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Hillslope runoff from an ice-cored peat plateau in a discontinuous permafrost basin, Northwest Territories, Canada

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Abstract:

Peat plateaus are important landscape features of many high-boreal, wetland-dominated drainage basins. Raised up to 2 m above the surrounding landscape and underlain by permafrost, these forested peatlands provide meltwater drainage to the surrounding wetlands, and to basin runoff. Understanding the factors that control the volume and timing of runoff from peat plateaus is the essential first step towards developing methods of accurately predicting basin runoff from wetland-dominated basins in the region of discontinuous permafrost, as well as understanding the basin response to hydrological changes brought on by the thermal degradation and thaw of permafrost peatlands. In this study, a water balance approach and the Dupuit–Forchheimer equation were used to quantify sub-surface runoff from a forested peat plateau at Scotty Creek, a small (152 km^2), wetland-dominated discontinuous permafrost basin in Northwest Territories, Canada. These two computations yielded similar results in both years of study (2004–2005), and showed that runoff accounted for approximately half of the moisture loss from the peat plateau, most of which occurred in response to snowmelt inputs. The melt of ground ice was also a significant source of water during the study periods, which was largely detained in soil storage. Soil moisture conditions prior to soil freezing were a major factor controlling the volume of runoff from the hillslope. Sub-surface drainage rates declined dramatically after the snowmelt runoff period, when the majority of water inputs went to soil storage and evapotranspiration. The minimal lag between rain events and hydrograph response in both years suggests that much of the runoff produced from rain events is rapidly transported to the adjacent wetlands. These results give insight into how current climate warming predictions for northern latitudes could affect the hydrological response of forested peat plateaus, and the basins which they occupy. Copyright © 2008 John Wiley & Sons, Ltd.

KEY WORDS snowmelt; water balance; Dupuit–Forchheimer; active layer; wetland; sub-arctic; forested bog

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INTRODUCTION

Permafrost underlies approximately 25% of the land surface in the Northern Hemisphere (Zhang *et al.*, 2000), and plays an important role in the northern hydrological cycle (Woo and Winter, 1993). A large portion of the permafrost regions of the world are covered by peatlands; as an example, one-half of the peatlands in Canada are found in permafrost-affected regions (Robinson *et al.*, 2003). The spatial extent of permafrost is discontinuous (covering 10–90% of the land surface) in the peatlands of western sub-arctic and northern boreal Canada, and the northern taiga of Europe and Russia. Permafrost in these regions often takes the form of large expansive areas of peatland, completely underlain by a perennially frozen core. These features are called peat plateaus in Canada (Zoltai, 1971; Zoltai and Tarnocai, 1975), and are similar to flat palsas in Russia and palsa plateaus in northern Norway [National Wetlands Working Group (NWWG), 1988]. Peat plateaus are treed in the discontinuous permafrost region of north-western Canada, whereas peat plateaus are commonly treeless in colder areas further north. This distinction is hydrologically important, as the presence of trees affects the amount of snow accumulation, evapotranspiration and runoff of peat plateaus.

This article focuses on treed peat plateaus. The surface of these peat plateaus rise 1–2 m above the regional water table because of the volumetric expansion of frozen peat, and are dominated almost exclusively by *Picea mariana* (in North America) (Camill, 1999; Quinton *et al.*, 2003). Their relatively high topographic position means that they are ombrotrophic, and as the hydraulic gradient is directed away from the plateaus, a high proportion of the hydrological inputs runs off to adjacent wetlands (Woo and Heron, 1987; Quinton *et al.*, 2003; Hayashi *et al.*, 2004). Surface flow is thought to be negligible due to the large water holding capacity (Dingman, 1971), high porosity (>0·7) (Hinzman *et al.*, 1991; Quinton and Gray, 2001) and high frozen and unfrozen infiltration rates of peat soils (Dingman, 1973; Slaughter and Kane, 1979). Thus, the dominant runoff mechanism from peat plateaus is sub-surface flow via the saturated layer between the water table and frost table (Hayashi *et al.*, 2004), as

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ice-rich permafrost confines the water flow to a relatively shallow, seasonally thawed active layer (Dingman, 1970; Roulet and Woo, 1988; McNamara *et al*., 1997). The depth of the frost table controls the magnitude and timing of hillslope drainage to stream channels, given the depth-dependency of saturated hydraulic conductivity (Quinton *et al*., 2000) and the relative impermeability of the upper boundary of the frozen saturated layer.

Hillslopes in permafrost regions are hydrologically important to basin runoff, because they accumulate a large proportion of the basin snow water prior to spring melt (Woo and Marsh, 1978; Woo and Heron, 1987; Metcalfe and Buttle, 1999), and contribute to streamflow generation long after other landscape types have become depleted of snow cover (Marsh and Pomeroy, 1996; Quinton and Marsh, 1998), even into the summer months (Woo and Heron, 1987). Quinton *et al*. (2003) found that in the wetland-dominated zone of discontinuous permafrost, drainage basins with a greater percentage of peat plateaus had greater total annual runoff than basins with a smaller percentage. The role of treed peat plateaus as runoff producers is particularly pronounced during the annual spring melt event, as there is a large amount of water input in a short amount of time, and flow during spring is restricted by shallow ground thaw (because the frozen soil is relatively impervious to water movement and storage), which results in a large amount of runoff to adjacent wetlands (Glenn and Woo, 1987).

Although peat plateaus are common in peatlands of the discontinuous permafrost zone [e.g. Zoltai and Tarnocai, 1975; National Wetlands Working Group (NWWG), 1988; Metcalfe and Buttle, 1999; Tarnocai *et al*., 2000], and appear to play an important role in basin runoff of these regions, very few studies have examined how runoff is generated from peat plateaus. As a result, there is a lack of knowledge on basin water cycling in a region that is undergoing rapid change due to climate warming, increased industrial development, and permafrost degradation (Rouse *et al*., 1997; Serreze *et al*., 2000). Increases in shallow ground temperature and active layer thickness in the southern boundary of discontinuous permafrost, and thermokarst development in recent decades, have been observed in North America (Brown *et al*., 2000; Osterkamp, 2005), Europe (Harris *et al*., 2003) and Russia (Frauenfeld *et al*., 2004), indicating degradation of warmer permafrost. In continental western Canada, there is evidence of extensive degradation and widespread disappearance of peat plateaus for some time (Robinson and Moore, 2000; Jorgenson *et al*., 2001; Beilman and Robinson, 2003; Camill, 2005). The implications of these changes on the future runoff production from the basins of this region are poorly understood.

Understanding the factors that control the volume and timing of runoff from peat plateaus is the essential first step toward developing methods of accurately predicting basin runoff from wetland-dominated basins in the region of discontinuous permafrost, as well as understanding the basin response to hydrological changes brought on by the thermal degradation and thaw of permafrost peatlands. Thus, the objective of this study was to determine the timing and volume of sub-surface runoff from a peat plateau hillslope during spring melt. A water balance approach was used to determine the magnitude of runoff, and to assess the relative contributions of snowmelt, ground ice melt, and soil storage to runoff generation from a permafrost peatland. A second estimation of runoff was also computed from hydraulic conductivity and gradient to corroborate the results from the water balance approach. Data sets such as these are useful in assessing the predictions of hydrological models, but are rare due to the logistics of northern research.

SITE DESCRIPTION

Scotty Creek is a 152 km^2 wetland-dominated drainage basin located in the lower Liard River valley (61°18′N, 121°18′W), 50-km south of Fort Simpson, Northwest Territories, Canada (Figure 1(a)). The stratigraphy of the region includes an organic layer of varying thickness (*ca* 0·5 to 8 m) that overlays a silt-sand layer, below which lies a thick (several metres) clay to silt-clay deposit of low permeability (Aylsworth *et al*., 1993). The Fort Simpson area is in the continental high-boreal wetland region of Canada, slightly south of the transition to low sub-arctic [National Wetlands Working Group (NWWG), 1988], and lies within the southern fringe of discontinuous permafrost (Heginbottom and Radburn, 1992). The region is characterized by a dry continental climate, with short, dry summers, and long cold winters. The 1964–2005 mean annual air temperature at Fort Simpson airport (169 m a.s.l.) is −3·2 °C, and the mean January and July temperatures are −25·9 and 17·1 °C, respectively [Meteorological Service of Canada (MSC), 2005]. The region receives an average of 363 mm of precipitation annually, of which 45% falls as snow [Meteorological Service of Canada (MSC), 2005]. Snowmelt usually commences in late March and continues throughout April, with small amounts of snow typically remaining on the ground during the early weeks of May.

The upper two-thirds of the Scotty Creek drainage basin is a peatland complex composed of peat plateaus, flat bogs and channel fens (Figure 1(b)). Spatial analysis of a *ca* 22 km^2 representative sample of this complex (Figure 1(b)), indicated that peat plateaus occupy the largest areal portion (43%), followed by flat bogs, and then channel fens. Most field measurements were made near the centre of this complex at the Study plateau (61°18′48·9″N, 121°18′22·7″W).

The Study plateau rises 0·9 m above the surrounding wetlands, and its active layer, measured over four consecutive years (2003–2006) extends to an average depth of $0{\cdot}64 \pm 0{\cdot}18$ m. The upper layer typically extends to an approximate depth of 0·2 m, and is composed of living vegetation and lightly decomposed fibric peat, while the lower layer contains denser, more decomposed sylvic peat with dark, woody material, and the remains of lichen, rootlets and needles (Quinton *et al.*, 2003). This two-layered peat profile is similar to that found in wetlands possessing acrotelm and catotelm layers (Ivanov, 1981). However, unlike a catotelm, the lower peat layer is not permanently saturated (water drains away from the peat profile as the soil thaws). The active layer mantles permafrost that is between 5 and 10 m in thickness (Burgess and Smith, 2000).

The Study plateau supports an open tree canopy, composed predominantly of black spruce (*P. mariana*). Maximum measured tree height is 9·3 m, and the mean tree height is $3{\cdot}1 \pm 2{\cdot}2$ m. Mean tree density is approximately 1 stem m^{-2}. Shrubs occupy 56% of the plateau and are dominated by Labrador tea (*Ledum groenlandicum*), small bog cranberry, Leatherleaf (*Chamaedaphne calyculata*) and bog birch (*Betula glandulosa*). *Cladina mitis, Cladina rangiferina* and other lichen species cover 65% of the forest floor, while the remaining 35% is occupied by *Sphagnum fuscum, Sphagnum capillifolium* and other moss species. The microtopographic variation on the ground surface of the peat plateau ranges from less than 10 cm to a few tens of centimetres per square metre. Although standing water occupied up to one-fifth of the plateau during the spring melt period, it accounted for <1% of the plateau ground surface during summer.

METHODS

Water balance equation

Field study was conducted in the spring of 2004 and 2005, to estimate the amount of spring runoff, and to assess the relative importance of the major hydrological elements of the Study plateau. The daily net runoff (R_1) was computed from the water balance:

$$R_1 = M + I_M + P - ET - \Delta S \qquad (1)$$

where snow melt (M), melt of ice in the active layer (I_M), rainfall (P), evapotranspiration (ET), and active layer moisture storage (ΔS) all have units of millimetre per day. Although some localized surface flows were discernable, once the snow cover on the plateau had melted, surface discharge on the plateau, measured from the velocity–area method, was relatively small (<1%). Thus, R_1 essentially represents sub-surface runoff, and will be treated as such in this study. Each of the terms in the right-hand side of Equation (1) were measured or estimated as follows.

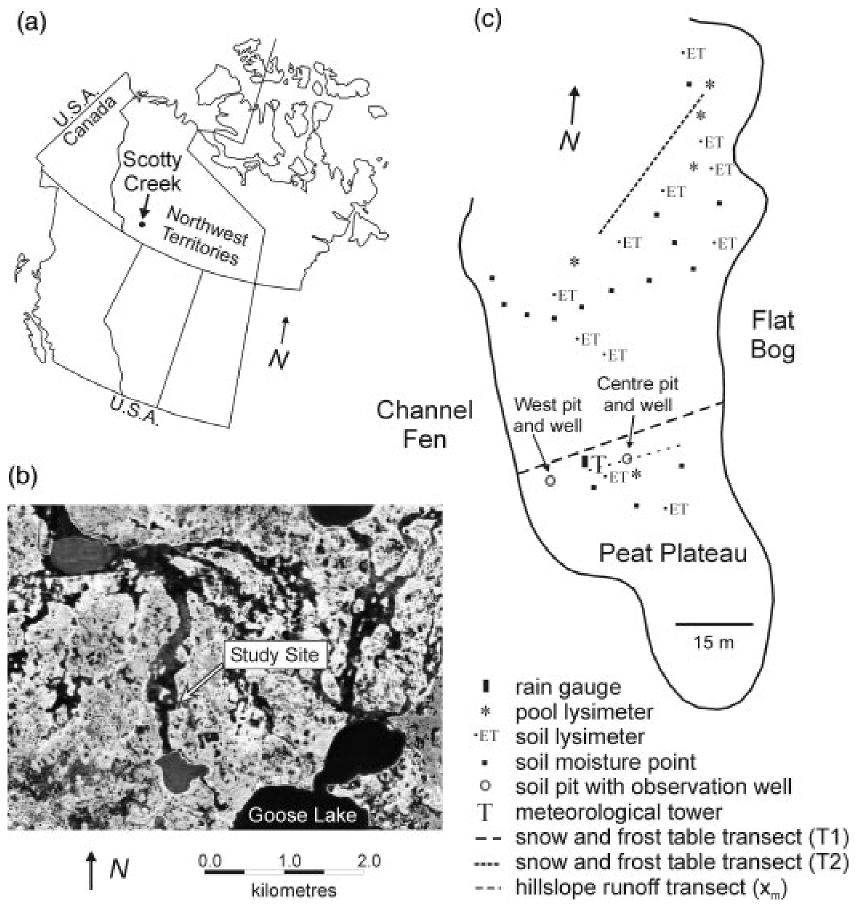

Figure 1. (a) The location of the Scotty Creek basin within north-western Canada. (b) A sample of the high-resolution (4 m × 4 m) IKONOS image showing a 22 km^2 section in the southern part of Scotty Creek basin where field studies are concentrated. The unclassified image has been converted from false colour to a grey scale. Channel fens appear relatively dark compared with the surrounding areas composed of flat bogs and peat plateaus. (c) The peat plateau study site with the location of instrumentation

Snow measurements and site instrumentation

Snow depth and snow water equivalent (SWE) measurements commenced at the height of the snow-accumulation season, and were made daily using an Eastern Snow Conference snow sampler (ESC-30) (Goodison, 1978) and calibrated (to $\pm$ 3 mm) scale. Snow depth was measured at 1 m, and SWE at 5 m intervals along a 41-m transect (T1) that traversed the Study plateau from channel fen to flat bog (Figure 1(c)). In 2005, a 28-m long transect (T2) was added, and like T1, was used for snow depth and SWE measurements following the same measurement intervals. The depth of melt M was calculated from the difference in successive daily measurements of SWE averaged along each transect.

Parallel to T1, were two polyvinyl chloride (PVC) observation wells (0·05-m i.d.), located 2·7 m (West well) and 15·9 m (Centre well) from the edge of a channel fen (Figure 1(c)). Each well was instrumented with a Global Water WL15 pressure transducer that measured water-table depth every minute, and averaged and recorded these measurements every 30 min. The water-table depth at each of the wells was also measured manually with a water level sounder and ruler on a daily basis, to ensure the accuracy of the continuous measurements. The Centre and West wells were each installed in a soil pit that was excavated to the frost table (a depth of 0·7 m) on August 20, 2001. Approximate locations of the two pits are shown in Figure 1(c), though only the data from the Centre pit were used in this study. The Centre pit was instrumented with soil temperature sensors (Campbell Scientific 107B) at 0·05, 0·1, 0·15, 0·2, 0·25, 0·3, 0·4, 0·5, 0·6, and 0·7-m depths; and soil moisture sensors (Campbell Scientific CS 615) at 0·1, 0·2, 0·3, and 0·4-m depths. A 92 cm^3 soil sample was taken next to each of the CS 615 water content reflectometers (WCR) at the time the sensors were installed, in order to measure the soil porosity and bulk density in the laboratory following the approach of Quinton *et al.* (2000).

At a meteorological station located between the Centre and West wells, upward and downward-directed short and long wave radiation at 2 m (Kipp & Zonen, CNR1), wind speed at 2·7 m (Met One 014A), relative humidity and air temperature at 2 m (Vaisala, HMP45C) and snow depth (Campbell Scientific, SR50) were measured above a moss-covered ground surface. A tipping-bucket rain gauge (0·2-m diameter, 0·35-m height) calibrated to 0·26 mm per tip was located next to the meteorological station. All sensors were connected to Campbell Scientific, CR10X dataloggers, programmed to measure every minute and averaged and recorded every 30 min.

Evapotranspiration

In 2005, five identical evaporation pans (colourless, plastic pails of 0·19-m diameter and depth) were installed in meltwater pools on the plateau to measure the evaporation rate of the standing water on the plateau. The rims of the pans were between 10 and 20 mm above the surrounding water surface. The decrease in pan water level below a fixed mark was recorded every 24 h. Daily evaporation was obtained by measuring the volume of water required to restore the water level to a fixed mark. Following the daily readings, water was added to each pan so that their water levels were brought back to that of the fixed marks. In addition, 10 soil lysimeters (0·19-m diameter; same as the evaporation pans) were used to measure the daily evapotranspiration from lichen and moss surfaces gravimetrically following the method of Lafleur and Schreader (1994). As drainage was not permitted, the monolith of soil and the surface vegetation in the lysimeters were carefully replaced after large rain events. During extended periods without rain, the near-surface soil moisture (0–0·2 m) inside and outside of the lysimeters was measured every few days with a portable soil water content probe (Campbell Scientific Inc., HydroSense), to ensure the soil in the lysimeters had not dried out, relative to the surrounding soil, because they were cut off from the vertical and lateral transport of water within the soil matrix. When the difference in volumetric water content (VWC) inside the lysimeter was more than 10% of the VWC outside, the soil monoliths were replaced. The daily rate of evapotranspiration from each of the moss and lichen soil-filled lysimeters (ET_{LM} in millimetre per day) was computed from:

$$ET_{LM} = [(\Delta W_w/\rho_w)/A] \cdot 1000, \qquad (2)$$

where ΔW_w is the daily change in weight of the soil-filled lysimeter (kg day^{-1}), ρ_w is the density of water (kg m^{-3}), and A is the cross-sectional area of the lysimeters (m^2). The total daily evapotranspiration for the plateau was computed as a weighted average based on the measured evaporation and percent cover of moss, lichen and pools on the plateau. The percent cover of each was estimated from the relative proportion of the three ground cover types measured in a 1 × 1-m grid, every metre along transects T1 and T2 (Figure 1(c)) on a weekly basis.

For 2004, and for days in 2005 when the lysimeter data were unreliable due to rainfall, evapotranspiration (ET in mm day^{-1}) from the plateau ground surface was estimated using the Priestley and Taylor (1972) method:

$$ET = \alpha E_{eq} = \alpha \left[\frac{s}{s+\gamma}\right] \frac{\lambda}{\rho_w} [Q^* - Q_g]\, 1000 \qquad (3)$$

where E_{eq} (mm day^{-1}) is equilibrium evaporation, α is a dimensionless coefficient to be determined (see below), s is the slope of the saturation vapour pressure—temperature curve (Pa °C^{-1}), γ is the psychrometric constant (0·066 kPa °C^{-1} at 20 °C), λ is the latent heat of vaporization (J kg^{-1}), ρ_w is the density of water (kg m^{-3}), Q^* (J m^{-2} day^{-1}) is the available energy from net radiation, and Q_g (J m^{-2} day^{-1}) is the ground heat flux. All meteorological variables were supplied by the measurements made at the meteorological tower. Ground heat flux was computed from the thermocalimetric method described in Quinton *et al.* (2005), using the temperature and moisture measurements at the Centre pit. Direct measurements of evapotranspiration

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from the pans and lysimeters in 2005 were plotted against E_{eq} to determine α for each of the three ground cover types (Table III).

The methods used in this study account only for evapotranspiration losses from the ground surface of the peat plateau. Although Equation (1) does not include the water loss evaporated from trees, omission of this is not thought to be significant to the water balance calculation, considering the relatively low water loss from black spruce stands in the boreal forest reported in the previous literature (e.g. Kimball *et al.*, 1997; Nijssen *et al.*, 1997; Pattey *et al.*, 1997; Arain *et al.*, 2003). On the basis of the evapotranspiration, rates of the black spruce stands published in these studies (which ranged from an average of 1·4–2·6 mm day^{-1}), total ET could be underestimated by as much as 50 mm in this study; however, the error is expected to be much less, because the peat plateau has an open canopy with a much lower stand density than those studied in the southern and northern boreal forest.

Active layer melt and soil water storage

The partitioning of snowmelt and rainfall input into soil storage and runoff is strongly affected by the soil moisture condition prior to freeze-up in the previous autumn, and by late winter moisture gains (Kane, 1980; and Kane and Stein, 1983; Gray and Granger, 1986). The available soil storage capacity (S_c) prior to freeze-up at the Centre pit was computed from the product of the thickness d_{uz}, and air-filled porosity of the unsaturated zone:

$$S_c = d_{uz} \cdot (\phi - \theta_L) \tag{4}$$

where ϕ is average porosity (= 0·8) of d_{uz}, derived from the single-valued function of porosity with respect to depth reported by Quinton and Hayashi (2005), and θ_L is the average liquid water content measured with all the CS615 sensors located above the water table.

The daily melt rate of ice in the active layer, I_M (mm day^{-1}) at the Centre pit was computed from

$$I_M = (\theta_T - \theta_L) \cdot \frac{\Delta z_f}{\Delta t} \cdot 0{\cdot}9 \tag{5}$$

where θ_T is the total volumetric moisture content (ice + water) below the cryofront, 0·9 accounts for the density difference between ice and water, and $\Delta z_f/\Delta t$ is the daily change in the depth, z_f (m) to the cryofront (i.e. the top of the frozen layer, which was measured with a graduated steel rod). The values of θ_T were not measured in this study, and hence had to be estimated from the liquid water content measured prior to freeze-up in the following manner. The water content profile at the Centre pit was measured in the fall of 2002, immediately before the freeze-up, and two 0·7-m deep frozen peat core samples were obtained near the Centre pit on April 6, 2003 for another study (Hayashi *et al.*, 2007). We assumed the difference between pre-freezing θ_L ($= \theta_T$) measured in 2002 and the average of θ_T of the two frozen peat cores in 2003 gave the over-winter addition of water to the peat profile, presumably due to mid-winter melt of snowpack and the capillary-driven, upward flow of water to the freezing front advancing from the surface. Assuming that the meteorological conditions in 2003–2004 and 2004–2005 winters were similar to 2002–2003 winters, we added the over-winter change of θ_T observed in 2002–2003 to the fall 2003 values of θ_L to estimate the spring 2004 values of θ_T, and to the fall 2004 values of θ_L to estimate the spring 2005 values of θ_T. The total amount of over-winter increase in θ_T was 99 mm over the 0·7-m peat profile.

To determine ΔS in Equation (1) the Centre pit was divided into soil layers, the boundaries of which were set at the mid point between the moisture sensors. The moisture content in each layer was assumed to be equal to the value measured by the sensor it contained. For each day, the liquid moisture content of the active layer was estimated as the sum of the liquid moisture in all layers. To determine how well the Centre pit represented the average soil moisture conditions on the Study plateau, ΔS estimations were also made daily (over 25 days) at 15 sampling points in 2005. Soil moisture was measured gravimetrically at 0·05-m depth increments to the water table at each of the 15 flagged sampling points (exact measurement locations changed to avoid destructive sampling, although all measurements were within 1 m of the flagged point), seven of which were over moss and eight over lichen (Figure 1(c)), using 162-cm^3 soil sample tins. The depth to the water table and frost table were also measured daily at each sampling point, with a ruler and graduated steel rod. The liquid moisture content of the thawed, saturated portion of the active layer at each flagged point was computed as the product of the thickness of layer between the frost and water tables, and the average porosity of the saturated layer, where the latter was derived from the single-valued function of porosity with respect to depth reported by Quinton and Hayashi (2005). For each day, the liquid moisture content of the thawed, saturated and unsaturated portions of the active layer at each sampling point were summed up. As the soil beneath the moss exhibited a significantly ($p = 0{\cdot}001$) greater volumetric water content at depths up to 0·15 m compared to lichen soils, the daily liquid moisture content of the plateau was obtained by averaging the liquid moisture content found at the two cover types of moss and lichen, and then weighting these average values by the proportion of moss (35%) and lichen (65%) found on the plateau.

Runoff computed from hydraulic conductivity and gradient (R_2)

To corroborate the runoff calculation based on the water balance (R_1), sub-surface runoff was independently estimated from the hydraulic gradient and conductivity using the Dupuit–Forchheimer approximation (Childs, 1971). The main assumptions for this method are as follows: (1) sub-surface runoff occurs only in the thawed, saturated zone between the water table and the impermeable frost table, (2) the sub-surface flow rate is equal to the amount of water added to the water table by snowmelt, precipitation minus evaporation, and storage release (Equation (1)), and (3) the flow is one dimensional and parallel to the slope. Our field observations indicate that these assumptions are justified as a first approximation. If the hydraulic conductivity, K (m day^{-1}) is constant, the rate of sub-surface runoff generation, r (m day^{-1}) on a hillslope can be estimated from the hydraulic head measured in two wells (Equation (A6) in Appendix):

$$r = -K y_m[(h_2 - h_1)/(x_2 - x_1)]/x_m \tag{6}$$

where y_m is the average thickness of the saturated zone between the two wells, $(h_2 - h_1)/(x_2 - x_1)$ is the hydraulic gradient between the two wells, and x_m (m) is the distance from the drainage divide ($x = 0$) to the mid point between the two wells. On the peat plateau in this study, the first well is located at Centre pit, the second well is at West pit (Figure 1(c)) and $x_m = 18$ m.

In reality, hydraulic conductivity is not constant on the hillslope requiring that K in Equation (6) be replaced by an average value representing the heterogeneous peat on the hillslope. Among many possible methods to calculate an average value, we chose the following because the depth-dependence of hydraulic conductivity is believed to be the most important factor controlling the hillslope flow in this environment (Quinton *et al*., 2000). First, saturated hydraulic conductivity, K_{sat} (m day^{-1}) was defined as a function of depth z (m) below the ground surface, based on the local data measured at various locations on the peat plateau by several methods (Figure 2), including *in situ* water tracing, constant-head well permeameter tests, and laboratory measurements (Quinton *et al*., 2008). The solid line in Figure 2 indicates the best-fit $K_{sat}(z)$ function representing the peat plateau (Quinton *et al*., 2008):

$$\log K_{sat}(z) = 0{\cdot}15 + 2{\cdot}41/[1 + (z/0{\cdot}15)^{4{\cdot}3}] \tag{7}$$

The function was integrated with respect to z to compute the vertically averaged conductivity, $\overline{K_{sat}}$ (m day^{-1}) of the thawed, saturated zone at the Centre well and the West well:

$$\overline{K_{sat}} = \frac{\int_{zw}^{zf} K_{sat}(z)dz}{z_f - z_w} \tag{8}$$

where z_w is the water-table depth and z_f is the frost table depth. Daily values of $\overline{K_{sat}}$ were computed from measured z_f and z_w. To represent the average conductivity of the hillslope between the two wells, a geometric mean $\overline{K_g}$ of the two $\overline{K_{sat}}$ values were computed daily. The geometric mean, rather than an arithmetic mean, was used because K_{sat} of peat soils is highly heterogeneous (e.g. Chason and Siegel, 1986) varying by orders of magnitude even within the peat plateau (Figure 2). Replacing K in Equation (6) by $\overline{K_g}$, the sub-surface runoff from the peat plateau (R_2) was computed from:

$$R_2 = -\overline{K_g} y_m[(h_2 - h_1)/(x_2 - x_1)]/x_m \tag{9}$$

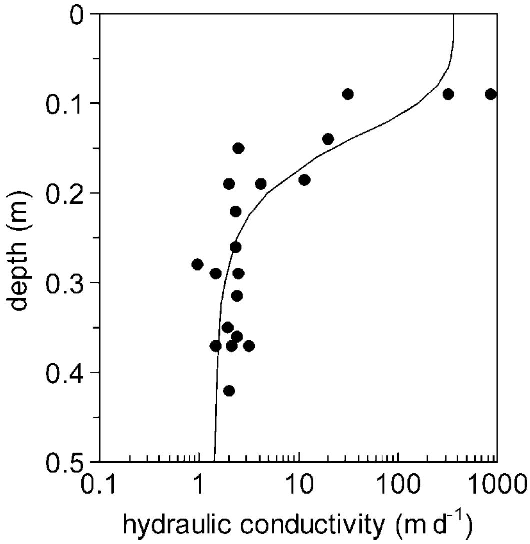

Figure 2. Saturated hydraulic conductivity with depth from the ground surface, measured at the peat plateau in Scotty Creek using various field and laboratory methods (modified from Quinton *et al*., 2008). The solid line indicates the best-fit function determined by Quinton *et al*. (2008)

RESULTS AND DISCUSSION

Snowmelt (M), rainfall (P), and evapotranspiration (ET)

The water balance (Equation (1)) was computed for the spring melt regime, from March 29 to June 4, 2004 and April 19 to June 8, 2005. The computations were split into two periods: the snowmelt runoff period, defined in this study as time from maximum SWE to when the snowpack completely disappeared from the Study plateau (period 1); and the following 3–4 week period from the day the Study plateau became snow-free to the end of the field campaign, during which time the spring freshet had ended (period 2) (Table I).

During the 2 years of study, the late winter snowpack contributed 63% (in 2004) and 46% (in 2005) of the total precipitation input to the Study plateau from late March to September. The SWE reached a maximum and the main phase of snowmelt started on March 29 in 2004, which was 20 days earlier than the maximum SWE in 2005 (Table I). The average daily air temperature during snowmelt was 0·67 °C in 2004 (Figure 3(a)) compared with 2·1 °C in 2005 (Figure 4(a)). This reflects the more

Table I. Measurement periods for the water balance computations. Period 1 represents the snowmelt runoff period, during which time the plateau was snow covered, and period 2 represents the 3–4 weeks from the day the plateau became snow-free to the end of the field campaign. Not all measurements started on the same day, due to frozen soils and/or snow cover; a summary of the start dates for various measurements of the water balance computations is listed in italics

	2004	2005
Period 1	March 29–May 12	April 19–May 9
Period 2	May 13–June 4	May 10–June 8
Ice melt	*May 2*	*April 27*
Soil moisture	*May 2*	*April 27*
Evaporation	*April 25*	*April 24*

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intensive melt pattern of 2005 (Figure 4(b)) compared to a relatively gradual melt, interspersed with sub-zero temperatures and additional snowfall events, in the preceding year (Figure 3(b)). So, although the snow pack yielded similar amounts of water in both years (202 mm in 2004 and 206 mm in 2005), it melted in about half the time in 2005 (Figures 3(b), 4(b)). An additional 20-mm SWE fell after the plateau became snow-free in 2004, which resulted in 16 mm more melt water input to the plateau in 2004 relative to 2005 (Table II).

Both study years had a higher than normal (from 1964 to 2005) amount of spring rainfall [Meteorological Service of Canada (MSC), 2005]. In 2004, 53 mm of rain fell during the 68-day measurement period for the water balance computations, 91% of which was deposited in the 3-day event of May 25–27 (Figure 3(c)). In 2005, 63 mm fell during the 51-day measurement period (Figure 4(c)).

Evapotranspiration from snow-free surfaces commenced on April 25 in 2004 and April 24 in 2005. The mean daily evapotranspiration rates varied dramatically between the three different cover types of moss, lichen, and melt water pools (Figure 5), with moss surfaces evaporating at much higher rates than lichen surfaces and pools located in low-lying areas (Table III). The greater evaporative loss from moss soils relative to those underlain by lichen was mostly due to the differences in near-surface moisture availability (as detailed in the following section). The difference in evaporation rates between moss and melt water pools was most likely due to the water wicking capability of moss species (Vitt, 2000), their greater aerodynamic roughness compared to pools, and to their higher landscape position relative to the pools (Lafleur and Schreader, 1994). Mean daily evapotranspiration for the plateau weighted according to the proportion of moss, lichen and pools was $1{\cdot}4 \pm 0{\cdot}5$ mm for the 2004 study period, and $1{\cdot}5 \pm 0{\cdot}5$ mm for the 2005 study period. The cumulative weighted evapotranspiration by the end of spring melt on June 4, 2004 was 56 mm, and 67 mm on June 8, 2005 (Table II).

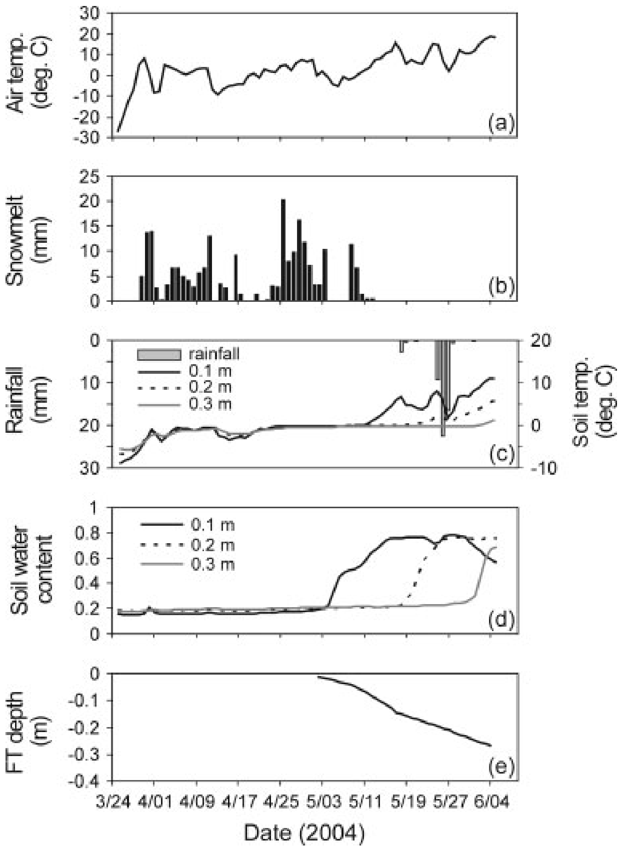

Figure 3. 2004 (a) air temperatures (Air temp.); (b) daily snow melt plotted from the day of maximum SWE; (c) rainfall depth plotted with soil temperatures (Soil temp.) measured at the Centre pit; (d) soil volumetric water content measured at the Centre pit; and (e) frost table (FT) depths measured manually at the Centre pit with a graduated steel rod

Active layer melt (I_M) and soil water storage (ΔS)

Rainfall recorded at the Study plateau from August to November of 2003 was 140 mm, approximately two and a half times the depth that fell over the same period in 2004 (55 mm). As a result, pre-freeze-up value of available soil storage capacity (S_c) was 53 mm smaller in the

Table II. Water balance computations for the 2004 and 2005 spring melt regime (period 1 + 2): March 29–June 4, 2004 and April 19–June 8, 2005, broken up into the snowmelt runoff period (period 1), and the following 3–4 weeks, during which time the Study plateau was snow-free and the spring freshet had ended (period 2). The runoff ratio in this instance is equal to $R_1/(M + P)$. The runoff rate (mm day^{-1}) is the average daily rate of runoff computed over the specified time period. All other values are expressed in mm

2004	Snowmelt M	Precipitation P	Active layer melt I_M	Evaporation ET	Soil moisture change ΔS	Computed runoff R_1	Runoff ratio	Runoff rate
Period 1	222	0	32	21	29	204	0·92	4·5
Period 2	0	53	130	35	114	34	0·64	1·5
Period 1 + 2	222	53	162	56	143	238	0·87	3·5

2005	Snowmelt M	Precipitation P	Active layer melt I_M	Evaporation ET	Soil moisture change ΔS	Computed runoff R_1	Runoff ratio	Runoff rate
Period 1	206	0	58	25	55	184	0·89	8·8
Period 2	0	63	105	42	97	29	0·46	1·0
Period 1 + 2	206	63	163	67	152	213	0·79	4·2

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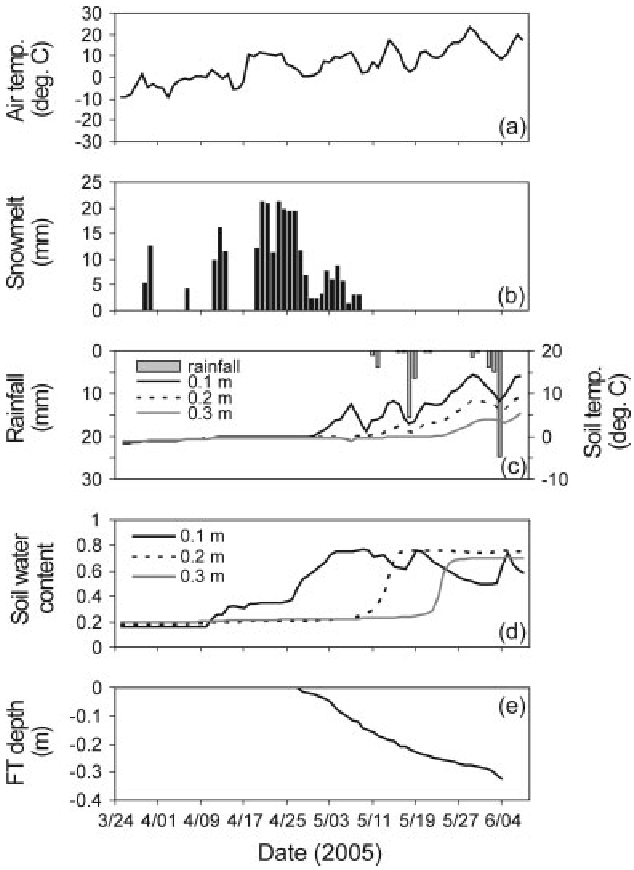

Figure 4. 2005 (a) air temperatures (Air temp.); (b) daily snow melt (note: water balance computations begin on the day of maximum SWE, April 19); (c) rainfall depth plotted with soil temperatures (Soil temp.) measured at the Centre pit; (d) soil volumetric water content measured at the Centre pit; and (e) frost table (FT) depths measured manually at the Centre pit with a graduated steel rod

Table III. Mean daily evapotranspiration and standard deviation (mm) for the dominant ground covers of moss and lichen on the plateau measured from May 2 to June 1, 2005, and the mean daily evaporation from meltwater pools from May 2 to May 14, after which the standing water had drained and/or evaporated. Average α represents the ratio of actual (lysimeter) to equilibrium evapotranspiration (mm) (Equation (3)). The average α of the peat plateau ground surface, based on the percent coverage of the three cover types, ranged from 0·68 to 0·91

	Moss	Lichen	Water
Mean ET	$2{\cdot}0 \pm 1{\cdot}0$	$0{\cdot}8 \pm 0{\cdot}3$	$1{\cdot}6 \pm 0{\cdot}5$
average α	$1{\cdot}2 \pm 0{\cdot}6$	$0{\cdot}4 \pm 0{\cdot}2$	$1{\cdot}1 \pm 0{\cdot}4$

autumn of 2003 compared to the drier autumn of 2004. If it is assumed that over-winter changes in soil moisture storage are similar in both years, then the capacity of the active layer to absorb melt and rainfall inputs would have been proportionately higher in the melt period of 2005 than during the same period of 2004. This seems reasonable, given the relatively small increase in soil water content prior to melt in 2005, as described below.

In 2005, the WCR at 0·1 m indicated an increase in water content on April 11 (Figure 4(d)), likely in response to meltwater infiltration from an early snow melt event (Figure 4(b)), as the soils were still largely frozen (Figure 4(e)) (Stein and Kane, 1983). By April 27, when soil thaw began in 2005, the liquid water content at this depth had increased by approximately 0·15 since April 11 (Figure 4(d)). This 15-mm increase in water content at 0·1 m potentially reduced the difference in the S_c, estimated at soil freeze-up between the 2 years, from 53 to 38 mm, as no similar increase in water content was indicated by the WCR at 0·1 m in 2004 until May 4 (Figure 3(d)), 2 days after soil thaw began (Figure 3(e)).

The timing of soil thaw initiation also played a role in the amount of active layer ice melt that occurred during the snowmelt runoff period (period 1) in both years. Active layer ice melt during this period was 26 mm higher in 2005 than in 2004, mostly due to the warmer air and soil temperatures (Figures 3(a), (c) and 4(a), (c)) and the earlier initiation of ice melt in 2005 (Figures 3(e) and 4(e)); while during period 2, 25 mm more ice melt was produced in 2004 relative to 2005 (Table II). The cumulative input of water from ice melt in the active layer at the Centre pit was 162 mm in the 34-day measurement period of 2004 and 163 mm in the 43-day period of 2005 (Table II), or 37% and 38% of the total water budget inputs during spring melt in 2004 and 2005, respectively.

Soil moisture conditions at the Centre pit, like ice melt, were similar in both years. Soil moisture storage at the Centre pit was computed from the day soil thaw was initiated to the end of the study period: May 2 to June 4, 2004 and April 27 to June 8, 2005. During this time, the total change in liquid water content ΔS at the Centre pit was 9 mm greater in 2005 than 2004 (Table II). ΔS was also measured at 15 sampling points for comparison with ΔS measured at the Centre pit. The daily standard deviation of soil moisture measured at the 15 sampling points varied from 31 to 96 mm, indicating a high spatial variability in moisture conditions on the peat plateau. When the average ΔS values of the 15 points are added for the 25 days when the data were available, the total ΔS was 47 mm, while the total ΔS for the Centre pit for the same 25 days was 44 mm. Thus, the total ΔS measured at the Centre pit was thought to represent the average moisture conditions on the Study plateau during spring melt reasonably well.

Runoff computed from the water balance (R_1)

The total runoff computed over the entire study period was 238 mm in 2004 and 213 mm in 2005 (Table II). As soil thaw at the Centre pit did not occur until May 2 in 2004 and April 27 in 2005, it cannot be concluded that all of the runoff computed from Equation (1) was indeed sub-surface runoff. Much of the runoff (84% in 2004 and 70% in 2005) had occurred before the soil began to thaw. However, many authors have found that snow meltwater can enter, percolate, and laterally drain through the organic layer, even under frozen conditions (Woo and Heron, 1987; Quinton and Marsh, 1999; Carey and Woo, 2000), as the near-surface soil layer is typically made porous by an upward flux of moisture to the overlying snowpack during winter (Smith and Burn, 1987).

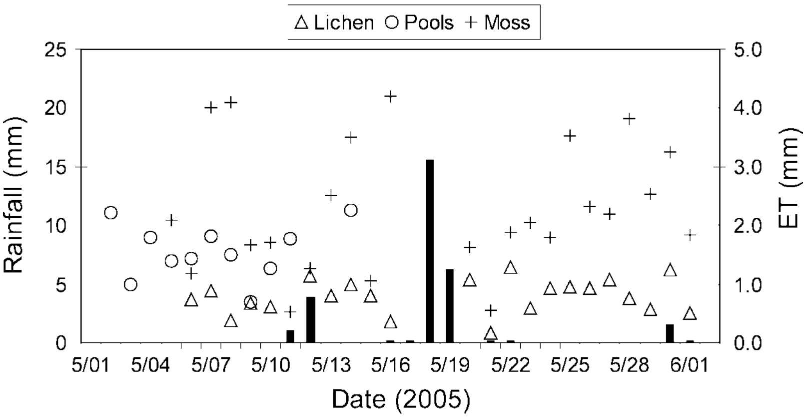

Figure 5. Daily evapotranspiration (ET) from moss and lichen-covered ground surfaces, and evaporation from melt water pools, measured on the Study plateau from May 2 to June 1, 2005, plotted with rainfall depth

A comparison of the water balances (Table II), showed that both years had significantly larger runoff totals and flow rates during snowmelt (period 1), compared to the following weeks (period 2). During the snowmelt period, the difference in runoff timing between the two years was mostly due to differences in the initiation of snowmelt and soil thaw. Sub-surface drainage rates decreased after the snowmelt runoff period (period 2), as ET and ΔS increased (Table II). For the most part, rainfall events produced a rapid response to lateral flow downslope; however, during larger rain events that occurred over more than one day, the runoff response was slower (Figure 6). For example, during the 18–19 May rain event in 2005, the hillslope runoff response was rapid on the first day of rainfall, however, the majority of rainwater inputs went to soil storage on May 19, and was laterally transported downslope over the following day (Figure 6(b)). The melt of ground ice resulted in a large source of water to the peat plateau; however, most of this input (88–93%) was detained in the soil as ΔS (Table II). The ratios of runoff to precipitation computed for the snowmelt runoff period (Table II) are similar to those published for other permafrost environments (Kuchment *et al.*, 2000).

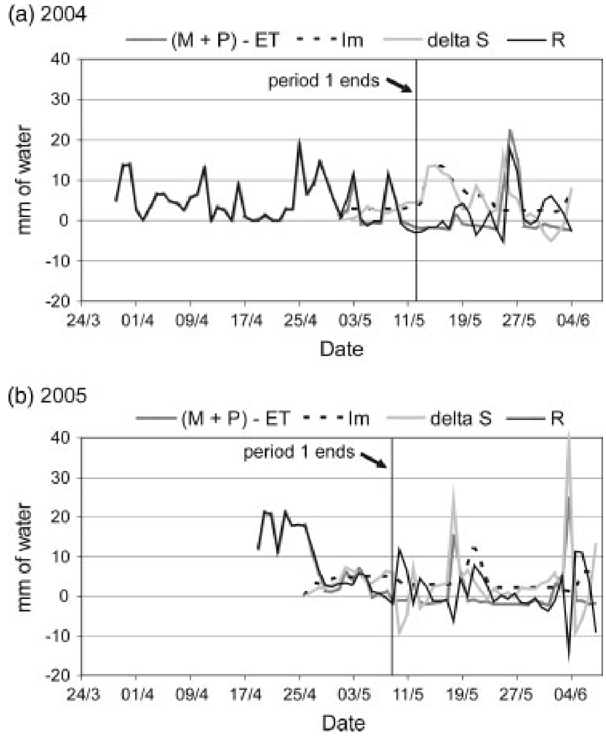

Figure 6. Daily values (in mm) of the water balance components. Runoff was computed using soil moisture measurements at the Centre pit in (a) 2004 and (b) 2005

Uncertainty in the water balance calculation

To estimate the accuracy of M in Equation (1), the standard deviation of all SWE measurements was calculated daily. The average of all daily SWE standard deviations was 11% of the maximum SWE in 2004 and 15% in 2005. The potential error in measuring evapotranspiration, taken to be equal to the areally weighted, mean daily standard error of the five pool evaporation pans, and the five moss and five lichen lysimeters, was $0{\cdot}15$ mm day^{-1}, or 9% of the mean daily evapotranspiration for the 24 days in which ET was measured in 2005. For those days in which ET had to be estimated from Equation (3), errors could be as high as 15% (Shuttleworth, 1993).

The accuracy of the computations of the melt of ice in the active layer is difficult to assess owing to the uncertainty of the performance of the CS615 water content metres in frozen conditions, which is not well known. For unfrozen conditions, their accuracy is expected to be $\pm$ 2%, as they were calibrated for the peat soils at the Study plateau. The time domain reflectometry (TDR) calibration equations of Spaans and Baker (1996) cannot be applied directly to CS615 data (Seyfried and Murdock, 1996). The use of TDR probes in measuring the dielectric permittivity to determine the liquid water content of unfrozen soil has been found to be highly accurate and relatively similar to CS615 WCR results under the same moisture conditions (Seyfried and Murdock, 2001; Yoshikawa *et al.*, 2004). Thus, the potential error of using the CS615 to measure the liquid water content in frozen soils is assumed to be $0{\cdot}02$ m^3 m^{-3} based on the relative errors published for

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TDR sensors (Spaans and Baker, 1995). It is difficult to estimate the magnitude of errors associated with the estimate of the total water content (θ_T) prior to snowmelt; however, we feel that an error of 10–20% is appropriate for this parameter based on the inter-annual variability in the fall water content.

Although Equation (1) includes all readily measurable water balance terms, it does not include a term that accounts for runoff inputs from upslope areas. These lateral inputs appeared to have an influence on the calculation of R_1 from Equation (1). For example, several instances were noted where the increase in soil moisture following a rain event exceeded the depth of rainfall. Ponding occurred at the Centre pit during snowmelt and following heavy rain, suggesting that the Centre pit was susceptible to runoff inputs. Consequently, Figure 6 shows instances of negative runoff, indicating that the sub-surface drainage to the soil pit was greater than the runoff from the pit. If it is assumed that the sum of the negative runoff computed for the study period was caused only by lateral inputs to soil storage, then the potential error of ΔS would be 13% in 2004 and 24% in 2005.

On the basis of the error estimates of individual water balance components, and noting that some errors may cancel each other over the entire period, we expect that the computed runoff (R_1) in Table II may have an error margin of 20–30%, although this only provides a rough estimate.

Runoff computed from hydraulic conductivity and gradient (R_2)

Sub-surface runoff computations commenced (when the water table was measurable in both the West and Centre wells) on April 26 in 2004, 28 days after maximum SWE (and the start of R_1 computations), and on April 22 in 2005, 3 days after maximum SWE (Figure 7). The total R_2 during spring melt was 175 mm in 2004 and 196 mm in 2005, whereas the total R_1 was 238 mm in 2004 and 213 mm in 2005. Thus, the two methods used to compute runoff (R_1 and R_2) from the Study plateau produced relatively similar runoff totals for the spring melt in 2005. To explain the difference between R_1 and R_2 for 2004, we note that sub-zero air temperatures occurred throughout the melt season in 2004 (Figure 3(a)), which may have caused the snowmelt water to re-freeze in the upper soil (depth <10 cm) and melt again at a later date, when air temperatures increased. The re-freezing of melt water was not accounted for in Equation (1) because the shallowest WCR was at 10 cm, and also because the WCR was insensitive to the change in ice

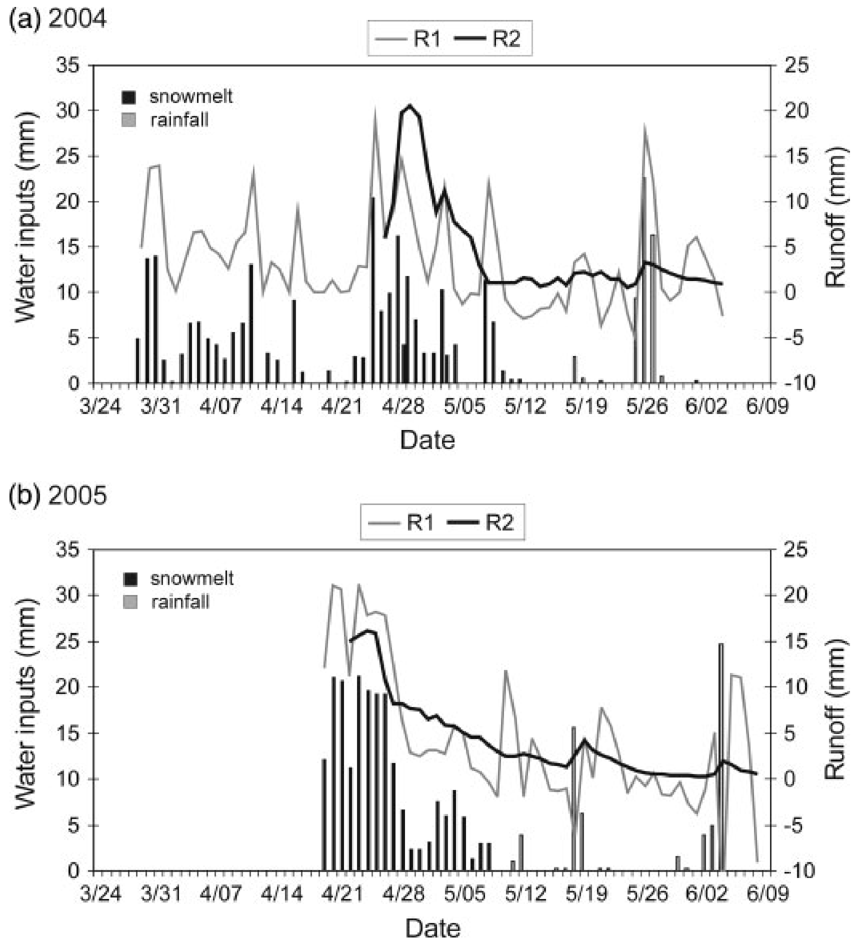

Figure 7. Daily snowmelt and rainfall depth, plotted with sub-surface runoff computed from the water balance (R_1) and from the hydraulic method (R_2) during the (a) 2004 and (b) 2005 study periods. Note that R_2 computed from Equation (9) was converted from metre per day to millimetre per day for comparison with R_1

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content. Therefore, Equation (1) may have overestimated runoff during the early snowmelt period, prior to April 26, 2004. It is also important to note that the R_2 computation missed 28 days of the snowmelt period in 2004, because of a delay in the water-table formation in the wells, possibly due to the sub-zero air temperatures that occurred throughout the melt season (Figure 3(a)). Although the peat normally transmits water readily when thawed, frozen icy peat may have inhibited lateral flow to the wells, which may have also caused some of the 63-mm difference in runoff amounts computed from the two methods (R_1 and R_2) in 2004. However, given the relative errors of the water balance approach, the runoff totals of the two methods were relatively similar in 2004.

The hydrographs for flows computed from the water balance and the hydraulic method are illustrated in Figure 7 for both years. The hydrographs for both years follow a similar pattern, with high flows in response to snowmelt when the water and frost tables were near the ground surface, followed by a general decline as the saturated layer became deeper with time and there were smaller water inputs to the Study plateau. The slow runoff recession of R_2 relative to snowmelt in 2005 implies that there was temporary storage in the soil (which is consistent with the larger pre-melt S_c in 2005, relative to S_c in 2004). The minimal lag between rain events and hydrograph response of R_2 in both years suggests that much of the runoff produced from rain events is rapidly transported to the adjacent wetlands.

CONCLUSIONS

Snow melt water constitutes a significant portion of the annual water input to the peat plateau, and was the largest contributor to runoff during spring melt. In both years, flow rates were significantly larger during the snowmelt runoff period, when there was more available water and frost table depths were shallow, than in the subsequent 3–4 weeks. During snowmelt, over half of the water inputs to the hillslope went to runoff. A major influence on the volume of runoff from the peat plateau during snowmelt was the soil storage deficit of the thawed layer, which had to be satisfied before water inputs could be transmitted downslope. After the snowmelt runoff period, sub-surface drainage rates declined dramatically, as the majority of water inputs went to soil storage. The minimal lag between rain events and hydrograph response in both years suggests that much of the runoff produced from rain events is rapidly transported to the adjacent wetlands. However, the timing of hillslope runoff response to rain water inputs is dependent on the timing and magnitude of the rain events. The melt of ground ice was a significant source of water to the peat plateau during the study periods; however, most of this water was detained in soil storage. Evapotranspiration from the plateau was relatively low and was of roughly equal magnitude to precipitation inputs. The results show the importance of correctly accounting for changing storage and storage capacity in hydrological models that are applied to permafrost hillslopes, specifically to accurately account for snow melt inputs and antecedent moisture conditions, as they are important in determining the runoff response of peat plateaus during spring melt.

Two methods (a water balance and Dupuit–Forchheimer approximation) were used to estimate spring runoff in this study, both of which yielded similar results. This suggests that the Dupuit–Forchheimer approximation may provide a simpler approach to accurately estimate the amount of runoff from peat plateaus compared to the field-intensive water balance approach; this assumes that an average K_{sat} value, representing the heterogeneous peat on the hillslope, can be obtained.

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APPENDIX

Dupuit–Forchheimer approximation of hillslope flow

Hillslope flow in a uniform material bounded by an impermeable sloping surface (e.g. permafrost table) can be modelled using the Dupuit–Forchheimer approximation (Childs, 1971). When the flow is driven by constant recharge flux r (m s^{-1}), the flow equation is given by (McEnroe, 1993, Equation (9))

$$rx = Ky\left(\tan\beta - \frac{\mathrm{dy}}{\mathrm{dx}}\right)\cos^2\beta \tag{A1}$$

where K (m s^{-1}) is hydraulic conductivity, y (m) is the saturated thickness (i.e. vertical distance between the water table and the impermeable boundary), β is the slope angle, and x (m) is the horizontal distance from the top of the slope (i.e. drainage divide). Suppose that two water-table wells are located on a slope transect at x_1 and x_2. Integrating Equation (A1) from $x = x_1$ to $x = x_2$ yields the following:

$$\frac{r({x_2}^2 - {x_1}^2)}{2\cos^2\beta} = K\tan\beta\int_{x1}^{x2} y(x)dx - K\frac{{y_2}^2 - {y_1}^2}{2} \tag{A2}$$

This can be written as follows:

$$\frac{rx_\mathrm{m}}{\cos^2\beta} = K\tan\beta y_\mathrm{m} - \frac{K({y_2}^2 - {y_1}^2)}{2(x_2 - x_1)} \tag{A3}$$

where $x_\mathrm{m} = \frac{x_1 + x_2}{2}$ and $y_\mathrm{m} = \frac{1}{x_2 - x_1}\int_{x1}^{x2} y(x)dx$ represent the mean values of x and y, respectively, in the interval $x_1 < x < x_2$. To simplify Equation (A3), we approximate y_m by an arithmetic mean of the two end values, $y_\mathrm{m} \cong (y_1 + y_2)/2$, and recall the definition of the slope angle,

$$\tan\beta = (\zeta_{\mathrm{f}2} - \zeta_{\mathrm{f}1})/(x_1 - x_2) \tag{A4}$$

where $\zeta_{\mathrm{f}1}$ and $\zeta_{\mathrm{f}2}$ are the elevation of the frost table (m) at x_1 and x_2, respectively. It follows from Equation (A3) that

$$\begin{aligned}\frac{rx_\mathrm{m}}{\cos^2\beta} &= K\frac{(\zeta_{\mathrm{f}1} - \zeta_{\mathrm{f}2})(y_2 + y_1)}{2(x_2 - x_1)} - \frac{K(y_2 - y_1)(y_2 + y_1)}{2(x_2 - x_1)} \\ &= K\frac{y_2 + y_1}{2}\frac{(\zeta_{\mathrm{f}1} + y_1) - (\zeta_{\mathrm{f}2} + y_2)}{x_2 - x_1}\end{aligned} \tag{A5}$$

Noting that water-table elevation h (m) is given by $h = \zeta_\mathrm{f} + y$, and that $\cos^2\beta \cong 1$ for gentle slopes (e.g. $\beta < 5°$), such as the slope of the peat plateau, Equation (A5) reduces to

$$r = -\frac{Ky_\mathrm{m}}{x_\mathrm{m}}\frac{h_2 - h_1}{x_2 - x_1} \tag{A6}$$

This equation can be used to estimate the rate of water input to a hillslope from the water-table elevation (i.e. hydraulic head) measured in two wells on the hillslope.