Journal of Hydrology, 103 (1988) 275–292
Elsevier Science Publishers B.V., Amsterdam — Printed in The Netherlands

[3]

STUDIES OF A SUBARCTIC COASTAL MARSH, I. HYDROLOGY

JONATHAN S. PRICE* and MING-KO WOO

Department of Geography, McMaster University Hamilton, Ont. L8S 4K1 (Canada)

(Received November 25, 1986; revised and accepted February 5, 1988)

ABSTRACT

Price, J.S. and Woo, M.K., 1988. Studies of a subarctic coastal marsh, I. Hydrology. J. Hydrol., 103: 275–292.

The hydrological processes of a subarctic coastal marsh undergoing rapid isostatic uplift were studied to assess the links between water inputs and outputs. A water balance was performed for the summer season. Precipitation was the major input, followed by surface inflow, but tidal inputs were unimportant. Streamflow and evaporation were major outputs, while surface flow losses were limited to the snowmelt period. Subsurface flows were insignificant. The water storage characteristics of the marsh regulated the flow processes and evaporation. Since they were strongly influenced by the uneven peat development and pronounced seasonality, water storage, and therefore the marsh hydrological system were highly variable in both time and space.

INTRODUCTION

Coastal salt marshes occur in different climatic regions ranging from the tropics to the subarctic and provide a special habitat for terrestrial and aquatic organisms. The hydrology of these marshes controls the spatial distribution of nutrient and ecological attributes (Tyler, 1971; Casey et al., 1986; Price and Woo, this volume). Apart from several detailed studies of limited scope (e.g. Hemond and Fifield, 1982; Hemond et al., 1984), there have been few comprehensive studies of the hydrological regime of coastal marshes. This is especially true of subarctic marshes, which are very extensive on a global scale (Chapman, 1974). Specific to the subarctic environment are processes associated with coldness, snow accumulation and melt, and a pronounced decrease in the vegetative growth rate. In some cases there are the added complications of isostasy which affects the development time of the marsh soils and vegetation. This has been observed in the subarctic marshes of North America (Sims et al., 1982) and northwestern Europe including the Baltic and Fennoscandia (Chapman, 1974), where isostatic uplift has led to a progressively more advanced stage of peat development inland from the coast. There, the presence of such coastal landforms as raised beach ridges affect the magnitude

* Present address: Department of Geography, Memorial University of Newfoundland, St. John's, Nfld. A1B 3X9 (Canada).

0022-1694/88/$03.50

and direction of water transfer, and vegetative succession modifies the snow and frost distribution and evapotranspiration rates.

While the hydrological processes of the marshes can be identified easily, the linkages which affect the variability of these processes need clarification. The present paper will focus upon the seasonal and spatial aspects of the hydrological processes, and demonstrate their complex interaction in an emerging subarctic coastal marsh. The area chosen for this study is located in the Hudson Bay Lowland, an area that attains special significance as a migratory bird breeding and staging ground because of its high plant productivity and fertility (Prevett et al., 1979).

STUDY AREA

The study site (51°10′N, 79°47′W) located at the southern end of James Bay (Fig. 1), has a continental subarctic climate, with mean January and July temperatures at Moosonee, Ontario, of −20.0 and 15.5°C respectively. Mean annual precipitation is 727 mm, 30% of which falls as snow (Environment Canada, 1982).

Paleozoic sedimentary bedrock (Hutton and Black, 1975) typically occurs at a depth greater than 20–30 m (Ontario Hydro, unpub. drill log). Sparsely fossi-

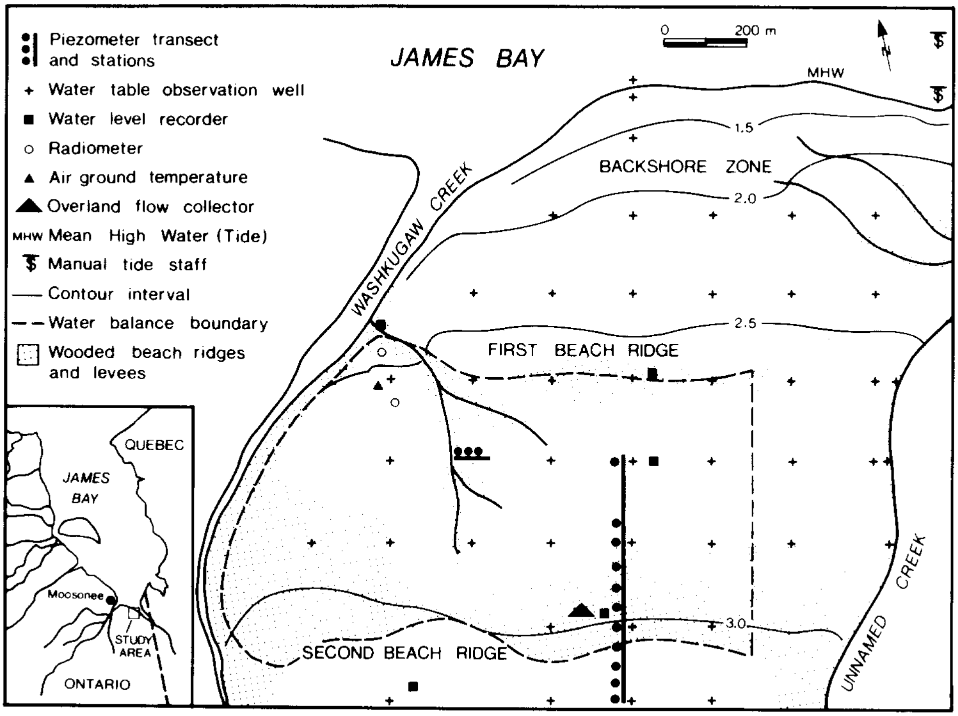


Fig. 1. Study area and instrumentation. Note that the present mean high water mark (MHW) is currently 500 m coastward of that shown on this 1957 aerial photograph. Marsh Creek drains the area within the water balance boundary.

liferous blue silty clay deposited during the post glacial Tyrrell Sea episode (Lee, 1960), is found at depths of 1.7–2.2 m and is overlain by more recently deposited marine sediments of silt and fine sand. The topography is characterized by very low gradients of 1.0 m km^{-1}. Subparallel raised beaches, formed of reworked sand and silt (Martini et al., 1980), are superimposed on the coastal plain (Fig. 1). At the average isostatic rebound rate of 0.7–1.25 m century^{-1} (Webber et al., 1970), the coast progrades one to two kilometres each century.

Raised beach ridges and interridge depressions form the two primary landscape units. These terrain types have different physical properties and are therefore hydrologically distinctive. The vegetation ranges from simple sedge (*Carex paleacea*) communities near the coast, to the more complex stands of woody vegetation inland (*Salix* sp., *Alnus* sp.), especially on the ridges (Ewing and Kershaw, 1986).

METHODS

Hydrological data were obtained between 1984 and 1986, mostly during the spring to late summer. Tidal data were also collected in July and August in both 1985 and 1986. The deployment of the requisite instrumentation is shown in Fig. 1.

A premelt snow survey was conducted in April 1984 using a method described by Adams and Roulet (1982). Daily snowmelt was monitored at several sites by measuring the daily lowering of the snow surface, and then multiplying it by the surface snow density (Heron and Woo, 1978). Summer rainfall was measured continuously with a tipping bucket rain gauge, supplemented by four manual rain gauges to determine the spatial variability of rain storms.

Evaporation was measured directly using a 0.3 m diameter clear plastic floating pan lysimeter, and by weighing soil-can lysimeters (138 × 98 mm diameter) at a ponded and a nonponded site. Evaporation was also calculated using the combination model of Priestley and Taylor (1972):

$$E = \alpha[\sigma/(\sigma + \gamma)]\,(Q^* - Q_g)/L_\rho \quad (1)$$

where E is the evaporation rate, L is the latent heat of vaporization, ρ is the density of water, σ is the slope of the saturated vapour pressure–temperature curve, γ is the psychrometric constant, Q^* is net radiation, Q_g is ground heat flux, and α is an empirical coefficient. To provide data for eqn. (1), net radiation, air temperature, and ground heat flux were measured. In the Hudson Bay Lowlands, Rouse et al. (1977) found $\alpha = 1.26$ for saturated surfaces and $\alpha = 1.0$ for drier surfaces. These values were acceptable for this study because the evaporation calculated compared favourably with the lysimeter data.

Streamflow was determined at a small creek draining the depression behind the first ridge (Fig. 1). Overland flow measurements were made in 1985 on the second beach ridge using metal eaves troughs to direct overland flow into a 20 l polyethylene bag attached to the apex of the gutters. The bag was allowed to fill for five minutes and the volume of water collected was measured in a graduated cylinder.

Water table elevation was measured periodically at 40 groundwater wells constructed from 20 mm i.d. PVC slotted pipes. Continuously recorded water table measurements were also made (Fig. 1). Piezometric heads were obtained using 6 mm i.d. piezometer tubes with 0.1 m slotted screened tips in 0.1 m diameter boreholes. A graded sand filter was placed around the piezometer tip and the tips were isolated with bentonite plugs. Each piezometer station shown in Fig. 1 consisted of a depth-integrated groundwater well and four or five piezometers with their bottom openings set at depths of ranging from 0.5 to 3.0 m. Ten selected piezometers were filled with kerosine in the autumn of 1985 to prevent freezing. The kerosine was removed before piezometric measurements were made in April 1986.

Hydraulic conductivity was assessed with bail tests in the piezometers using the method of Hvorslev (1951). Hydraulic conductivity of the organic layer was computed by the same method, but using a 50 mm i.d. tube slotted over the thickness of the entire organic layer. Flow net analysis (Freeze and Cherry 1979) was used to determine the magnitude and direction of groundwater fluxes.

RESULTS

A comparison of the air temperature, precipitation, and net radiation for 1984 with the 30 year mean at Moosonee, 100 km east of the study site, indicates the representativeness of the 1984 data (Fig. 3). The water inputs to the marsh were snowmelt and rain, and surface and subsurface inflow from the inland marshes, whereas the major outputs were evaporation and streamflow.

Tidal effects

Tidal inputs were limited to the backshore zone in 1984 and most of 1985 because the maximum tidal elevation did not exceed 1.9 m a.s.l. (Fig. 2). A storm surge in October 1985 reached an elevation of 2.9 m and was the only major tidal inundation during the study period. Tidal surges occur most frequently in autumn (Glooschenko and Clarke, 1982). Given the low frequency and magnitude of tides with respect to the emerged wetland in this location, they were not significant in terms of marsh hydrology beyond the mean high water (MHW) zone. In the backshore zone above MHW, where occasional flooding occurred, surface drainage was rapid, although infiltration of 60 mm h^{-1} (measured with a ring infiltrometer) indicated that total saturation of the soil would have occurred during tidal inundation. The low-permeability sediments preclude the intrusion of tidal water much beyond MHW.

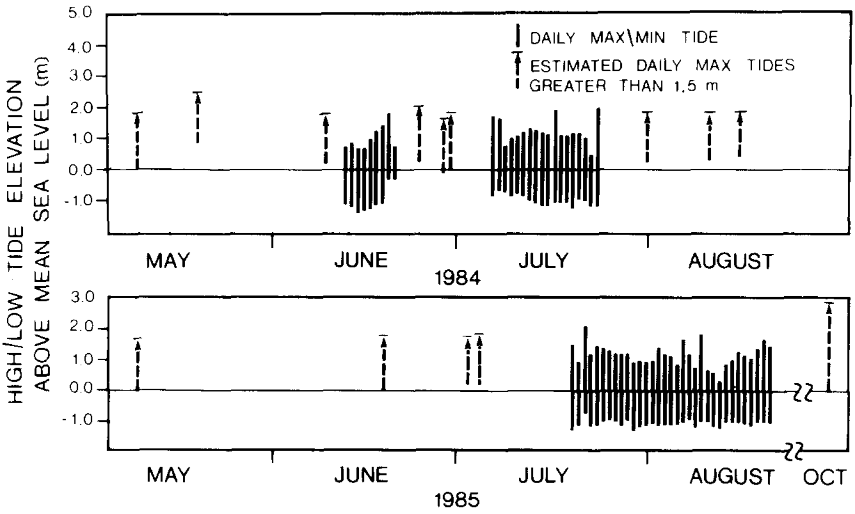


Fig. 2. Highest and lowest daily tides at the study site.

Snow accumulation and melt

A premelt snow survey (April 6, 1984) indicated a considerable redistribution of snow by wind during the winter. Raised beach ridges and stream banks with willow and alder vegetation trapped 0.6–2.0 m or more of snow [0.13–0.45 m water equivalent (w.e.)] but the snow on the open interridge depression was scoured to 0.14 m (0.05 m w.e.). The areally weighted average was 0.5 m (0.15 m w.e.).

Snowmelt was concentrated within the spring period, with few major melt events during the winter. The snowmelt rates at open and willow/alder sites were similar, ranging from 4 to 30 mm d^{-1} during the main melt period. By late April the melt rates for the residual snowbanks reached as high as 40 mm d^{-1}. Most of the snow in the depression melted by 14 April and meltwater inundated the low lying areas. Snow remained under the willow and the alder until 21 April, though late-lying deep snowbanks persisted until 6 May. Snow meltwater quickly satisfied the depression storage requirements of the wetland, and since the ground was frozen in the saturated state, little infiltration occurred.

Rainfall

After the melt period, rainfall became the major source of water input to the marsh. In April, 20 mm of rain fell and this, with 150 mm of snowmelt, was the largest monthly water input (Fig. 3). At this time the ground was still frozen and evaporation minimal, thus this water had little interaction with the environment and little was stored. Rainfall in May was low (90 mm) and the intensity was under 10 mm d^{-1}. June and July had above average rainfall, with

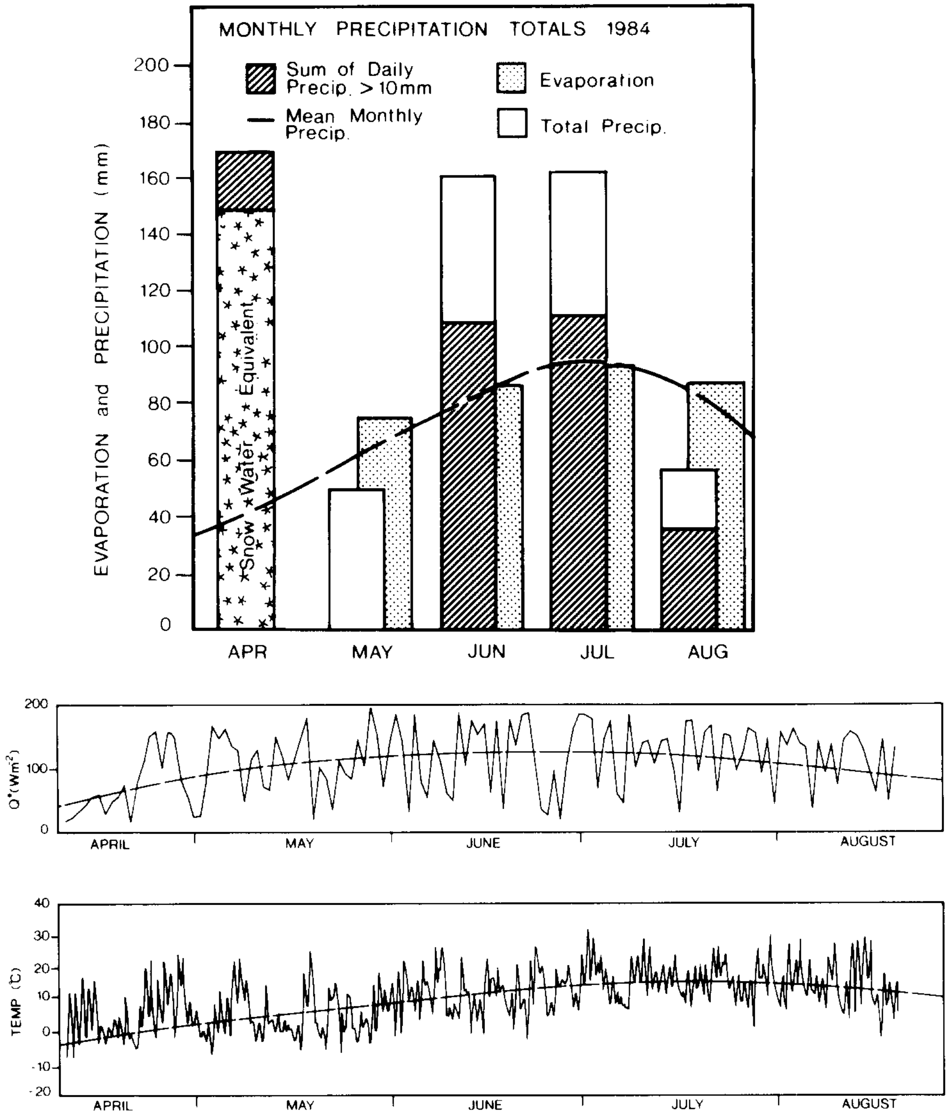


Fig. 3. Meteorological data for the 1984 field season at the study site, compared with the 30 year climatic averages for Moosonee (shown as dashed lines). Precipitation and evaporation (top), net radiation (middle), and air temperature (bottom).

75% of rain falling in events exceeding 10 mm d^{-1}. August rainfall was low, and 64% occurred in events greater than 10 mm d^{-1}. Records from the manual rain gauges did not reveal any noticeable spatial pattern.

Evaporation

Evaporation increased steadily from May to July and then declined in

August (Fig. 3). This overall trend reflects the net radiation regime. The average daily evaporation from saturated surfaces between 4 May and 22 August was 2.8 mm, with a range of 0.7–5.0 mm, for a total of 303 mm. From the first beach ridge to the coast the surface was normally unsaturated and the evaporation ranged from 0.3 to 4.4 mm d^{-1}, totalling 246 mm for the season (107 days). Local water storage characteristics affect the evaporation rates through its control on the surface moisture supply. Thus evaporation is self-limiting on the first beach ridge and backshore zone because it causes the water table to drop below the surface. At more inland locations surface saturation is maintained, allowing evaporation to proceed at the potential rate at all times.

Storage

Storage provides the link between the water inputs and losses of the marshes and different storage mechanisms become more prominent during different times of the year. Temporary water storage occurs as snow accumulation in winter and as ponded water and soil water in the near surface zone during the thawed season. Throughout the winter the surface layer of the marsh remains frozen and surface and subsurface storage changes are small or negligible.

Snowmelt water was released from storage when the ground was frozen. Surface depression storage was gradually replenished and considerable outflow followed. Depression storage was ephemeral on the ridges but persisted throughout the season in the interridge troughs. Between May and June, the water storage level in the troughs increased (Fig. 4) despite low precipitation and large evaporation rates. This rise was associated with the increased hydraulic resistance to surface flow as the vegetation developed in the marsh, thus drainage of the ponds and puddles was retarded. Evaporation had little noticeable effect on the water table (i.e. storage) in May (Fig. 4). In June, slight diurnal lowering of the water table on the first ridge was discernible; it became prominent in July and extreme in August when high evaporation was coupled with low rainfall. The exaggerated water table response on the first ridge reflected the low specific yield of its soil (S_y = 0.04) compared to the ponded water in the interridge depression ($S_y \cong 1.0$), because the water table change is $\Delta S/S_y$, where ΔS is the change in storage at a particular site.

Topography and peat development exert a strong control over the water storage characteristics of the marsh. The water storage capacity of the first ridge is small, as indicated by the large elevation range of the daily water table positions (Fig. 5). Although the water table there rose into the organic soil layer during rain events, it drained rapidly and evaporation reduced it further. The water table of the first ridge therefore resided primarily in the mineral soil. Further inland, peat accumulation has resulted in a water table which is sustained above the mineral soil. Because peat grows faster in the interridge areas, the topographic advantage of the ridges in shedding water is eventually eliminated. This enables surface and subsurface flow to be maintained across the second ridge, and those ridges located further inland.

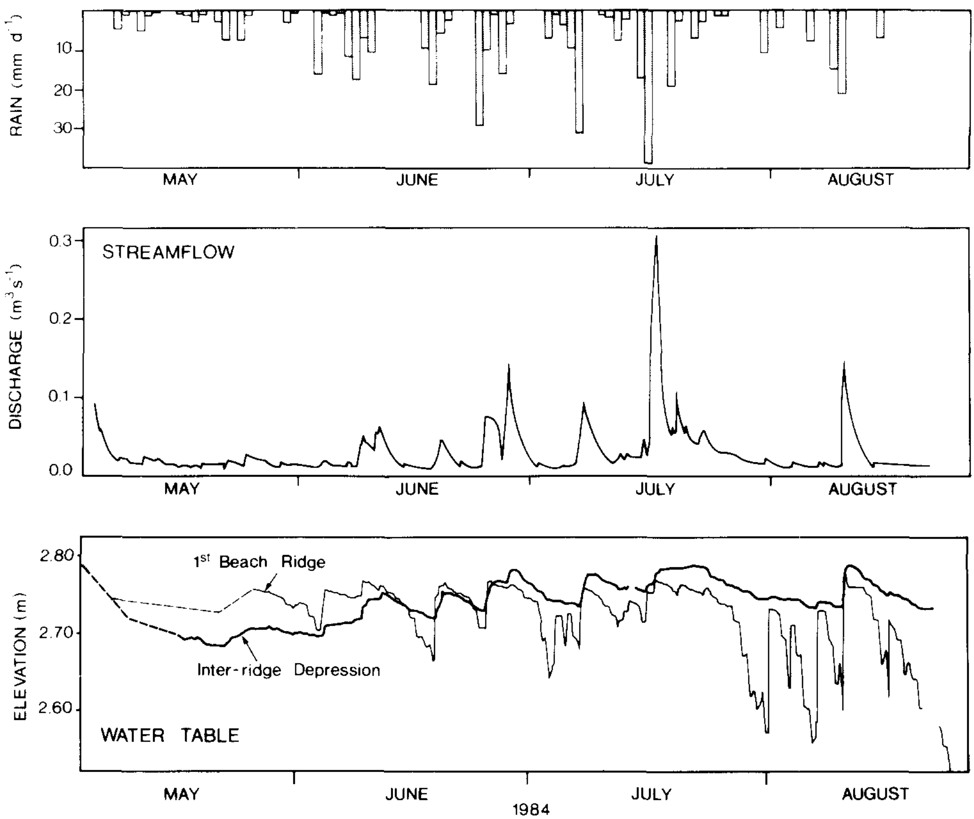


Fig. 4. Precipitation, streamflow, and water table elevation from May to August 1984. There is no record for the snowmelt period.

Surface and subsurface flow

Surface flow occurred over the entire marsh during the snowmelt period when the water table was at its maximum observed elevation. The high water table caused the first ridge to be completely inundated in some locations; hence water flow followed the regional gradient towards the coast (Fig. 6a). After the snowmelt period, rapid runoff and a diminished water supply resulted in a general recession of the water table. The hydraulic gradients and hence flow directions were then controlled by the local ridge and trough topography (Fig. 6b); this caused back-flow into the depression from the first beach ridge. However, the groundwater "ridge" disappeared during dry periods, permitting subsurface flow to resume toward the coast (Fig. 6c). At these times (e.g. Figs. 6b and c), surface flow could not cross the beach ridge but drained perpendicularly to the regional gradient toward the creek. The groundwater "ridge" was re-established following rain events of sufficient magnitude (Fig. 4), causing water flow back into the marsh. This process is analogous to the

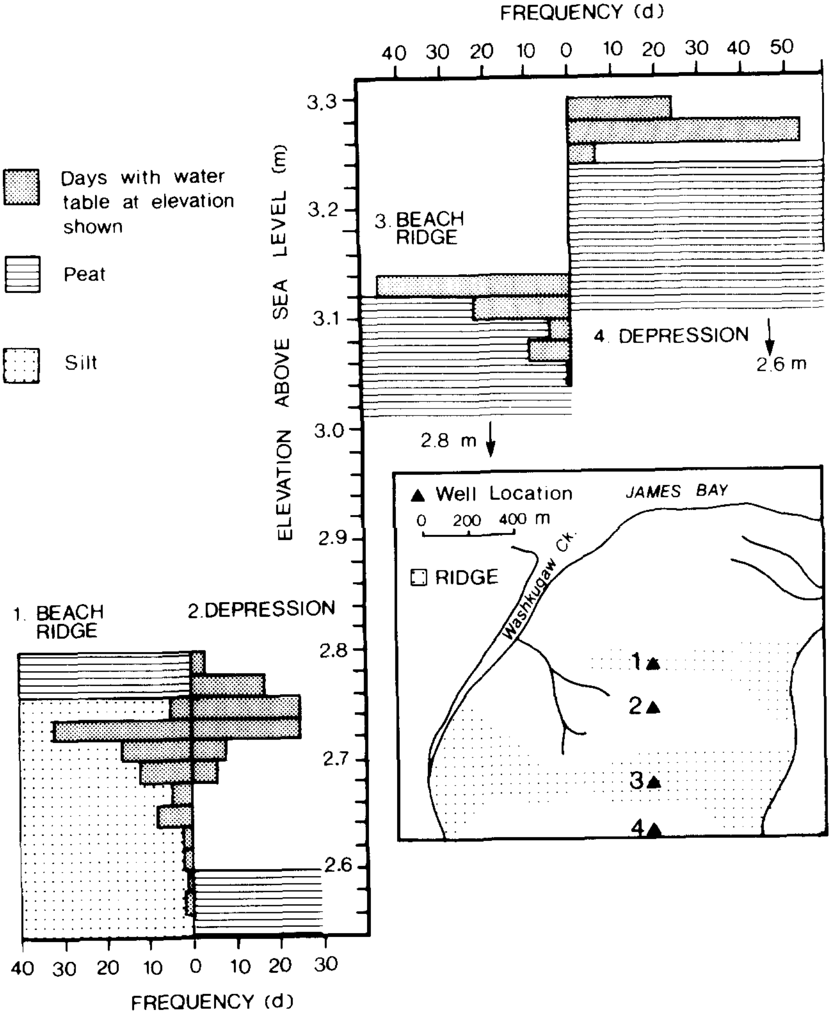


Fig. 5. Daily frequency of mean daily water table elevation at four locations along a transect perpendicular to the coast. The contrast between ridge and depression can be seen, as can the effect of increasing peat thickness.

variable source area concept (Dunne and Black, 1970). On the second and subsequent beach ridges, continuous surface and subsurface flow occurred from the adjacent landward marsh where thicker peat sustained a higher water table (Fig. 5). In 1985 surface flow was measured here and a rating curve was developed. This was applied to the 1984 water level data to estimate 1984 surface flow, which averaged $0.25\,m^3\,d^{-1}$ per m of ridge. At times of high water level it reached $0.75\,m^3\,d^{-1}\,m^{-1}$. The average subsurface flow through the 0.3 m organic layer was calculated using Darcy's law to be $0.001\,m^3\,d^{-1}\,m^{-1}$. The average hydraulic conductivity of the peat was $10^{-5}\,m\,s^{-1}$.

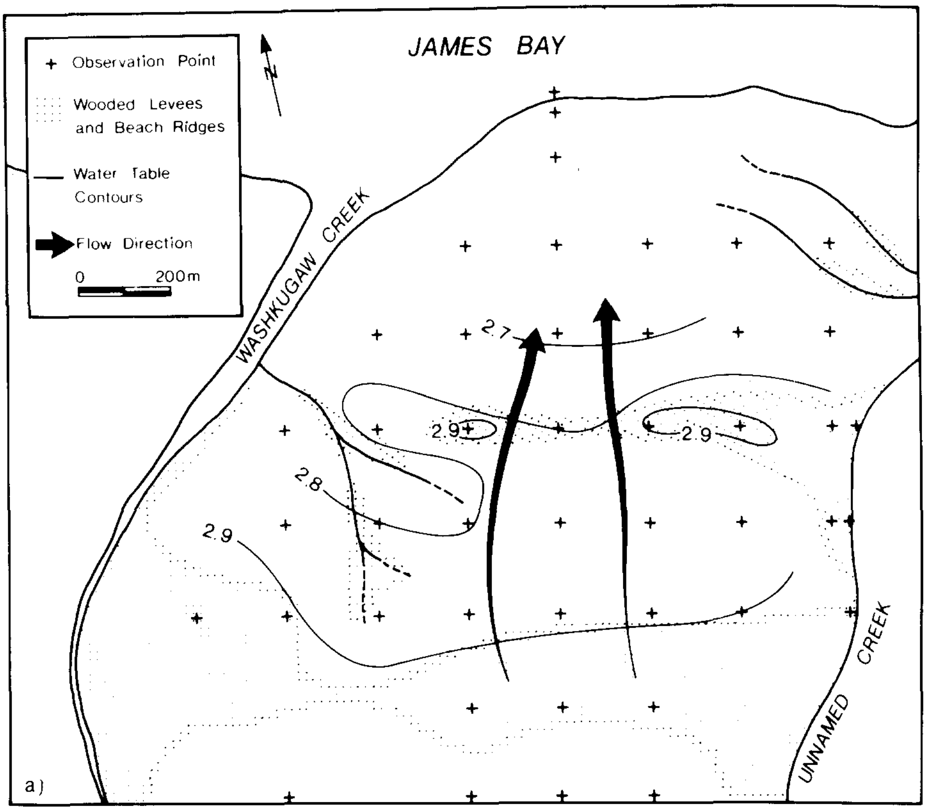
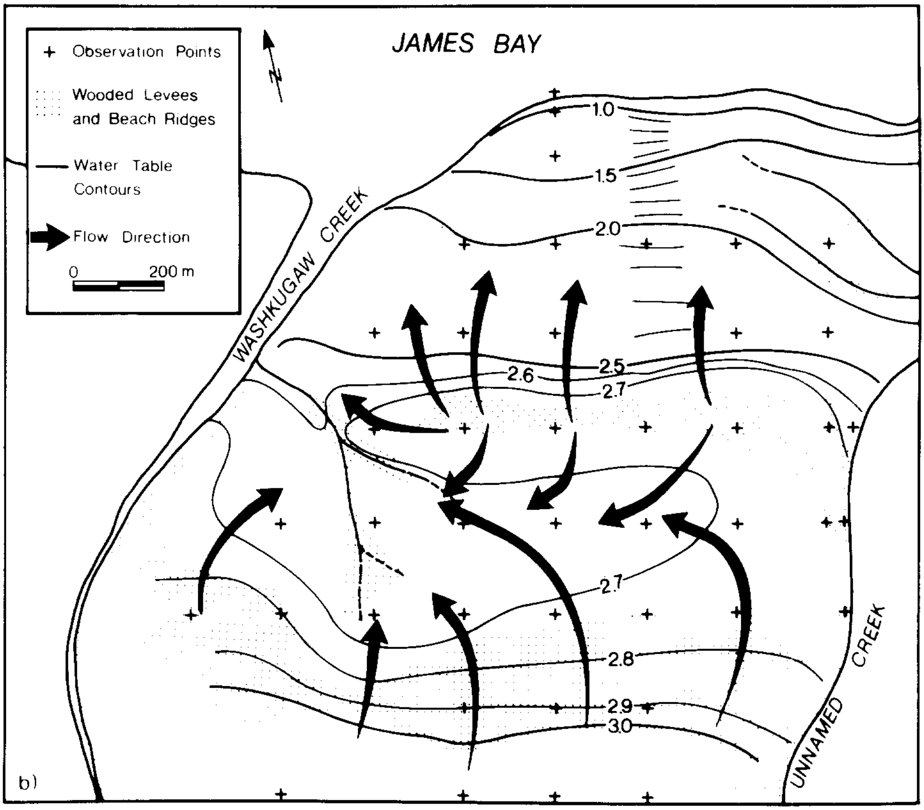


Fig. 6. Water table elevation and water flow directions for the marsh during (a) snowmelt; (b) early summer; and (c) summer dry period.

Deep groundwater

Deep groundwater refers to the water in the marine sediments underlying the peat. The average hydraulic conductivity of the silt and sand of the upper several metres is $10^{-7}\,m\,s^{-1}$, whereas the underlying clay of the Tyrrell Sea unit has a value of $10^{-10}\,m\,s^{-1}$. The piezometer transect across the second beach ridge indicated a general flow toward the coast (Fig. 7, top). However, there was a slight downward flow componenet on the seaward side of the ridge and an upward component where it joins the trough. Wet or dry periods in summer did not significantly alter the flow direction or the piezometric gradients. The water table mound of the first beach ridge suggests that downward flow also occurs there except in the dry periods. Deep groundwater flow toward the coast in summer was estimated to be $3.5 \times 10^{-5}\,m^3\,d^{-1}\,m^{-1}$.

A different hydraulic head distribution was encountered in winter (Fig. 7, bottom). The windswept interridge depression was relatively free of snow and was frozen more deeply (1.7 m) than beneath the ridges (0.35 m). A desaturated



zone developed beneath the frozen layer of the depression because soil freezing induced an upward migration of moisture from the lower layers to the freezing front (Smith, 1984; Kingsbury and Moore, 1987). In Fig. 7b it was assumed that the piezometric head of the unsaturated zone beneath the frozen layer was equal to the elevation head and that the pressure head was zero. Under the winter condition, coastward flow of groundwater was limited to a 0.4 m unfrozen unsaturated layer. The rate of flow under these conditions is at least several orders of magnitude smaller than during summer. From the observations made in summer and winter, it is further concluded that the transport of deep groundwater throughout the marsh is insignificant compared with surface and near surface flows.

Streamflow

Streams fed primarily by surface and near-surface water were observed (12 November 1985) to dry up at the onset of freezing of the marsh. Stream channels received a larger snow accumulation than the surrounding marsh during the winter because windblown snow was trapped by riparian vegetation and in-

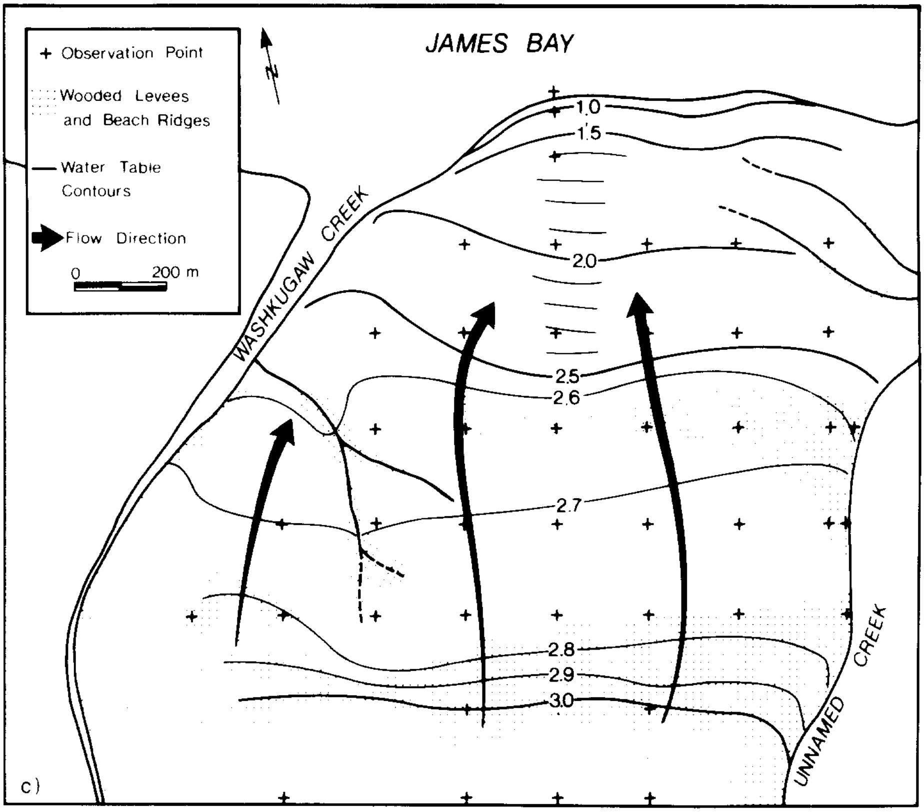


Fig. 6. (continued).

filled the channels. These snow drifts had a high density (over 300 kg m^{-3}) and became barriers to streamflow in spring. Such snow jams were found in the Washkugaw River and in the smaller creeks draining the marsh (Woo and Heron, 1987), causing meltwater to flood the entire marsh. When the jams were eroded the ponded water was released rapidly to produce the highest annual discharge.

After the spring freshet, low flows of 5–8 l s^{-1} were frequently encountered in the small stream during extended dry periods (Fig. 4). Low flows were maintained partially by surface drainage from the depression (Fig. 6) and partly from groundwater discharge towards the stream channel. The latter was facilitated by local increases in the piezometric gradient adjacent to the stream, which were an order of magnitude greater than elsewhere in the marsh. During the wet periods both overland flow and flow within the organic layer fed the stream directly (Fig. 8, left). In the dry periods the water level receded below the organic layer, halting the surface and near surface flows to the stream (Fig. 8, right). However, the piezometric gradient beneath the stream increased, yielding higher groundwater discharge rates, thus maintaining streamflow throughout summer. Groundwater flow to the streams in this area is relatively small, representing only 1% of the seasonal flow (DiCenzo, 1987).

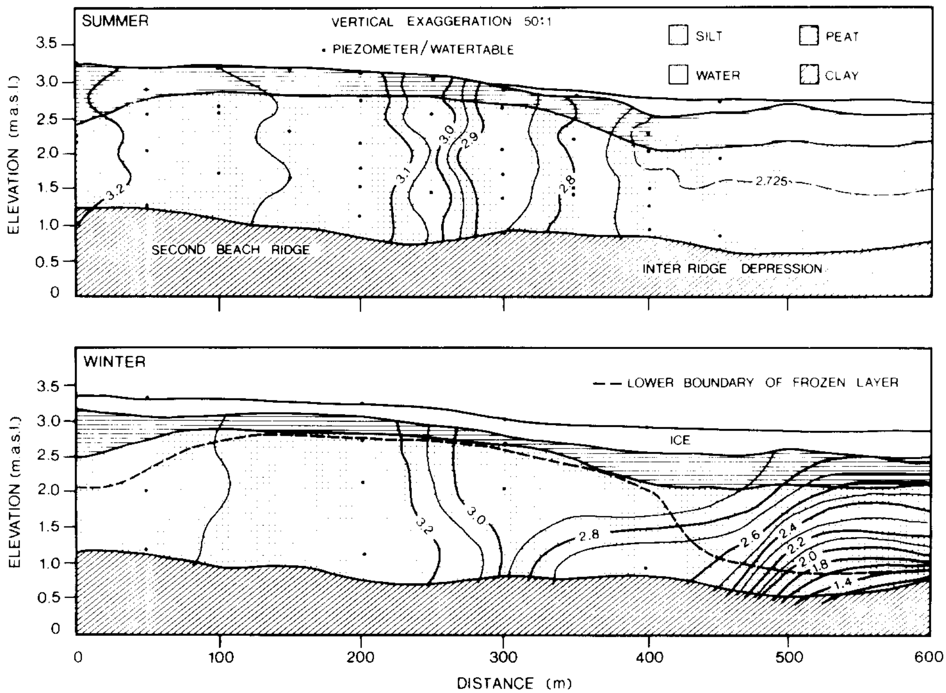


Fig. 7. Piezometric isopotential across the second beach ridge for the summer (top), and winter (bottom).

WATER BALANCE

Water balance relationships indicate the relative importance of different hydrological processes through time. The water balance equation for the marsh can be written as:

$$P + T + Q_{si} + Q_{ssi} - Q_{so} - Q_{sso} - Q - E = \Delta S + \xi \qquad (2)$$

where P is precipitation, T is tidal inflow, Q_{si} and Q_{ssi} are surface and subsurface inflow from interior marshes respectively, Q_{so} and Q_{sso} are surface and subsurface outflow respectively, Q is streamflow, E is evaporation, ΔS is change in storage, and ξ is the residual error term.

The water budget was calculated for an area of the marsh (0.91 km^2) bounded by the crests of the first and second beach ridges, by the Washkugaw River to the west and by a subdued drainage divide to the east (Fig. 1). Water balance computations were made for the period 4 May–22 August 1984 which excludes the snowmelt season. The width of the second beach ridge that forms the water balance boundary was assumed to be the input face (1340 m) along which surface and subsurface inputs occurred. Results presented in the previous section show that Q_{ssi}, Q_{so}, Q_{sso}, and T were negligible. For the calculation of evaporation, 15% of the area was unsaturated at the surface most of the time,

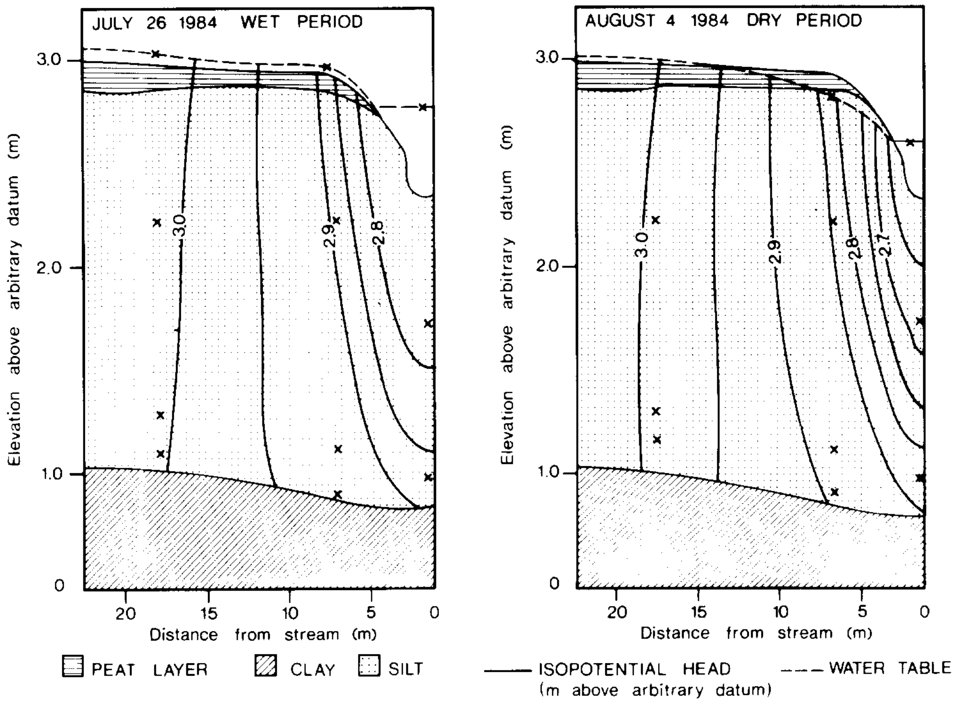


Fig. 8. Piezometric isopotential across a small creek draining the depression during a wet period (left) and a dry period (right).

$\alpha = 1.0$ for eqn. (1), and the remainder was always saturated ($\alpha = 1.3$). Storage change (ΔS) was calculated using the specific yield (S_y) and change in water level elevation (Δh), such that $\Delta S = \Delta h\ S_y$. In the ponded areas $S_y \cong 1.0$, and was 0.04 in the mineral soil of the first beach ridge.

Over the water balance period the largest single component was precipitation (429 mm) (Fig. 9). Rainfall was low in May, with no event greater than 10 mm d^{-1}. This created a large storage deficit even though evaporation had not yet reached its peak. Heavy precipitation during June and July replenished the storage, while August experienced a large water deficit because of high evaporation and low precipitation (Fig. 3). Surface inflow was only 15% of precipitation, but this was equivalent to 22% of evaporation, and was therefore a significant water source which partially offset storage losses.

Streamflow was the largest output (71% of precipitation) followed closely by evaporation (68% of precipitation). The response of streamflow to rainfall was governed by storage. For example, when storage was low, a 15.9 mm rainfall on 3 June 1984 yielded in a maximum streamflow of 1.8 mm d^{-1}, whereas 15.9 mm of rain on 27 June produced a peak flow of 10.7 mm d^{-1} because of wet antecedent conditions. Although the flat marshy area contains a considerable volume of water, it has little dynamic storage capacity; thus it cannot retain much water to sustain prolonged recession flow.

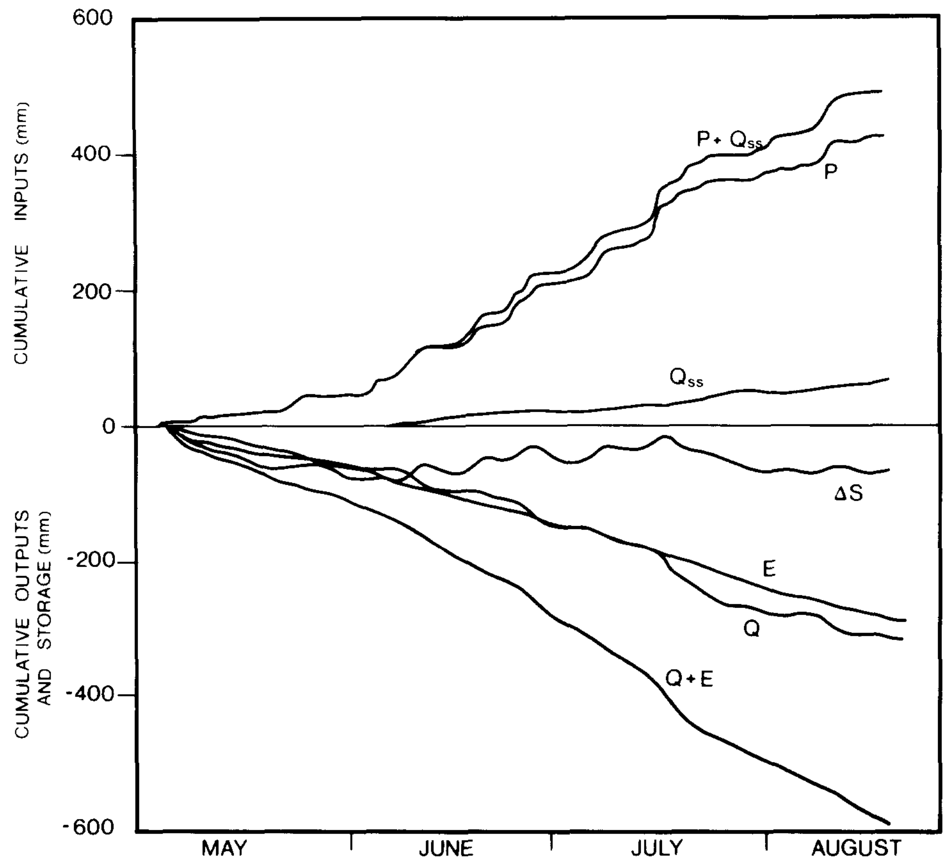


Fig. 9. Components of the marsh water balance for the period 4 May–22 August 1984. Numbers in brackets represent seasonal totals (mm).

Evaporation loss during the water balance period was comparable to streamflow in magnitude but was less variable through time. Even at low flow periods, evaporation continued to withdraw water from wetland storage and this deficit had to be eliminated before rainfall could raise the flow again. A similar phenomenon was observed in an arctic wetland by Roulet and Woo (1986a).

Net storage change over the water balance period was small, representing only 12% of precipitation. The estimated storage change (−50 mm) differed from storage change calculated as a residual (−106 mm). The difference (56 mm) is the error term (ξ) in eqn. (2) and is 13% of precipitation. Some of this error is attributable to the method of determining surface and subsurface inflow which may not have been representative of the entire ridge. There were preferred flow zones crossing the ridge which were not been accounted for, so that the value of Q_{si} is probably underestimated. However, as this component is small in magnitude, the error introduced is small relative to the other components. Measurements of precipitation, evaporation, and streamflow are approximately $\pm$ 15% under ideal conditions (Gray, 1970; Stewart and Rouse, 1976; Ackers et al., 1978). Thus the error in the water balance lies within the acceptable limits based on the measurement techniques used.

The water budget indicates that streamflow and evaporation are of equal importance as water sinks. Similar results were reported by Brown et al. (1968) in an Alaskan wetland. Runoff, however, is more important in spring and during flood events, while evaporation plays a more prominent role later in the summer. Surface inflow from inland marshes partially offsets evaporation losses during dry periods and thereby reduces the storage deficit that must otherwise be eliminated by rainfall. Streamflow can therefore respond more quickly. Surface inflow is less important compared with some wetlands in arctic Canada where surface inflow cascades from lakes, snow meltwater and stream overflow to maintain a high water table throughout summer (Roulet and Woo, 1986b).

DISCUSSION AND CONCLUSIONS

Subarctic coastal marshes have several characteristics distinguishable from other types of wetlands. Water may be introduced to the marsh tidally, although tidal inputs to the study area were largely restricted to autumn storm surges because isostatic rebound has raised the marshes well above the normal tidal range. These surges contribute little to the water storage because the peat is already saturated during this period of low evaporation and frequent rain events. Additional tidal water will merely run off. Coastal emergence due to isostatic rebound imparts other special properties to these wetlands. Raised beach ridges control the direction and velocity of the surface and groundwater flow, while the storage level in the depressions determines the flow magnitude. As peat thickness increases inland, the water storage capacity is enhanced, affecting both evaporation (saturated or unsaturated conditions) and runoff processes (surface, subsurface, or deep groundwater flows).

The temporal variability of the marsh hydrology is regulated by seasonal changes in water inputs and storage. In the spring, meltwater in the marsh and inflows from interior wetlands release water which cannot be absorbed by the low storage capacity of the frozen ground. Total inundation is accompanied by rapid surface flow in the regional direction toward the coast. The presence of snow jams along the coastal section of the stream leads to overflowing of the marsh creeks and rivers, resulting in an uninhibited exchange of the stream and marsh waters. Following snowmelt, the high water table in the ridge acts as a drainage barrier and surface and subsurface flows converge in the depression and then discharge as streamflow. This barrier is removed during periods of high evaporation and low rainfall, so that regional groundwater flow can resume. At this time storage declines and streamflow recedes to a baseflow level but evaporation remains important. In winter there is no surface or near-surface flow, and groundwater flow in the regional direction is minimal.

The integrated effects of spatial and temporal changes on marsh hydrology are reflected in the pattern of streamflow. This is the major output from the hydrological system controlled by precipitation, inflow, storage and evaporation loss. Streams of the subarctic coastal marsch do not exhibit the muskeg regime proposed by Church (1974) in which high flows are attenuated by the wetlands. In spite of the large total water storage in these wetlands, the small dynamic storage capacity means that snowmelt and rainfall are transmitted quickly through the system unless preceded by a long dry period when evaporation has created a storage deficit.

ACKNOWLEDGEMENTS

The financial assistance of the Natural Sciences and Engineering Research Council collaborative grant, and of the Department of Indian and Northern Affairs' Northern Training Grant is gratefully acknowledged. Invaluable help in the field was received from Drs. R. Heron and K. Ewing; and P. Steer, P. diCenzo, P. Lafleur, and S. Hardill. Ground heat flux data was generously provided by Dr. W.R. Rouse, and advice on the piezometer network was kindly given by Dr. J.A. Cherry.

REFERENCES

Ackers, P., White, W.R., Perkins, J.A. and Harrison, A.J.M., 1978. Weirs and flumes for flow measurements. Wiley, New York, N.Y., 252 pp.

Adams, W.P. and Roulet, N.T., 1982. Areal differentiation of land and lake snow cover in a small subarctic drainage basin. Nord. Hydrol., 13: 139–156.

Brown, S., Dingman, S.L. and Lewellen, R.I., 1986. Hydrology of a drainage basin on the Alaskan coastal plain, U.S. Army, Corps Eng., CRREL, Res. Rep. 240, 18 pp.

Casey, W.H., Guber, A., Bursey, C. and Olsen, C.R., 1986. Chemical controls on ecology in a coastal wetland. Eos, 67, 1305: 1310–1311.

Chapman, V.J., 1974. Salt Marshes and Salt Deserts of the World. Cramer, Lehre, 392 pp.

Church, M., 1974. Hydrology and permafrost, with reference to Northern North America. In: Permafrost Hydrology. Proc. Workshop Semin. Can. Natl. Comm. Int. Hydrol. Decade, pp. 7–20.

DiCenzo, P., 1987. Hydrology of subarctic wetland. M.Sc. Thesis, Geogr. Dep., McMaster University, Hamilton, Ont.

Dunne, T. and Black, R.D., 1970. An experimental investigation of runoff production in permeable soils. Water Resour. Res., 6: 478–490.

Environment Canada, 1982. Canadian climatic normals. Temperature and precipitation, Ontario, Ottawa. Environ. Can., Ottawa, Ont., 254 pp.

Ewing, K. and Kershaw, K.A., 1986. Vegetational patterns in James Bay coastal marshes. I. Environmental factors on the south coast. Can. J. Bot., 64: 217–226.

Freeze, R.A. and Cherry, J.A., 1979. Groundwater. Prentice-Hall, N.J., 604 pp.

Glooschenko, W.A. and Clarke, K., 1982. The salinity of a subarctic salt marsh. Naturaliste Can., 109: 483–490.

Gray, D.M., 1970. Handbook on the Principles of Hydrology. Can. Natl. Comm. Int. Hydrol. Decade, Ottawa, Ont.

Hemond, H.F. and Fifield, J.L., 1982. Subsurface flow in salt marsh peat: A model and field study. Limnol. Oceanogr., 27: 126–136.

Hemond, H.F., Nuttle, W.K., Burke, R.W. and Stolzenbach, K.D., 1984. Surface infiltration in salt marshes: Theory, measurement and biogeochemical implication. Water Resour. Res., 20: 591–600.

Heron, J.R. and Woo, M.K., 1978. Snowmelt computation for a High Arctic site. Proc. 35th Eastern Snow Conf., Hanover, N. H., pp. 162–172.

Hutton, C.L.A. and Black, W.A., 1975. Ontario arctic watershed. Environ. Canada, Ottawa, Ont., 107 pp.

Hvorslev, M.J., 1951. Time lag and soil permeability in groundwater observations. U.S. Army Corps Eng., Waterways Exp. Stn., Vicksburg, Miss., Bull. 36: 50 pp.

Kingsbury, C.M. and Moore, T.M., 1987. The freeze–thaw cycle of a subarctic fen, northern Quebec, Canada. Arct. Alp. Res., 19: 289–295.

Lee, H.A., 1960. Late glacial and postglacial Hudson Bay Sea episode. Science, 131: 1609–1611.

Martini, I.P., Cowell, D.W. and Wickware, G.M., 1980. In: S.B. McCann (Editor), The Coastline of Canada. Geol. Surv. Can., Pap. 80-10: 293–301.

Prevett, J.P., Marshall, I.F. and Thomas, V.G., 1979. Fall foods of lesser snow geese in the James Bay region. J. Wildl. Manage., 43: 736–742.

Price, J.S. and Woo, M.K., 1988. Studies of a subarctic coastal marsh, II. Salinity, J. Hydrol., 103: 293–307 (this volume).

Priestley, C.H.B. and Taylor, R.J., 1972. On the assessment of surface heat flux and evaporation using large scale parameters. Mon. Weather Rev., 100: 81–92.

Roulet, N.T. and Woo, M.K., 1986a. Hydrology of a wetland in the continuous permafrost region. J. Hydrol., 89: 73–91.

Roulet, N.T. and Woo, M.K., 1986b. Low arctic wetland hydrology. Can. Water Resour. J., 11: 69–75.

Rouse, W.R., Mills, P.F. and Stewart, R.B., 1977. Evaporation in high latitudes. Water Resour. Res., 13: 909–914.

Sims, R.A., Cowell, D.W. and Wickware, G.M., 1982. Classification of fens near southern James Bay, Ontario, using vegetation physiognomy. Can. J. Bot., 60: 2608–2623.

Smith, M.W., 1984. Observation of soil freezing and frost heave at Inuvik, Northwest Territories, Canada. Can. J. Earth Sci., 22: 283–290.

Stewart, R.B. and Rouse, W.R., 1976. Simple models for calculating evaporation from dry and wet surfaces. Arct. Alp. Res., 8: 263–274.

Tyler, H., 1971. Hydrology and salinity of Baltic sea-shore meadows. Oikos, 22: 1–20.

Webber, P.J., Richardson, P.W. and Andrews, J.T., 1970. Postglacial uplift and substrate age at Cape Henrietta Maria, southeastern Hudson Bay, Canada. Can. J. Earth Sci., 7: 317–325.

Woo, M.K. and Heron, R., 1987. Breakup of small rivers in the subarctic. Can. J. Earth Sci., 24: 784–795.