by solution of [4.4] is linear with respect to the net change in head, which is determined by the mean storativity. The unit head response is fully determined by the sediment property \( \langle \alpha \rangle \) and a parameter introduced to account for the change in the boundary condition at the sediment surface at the end of the infiltration phase of the response:

\[
\Gamma = \frac{\psi_f - \psi_i}{\psi_i}
\]  

[4.5]

in which \( \psi_i \) is the initial pressure head at the sediment surface and \( \psi_f \) is the value at equilibrium. Water will not infiltrate into the sediment under equilibrium pressure conditions when the pressure head at the surface has its maximum value \( \psi = 0 \). When \( \Gamma = 1.0 \) then the amount of water added to the lysimeter was sufficient to maintain infiltration across the upper boundary through the entire period of head response. The other extreme,
\( \Gamma = 0 \), is approached by conditions in which the infiltration phase of the response accounts for a small portion of the total response time because the volume of water added was small, resulting in a net change in head which is small with respect to the magnitude of \( \psi_i \).

Equation [4.4] was solved numerically for the unit head response at each tensiometer to generate a set of type curves, shown in Figure 7 for the upper and lower tensiometers. The unit head response is given by

\[
h^*(t) = \frac{h(t)-h_i}{h_f-h_i} \tag{4.6}
\]

The solid curves in Figure 7 are solutions for the limiting cases of the upper boundary condition. Analytical solutions for the head distribution are presented in the appendix for these cases.

The observed head responses were normalized, [4.7], and plotted against \( \ln(t) \), \( t \) being the time since the addition of the water pulse. A fit was then established between the data and the type curve for the appropriate value of \( \Gamma \), Figure 8. The correspondence between data and model results is not particularly good as a consequence of the assumption that the hydraulic properties are constant and uniform. The interest is in matching the time scale of the model to the head transient, so attention was paid to matching the point of convergence to equilibrium conditions in establishing a fit between model and the data. Once a fit had been made then the difference between the superimposed values at any point on the abscissa is
\[ \ln(\langle \alpha \rangle t)_{\text{type curve}} - \ln(t)_{\text{data}} = \ln(\langle \alpha \rangle)_{\text{data}} \]  

[4.7]

by virtue of the fact that the normalized head at a given depth is a function only of the product \( \langle \alpha \rangle t \) and the boundary condition parameter \( \Gamma \).

Values of the fitting parameter \( \ln(\langle \alpha \rangle) \) are plotted in Figure 9. The curve fitting was done by eye with an estimated standard error of \( \pm 0.2 \) ln-units.

The value of the fitting parameter \( \langle \alpha \rangle \) varies with head in two ways which reflect the true non-uniform, temporally variable nature of the hydraulic properties of the sediment. First \( \langle \alpha \rangle \) is seen to depend on the mean value of head in the lysimeter during the response. This is a result of pressure dependence of the depth average \( \langle \alpha(\psi) \rangle \) in which \( \psi(z) \) is determined to a large degree by the static distribution of the pore pressure. The addition of small volumes of water (500 ml) limits the degree to which pressures differed from hydrostatic during the response, so the filled shapes in Figure 9 are the best estimates of \( \langle \alpha(\psi) \rangle \).

Second, the estimates of the fitting parameter depend on the net changes in head during the response. The net change in head was limited to the range 3–10 cm for the case of the 500 ml water additions (●), but the change in head ranged from 15–40 cm when the amount of water added to the lysimeter exceeded the infiltration capacity of the sediment (○). When an unrestricted amount of water is allowed to infiltrate into sediment that is partially saturated at the surface, the partially saturated sediment quickly becomes fully saturated and remains saturated. This results in a large increase in the hydraulic conductivity and the head response is faster than
one would estimate based on the bulk average hydraulic conductivity corresponding to mean head conditions in the sediment.

Near saturation in the lysimeter (mean head greater than 40 cm) the theoretical response based on the linearized mass conservation equation matches the actual response fairly well, and the pressure dependence of the fitting parameter observed in other situations is mitigated. There is a factor of 2 increase in the fitted value of \( \langle \alpha \rangle \) for unlimited infiltration relative to that for a 500 ml pulse response, which is about the same magnitude as the variability in estimates of hydraulic conductivity obtained by current laboratory and field methods (Chen 1986, Knott et al. 1984). Therefore the error introduced by applying the linearized model to predict the head response to infiltration under conditions of partial saturation is comparable to the uncertainty introduced by the variability in hydraulic conductivity estimates.

\[ \text{Hydraulic Conductivity} \]

Estimates of mean hydraulic conductivity, Figure 10, were obtained from estimates of \( \langle \alpha \rangle \) and the corresponding estimates of specific storativity for the 500 ml water additions. Mean hydraulic conductivity declines with decreasing water table elevations due to the desaturation of the surface layer of sediment. The value of hydraulic conductivity for mean hydraulic head greater than 43 cm is taken as the mean saturated hydraulic conductivity in the lysimeter; \( \ln(\langle K_{\text{sat}} \rangle) = -7.6, \langle K_{\text{sat}} \rangle = 5.0 \times 10^{-4} \). This value compares well with estimates of hydraulic conductivity by the falling head procedure. Chen (1986) determined the hydraulic conductivity
of 40 sediment samples obtained from cores over the 0–80 cm depth interval in Belle Isle marsh and found that the values were lognormally distributed with mean $\ln(K_{sat})$ of $-7.58$, $K_{sat} = 5.1 \times 10^{-4}$ cm s$^{-1}$, and a standard deviation of 0.94 ln-units.

**Numerical Simulations**

The basis for estimating the hydraulic conductivity from the transient pressure response is the assertion that the time scale of the pressure response is accurately represented by solutions to the linearized mass balance equation [4.4] in which the uniform, constant values of $K$ and $S_s$ are taken to be the spatial averages of the variable hydraulic properties of the sediment. This is demonstrated to be true by applying the fitting procedure used with the lysimeter data to estimate the mean hydraulic conductivities from a series of numerically simulated lysimeter head responses in which a known variable hydraulic conductivity $K(\psi)$ is used. The simulated head responses were calculated using a numerical model of flow in saturated and partially saturated porous media (Narasimhan and Witherspoon 1977; Narasimhan, Witherspoon, and Edwards 1978). The simulation results are in effect solutions to the full non-linear mass conservation equation [4.1].

Simulations were run using two different forms of the function $K(\psi)$, an exponential decay in hydraulic conductivity with increasingly negative pressure heads and a step decrease in conductivity at $\psi = -10$ cm, Table 2. The conductivity functions were parameterized so that the variation of the depth-averaged conductivities with hydraulic head was in reasonably good
agreement with the values determined for the lysimeter, Figure 10. Depth averages were calculated for this purpose by the relation

\[
\langle K \rangle = e^{F_m} ; \quad F_m = \frac{1}{D} \int_{-D}^{0} \ln(K(\psi)) \, dz
\]  

[4.8]

This is the geometric mean of K that has been determined to be the pertinent average K to apply to a soil in which the hydraulic conductivity is randomly distributed (Bouwer 1969). The distribution of pressure head was taken to be hydrostatic, so the mean head defined by [4.8] is a function only of the mean hydraulic head in the lysimeter, as in Figure 10.

The specific storativity was represented as spatially uniform and constant, and its value was determined from the lysimeter storativity estimates in Figure 5 and the mean hydraulic head during each simulated water addition.

The results for a series of five 500 ml water additions are shown in Table 2. The agreement between data and simulation results is reasonably good, indicating that the estimation of hydraulic conductivity is not unduly biased by using the approximate model, Equation [4.4]. The discrepancy between the fitted K estimates and calculated mean K is on the order of 50 percent of the estimated value, somewhat higher than the errors associated with the fitting procedure itself.
APPENDIX

The approximate response of hydraulic head to water additions to the lysimeter is given by solutions to the equation of conservation of mass [4.3] subject to the following conditions. The flow of water is vertical and the bottom of the lysimeter is impermeable.

\[ t > 0 \quad , \quad z = -D \quad ; \quad \frac{dh}{dz} = 0 \]  \hspace{1cm} [A.1]

where \( D \) is the depth of sediment. The initial distribution of head is uniform and is fully described by specifying the pressure head at the sediment surface \( \psi_1 \big|_{z=0} \)

\[ t \leq 0 \quad h_0 = \psi_1 + z \]  \hspace{1cm} [A.2]

This establishes the maximum depth of water that can be added to storage in the sediment \((-DS_\psi \big|_{z=0})\) because the ultimate storage capacity of the sediment is reached at static pressure conditions when \( \psi \big|_{z=0} = 0. \) At time \( t = 0, \) a depth of water \( I^w \) is applied to the sediment surface. While water is ponded on the surface, the pore pressure at the surface is held at atmospheric pressure \( \psi \big|_{z=0} = 0 \) and water infiltrates into the sediment. If the amount of water added to the lysimeter is less than the amount that can be accommodated by the sediment then the boundary condition at the surface must be changed to a zero flux condition when the cumulative infiltration depth \( I(t) \) is equal to the depth of water added \( I^w. \)
\[ t > 0 , \ z = 0 ; \quad h = 0 \quad I(t) < I^* \]  \[ \frac{dh}{dz} = 0 \quad I(t) < I^* \]  \[ [A.3] \]

The rate of infiltration when water is ponded on the surface is a function of the initial change in head at the sediment surface and the properties of the sediment. The depth of water added \( I^* \) determines the maximum net change in head that can occur \( \Delta \psi \leq -\psi^* \); \( -\psi^* = \frac{I^*}{S} \). Therefore, the surface boundary condition is fully specified by the ratio \( \Gamma = \frac{\Delta \psi^*-\psi_1}{S} |_{z=0} \).

There are two cases of the surface boundary condition \([A.3]\) for which a simple analytic solution for the head response is possible. In the first case the amount of water added to the lysimeter is equal to or exceeds the unfilled storage capacity of the sediment. Under these conditions \( \Gamma = 1 \) and the boundary condition \([A.3]\) becomes

\[ t > 0 \quad z = 0 ; \quad h = 0 \]  \[ [A.3.1] \]

The solution for the head response for this case can be obtained from solutions presented by Carslaw and Jaeger (1959).

\[ h(z, t) - h_0 = \psi_1 \left[ -1 + 4 \sum_{n=0}^{\infty} \frac{1}{(2n+1)} \exp \left( -\frac{a(2n+1)^2 \pi^2 t}{4D^2} \right) \sin \left( \frac{(2n+1)\pi z}{2D} \right) \right] \]  \[ [A.4] \]

in which \( \alpha = \frac{K}{S} \).

In the second case the net change in head response to the water addition is small with respect to \( -\psi_1 \) and \( \Gamma \sim 0 \). This corresponds to the
situation in which the water added to the lysimeter is instantly infiltrated into the sediment. The solution can be obtained as the Green's function for an instantaneous plane source of head at the sediment surface.

\[
 h(z,t) - h_0 = \frac{I_w}{D^2 S_s} \left[ 1 + 2 \sum_{n=1}^{\infty} \exp\left(-\frac{n^2 \pi^2 t^2}{D^2}\right) \cos \frac{nn\pi(-z)}{D} \right] \tag{A.5}
\]
REFERENCES


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Table 1

Specific Storativities from Tension Plate Tests on Belle Isle Marsh Sediment (Chen 1986) (storativity values in $10^{-4}$ cm$^{-1}$)

<table>
<thead>
<tr>
<th>Test</th>
<th>$S_s^{(1)}$</th>
<th>$S_s^{(2)}$</th>
<th>$\psi_{crit}$ (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$\psi &gt; \psi_{crit}$</td>
<td>$\psi &lt; \psi_{crit}$</td>
<td></td>
</tr>
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Table 2

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Exponential model

\[ K = K_s e^{\alpha \psi} \quad ; \alpha = 0.3 \]

\[ K_s = 5.53 \times 10^{-4} \text{ cm/s} \]

\[ \psi = \text{cm of water} \]

Step model

\[ K = K_s \quad \psi > \psi_B \quad K_s = 5.53 \times 10^{-4} \text{ cm/s} \]

\[ K = K_u \quad \psi \leq \psi_B \quad K_u = 1.37 \times 10^{-4} \text{ cm/s} \]

\[ \psi_B = -10 \text{ cm} \]
Figure 1

Cutaway view of the lysimeter showing placement of the tensiometers.
Response of hydraulic head in lysimeter following the addition of a water pulse. A) Initial pressure distribution is hydrostatic and hydraulic head is uniform. B) Water infiltrates into the sediment and increases head near the surface. C) Infiltration stops, but water continues to flow in the sediment until a new uniform hydraulic head is obtained. The difference in head $h_f - h_i$ is determined by $S_s$, and the time to reestablishment of equilibrium conditions depends on the ratio $\alpha = K/S_s$. 
A) INITIAL CONDITIONS

B) INFILTRATION

C) REDISTRIBUTION
Figure 3

Record of tensiometer data for a series of water additions. Head response of the tensiometer at 7 cm depth leads the response of the others. All water additions in the series were 500 ml. Differences in the net change in head and in the time scale of the responses result from changes in the hydraulic properties of the sediment.
Infiltration of 1.5 l of water resulted in a small increase in the bulk volume of the sediment in the lysimeter. The change in sediment volume is only 117 ml.
Figure 5

Mean specific storativity of the lysimeter determined from the net change in hydraulic head in response to an addition of 500 ml of water. Data are for repeated series of water additions; each shape is for a different series.
Cumulative water loss during a tension plate experiment on an 8 cm-deep surface sample of Belle Isle sediment. The rate of change of water content with pressure head is the specific storativity. The change in storativity at $\psi_{\text{crit}}$ is characteristic of all tests. (Chen 1986).
Figure 7

Type curves for upper and lower tensiometer response. $\Gamma$ is related to the proportion of the time that infiltration takes place during the head response. Normalized head is the change in head relative to initial conditions, divided by the net head change $h_f - h_i$. Solid curves correspond to the limiting cases for the surface boundary condition described in the text.
Figure 8

Example of fitting data to type curve for a pulsed water addition. Each curve is for the corresponding tensiometer, top to bottom. Data is shifted horizontally until the best match is obtained at the point of convergence to equilibrium. $\langle \alpha \rangle$ is determined by comparing abscissa values of the type curve, ln($\langle \alpha \rangle t$), and the data, ln(t).
Values of the fitting parameter, $\ln(\langle \alpha \rangle)$, depend on the mean hydraulic head in the lysimeter. Error bars are the estimated error of fitting the data to the type curves. The filled circles are for head response to 500 ml water additions in which the net change in head was limited to 3-10 cm. The open circles are for head response under "natural" infiltration conditions in which the sediment surface was flooded by more water than could be infiltrated. The net change of head in these experiments ranged from 15-40 cm.
Figure 10

Hydraulic conductivities determined from the pulsed addition data in Figure 9.
PEAT ROSE!

William Nuttle
Massachusetts Institute of Technology
May 1986
ABSTRACT

Salt marsh sediments expand and contract in response to surface loading and to changes in water content while remaining near full saturation. In Sippewissett marsh, the amount of compression of the sediments under the weight of tidal inundation indicates that the sediment is at most 1-2% gas by volume. In both Sippewissett and Belle Isle marshes, changes in the bulk volume of the sediments that occur over several days time are correlated with changes in the mean pore pressure and with the dept of sediment. Change in bulk volume change is a significant mechanism for water storage and accounts for 20% of the total change in water content in Belle Isle marsh. As sediment accumulates and the amount of water available from changes in bulk volume increases, the amount of desaturation necessary to satisfy the demands of evapotranspiration decreases. Thus, large changes in the bulk volume of the sediment that occur in the older regions of a marsh, where the sediment is deep, reduce the frequency of air entry, which has implications for the productivity of the grasses there.
I INTRODUCTION

Variation in the bulk volume of the sediment is one of several storage mechanisms active in salt marsh sediments, along with saturation–desaturation of the pores and compression of roots and trapped pockets of gas (Nuttle 1986a). The total storage coefficient of the sediment (the change in water content per unit area per unit change in hydraulic head), a parameter useful for modeling water flow in the sediment, can be determined by water balance techniques (Dacey and Howes 1984, Nuttle 1986b, 1986d). However, because of the biogeochemical implications of the desaturation of the sediment and subsequent availability of oxygen in the normally anaerobic sediments (Howes et al. 1981, Hemond and Fifield 1982), it is important to know what proportion of the total change in water content is attributable to each mechanism.

The amount of water storage by changes in bulk volume can be determined from observations of changes in surface elevation in the marsh. This paper presents observations of sediment volume change in two New England salt marshes during daily tidal inundation and over the spring-neap cycle of flooding and non-flooding periods. The magnitude of the associated changes in water content is compared to that due to other mechanisms of water storage in the sediment. Movement of the soil surface has been observed outside of salt marshes in response to water withdrawal from a well (Davis et al. 1969) and meteorologic forcing (Ward 1953), and Gates (1940) has reported a 66 cm displacement of the surface of a bog. The compressibility of salt marsh sediments has been assumed to be the dominant
mechanism of water storage in hydrologic models by Hemond and Fifield (1982) and Hemond et al. (1984), but until now, no study has been made specifically of the importance of changes in sediment volume to marsh hydrology. Changes in surface elevation have been observed in salt marshes at the time of rising tides (Harrison 1975).

II METHODS

Study Sites

The Belle Isle marsh is the last remnant (100 ha) of the once extensive marshes in Boston, Massachusetts. The sediment at the study site was 1.7 m deep and consisted of a clayey peat (90% ash-free dry weight) underlain by clay (Chen 1986). Surface movement associated with changes in the volume of the sediment was observed during August and September 1984 in conjunction with a study of the sediment water balance (Nuttle 1986b). The observations were repeated in 1985 with duplicate instruments 8 m apart, as a check on the previous year's results and on the variability of the phenomenon of bulk volume change. Observations in Belle Isle were made approximately 20 m from the nearest creek.

Surface movement was observed at two sites in Sippewissett marsh in Falmouth, Massachusetts. The sediment at the first site, monitored during July 1984, was 1 m deep, and the sediment at the second site, monitored during July and August 1985, was 4.5 m deep. The sediment at both sites was peaty (70% ash-free dry weight; Knott et al. 1984) and underlain by a thick deposit of sand. Core sampling revealed that the lower 2 m of sediment at the 4.5 m site was fresh water peat.
Apparatus

Simple piezometers were used to monitor hydraulic head over the depth of the sediment. The piezometers were constructed of 0.5 in (nominal) PVC pipe slotted over the lower 15 cm. At each site, three piezometers were driven to different depths to monitor changes in the mean head in the sediment. Water levels in the piezometers were read to 0.5 cm with an electrode-tipped dipstick. Contact of the electrode with the water surface causes a change in the resistance between the electrode and the sediment that can be detected with an ohmmeter. The piezometers were used to characterize the day to day variation in head so fast response time was not critical to their design.

Vertical displacements of the marsh surface were measured using the apparatus shown in Figure 1. A rigid metal frame 3 m long was supported 10 cm above the surface by 1 in (nominal) pipes driven through the marsh sediments and well into the underlying material. The frame supports were isolated from the sediment by loose fitting sleeves of PVC pipe. The distance between the frame and metal targets mounted on the marsh surface was measured with a dial gage accurate to 30 μm.

Surface movement was observed on two time scales, continuously during the period around high tide (only at the 4.5 m Sippewissett site) and daily over the monthly spring-neap cycle. Continuous measurements were made with the dial gauge permanently attached to the frame and read from a distance through binoculars to prevent disturbance by the observer walking on the marsh. Daily observations were made with a portable dial gauge, taking the average of measurements to three different targets on the surface. Dis-
turbance of the surface was minimized by working from boardwalks when taking the daily readings.

III RESULTS

Long Term Variation in Surface Elevation

Surface displacements were observed over periods of weeks (Sippewissett, Figure 2) to months (Belle Isle, Figure 3). The sediment of both marshes remained near full saturation during the period of observation. The slope of the data is an estimate of the storage coefficient (cm$^3$ per cm$^2$ per cm change in head) due to the mechanism of bulk volume change. The differences in the storage coefficient between the data sets in Figure 3 are not statistically significant. If the storage coefficients are divided by the sediment depth then an intensive property of the sediment is obtained, the specific storativity due to bulk volume change (Jacob 1949, Nuttle 1986a)

$$S_s = \frac{1}{V_B} \frac{dV_B}{dh} \quad [1]$$

where $h$ is the hydraulic head, the level of the water in the piezometers, and $V_B$ is the bulk volume of the sediment. The specific storativities are about $1.5 \times 10^{-4}$ cm$^{-1}$ for Belle Isle and the 1 m site in Sippewissett marsh, and $4.5 \times 10^{-4}$ cm$^{-1}$ for the 4.5 m site in Sippewissett.
Surface Displacement during Inundation

Displacements of the marsh surface at the 4.5 m site in Sippewissett were observed during high tide, Figure 4. The long term variations in surface elevation (above) have been removed from these data. In all cases the marsh surface responded by rising at the time of high tide, peaking somewhat later than the water level in a creek nearby, 20 m away. In one case (A) the marsh surface at the instrument did not flood, although the tide overflowed the creek bank, and the record of surface displacement is a symmetric peak. In another case (B) the marsh surface was flooded by 10 cm of water and the record of surface displacement is a clipped peak with a plateau beginning at the time of surface flooding. In the third case (C) the surface was also flooded by 10 cm of water. The variation of the water level in the creek was identical for B and C, but the surface displacement was much larger in case C. The difference between the two cases was that the pressure head ($\psi = P/\gamma_w$; $P =$ pore pressure and $\gamma_w =$ unit weight of water) in the pores at the surface was near zero in the former case and $-6$ cm in the latter. As a result, infiltration occurred in case C, but little of no infiltration could occur in case B (Hemond et al 1984).

IV DISCUSSION

Flooding and Infiltration

Hemond et al. (1984) present the following relation for the cumulative water flux into a saturated sediment during a period of inundation:
\[ I = -2\psi_o \left( \frac{K S_s t}{\pi} \right)^{-\frac{1}{2}} \]  

in which \( \psi_o \) is the pressure head at the sediment surface, \( K \) is the hydraulic conductivity and \( S_s \) is the specific storativity of the sediment, and \( t \) is the time since the beginning of inundation of the surface. This is equivalent to an expression developed by Smiles (1974) along different lines of reasoning. Nuttle (1986d) has shown that the infiltration model of Hemon et al. describes the dynamics of the pore pressure change fairly well for infiltration into nearly saturated salt marsh sediments.

If sediment volume change is a significant mechanism for water storage then it follows that Equation [2] must also describe the change in surface elevation during inundation. This proposition was tested by plotting the surface displacement due to infiltration alone, the difference between the curves B and C in Figure 4, against the square root of time of surface inundation (Figure 5). The infiltration response is proportional to \( t^{\frac{1}{2}} \), as expected, and the slope is consistent with estimates of the hydraulic properties of the peat at the 4.5 m site (\( K = 2 \times 10^{-3} \) cm/s and \( S_s = 9 \times 10^{-4} \) cm\(^{-1}\); Nichols 1985) and the observed value of \( \psi_o = -6 \) cm.

The cause of the rise and fall of the surface during peak tide, as in curves A and B, is not known. This could be the result of a tidally induced water flux across the sand-sediment interface (Hemon and Fifield 1982), or it could be the result of lateral deformation of the sediment due to surface loading by tidal flooding in adjacent areas.
Bulk Volume Change and Water Storage

The relation between surface elevation and hydraulic head (Figures 2 and 3) is linear, as would be expected for a perfectly elastic soil (Verruijt 1969, Jacob 1949). The difference between the storage coefficients due to bulk volume change at the Sippewissett sites is due in part to the difference in the depth of the sediment. The more sediment there is, the larger the volume change for a given change in head. The remaining difference is due to a difference in the specific storativity of the sediment at the two sites. The larger value of specific storativity at the 4.5 m site may reflect the influence of the fresh water wetland sediments, or it may be the result of a gradual increase in the compressibility of long buried sediments due to decomposition (Boelte 1969).

The water content in the sediments at the Belle Isle site is determined by the balance between water loss due to evapotranspiration and replenishment by tidal inundation and precipitation (Nuttle 1986b). Water content, hydraulic head and surface elevation follow a monthly cycle of drying periods, falling head and sediment shrinkage; and wet periods, rising head and sediment expansion. The elevation and head data for 1984 in Figure 3 were collected every other day or so through two flooding, non-flooding cycles in conjunction with a water balance study (Nuttle 1986b). Estimates of evapotranspiration and precipitation were used to estimate changes in the total water content of the sediment, and the total storage coefficient of the 1.7 m of sediment in Belle Isle marsh was found to be 0.11–0.16 cm/cm. The storage coefficient due to bulk volume change in Belle Isle marsh was found to be 0.024 in this study, so bulk volume change
of the sediment accounts for about 20% of the total change in water content.

The specific storativity due to the compressibility of gas pockets is

$$\frac{\gamma_w}{V_B} \frac{dV_G}{dP}$$  \[3\]

in which $V_G$ is the volume of gas contained in a volume $V_B$ of sediment (Verruijt 1969, Nuttle 1986a). This quantity can be estimated from the change in bulk volume immediately following a change in stress in the sediment during which the water content of the sediment remains constant, the so-called "undrained" response to loading (Nuttle 1986a). These conditions are satisfied for the surface loading by 10 cm of water at high tide in which no infiltration occurred (curve B, Figure 2). If the 0.1 cm rebound in the surface from high tide to the time that the surface was completely drained is attributed to the expansion of gas in the sediment then the specific storativity due to the gas pockets is $2.2 \times 10^{-5}$ cm$^{-1}$, and the associated component of the total storage coefficient is 0.01 at the 4.5 m site. This implies a gas content of 1-2% by volume.

**Effects on Sediment Aeration**

As mentioned above, the water content of salt marsh sediments is determined principally by cycles of evapotranspiration, precipitation and tidal inundation. The amount of water that is drawn from storage in the sediment during a dry period depends on the rate of evapotranspiration and the length of the dry period. These factors depend on the local climate.
season of the year, characteristics of the tides, and elevation of the marsh surface relative to the height of high tide. Therefore, the magnitude of changes in water content of the sediment are imposed by conditions external to the sediment and are roughly the same throughout the marsh, except in low areas that are flooded more than the rest of the marsh and near the creeks where additional water is lost by horizontal drainage into the creeks.

The change in head, and therefore in pore pressure, that occurs in response to a change in water content is controlled by the total storage coefficient of the sediment deposit. The storage coefficient is the sum of four components, at least one of which, bulk volume change, varies systematically with the depth and age of the sediment. For a given change in water content, the smaller the storage coefficient, the larger the change in head and in pore pressure. Large changes in pore pressure at the marsh surface increase both the likelihood that the sediment will desaturate and the depth to which air will penetrate if desaturation occurs. Large values of the storage coefficient decrease the variation in pore pressure and decrease the probability that air will enter to a given depth. Limited air entry is not simply the result of "poor drainage" in the sediment. The same amount of water can be removed from the sediment in two areas of a marsh, and air will enter at one but not the other because of differences in the total storage coefficient of the sediment.

Kaye and Barghoorn (1964) propose that the age of a salt marsh is limited by the rate at which the sediment compresses under its own weight, a process known as autocompaction. For some depth of sediment, the rate of
autocompaction plus the rate of sea level rise exceeds the accretion rate and the marsh drowns. It is suggested here that sediment compressibility has another effect with essentially the same result. Increased sediment depth increases the storage coefficient due to bulk volume change and possibly the effect of gas compressibility as well. This increases the total storage coefficient, other factors held constant, and reduces the probability of air entry. Aeration of the sediments is linked to primary production, so the effect of sediment compressibility on air penetration into the sediments offers an explanation for the lower productivity and the formation of pannes observed in the older, interior regions of salt marshes (Redfield 1972). The marsh does not so much drown by increased inundation as it suffocates.
REFERENCES


Changes in the volume of salt marsh sediments cause the surface to move relative to a rigid frame supported on the substratum. A dial gauge on the frame measures the distance to the marsh surface with an accuracy of 30 mm. PVC sleeves isolate the frame supports from the marsh sediment.
The relation between surface elevation and head varies significantly between two widely separated locations in Sippewissett marsh. The lower data is from a shallow (1 m) deposit and the upper data is from a deep (4.5 m) deposit of peat. The slopes are 0.014 (□) and 0.185 (◯) cm/cm.
Surface elevation varies linearly with depth-averaged hydraulic head in the sediment in Belle Isle marsh. Data is for adjacent locations 18 m (□, ○) and 26 m (Δ) from the nearest creek. The lowest data (□) is for the period August-September 1984, while the other data are for August 1985. An arbitrary offset has been added to the surface elevation for the purpose of separating the data in the graph. The slopes are not significantly different; □ - 0.027, ○ - 0.023, Δ - 0.019 cm/cm.
Changes in surface elevation around high tide are affected by loading of the sediment by flooding tides and by infiltration. Data are for three tides at the 4.5 m site in Sippewissett marsh with the long term variation in surface elevation removed. The upper pair of curves is for identical tides and diverges at the time of surface inundation due to differences in the amount of infiltration that occurs. The surface did not flood at the instrument during the period of observation for curve A.
SURFACE DISPLACEMENT (mm)

PERIOD OF SURFACE INUNDATION (B,C)

HOUR SINCE HIGH TIDE

A

B

C
Figure 5

Sediment volume increases in proportion to the square root of time during inundation of the surface and infiltration of water into the sediment. Data are the difference between B and C in Figure 4.