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STUDIES OF A SUBARCTIC COASTAL MARSH, II. SALINITY

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ABSTRACT

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The salinity of James Bay coastal marshes is more strongly affected by fossil salt from the underlying sediments than from tidal sources. A numerical model simulates the transport of this salt towards the marsh surface by molecular diffusion. However, the more permeable sediments near the surface transmit salt by advection in the groundwater flow, notably at the beach ridges, where the vertical hydraulic gradients are the strongest. The diffusion and advection transport processes supply chloride (Cl^-) to the surface at rates of 0.4 and $6.0 \text{ gm}^{-2} \text{ d}^{-1}$ respectively. Although the diffusion transport rate is low, this process contributed nearly half as much chloride to the marsh surface as advection did, because diffusion operates over a much larger area. These estimated chloride fluxes were within 15% of the measured average Cl^- export in the stream channel which drains the marsh. A model simulating the daily rate of Cl^- export indicates that its day to day variability is due to the near-surface hydrological processes, but over the long term is limited by the vertical advection and molecular diffusion processes that bring the salt to the surface.

INTRODUCTION

Coastal salt marshes have a broad global distribution (Chapman, 1974), forming a rich biological ecosystem because of their interaction with the flow of nutrients between the terrestrial and marine environments. Water is the primary conveyor of nutrient and salt in the marsh and a thorough knowledge of marsh hydrology is vital to understanding the transport mechanisms involved. These transfer processes control the surface salinity, which adversely affects vegetation productivity (Percy and Ustin, 1984).

The salt transport mechanisms include molecular diffusion across the concentration gradient and advection by the flowing water. Advection occurs in marshes underlain by permeable sediment but is controlled by: (1) tidally induced changes in water pressure (Hemond and Fifield, 1982); (2) changes in surface piezometric pressure due to precipitation or evaporation (Hemond et al., 1984); and (3) horizontal groundwater flow along the regional hydraulic gradient (Lindberg and Harriss, 1973). Advective exchange is limited in marsh

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sediments of low permeability so that salt transport is dictated by molecular diffusion (Casey and Lasaga, 1987).

Salt is typically introduced to coastal marshes through tidal flooding at the surface (Valiela et al., 1978), or may enter through the creek banks when the hydraulic gradients are reversed during high tides (Gardner, 1973; Jordan and Correll, 1985). Salt is concentrated by evaporation (Casey and Lasaga, 1987) but is diluted by snowmelt and rainfall. These processes normally lead to decreased salinity with depth (Tyler, 1971; Lindberg and Harriss, 1973). The seasonal precipitation and evaporation regimes will therefore cause the salinity of the marsh surface to vary considerably.

In James Bay marshes the salinity decreases with distance inland but increases with depth (Price and Woo, 1988), which contradicts the previous arguments and points to the need for alternative explanations. Price and Woo (1988) demonstrated that the salt is relict in origin, and is now being released from the sedimentary deposits of the former post-glacial Tyrrell Sea. The present paper will relate the surface and near surface salinity of the marsh to the sources of salt; and will examine how the surface hydrology affects the export of salt from the marsh system. Knowledge of the magnitude of these salt fluxes is important to understanding the marsh ecology.

STUDY AREA AND METHODS

The study area has been described fully in a previous paper (Price and Woo, this volume). The present study is focussed on a narrow coastal strip at the southern end of James Bay. The site extends from the mean high water (MHW) mark is about 3 km inland. Landward of the MHW mark is the backshore zone which is periodically saturated, but which is inundated only by very high tides. This zone stretches for about 600 m at which distance is located the first beach ridge. The ridge crest is 1.8 m above the MHW level, varying from a moist to dry condition and with little or no peat cover. Its elevated position causes the ponding of marsh water in the interrIDGE depression area. The second beach ridge, approximately 0.5 km further inland, allows water to continuously flow over its surface during the summer, draining the water from the interior depression where peat growth has in-filled the depression sufficiently to raise the water table. Some depressions are also drained by small creeks, an example of which is Marsh Creek (unofficial name) (see fig. 1 in Price and Woo, p. 276, this volume).

This study was carried out between 1984 and 1986. Water samples were collected at the surface and at depths of 0.3, 0.5, 0.75, 1.0, 1.5, 2.0 and 2.5 m from 20 mm i.d. PVC slotted, screened pipes 24 h after the pipes were bailed to remove stagnant water. The sampling tube nests were situated at 200 m intervals in a square grid covering 2.5 km² of the marsh and at 50 m intervals along the transect (fig. 1, Price and Woo, this volume). Total salinity and chloride were measured within 24 h by an optical salinometer and a combination chloride electrode respectively; and electrical conductivity was measured with a conductivity meter. The electrical conductivity of Marsh Creek was measured

directly with a conductivity probe attached to an electronic data logger. All measurements were then adjusted to conductivity at 25°C. Local water samples with electrical conductivity (EC) less than 5000 μS were used to determine the chloride (Cl^-) concentration by regression, so that stream water chlorinity could be estimated. The relationship is: $\text{Cl}^- (\text{mg}^{-1}\text{l}) = \text{EC}/4.44 - 9.3$, with an accountable variance of 0.98.

Soil cores were retrieved in the frozen state with a CRREL (U.S. Army Cold Regions Research and Engineering Laboratory) corer, or in the unfrozen state using a 22mm i.d. agricultural soil sampler. Pore water was extracted by centrifuge and then the chloride concentration was measured with an electrode. In this paper, the terms "salt" and "salinity" are used loosely to describe the total salinity, unless otherwise specified. Chloride values are used in all calculations and modelling applications.

RESULTS

Source of salt

James Bay is a shallow, semi-enclosed sea into which many large rivers discharge, resulting in water salinity considerably less than that of seawater. Off-shore and tidal zone salinity ranged from 0–4 kg m^{-3} , which compare favourably to similar areas in James Bay (McCrea et al., 1984; McCrae and Wickware, 1986). Thus tidal water of relatively low salinity is available to the coastal marshes, depending on the local frequency of inundation. Tide data for this location indicate that less than 4% of the tides surpass an elevation of 1.5 m above mean sea level and seldom penetrate 200 m beyond the MHW line. Only one storm tide, in October 1985, inundated the intensive study section of the marsh, reaching 1200 m beyond MHW. With infrequent inundation by this low-salinity tidewater, there is little opportunity for salt to be introduced tidally. The lower salinity in the upper layer of the marsh attests to this.

The higher salinity of the relatively impermeable Tyrrell Sea clay underlying the marsh suggests that the salt is relict in origin. This was confirmed by Price and Woo (1988) who showed that $\delta^{18}\text{O}$ and δD of pore water was isotopically heavier than the meteoric and tide water (Table 1). Thus the

TABLE 1

Average oxygen-18, deuterium, and chloride concentrations in soil water, rain and tides; pore water, especially that from deeper than 3 m is less depleted in the heavy isotopes than tide or rain water, indicating a different origin

| | Pore water > 3 m | Pore water < 3 m | Tide water | Rain water |
|-------------------------------------|---------------------|---------------------|------------|------------|
| $\delta^{18}\text{O}^*$ (per mille) | - 7.2 | - 10.0 | - 11.6 | - 11.8 |
| δD^* (per mille) | - 68.4 | - 78.0 | - 90.0 | - 99.1 |
| $\text{Cl}^- (\text{kg m}^{-3})$ | 12.2 | 4.6 | 0.6 | > 0.01 |

* Expressed as per mille difference from SMOW.

salt in the deeper pore water was not introduced by rain or tide water, which is relatively depleted in the heavier isotopes, but originated within the sediments laid down by the postglacial Tyrrell Sea.

The Tyrrell Sea sediments have a low permeability ($10^{-10} \text{ m s}^{-1}$), so that molecular diffusion is probably responsible for delivering the salt from there to the surface. Molecular diffusion is a slow process and cannot be observed during the course of a short-term study. Nevertheless, due to the isostatic rebound of this region, long-term changes are translated into variations on a spatial scale. A site located 3 km inland emerged from the sea 200–270 years before a site 1 km inland from the shore. These figures are obtained by counting beach ridge sequences, or by finding the product of the regional gradient, the distance inland and the inverse of the rate of isostatic uplift. The vertical salt distribution at the inland site should have been similar to that of the coastal site when the former site occupied an equivalent position relative to the coast. A direct comparison of their vertical salt concentration profiles can therefore be made to assess the effect of molecular diffusion over this time span (Fig. 1). Today, both sites have near-zero surface salinity (because the surface is continuously refreshed by meteoric water) and they have similar salinity at a depth below about 3 m due to the common origin of salt. Between these depths, the inland site experienced a larger amount of salt loss because the upward transport of salt has continued there for an additional 200–270 years.

The difference in the chloride concentration profile at these sites, which approximates a change in concentration through time, can be modelled numerically as a diffusion process, using Fick's second law:

$$\frac{\partial C}{\partial t} = D^* \frac{\partial^2 C}{\partial z^2} \quad (1)$$

where C is concentration, t is time and D^* is the molecular diffusion coefficient. The initial condition was assumed to be equal to a best-fit line of the actual vertical chloride distribution observed at the coastal site and:

$$C(z, 0) = C_0(z), \quad z \geq 0 \quad (1a)$$

$$C(0, t) = 0, \quad t \geq 0 \quad (1b)$$

$$C(z_m, t) = C_m, \quad t \geq 0 \quad (1c)$$

where C_0 is the initial concentration and C_m is the constant concentration at the maximum depth (z_m). The diffusion coefficient (D^*) in the clay was estimated in other marine clays of similar origin to be about $3 \times 10^{-10} \text{ m}^2 \text{ s}^{-1}$ (Desaulnier, 1986). The diffusion coefficient in the peat layer was assumed to be $4\text{--}6 \times 10^{-10} \text{ m}^2 \text{ s}^{-1}$ (Loxham and Burghardt, 1983). A moving upper boundary was incorporated into the model because an additional 0.5 m of peat exists at the inland site. The upper boundary elevation (z_0) rises at the average rate of peat growth over the simulation period such that:

$$z_0(t) = z_0(0) + t \left(\frac{dp}{dt} \right) \quad (2)$$

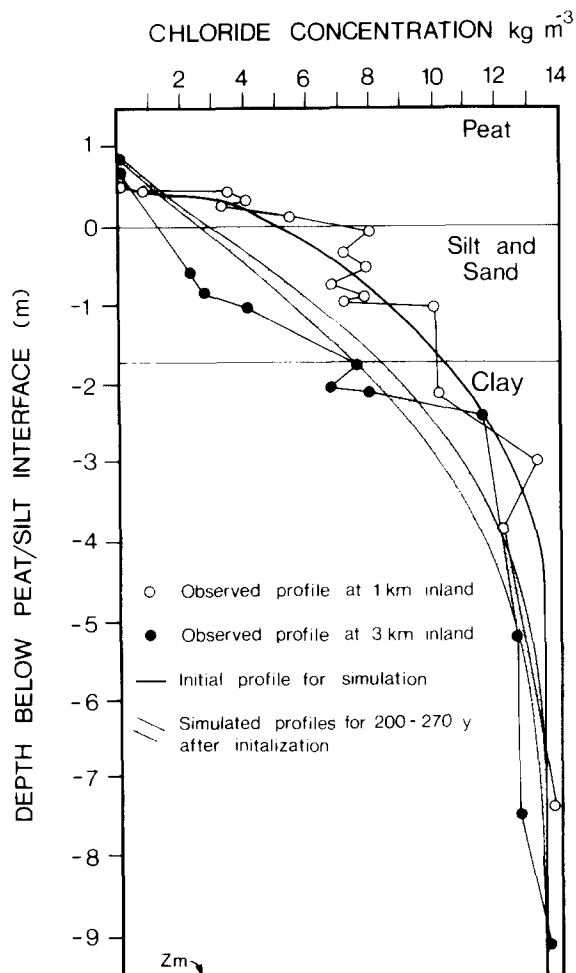


Fig. 1. Observed and simulated vertical profiles of chloride concentration located at 1 and 3 km inland from the present day shoreline.

where dp/dt is the average rate of peat accumulation, locally estimated to be 0.001 m yr^{-1} .

The results of this diffusion model compare well with the general vertical trend exhibited by the field data obtained at 3 km inland, indicating that salt migration via diffusion is a plausible mechanism (Fig. 1). Varying the D^* value for peat within the given range made no appreciable difference to the results. However, a better fit could be achieved by increasing the diffusion coefficient for the clay, or the time elapsed. The simulated values (Fig. 1) deviate from the observed results in the top 2 m zone where there are more recent silt and sand deposits. This suggests that the advection of freshwater into and through the more permeable upper layers removes salt more rapidly than is supplied by molecular diffusion.

Salt transport processes in the marsh

While molecular diffusion is responsible for delivering salt to the marsh from the underlying marine sediments, the upper marsh sediments are more permeable permitting both diffusion and advection to take place. Knowledge of these processes can be used to estimate the salt fluxes which are primarily responsible for the loss of salt from the marsh.

Diffusive salt transport

The mass flux of salt by diffusion can be calculated from Fick's first law as:

$$f_d = D^* \frac{dC}{dz} \quad (3)$$

where f_d is the mass flux of solute per unit area per unit time due to molecular diffusion and dC/dz is the concentration gradient. The chloride concentration gradient was measured at 26 locations in the marsh between 0.3 m depth and the peat surface and averaged $9.3 \text{ kg m}^{-3} \text{ m}^{-1}$. The salt passes from silt, through the peat, to the ponded water on the surface. The apparent diffusion coefficient of chloride was estimated to be $4\text{--}6 \times 10^{-10} \text{ m}^2 \text{ s}^{-1}$ based on values in the literature for peat and geologic material, as described earlier. The mean daily diffusive chloride flux to the surface was thus estimated to be $3.2\text{--}4.8 \times 10^{-4} \text{ kg m}^{-2} \text{ d}^{-1}$.

Advective salt transport

Groundwater advects salt along its flow paths. Flow was measured along the piezometer transect which traversed a groundwater "mound" caused by the beach ridge topography. The mound is a zone of downward recharge, while upward flow (discharge) occurs at the foot of the ridge where it meets the interridge depression (Price and Woo, this volume). Recharge introduces fresh surface water into the marsh sediments and discharge results in the loss of salt to the marsh surface.

The average vertical water flux (\bar{q}_v) was calculated at each piezometer nest. The average head at each piezometer (\bar{h}) was determined from eleven sets of measurements between 9 May and 9 August 1985. The \bar{h} value at piezometer i in a nest of n piezometers allows the mean hydraulic gradient $(d\bar{h}/dz)_i$ to be obtained. The discharge at the piezometer nest is then obtained by:

$$\bar{q}_v = \frac{1}{n} \sum_{i=1}^n \left(K \frac{d\bar{h}}{dz} \right)_i \quad (4)$$

where K_i is the hydraulic conductivity determined at piezometer i . The water table elevation along the transect varied by only 0.073 m during this period, with little measurable difference in the hydraulic gradient between piezometers, suggesting a steady groundwater flow regime throughout the summer. This is not unexpected considering that the water table configuration reflects the surface topography regardless of high or low water levels. The trend was for downward water movement (recharge) on the ridge, typically 0.1–

0.5 mm d⁻¹. Discharge into the depression was small except at the foot of the ridge, where it reached 0.4–0.9 mm d⁻¹. The error associated with these estimates is primarily due to the determination of hydraulic conductivity. The average hydraulic conductivity at 30 locations was $4.6 \times 10^{-7} \text{ m s}^{-1}$ and the standard deviation was $2.4 \times 10^{-7} \text{ m s}^{-1}$. If the standard deviation is assumed to encompass the error, these measurements are approximately within 50% of their true value.

The advective transport of chloride (f_a) can be obtained as:

$$f_a = \bar{c} \bar{q}_v \quad (5)$$

where \bar{c} is the average concentration of the advected chloride measured at the piezometers where the water flux was obtained.

Based on the average chlorinity of the pore water measured in the discharge zone (9.3 kg m^{-3}), the average daily flux to the surface by advective transport was estimated to be $5.8 \times 10^{-3} \text{ kg m}^{-2} \text{ d}^{-1}$. Another set of piezometers 200 m west of the transect had a daily discharge of 1.0 mm and an average chloride concentration of 7 kg m^{-3} , yielding an advective flux of $7.0 \times 10^{-3} \text{ kg m}^{-2} \text{ d}^{-1}$. Compared with diffusion, therefore, advective salt transport is one order of magnitude larger.

Spatial variation of salinity

Salt transport by diffusion and advection can explain the distinct spatial distribution of salinity in the marsh sediments. A section across a beach ridge–depression sequence (Fig. 2) shows that the salinity in the ridge sediments is proportional to the magnitude of local recharge. Recharge on the ridge suppresses the salinity by leaching away the salt that diffuses upward from the underlying clay. This water and the salt are eventually discharged to the surface at the foot of the ridge. Advection in groundwater is insignificant in the depression because of its flat surface and low hydraulic gradient, so that salt transport occurs primarily by diffusion. There, the salt diffused to the surface is not removed from the sediments as quickly and areas with a relatively high salt content develop. The alternating pattern of lower salinity on the beach ridges and higher salinity in the depressions is repeated at least as far as 6 km inland.

The transect represents a typical spatial pattern of salt distribution across the marsh. This is confirmed in Fig. 3, which shows that the horizontal changes in salinity are mostly perpendicular to the coast. Figure 3 also shows that salinity is reduced in riparian zones. This is due to the local increase in hydraulic gradients near the streams (Price and Woo, this volume) that enhances advective transport of salt to the channel. Groundwater effluence along Marsh Creek reaches 0.6–1.3 mm d⁻¹. Over the entire marsh, however, the low drainage channel density renders the horizontal export of salt by groundwater toward the streams much smaller than the vertical transfer of salt to the marsh surface.

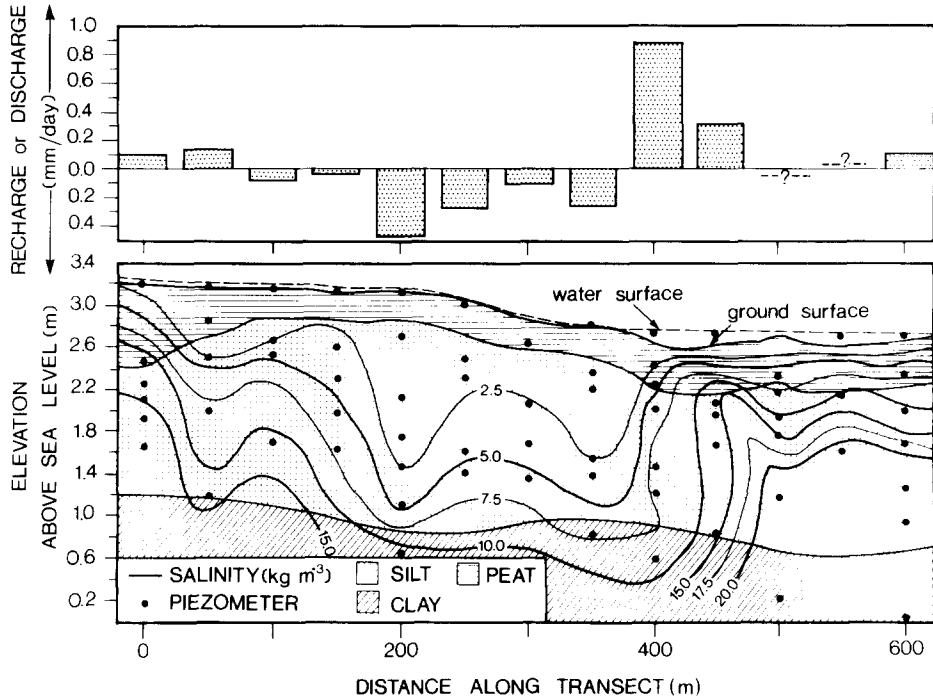


Fig. 2. Isohalines of total salinity along the transect, and the rates of vertical recharge and discharge.

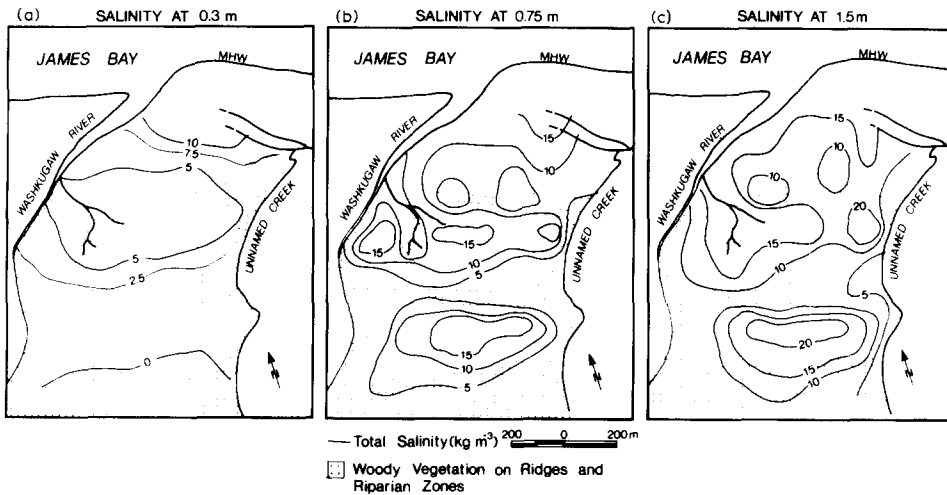


Fig. 3. Isohalines showing total salinity distribution on 11 August 1984, at depths of: (a) 0.3 m; (b) 0.75 m; and (c) 1.5 m below the surface.

Temporal variability of salinity

The short-term changes of chloride concentration in the pore water was measured at a ridge and a depression location along the transect. Water samples were drawn every other day between 26 July to 15 August 1985 from depths of 0, 0.05, 0.10, 0.20, 0.30, 0.40 and 0.50 m. The results show little variation in the salinity pattern through time (Fig. 4). This is because molecular diffusion is a slow process and there is a steady flow regime in the zones of advection. These data confirm that the estimated flux of salt to the surface by diffusion and groundwater advection are representative of the entire summer period; and that salinity changes become noticeable only over a much longer time span.

In contrast to the slow change of salt concentration in the pore water, the surface water exhibited high variability, as is indicated indirectly by its electrical conductivity (Fig. 5). Electrical conductivity was low in spring because of salt dilution by the abundant snowmelt water. Afterward, conductivity of the water in the marsh and the stream increased and became more variable as the summer progressed. Conductivity of the surface water peaked in May when precipitation was low and evaporation was high, but during wetter periods when precipitation dominated over evaporation, conductivity was lower. These tendencies were paralleled by the stream electrical conductivity, which peaked in low flow periods and decreased during storm flow,

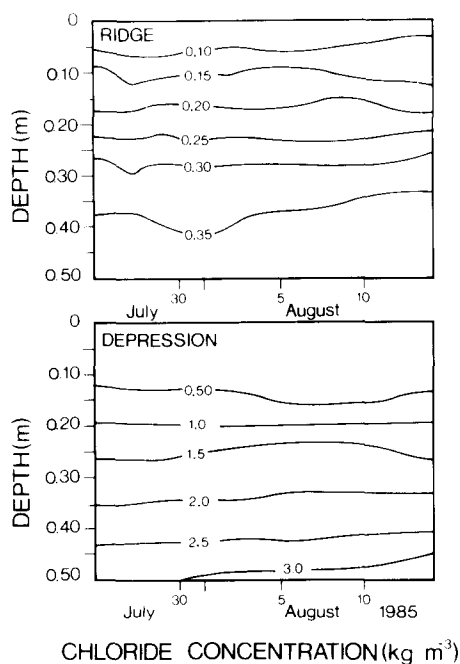


Fig. 4. Isochrones of chloride concentration at 0.05, 0.1, 0.2, 0.3, 0.4, and 0.5 m below the surface of the second ridge and the depression.

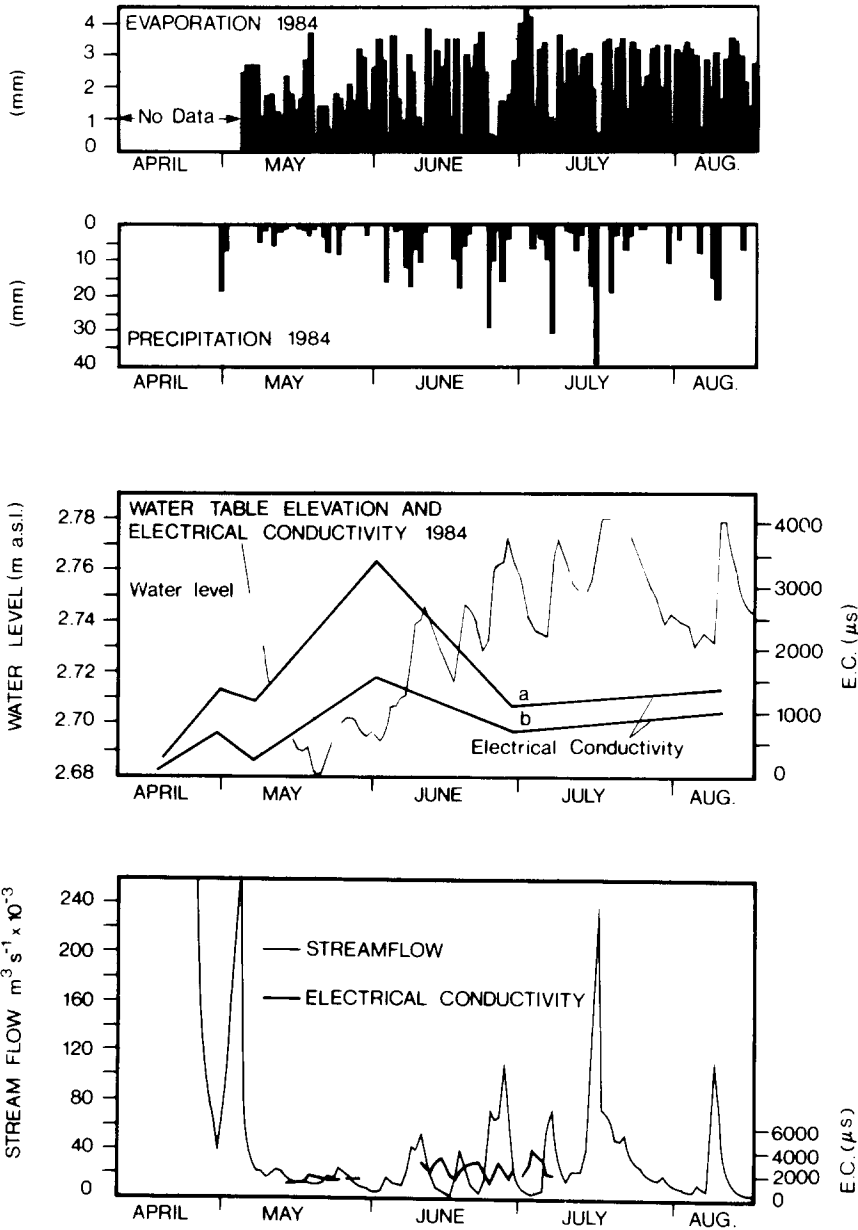


Fig. 5. From top to bottom; daily values for evaporation; precipitation; water level in the interridge depression; electrical conductivity at (a) the centre and (b) the edge of the depression respectively; and streamflow and electrical conductivity in Marsh Creek.

indicating that the marsh and stream salinity are commonly linked to the surface hydrological processes.

Evaporation, precipitation and surface inflow to the marsh control the volume of surface water storage (V) and thus its salt concentration (C)

through the relationship:

$$C = M/V \quad (6)$$

where M is the mass of salt present in the surface water. While C is inversely related to V , runoff generation from the marsh increases as the level of storage increases (Price and Woo, this volume). Thus, the water storage level controls the rate of salt export from the system through its effect on the salt concentration in the marsh water and on the water (hence solute) discharge itself.

Salt budget

Salt enters the marsh through diffusion, advection by groundwater and surface inflow and from precipitation. Streamflow is the primary means of salt export from the marsh. Salt export by groundwater is negligible because of the low rate of lateral groundwater discharge ($3.5 \times 10^{-5} \text{ m}^3 \text{ d}^{-1} \text{ m}^{-1}$). Local salt transport in groundwater to streams may be of limited importance. Horizontal salt diffusion is insignificant because of the small lateral salinity gradients. Estimates of the major fluxes can be used to determine the chloride budget in the marsh for the period (T) between 12 June and 8 July 1984, based on the inputs and outputs from the 0.91 km^2 drainage area used for the water balance study (Price and Woo, this volume):

$$\int_T (F_a + F_d + F_p + F_s) dt = \int_T QC_s dt + \Delta M \quad (7)$$

where F_a , F_d and F_p are the average daily chloride influxes by advection, diffusion and precipitation integrated over the drainage area and F_s is the average daily chloride flux in surface inflow into the area. The outputs are via streamflow (Q) at the average chloride concentration (C_s) and ΔM is the change in salt storage in the marsh surface water.

The mass flux of chloride to the surface can be estimated from the calculated flux per unit area from diffusion ($4 \times 10^{-4} \text{ kg m}^{-2} \text{ d}^{-1}$) and advection ($6 \times 10^{-3} \text{ kg m}^{-2} \text{ d}^{-1}$), by integrating the flux over the area of the marsh influenced by each process. Diffusion was assumed to occur everywhere in the marsh (0.91 km^2) at the average flux rate and was thus estimated to be 364 kg d^{-1} . The daily chloride flux to the marsh surface by advection in the area in the marsh where it occurs (0.12 km^2) was 690 kg d^{-1} . The zone of groundwater discharge was estimated on the basis of the distinctive vegetation association of *Scirpus Americanus* and *Typha* sp. The remaining fluxes of salt to the depression are comparatively small. For example, precipitation on 27 June 1984 had a very low chloride concentration of $9.3 \times 10^{-3} \text{ kg m}^{-3}$. Surface flow over the second ridge which drained into the study area also had a low chloride concentration of $9.6 \times 10^{-2} \text{ kg m}^{-3}$. The chloride fluxes to the marsh due to precipitation (F_p) and surface inflow (F_s), obtained as concentration times the daily water flux, averaged 14 and 16 kg d^{-1} for the study period. These values are small compared to diffusion and advection by groundwater. The sum of the fluxes on the left

side of eqn. (7) is 1084 kg d^{-1} . The average chloride export measured in Marsh Creek between 12 June and 8 July 1984 was 1283 kg d^{-1} . The change in salt storage (ΔM) was probably small because there was no persistent drying or wetting spell during this period, as suggested by the small net change in water level (0.01 m). The error associated with the salt budget is 199 kg d^{-1} , which is small relative to the major inputs.

A Salt export model

Streamflow in Marsh Creek is mainly fed by the direct surface runoff from the marsh. The daily chloride export [$E(t)$] in Marsh Creek is therefore:

$$E(t) = Q\hat{C}_s \quad (8)$$

where \hat{C}_s is the predicted chloride concentration, obtained from eqn. (6) to give:

$$\hat{C}_s = \frac{M_0}{V_0} + \frac{d}{dt} \left(\frac{M}{V} \right) \quad (9)$$

where M_0 and V_0 are the initial quantities of chloride and water present at time $t = 0$. The second term on the right-hand side can be approximated by eqn. (7) to yield:

$$\hat{C}_s = \frac{M_0}{V_0} + \frac{d}{dt} \left(\frac{M}{V} \right) = \frac{M_0}{V_0} + \frac{F_a(t) + F_v(t) + F_p(t) + F_s(t) - E(t - \Delta t)}{V(t) - V(t - \Delta t)} \quad (10)$$

where Δt indicates the time interval between calculations. The change in water storage in the marsh due to the various hydrological processes [$V(t) - V(t - \Delta t)$] can be lumped by using the marsh water level as a model input. The initial conditions in this study are those which were measured on June 12, 1984, with M_0 being $3.7 \times 10^4 \text{ kg}$ and V_0 being $6.2 \times 10^4 \text{ m}^3$. Δt was taken as one day.

For a first approximation the daily values of $F_a + F_d + F_p + F_s$ were held constant at 1080 kg d^{-1} because both diffusion and advection varied little and the salt fluxes from rainfall and surface inflow were small. Simulation was performed for the period 13 June–9 July, 1984 (Fig. 6) and the simulated and the measured export values agree closely. Since daily chloride input was held constant during the simulation, the results suggest that chloride export is controlled mainly by V alone (which reflects the combined effect of the surface hydrological processes).

A sensitivity analysis was performed by varying the salt input values. It was found that the daily salt export was fairly insensitive to changes of salt input compared to changes in storage, because the input is small relative to the initial mass (M_0). With a salt input of only 750 kg d^{-1} , the results are not appreciably different from that of the previous simulation, suggesting that inaccuracies in the estimates of the salt input do not seriously affect the results of this short-term simulation.

The most notable deviation between simulated and predicted results

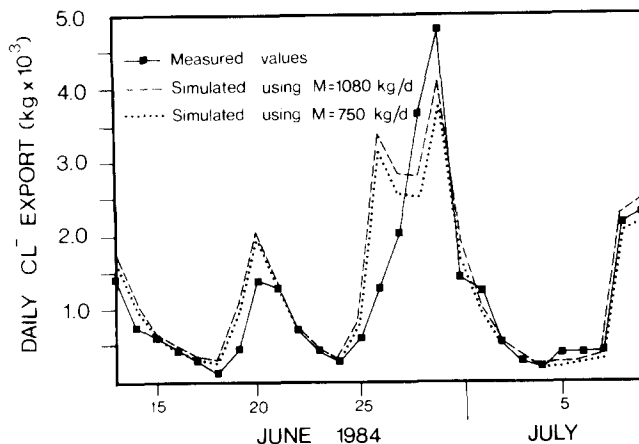


Fig. 6. Measured and simulated daily salt export in Marsh Creek.

occurred on 28 June when 14.2 mm of rain fell. There was a substantial increase in discharge and a corresponding reduction in measured salinity, whereas the model predicted a sharp increase in salt export due to the large increase in streamflow. The actual concentration of salt in runoff water was lower than calculated because of rapid runoff during such heavy rainstorms. This runoff water did not mix completely with water in the marsh, but drained directly into the channel as overland flow with low salinity (Price and Woo, 1988). The necessity of complete mixing is implicit in the model, which requires that salt concentration in stream runoff equals the average salt concentration in the marsh.

CONCLUSION

Salt marshes occur at distances well beyond the upper tidal limit of southern James Bay due to upward molecular diffusion of relict salt from marine sediments laid down during deglaciation. The surface topography of the overlying recent deposits comprises a recurring sequence of raised beaches and interridge depressions. The depressions retain surface water in summer which, when drained, carries with it the salt brought to the surface by diffusion and upward advection of groundwater. The beach ridges receive recharge of relatively fresh surface water to leach the salt downward and there is a direct relationship between the magnitude of recharge and diminished salt concentration. An upwelling of groundwater at the foot of the beach ridges brings the salt to the surface to be discharged into the marsh depressions. While this transport process proceeds much faster than diffusion, it is restricted to a relatively small area, so that its net effect is limited. The diffusion of salt and its advection by groundwater occur steadily throughout the summer, but the hydrological processes at the surface may change greatly within a short time span. Salt brought to the surface from beneath is flushed away at a rate dependent upon

the surface hydrological processes which control the water storage condition of the marsh. Both field observations and the model results demonstrate that salt export is less sensitive to the daily salt input than to the daily water storage change. This storage change variously concentrates salt (by evaporation) and dilutes it (by precipitation and by inflow from the interior wetlands). However, short-term salt export is ultimately limited by the rate at which salt is supplied to the surface by the slower processes of molecular diffusion and vertical advection.

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