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MECHANISMS OF WATER STORAGE IN SALT MARSH SEDIMENTS: THE IMPORTANCE OF DILATION

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ABSTRACT

Direct observation of surface displacement in two New England salt marshes shows that shrinking and swelling of the sediment is an important mechanism for water storage. This mechanism accounts for 20 per cent of the total change in water content of the sediment in Belle Isle marsh, Massachusetts, and for as much as 36 per cent and 86 per cent of the total at separate sites in Sippewissett Marsh, Massachusetts. Swelling in response to infiltration appears to follow the infiltration model of Hemond *et al.* (1984). Shrinkage decreases the aeration of the sediment compared to soils that do not shrink, and reduced aeration represents a physiological stress to vegetation. For a given depth of water loss, e.g. one day's evapotranspiration, more shrinkage, less aeration, and higher stresses on vegetation will occur in deeper sediment deposits.

KEY WORDS Water storage Specific yield Storage coefficient Compressibility

INTRODUCTION

Water storage in salt marsh sediments has become an important factor in studies of nutrient dynamics and plant physiology. Changes in the amount of water stored have been used to estimate pore water fluxes and the advective transport of solutes (Howarth *et al.*, 1983; Agosta, 1985; Yelverton and Hackney, 1986; Harvey *et al.*, 1987; Nuttle and Hemond, 1988). The mechanics of water storage and gas exchange driven by infiltration and drainage are also of interest. Salt marsh sediments are periodically inundated by tides and the watertable is typically within 20 cm of the surface. Aeration, to the extent that it does occur in these nearly saturated sediments, has been found to be one of the most important factors related to the productivity of the vegetation (Howes *et al.*, 1986). In this paper we report the results of our investigations into the mechanics of water storage in salt marsh sediments and suggest some ecological consequences.

Two things occur when water infiltrates into salt marsh sediments: gases are displaced from the pores and the sediment swells. We define the volume of water added to (or removed from) *dilation storage* as being equal to the associated change in the bulk volume of the sediment, and the volume of water added to (or removed from) *saturation storage* as being equal to the volume of gas lost (or gained) by displacement by water. Saturation storage is limited to the sediment above the watertable where pressures in the 'free' (e.g. unbound) water are low enough to allow atmospheric gases to enter the sediment. Dilation storage can occur throughout the sediment. As more water is incorporated into already saturated sediment below the watertable, the sediment must dilate, i.e. the bulk volume of the sediment increases by an amount equal to the volume of water added to storage. The occurrence of dilation storage in unsaturated soils is well known in the case of 'swelling' clays. For both mechanisms, an increase in the amount of water held in storage is associated with an increase in the pressure of 'free' water in equilibrium with the water in the pores, as measured with a piezometer or similar apparatus.

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Saturation storage is the predominant storage mechanism in most soils and in phreatic aquifers, and it is an important mechanism in salt marsh sediments. Desaturation is evident from the effects of air entry on Pt electrode potentials (Howes *et al.*, 1986; Teal and Kanwisher, 1961), bubbling during tidal flooding and infiltration and the quantitative observations of Morris and Whiting (1985). The change in water content above the watertable by desaturation, measured by weight change, has been used by Howarth *et al.* (1983) and Agosta (1985) to estimate the total change in water content by drainage through the sediment. The possible significance of dilation storage has been ignored in these and other studies involving salt marsh hydrology.

Salt marsh sediments are distinguished from most other soils by an unusually high degree of compression under a load and by significant shrinkage when dried. These phenomena are related to the mechanics of dilation storage, and their occurrence argues for significant dilation storage in these sediments. Modelling studies by Hemond and Fifield (1982) and Hemond *et al.* (1984) suggest that dilation storage controls the dynamic of pore water fluxes in response to tidal forcing of underlying aquifers and tidal inundation of the marsh surface. Direct evidence for significant dilation storage is offered by observations of vertical displacement of the marsh surface (Harrison, 1975; Nuttle and Hemond, 1988), theoretical analysis of the response of piezometers to perturbations (Nichols, 1985), laboratory measurements of sediment compressibility (Knott *et al.*, 1987), and volume change observed in response to the addition of water to a lysimeter (Nuttle, 1988).

The purpose of this paper is to report further evidence for dilation storage in salt marsh sediments and to compare the magnitudes of dilation storage and saturation storage. Observations of sediment expansion and shrinkage were carried out in two Massachusetts salt marshes to test the following hypotheses: (1) Salt marsh sediments experience measurable expansion and shrinkage associated with increases and decreases in water content, respectively; (2) Changes in water content due to sediment dilation are comparable to concurrent changes in water content due to changes in the degree of saturation of the sediment above the watertable; and (3) The infiltration model proposed by Hemond *et al.* (1984) accurately describes the kinetics of infiltration into dilation storage during tidal inundation.

BACKGROUND

The interpretation of our observations are based on the assumption, widely held in the fields of hydrology and soil science, that the amount of water contained in a given mass of soil (dry weight) is a unique function of piezometric head. Piezometric head h is related to the pressure P of 'free' water in equilibrium with the soil-water mixture by the following relation

$$h = \frac{P}{\rho_w g} + z \tag{1}$$

in which ρ_w is the density of water, g is the acceleration due to gravity, and z is elevation above some common datum. Piezometric head is measured as the elevation of the water surface in a piezometer. Water content increases with increasing head. The change in water content per unit increase head per volume of soil defines a property related to the soil's ability to store water. We refer to this property as the specific storage of soil, S_s; although it is also known as the differential moisture capacity. Water storage mechanisms can be described through a mathematical expression that relates specific storage to the structure and mechanical properties of the soil. We are interested in comparing the relative magnitude of terms related to saturation and dilation storage in such an expression. A general understanding of water storage has been pursued along thermodynamical and mechanical lines more or less independently. We briefly summarize these approaches to demonstrate common themes and to clarify our terminology.

The thermodynamical approach presumes that properties of the soil–water mixture are functions of the moisture potential, ψ. In swelling soils, the moisture potential is a component of the total potential Φ, defined as the sum (Philip, 1969b, 1970)

$$\Phi = \psi + z + \Omega \text{ (units of length)} \tag{2}$$

in which Ω is the overburden potential. Philip (1969b) defines the overburden potential as 'the work per unit weight of water added to realize the required (vertical) soil movement against the (weight of the overlying material).' The total potential can be measured as the piezometric head of free water in equilibrium with the soil-water mixture.

Properties of the soil–water mixture of interest here are the void ratio, $e(\psi)$, the ratio of the volume of void space in a sample to the volume of the solid material, the moisture ratio, $v(\psi)$, the ratio of the volume of water to the volume of the solid material, and the degree of saturation, $S(\psi)$, the proportion of the void space occupied by water. These properties are interrelated as follows.

$$\mathrm{d}v = S\,\mathrm{d}e + e\,\mathrm{d}S \tag{3}$$

An increase in the moisture ratio is the sum of components due to an increase in void space (dilation) and an increase in saturation, the first and second terms in Equation 3 respectively. All three increase with increasing ψ. If the weight of the overburden is constant and vertical displacements due to swelling are negligible, then e, v, and S are also functions of Φ, i.e. piezometric head. Under these conditions, the total specific storage of the soil is given by

$$S_s = \frac{\mathrm{d}}{\mathrm{d}\phi}\left(\frac{v}{1+e}\right)\Bigg|_{Z,\,\Omega} \tag{4}$$

and by Equation 3, is the sum of components related to dilation storage and saturation storage.

The mechanical approach is based on the following relation for the distribution of total normal stress σ_T in a soil containing air at atmospheric pressure, P_a (see review by Brutsaert and El-Kadi, 1984)

$$\sigma_T = \sigma' + \chi P + (1 - \chi)P_a \tag{5}$$

σ' is the normal stress carried by intergranular stresses within the soil, the effective stress, and P is the pressure component of head in Equation 1. Atmospheric pressure is assumed to be constant and is usually taken as the reference normal stress, $P_a = 0$. χ is a factor that accounts for the effect of the degree of saturation; χ is one for fully saturated soils and decreases to zero at zero saturation. The degree of saturation is assumed to be a function of P, and this function is characteristic of a particular soil. Dilation of the soil matrix is assumed to only be a function of σ' (Terzaghi, 1943; Biot, 1955). When σ_T is constant (constant overburden weight) and vertical displacements are negligible, σ', P and head are related by

$$-\,d\sigma' = d(\chi P) = d(\chi\rho_w g h) \tag{6}$$

Under these conditions, the total specific storage of the soil is given by (Brutsaert and El-Kadi, Equation 26)

$$S_s = \rho_w g\left[S\alpha\left(\chi + P\frac{\mathrm{d}\chi}{\mathrm{d}P}\right) + Sn\beta + n\frac{\mathrm{d}S}{\mathrm{d}P}\right] \tag{7}$$

in which n is the porosity, α is the compressibility of the soil matrix, with respect to changes in σ' and β is the compressibility of water. The first and last components are the contributions of dilation storage and saturation storage respectively. The remaining component is not significant for the range of pore pressures encountered in surficial soils.

The thermodynamical and mechanical approaches both lead to expressions in which the total specific storage of a soil is the sum of independent components related to dilation storage and saturation storage. Water storage is a function only of head in the case of (1) constant weight of overburden and (2) changes in elevation that are small compared to the associated changes in head. The expressions for specific storage derived above are for a volume of soil that contains a constant mass of solid material. In general, specific storage at a point will vary strongly with pressure, P.

Water storage in a horizontally extensive soil layer is characterized by the layer's *storage coefficient*, the increase in volume of water stored per unit area of the deposit per unit increase in head. The storage coefficient is defined only for the condition of a uniform distribution of head through the thickness of the soil; in other words, P is hydrostatically distributed. The storage coefficient is the integral of the specific storage over the thickness of the soil deposit. Therefore, storage coefficient is also a sum of components related to dilation storage and saturation storage. The component of the storage coefficient related to saturation storage is called the specific yield, S_Y. Volume change due to dilation of a horizontally extensive soil is manifested entirely as a change in the soil's thickness, and thus in the elevation of its surface. The dilation component of the storage coefficient is the increase in soil thickness per unit increase in head. Note that if the compressibility of the soil, α in Equation 7, is independent of depth, then the magnitude of dilation component of the storage coefficient will increase with increasing thickness of the soil deposit. In contrast, specific yield depends only on the depth to the watertable (Philip, 1969a) and is independent of thickness.

METHODS

Changes in sediment thickness (surface elevation) and piezometric head were measured periodically for periods of weeks to months at one site in Belle Isle marsh (BI) and two sites in Sippewissett marsh (S-shallow, S-deep). These data are used to test our hypotheses 1 and 2. The period of measurement coincided with a water balance study at BI (Nuttle and Hemond, 1988) so that changes in thickness and head could be correlated with periods of net gain or loss of water storage due to precipitation, tidal inundation, evapotranspiration, and drainage. The storage coefficient due to dilation storage is the slope of the thickness versus head relation, and these values are compared with either the total storage coefficient (BI) or the specific yield (S-shallow and S-deep) to assess the relative importance of dilation storage.

Changes in sediment surface elevation were monitored continuously during tidal inundation at S-deep for comparison with the infiltration model proposed by Hemond *et al.* (1984) for salt marsh peat (hypothesis 3). This model was derived from mechanical arguments for infiltration into saturated soils. Smiles (1974) presents an infiltration model derived from thermodynamical arguments for which the model of Hemond *et al.* (1984) represents a limiting case. The load imposed on the sediment by tidal inundation is presumed to be unform throughout the marsh, but differences in the depth and duration of flooding at different points in the marsh will result in a non-uniform loading. Observations of surface displacement during inundation without infiltration are used to account for the effects of non-uniform surface loading.

Study sites

The Belle Isle marsh (N 42° 22′, W 71° 02′) is the last remnant (100 ha) of the once extensive marshes in Boston, Massachusetts. The sediment at BI is 1·7 m deep and consists of a clayey peat (10 per cent 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, 1986), and observations were repeated in August 1985 with duplicate instruments 8 m apart. Observations were made at 18 m (1984, 1985) and at 26 m (1985) from the nearest creek.

Surface movement was observed at two sites in Great Sippewissett marsh (N 40° 35′, W 70° 38′) in Falmouth, Massachusetts. The sediment at S-shallow, monitored during July and August 1984, is 1 m deep, and the sediment at S-deep, monitored during July and August 1985, is 4·5 m deep. The sediment at both sites is peaty (30 per cent ash-free dry weight; Knott *et al.*, 1987) and underlain by a thick deposit of sand. The 1 m depth site is 10 m from a creek and the 4·5 m site is 20 m from a creek. Pieces of wood retrieved by core sampling suggest that the lower 2 m of sediment at the 4·5 m site is probably freshwater peat. Apparently, this site is located over one of several ancient kettle hole marshes present within the bounds of Sippewissett marsh (Treggor, 1983).

Apparatus

Vertical movement of the marsh surface was measured relative to a rigid metal frame 3 m long supported 10 cm above the surface by 1 in (nominal) pipes driven through the marsh sediments and well into the underlying material (Figure 1). 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 gauge accurate to 0·03 mm. Periodic observations were made with a portable dial gauge, and observations were always made when the marsh was not flooded. Three measurements were made on separate targets and averaged to give one observation of surface elevation. Continuous measurements were made only at S-deep with the dial gauge attached to the frame and read from a distance. Disturbance of the surface was minimized by working from boardwalks.

Piezometers, constructed of 0·5 in (nominal) PVC pipe slotted over the lower 15 cm, were used to characterize the day-variation in head. Fast response time was not critical to their design. Water levels in the piezometers were read to ± 0·25 cm with an electrode-tipped dipstick. Hydraulic head was measured at three depths at each site, evenly distributed through the depth of the sediment. The measurements of head were referenced to the sediment surface ($z = 0$). Errors introduced by changes in surface elevation were, for the most part, within the error of reading the piezometers.

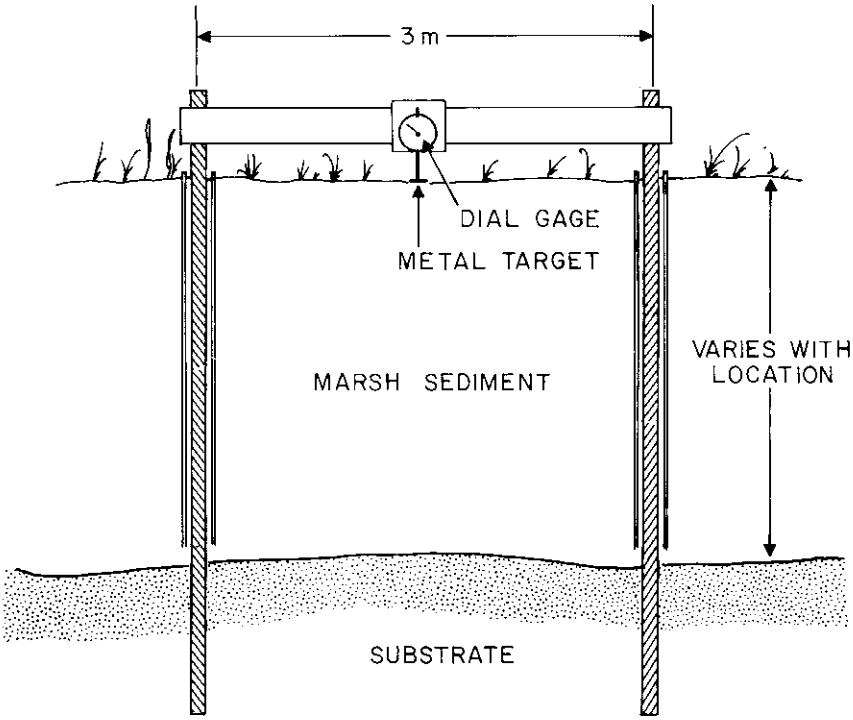

Figure 1. Changes in the volume of salt marsh sediments cause the surface to move relative to a rigid frame supported on the substratum. A dial gauge on the frame measures the distance to the marsh surface with an accuracy of 0·03 mm. PVC sleeves isolate the frame supports from the marsh sediment

Sources of Error

Experimental errors are introduced by inaccuracy in reading the dial gauge (± 0·015 mm), and compression of the surface by the observer's weight (< 0·2 mm). Vertical deflections of the rigid frame by bending were undetectable. Surface compression is not a factor in the short-term data collected during high tides because the dial gauge was read from 20 m away through a pair of binoculars at these times.

The added stress of an observer's weight depresses the marsh surface and increases hydraulic head locally. Disturbance by the observer was minimized by the use of boardwalks to (1) carry the weight of the observer away from the location of the surface elevation measurement and (2) distribute the weight of the observer over a large area, thus reducing the magnitude of the added stress on the surface. We controlled for the amount of surface depression that did occur by measuring it. The reported value, 0·2 mm, is an upper bound on these measurements. The slight depression of the surface by the observer introduces a bias into the elevation data because, with only a few exceptions, all measurements at a site were made by the same observer. A bias does not affect our estimates of the storage coefficient. The effect of the observer on measured head was controlled for by reading the piezometers before they could respond to changes in head.

We tested for the possibility that settlement of the frame contributed to the apparent relation between surface elevation and head by repeating the 1984 Belle Isle measurements after the frame had been in the marsh for a year. At the same time changes in the surface elevation were measured using a newly installed frame 8 m away.

RESULTS AND ANALYSIS

Occurrence of dilation storage

Swelling and shrinkage of the sediment was observed at all three sites, Figures 2–4. Variations introduced by experimental errors are much smaller than the observed variation in surface elevation, which ranges from 2·7 mm to 24 mm, Figure 4. Increases in sediment thickness and head at the BI correspond with periods of net influx of water into the sediment, and decreases in thickness and head corresponded with periods of net loss of water, Figure 2. The three values of piezometric head at each site were always within 1 cm of each other, except as noted below; therefore pore pressures were nearly hydrostatic at all times. The three head measurements have been averaged to obtain an estimate of the vertically-averaged head, i.e. position of the watertable. Changes in surface elevation (sediment thickness) were highly correlated with changes in mean head both in Belle Isle marsh (Figure 3) and in Sippewissett marsh (Figure 4). There is no significant difference between the slopes of the elevation versus head relation between years on the same frame or between the two frames in 1985 (Figure 3). Therefore, changes in sediment water content resulted in measurable changes in sediment volume at all three sites, hypothesis 1.

Importance of dilation storage

We assume that the depth of water incorporated into the sediment by dilation storage (volume added per unit area) is equal to the measured increase in elevation of the surface. The slopes inferred from the data in Figures 3 and 4 are estimates of the component of the storage coefficient associated with the mechanism of dilation storage. These values can be compared directly to estimates of specific yield and the total storage coefficient of the sediment to judge the relative importance of sediment compressibility as a mechanism of water storage.

The total storage coefficient of the sediment in Belle Isle marsh is estimated to be in the range 0·11 to 0·16 by Nuttle and Hemond (1988). The storage coefficient due to sediment dilation is 0·024, the mean of the slopes in Figure 3. Therefore, dilation storage accounts for about 20 per cent of the changes in water storage in Belle Isle marsh.

In Sippewissett marsh, Dacey and Howes (1984) estimate that the specific yield (saturation storage) of the sediment is between 0·02 and 0·03. The total storage coefficient is the sum of specific yield, the

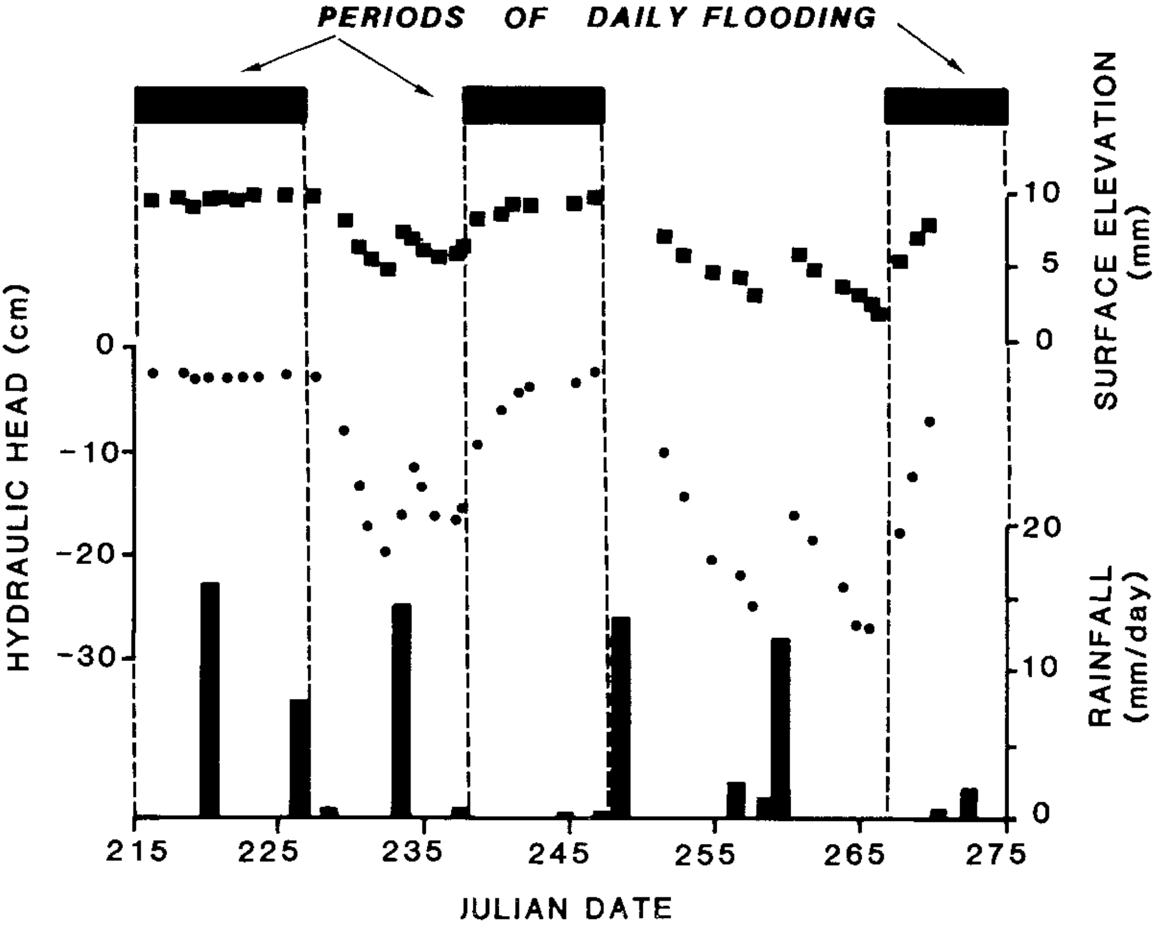

Figure 2. Changes in surface elevation (■) and depth-averaged hydraulic head (●) observed in August and September 1984 in Belle Isle marsh are highly correlated with hydrologic conditions. Daily flooding of the marsh by the highest spring tides maintains the hydraulic head at levels near the marsh surface. The sediment shrinks and hydraulic head declines as a result of a net loss of water from the sediment by evapotranspiration during periods in which the marsh does not flood. Infiltration of rainfall and the resumption of daily flooding correspond to periods of swelling of the sediment and increasing hydraulic head

storage coefficient due to dilation storage compressibility, and possibly a component due to other mechanisms as yet not described, such as the compression of air-filled roots and gas bubbles trapped in the sediment. Neglecting the contribution of other storage mechanisms, we calculate a lower bound on the total storage coefficient for the two sites in Sippewissett marsh as the sum of the specific yield (using 0·03) and the storage coefficient due to dilation storage, the slopes from Figure 4. The total storage coefficients are 0·047 and 0·207 for the 1 m and 4·5 m sites, respectively. Dilation storage accounts for 36 per cent of the total storage coefficient at the 1 m site and 86 per cent of the total at the 4·5 m site in Sippewissett marsh. These percentages are upper bounds as the contributions of all possible mechanisms for water storage have not been accounted for in the above calculations. Therefore in both marshes dilation storage accounts for a significant fraction of the total change in water content, hypothesis 2.

The storage coefficient divided by the sediment depth is an estimate of the depth-averaged specific storage, a material property. Values for specific storage are about $1{\cdot}4 \times 10^{-4}\,\mathrm{cm}^{-1}$ for BI, $1{\cdot}7 \times 10^{-4}\,\mathrm{cm}^{-1}$ for S-shallow, and $3{\cdot}9 \times 10^{-4}\,\mathrm{cm}^{-1}$ for S-deep.

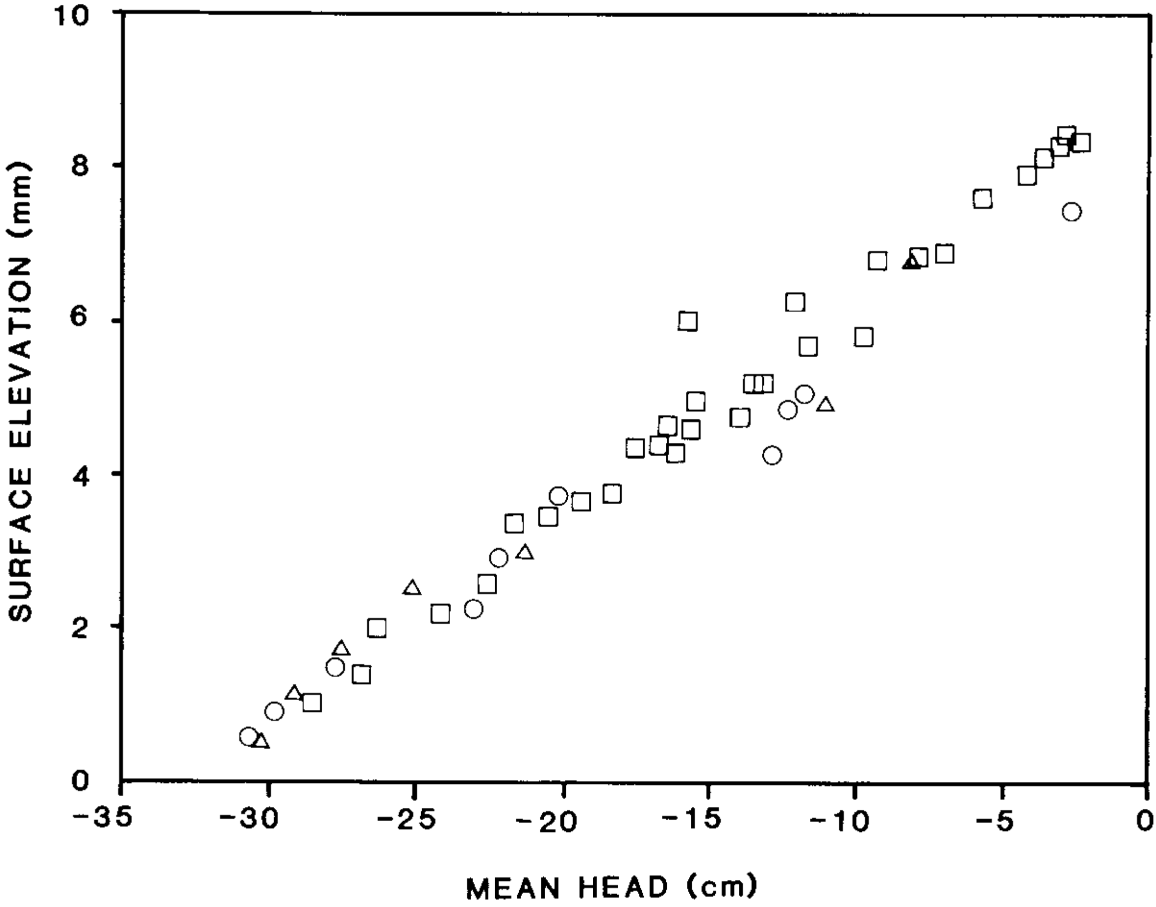

Figure 3. Surface elevation varies linearly with mean head in the sediment in Belle Isle marsh. Values of head are referenced to the sediment surface. Data are for adjacent locations 18 m (□, ○) and 26 m (△) from the nearest creek. The data (□) are for the period August–September 1984, while the other data are for August 1985. The slopes are not significantly different (80 per cent confidence interval = ± 0·005); □ — 0·027, ○ — 0·023, △ — 0·019. Error bars on surface elevation are smaller than the symbols

Surface displacement during infiltration

Continuous records of displacement of the sediment surface at S-deep site are shown for three tidal flooding events, Figure 5. Larger changes in elevation occurred *between* the flooding events as the sediment experienced a net loss of water by evapotranspiration and drainage. In case A the surface of the marsh in the vicinity of the instrument was not flooded, although flooding occurred nearby. In case B the surface flooded to a depth of 10 cm, but infiltration could not occur because, with the watertable at the surface, no additional water could be added to storage. This folowed a period of soaking rains. The average of the piezometric head readings corresponding to case B was 1·8 cm, Figure 4, but this was the result of an anomalously high reading from the lowest piezometer. This reading could be in error, and the other two piezometers indicated a piezometric head of zero.

There are two causes for the observed fluctuations in surface elevation. These are (1) infiltration and consequent dilation of the sediment as the amount of water held in storage increases, and (2) deformation of the sediment layer by the non-uniform load imposed by water ponded on the surface during flooding. Water floods the marsh to different depths, depending on surface topography, resulting in larger vertical loading in areas of deeper flooding. Horizontal variations in loading may cause lateral movement and

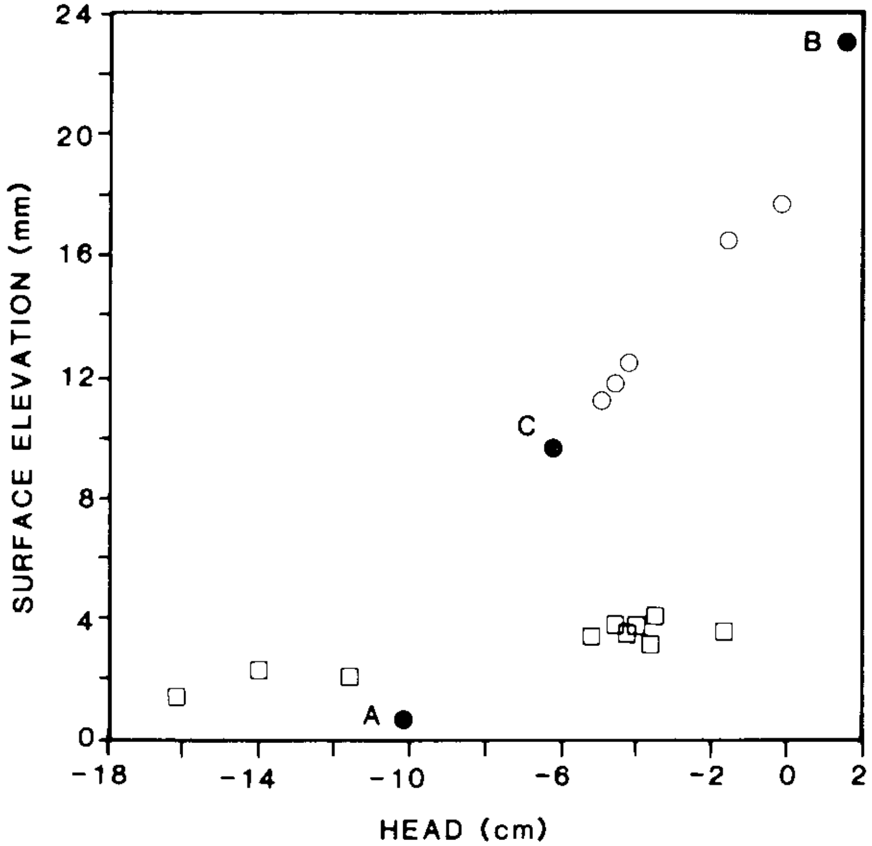

Figure 4. The relation between surface elevation and mean head varies significantly between two widely separated locations in Sippewissett marsh. Values of head are referenced to the sediment surface. The data from a shallow (1 m) deposit (□) have a slope of 0·017 cm cm^{-1} (80 per cent confidence interval = 0·002) and the slope of the data from a deep (4·5 m) deposit (○) is 0·177 cm cm^{-1} (80 per cent confidence interval = 0·01). The points A, B, and C correspond to the initial points on the corresponding curves in Figure 5. Error bars on surface elevation are smaller than the symbols

uplift in some areas. No infiltration, and therefore no dilation, occurred in cases A and B. Changes in surface elevation in cases A and B therefore the result of the complex deformation of the sediment under the non-uniform loading by flooding tides. Both causes were active in case C; the surface was flooded to a depth of 10 cm and infiltration occurred, resulting in a net increase sediment depth during the period of flooding.

Depths and duration of tidal flooding were nearly identical for cases B and C. If it assumed that the deformation of the surface due to non-uniform loading was identical in these cases, then the change in surface elevation associated with infiltration and dilation in case C can be computed as the difference between the record of surface elevation versus time in cases B and C, Figure 6. Elevation increases as the square root of time, as expected given the expression for the cumulative depth of infiltration derived by Hemond *et al.* (1984)

$$I = -\,Z\,z_O\left[\frac{K\,S_s\,t}{\pi}\right]^{1/2} \tag{8}$$

in which z_O is the initial elevation of the watertable, K is the hydraulic conductivity, S_s is the specific storage of the sediment, and t is the time since the beginning of inundation of the surface. The curve

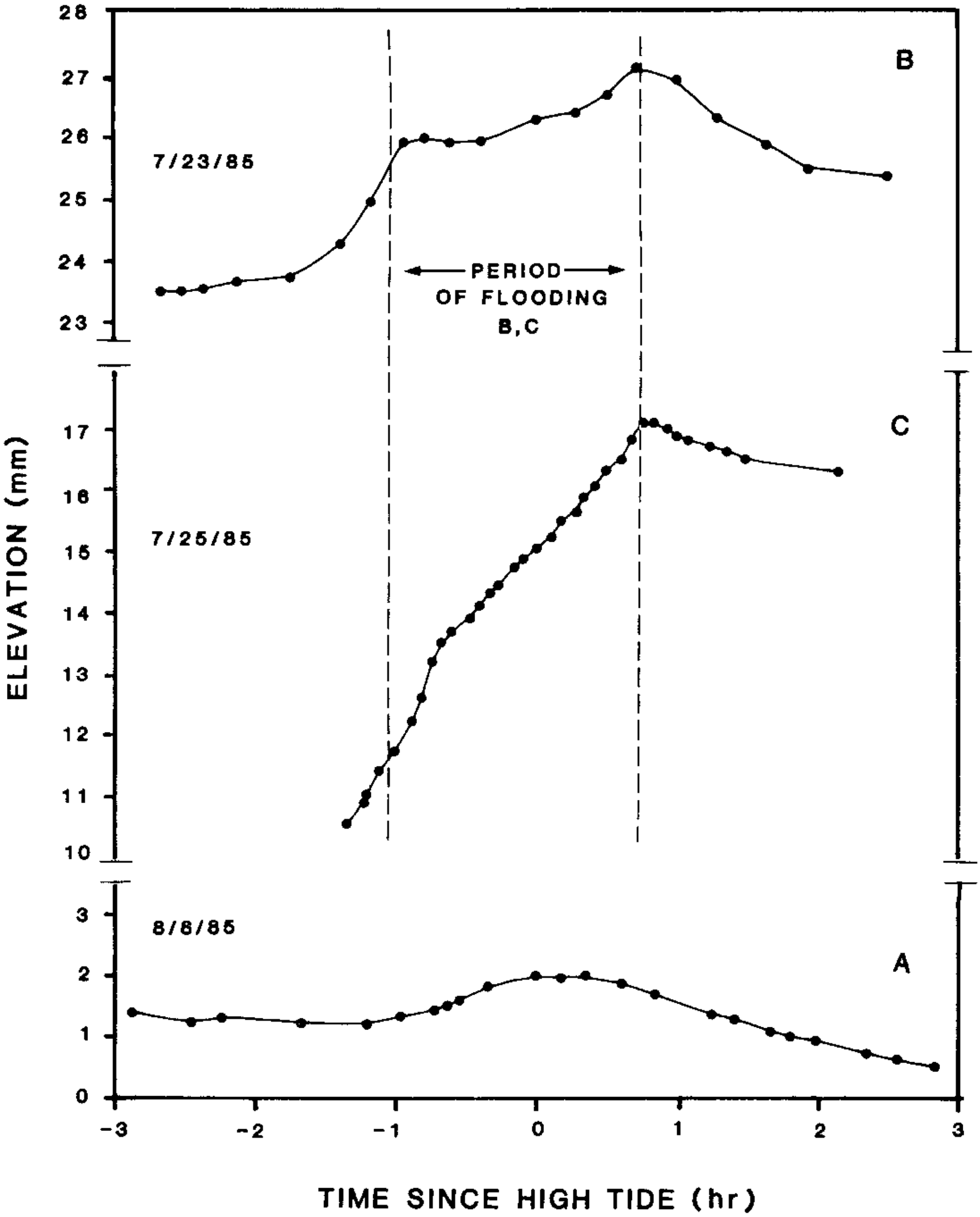

Figure 5. Changes in surface elevation around high tide are affected by loading of the sediment by flooding tides and by infiltration. Data are for three tides at the 4·5 m site in Sippewissett marsh. The curves, B and C, are for identical tides. The initial pressure head at the sediment surface, which controls infiltration, was 0 cm in the case of B and −6·2 cm for C. The surface did not flood at the instrument during the period of observation for curve A. Error bars on surface elevation are smaller than the symbols

passing through the data in Figure 6 was calculated from Equation 8 without fitting. The parameter values are $z_0 = -6{\cdot}2$ (observed, Figure 4), $S_s = 3{\cdot}9 \times 10^{-4}\,\text{cm}^{-1}$ (estimated above) and the mean hydraulic conductivity for Sippewissett marsh sediment, $10^{-3}\,\text{cms}^{-1}$, reported by Knott *et al.* (1987). Therefore, sediment swelling during infiltration follows the infiltration model of Hemond *et al.* (1984), hypothesis 3.

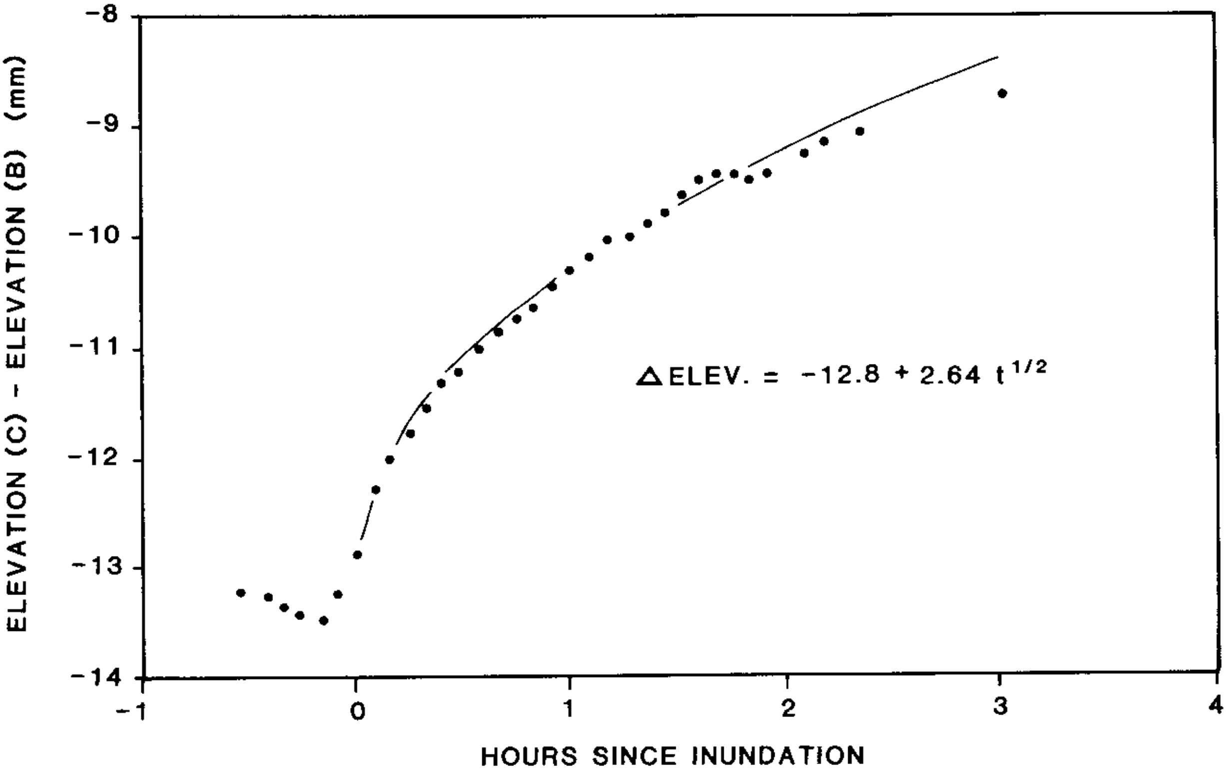

Figure 6. The displacement of the surface in curve C relative to that in curve B (elevation(C) − elevation(B)), at the same time since the initial flooding of the surface during identical tides, is related to differences in the amount of infiltration that took place. The curve is the theoretical swelling curve based on the model of Hemond *et al.* (1984)

DISCUSSION

Dilation storage is an important mechanism for water storage in the two marshes we have studied. The data support all three of the hypotheses that guided this study. Measurable changes in sediment volume occur and these changes were associated with changes in the water content of the sediment. Dilation storage accounts for a significant fraction of the change in sediment water content; 20 per cent in Belle Isle marsh and up to 36 per cent and 86 per cent at the two sites in Sippewissett Marsh. The remainder is accounted for by saturation storage and perhaps by other mechanisms not yet described. Incorporation of water into dilation storage during infiltration appears to follow the infiltration model of Hemond *et al.* (1984) based on our interpretation of one event. More data are needed before a definitive statement can be made about the general applicability of this model. The infiltration model does not account for the infiltration of water involved in displacing air from the pores.

Dilation storage is likely to be significant wherever soils are near saturation. Gillham (1984) reviews phenomena related to the occurrence of tension saturaton near the soil surface. Under these conditions saturation storage is greatly reduced and dilation storage becomes relatively more important. Changes in surface elevation associated with dilation have been observed in sphagnum bogs (Almendinger *et al.*, 1986; Gates, 1940) but not in sedge peats (Hammer and Kadlec, 1989). In more typical soils, saturation storage is three to four orders of magnitude greater than dilation storage (Brutsaert and El-Kadi, 1984).

We have found that dilation storage in salt marsh sediment is comparable to saturation storage. Therefore, it is incorrect to assume that water loss from the sediment requires an equal volume of air to enter the pores. Furthermore, water fluxes across the sediment surface, e.g. evapotranspiration and infiltration, cause vertical water movement below the watertable because dilation occurs throughout the depth of the sediment.

Howes *et al.* (1981, 1986) have proposed that increased aeration of the sediment increases plant growth, which increases transpiration losses from the sediment causing greater rates of air entry into the sediment and increased aeration. If the loss of water results in significant sediment shrinkage, then loss of pore volume occurs instead of air entry into the sediment, and the feedback loop linking vegetative growth and sediment aeration will be less effective. Water removed from a sediment deposit, e.g. by daily evapotranspiration, is increasingly acounted for by the collapse of pore space as the depth of the sediment increases, increasing the proportion of the total storage coefficient accounted for by dilation storage (36 per cent at the 1 m site in Sippewissett marsh versus 86 per cent at the 4·5 m site). This offers a possible explanation for lower plant productivity and the formation of unvegetated panne areas observed in the older, interior regions of salt marshes (Redfield, 1972) where presumably the depth of sediment is greatest.

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