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Dynamics of evapotranspiration from a riparian pond complex in the Western Boreal Forest, Alberta, Canada

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Abstract:

The relative contributions to total actual evapotranspiration (AET) from pond and riparian areas in a pond-wetland complex in the Western Boreal Plain (WBP) of northern Alberta are measured using the Bowen ratio energy balance technique. Measurements show that a pond typical of the WBP evaporates at a rate more than twice that of the adjacent riparian peatland. Relating the actual to potential evapotranspiration over both surfaces yields Priestley–Taylor α coefficients of 0·69 and 1·11 for the peatland and pond respectively. Further results demonstrate that the sheltering and turbulent influences of the adjacent forested areas must be considered in the processes governing the permanence of WBP ponds. That is, forestry practices may inadvertently enhance the evaporative losses from the ponds, over and above the controls exerted by the regional climate. Copyright © 2006 John Wiley & Sons, Ltd.

KEY WORDS evapotranspiration; potential evapotranspiration; Bowen ratio; Priestley-Taylor model; Western Boreal Forest; pond evaporation; wetlands

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BACKGROUND

The Western Boreal Forest (WBF) is a mosaic of vegetation types (upland forest, peatland and ponds) superimposed on three main landforms (coarse-grained outwash, and fine-grained disintegration moraines and low-lying plains), each of which produces different hydrologic responses through topographic and soil controls on hydrologic processes. Wetlands on the Canadian Boreal Plains cover 25 to 50% of the landscape (Kurhy *et al.*, 1993; Vitt *et al.*, 1995) and provide one of the most important waterfowl habitats in North America (Ducks Unlimited Canada, 2000). The Boreal region also represents the largest carbon pool in Canada (Gorham, 1991). However, the Boreal Plain region is among the most threatened. Increased industrial, agricultural and recreational developments have all contributed to the vulnerability of this region's habitat and resources (Alberta Environmental Protection, 1998; Global Forest Watch, 2000).

The pond–riparian–forest mosaic in the WBF is sustained by infrequent wet years within periods of drought, and although mild within a geological context, potential evapotranspiration (PET) exceeds precipitation P in most years, with infrequent wet years occurring on a 10–15 year cycle (Marshall *et al.*, 1999; Devito *et al.*, 2005). Thus, the wetlands and ponds within this region are vulnerable to any climatic change that may alter patterns of P and actual evapotranspiration (AET). Indeed, aerial photographs show that vegetation succession within changing pond surface areas has varied significantly over the past 60 years (Devito, unpublished data).

An ecohydrological approach can be adopted to quantify the hillslope–pond connections in the sub-humid WBF to shed light on these changes and the factors regulating development (and sustainability) of WBF wetlands and ponds. This information is essential to manage the valuable surface and groundwater resources and associated wildlife habitat, and to predict the impact of climatic variability and resource development in the region. For instance, Boreal wetland ponds are sustained by a balance between pond and peatland evaporation and hydrologic connectivity with the surrounding forested uplands. If hydrologic connectivity between these ponds and upland areas is disrupted, then the pond water levels could lower and eventually dry up, exposing the carbon stored in these pond systems to conditions conducive to oxidation. Such breaks in connectivity could result from natural climatic cycles (drought) or due to the impact of some anthropogenic disturbance (access roads, logging, etc.).

PET is the evapotranspiration rate under prevailing solar inputs and atmospheric properties if moisture at the surface is not limiting (Hornberger *et al.*, 1998).

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Given that PET exceeds P during most years in the WBF (Marshall *et al*., 1999), groundwater flow within wetland pond catchments is often from the pond to the hillslope for most of the time within moraine systems, and ponds may represent a significant water source for adjacent peatlands and forests (Ferone and Devito, 2004). In addition, precipitation events can result in dramatic changes in runoff, and potentially in pond levels (Ferone and Devito, 2004). Increased climatic variability will likely influence the duration of drought cycles and pond permanence. Although well documented in prairie systems, the processes influencing drying cycles in western Boreal ponds are poorly understood. For instance, the Utikuma Lake area in northern Alberta is drier than it has been in the previous 50 years based on disappearance of pond features in serial aerial photographs (Devito, unpublished data).

The spatial and temporal variability of AET and factors controlling vapour fluxes in this complex landscape are not adequately understood to provide insight into the impact of climatic and land-use changes. Within complex landscapes like the Western Boreal Plain (WBP), large ranges in the ratio of AET to PET can accompany fine-scale variations in vegetation and microtopography (Petrone *et al*., 2004). Further, temporal variations in the ratio of AET to PET can be expected in riparian peat systems with dramatic changes in water levels influenced by wet and dry cycles (Morton, 1983). Microclimatological variations may result in differences in AET among pond wetland systems as advective influences change (Jacobs *et al*., 2002). An understanding of how large-scale climatic cycles and microclimatic processes influence pond permanence are needed to assess how these wetland pond systems may respond to large-scale climatic variability and human activities in the surrounding upland areas (such as logging and seismic exploration).

OBJECTIVES

The first objective of this study is to determine what the relative AET contributions are from the pond and surrounding riparian peatlands in a pond–wetland complex representative of the highly heterogeneous moraine systems in the WBP. The second objective is to examine the effects of the surrounding forested uplands on sheltering, atmospheric vapour demand and AET in these small wetland ponds, and to infer how this may be impacted by development in these upland areas.

STUDY AREA

Measurements were made during the summer of 2003 within a wetland–pond complex in the WBP of north-central Alberta, Canada (Figure 1). This area is located within the Mixedwood Boreal Plains Ecozone, at the transition between the Mid and High Continental Boreal Subregion (National Wetlands Working Group, 1988). Mean normal summer (July) and winter (January) temperatures for the region are 15·7 °C and −14·6 °C respectively (Marshall *et al*., 1999). Normal annual precipitation and potential evapotranspiration in the region nearly balanced at 515 mm and 517 mm respectively (Marshall *et al*., 1999). Some 50–60% of the annual precipitation occurs on average between June and August, followed by drier autumn months (Marshall *et al*., 1999).

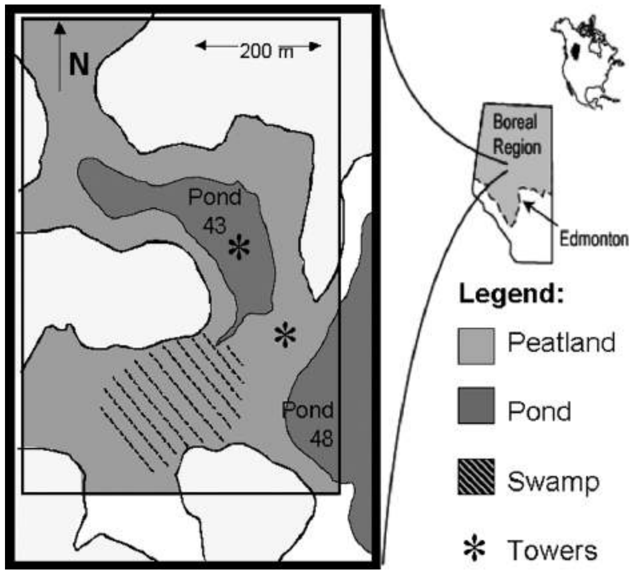

Figure 1. Map showing location of study pond, illustrating the distribution of peatland, pond and swamp and the locations of the Bowen ratio towers (∗). The location of the study area in relation to the Boreal region and North America is also shown

The study area is located approximately 50 km north of Utikuma Lake, 300 km north of Edmonton, Alberta (Figure 1), and occurs along a geological transect comprised of a low-lying clay plain (southeastern portion), a moraine (central portion), and a glacial outwash plain (northwestern portion). Pond 43, a wetland–pond system located on a topographic high stagnant ice moraine (56°20′N, 115°30′W) has a recharge function and represents systems most susceptible to climate changes (Petrone *et al*., 2005) and was chosen to study the relative contributions of riparian peatland and pond surfaces to larger landscape ET. Pond 43 is a small shallow pond with a mean water depth less than 1 m overlaying 2–3·5 m of gyttja. The distance from the shore to the location of the Bowen ratio energy balance tower was about 20 m. The pond is located in the middle of a densely forested area extending several kilometres in all directions. This catchment is 17·4 ha, of which the pond, peatland and upland area constitute 1·4 ha (8%), 5·6 ha (32%) and 10·4 ha (60%) of the catchment area respectively (Ferone and Devito, 2004).

The dense aspen-dominated forests extend down to within approximately 40 m of the pond shore, which is predominately riparian peatland (with characteristics ranging from bog to intermediate or poor fens), along the north, west and east sides. To the south, Pond 43 is bordered by a bog peatland (for ∼100 m) that separates it from Pond 48 (Figure 1). This transition from

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forest to bog and pond causes abrupt changes in aerodynamic roughness for winds crossing the shoreline from all directions. The height of the surrounding upland forests is about 30 m above the pond surface, and the elevation of the terrain in the riparian peatland increases a few metres behind the shoreline. Generally, the vegetation in the Pond 43 catchment consists of aspen (*Populus tremuloides*), balsam poplar (*Populus balsamifera*) and white spruce (*Picea glauca*) forests in the upland areas, and black spruce (*Picea mariana*) in poorly drained areas (National Wetlands Working Group, 1988). The riparian treed bog/fen areas are dominated by black spruce, with some larch (*Larix laricina*), and an understorey of leather leaf (*Chamaedaaphne calyculata*), Labrador tea (*Ledum groenlandicum*) and mosses (*Sphagnum* spp.) (Ferone and Devito, 2004). During the summer period, submergent macrophyte (*Ceratophyllum demersum*, *Potamogetan richardsoni*, *Potamogetan zosteriformis*, and *Myriophyllum exalbescens*) and floating macrophyte growth is typical in the ponds (Ferone and Devito, 2004). The treed bog/fen areas of this catchment consist of fibric peat in the top 30–40 cm and mesic peat to a depth of approximately 1·5 m, overlying mesic-humified peat to a depth of up to 3·5 m (Ferone and Devito, 2004).

METHODS

To assess the distribution of AET between the pond and riparian peatlands two Bowen ratio energy balance systems, and associated instruments, were installed on Julian day (JD) 193 over the pond surface and riparian peatland (Figure 1). The energy balance of any surface is given by

$$Q^{*} = Q_{\mathrm{G}} + Q_{\mathrm{H}} + Q_{\mathrm{E}} \quad (1)$$

where Q^{*} is the net radiative flux at the surface, Q_{G} is the ground heat flux, Q_{H} is the sensible heat flux and Q_{E} is the latent heat flux. Vegetated surfaces generally have a mean daily soil heat flux Q_{G} one or two orders of magnitude smaller than the major terms in the surface energy balance (Brutsaert, 1982). However, on shorter temporal scales, Q_{G} can be quite important. The subsurface soil temperatures, and Q_{G}, are a function of solar radiation, soil texture, soil moisture content and state, in addition to surface vegetation cover and weather conditions (wind, air temperature, humidity) (Williams and Smith, 1989). When considering the exchange of energy from a body of water the amount of heat energy stored within the water column Q_{W} must be considered instead of a substrate (ground) heat flux.

A boardwalk and tower were used as instrument platforms over the pond surface. The instruments for the profile measurements of air temperature and vapour pressure were installed on a 1·5 m tower at the end of the boardwalk, roughly in the middle of the largest portion of the pond. The tower consisted of a Bowen ratio system utilizing a chilled-mirror hygrometer (Campbell Scientific, Utah) and measurements of air temperature (using chromel–constantan thermocouples, 25 µm diameter) at 0·25 and 1·5 m above the water surface, cup anemometer (014A-over the pond surface, Met One, Oregon), net radiometer installed 0·65 m above the water surface (NRLite, Kipp and Zonen, The Netherlands), and wind direction (Wind Sentry set (03 001-L) over the peatland surface, R. M. Young, Michigan). In addition to the energy balance measurements, at the location of the tower temperatures were also measured at the water surface, within the water column (at depths of 0·15 and 0·30 m), the sediment surface and at depth (0·40 m) within the sediment using HOBO thermistor probes (Hoskin Scientific, Vancouver). The water surface temperature was measured using a thermistor affixed to the bottom of a styrofoam float with a small weight suspended from its bottom side to prevent flipping due to wind and waves. This float was tethered to the boardwalk to prevent drifting distances greater than 3 m in any direction. These data were then compared with other thermistor data and different locations in the pond. All of the surface temperature measurements were found to vary on average less than 5% from each other over the course of the season.

An identical Bowen ratio system (Campbell Scientific, Utah) was used over the riparian peatland on the south end of the pond. At this site, as above, vapour pressure and temperatures were measured at 1 and 3 m above the peat surface. Soil moisture was measured using continuously logging time-domain reflectometry probes (CS616, Campbell Scientific, Utah) logged on a CR10X data logger every 20 min (Campbell Scientific, Utah). The ground thermal regime was obtained using thermocouple arrays and heat flux plates (HFT03, Campbell Scientific, Utah) in the peat. Peat surface temperature was measured using a thermistor (CSI-107B) carefully inserted into the surface vegetation cover, which was sampled every 10 s and averaged every 20 min on a CR10X data logger (Campbell Scientific, Utah). The ground temperature profile and ground heat flux were measured in the peat using two ground heat flux transducers (CSI-HFT03; Campbell Scientific, Utah) and four parallel thermocouples (CSI-TCAV; Campbell Scientific, Utah). The CSI-TCAV parallels four thermocouples together to give an average temperature such that two thermocouples can be used to obtain the average temperature of the soil layer above one heat flux transducer, and the other two above the second transducer. The depths of the thermocouples at each transducer were 0·02 and 0·06 m below the surface. The depths of each transducer were 0·08 m below the surface (0·02 m below the lowest thermocouple). The thermocouple arrays and associated transducers were installed approximately 2 m apart to give measurements from a moss-dominated surface area and a vascular-dominated surface patch. Published values for heat capacities of peat soils under a range of moisture conditions were used to determine the heat capacity values to be used in the ground heat flux and storage calculations (Oke, 1987). Ground heat-flux plates provide a good account of the diurnal behaviour of the subsurface heat flux in mineral

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soils, but previous studies indicate that they can significantly underestimate the flux in organic soils due to poor contact between the plate and peat, and the disruption of vapour flow (Rouse, 1984; Halliwell and Rouse, 1987; Rouse *et al*., 1987). Thus, the ground heat flux was measured with two soil heat flux plates as described above, but also calculated using the calorimetric method (Halliwell and Rouse, 1987) using the ground temperature profile and heat capacity calculations for each soil layer accounting for changes in moisture amount and state. The ratio between the seasonal mean heat flux plate measurements and the calorimetric calculations of the total seasonal Q_G was 0·79.

The upper and lower sensors on both Bowen ratio systems were switched every 2 min. After each switch, 40 s was allowed for the mirror to equilibrate to the new dew-point temperature. This is then followed by actual measurements for 1 min 20 s. The dew-point temperature is sampled every second and the vapour pressure calculated online every 20 min. Some Bowen ratio energy balance data were discarded following the rejection criteria based on range limits, visual inspection of plotted net radiation, temperature and vapour pressure values to eliminate periods when sensors were clearly malfunctioning, and on Ohmura's (1982) criteria. Ohmura (1982) suggested that flux calculations be rejected if: (1) the calculated latent heat flux is not in the opposite direction from the observed vapour pressure gradient; (2) the temperature and vapour pressure gradients are at or less than sensor resolution limits, i.e. 90·013 °C and 0·003 kPa. In addition, periods near sunset and sunrise (due to the frequency of near-neutral conditions) and during rain events were rejected. This quality controlling of Bowen ratio data eliminated approximately 40% of the recorded raw data, primarily because of resolution limits or flux directions during night-time periods. This is because energy inputs, temperature and vapour pressure gradients are all relatively low at night. Missing data were filled using a protocol outlined by Crago and Brutsaert (1996), which estimates the latent heat flux using the measured available energy and the relative evaporation from a nearby day with a similar soil moisture status and good Bowen ratio data (Crago and Brutsaert, 1996; Bremer *et al*., 2001).

To ensure accurate comparisons between the pond and peatland sites, net radiometers were tested side by side over a turf surface prior to field installations. Factory calibrations were then adjusted accordingly. The AET calculations determined using the Bowen ratio energy balance method were then compared with PET estimates obtained via the Priestley–Taylor equation.

The Priestley–Taylor equation is classified as a radiation-based approach to estimating PET, using net radiation and air temperature, to evaluate equilibrium evaporation, which assumes that an air mass moving over a homogeneous, well-watered surface would become saturated (Priestley and Taylor, 1972). Under these ideal conditions AET would eventually reach a state of equilibrium (equilibrium potential evapotranspiration PET_{eq}; Priestley and Taylor, 1972). Radiation is a very effective parameter to use in measuring equilibrium evaporation or PET. In a review of 30 studies it was commonly found that approximately 95% of the annual evaporative demand was supplied by radiation in vegetated areas with very small, or no, water deficits (Stagnitti *et al*., 1989). The Priestley–Taylor model obtains PET_{eq} (herein referred to only as PET) via

$$\mathrm{PET_{eq}} = \frac{\Delta}{\Delta - \gamma}(Q^* - Q_G) \tag{2}$$

where Δ (°C kPa^{-1}) is the slope of the saturated vapour pressure curve, γ (Pa K^{-1}) is the psychrometric constant, Q^* (W m^{-2}) is net radiation, and Q_G (W m^{-2}) is the soil heat transfer. PET is related to AET via the Priestley–Taylor coefficient α. The term α is generally solved using

$$\alpha = \mathrm{AET/PET} \tag{3}$$

where AET is the total measured evaporation and PET is the total equilibrium evaporation (Wilson and Baldocchi, 2000). However, equilibrium rarely occurs, as there is almost always horizontal advection and deviations from a 'wet' surface (Wilson and Baldocchi, 2000).

The roughness of a surface, in terms of its aerodynamic properties, is generally described by the surface roughness length z_o:

$$z_o = \frac{z - d}{e^{k(u_z/u^*)}} \tag{4}$$

where u_z is the mean wind speed at height z (1·5 m), k is von Karman's constant (∼0·40), d is the displacement height (approximated as $2h_o/3$, where h_o is the mean height of the vegetation), and u^* is the friction velocity. The surface friction velocity u^* is determined via

$$u^* = \frac{k\overline{u}_z}{\ln(z/z_o)} \tag{5}$$

where $\overline{u}_z$ (m s^{-1}) is the mean wind speed at height z (1·5 m), k is von Karman's constant (∼0·40), and z_o (m) is the roughness length (Oke, 1987).

However, Equation (4) only applies to neutral conditions. For stable or non-stable conditions the above relationship must be corrected with an appropriate profile function (Brutsaert, 1982). Thus, for non-neutral conditions, Equation (4) becomes

$$z_o = \frac{z - d}{e^{k(u_z/u^*)+\psi(\zeta)}} \tag{6}$$

where $\psi(\zeta)$ is the profile function that depends on Q_H and Q_E. For unstable conditions $\psi(\zeta)$ is given by

$$\psi(\zeta) = 2\ln\left(\frac{1+x}{2}\right) + \ln\left(\frac{1+x^2}{2}\right) - 2\tan^{-1}x + \frac{\pi}{2} \tag{7}$$

where

$$x = (1 - \gamma\zeta)^{1/4} \tag{8}$$

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and γ is an empirical constant in the Monin–Obukhov equations and

$$\zeta = \frac{z-d}{L} \tag{9}$$

in which L is the stability length given by

$$L = \frac{-u^{*3}\rho}{kg\left[\left(\frac{Q_{\mathrm{H}}}{T_{\mathrm{a}}c_p}\right) + 0{\cdot}61T_{\mathrm{a}}Q_{\mathrm{E}}\right]} \tag{10}$$

where the sign of L indicates whether conditions are stable or unstable, and varies diurnally, ρ is the density of moist air, g is the acceleration due to gravity, and c_p is the specific heat for a constant pressure (Brutsaert, 1982; Garratt, 1992). For stable conditions, the profile function is given by

$$\psi(\zeta) = -\beta\zeta \tag{11}$$

where β is also an empirical constant from the Monin–Obukhov equations (Garratt, 1992).

RESULTS AND DISCUSSION

General climate and hydrology

The study site received approximately 412 mm of precipitation during the period of study in 2003, with the largest storm having occurred on 13 August delivering 37 mm. The average event received during the 2003 study season was approximately 1·3 mm. Although precipitation events in the area of Pond 43 were generally evenly distributed during the study season, larger magnitude events occurred more frequently from 19 June to 6 August (JD 170–218; Figure 2). Historically June–August represents 70% of the annual precipitation (Devito *et al*., 2005).

Mean daily air temperatures ranged between 20·9 °C (JD 213) and −13·9 °C (JD 304) during the study. Over the course of the season, the average daily air temperatures at Pond 43 were similar from June to August, decreasing after JD 210 (Figure 2). Near-surface air temperatures were similar over the pond and peatland surface, with values of $16{\cdot}2 \pm 6{\cdot}9$ °C and $12{\cdot}2 \pm 4{\cdot}7$ °C respectively. Further, the pond demonstrated a larger range in air temperature over the course of the season than did the peatland (24·5 °C and 15·7 °C respectively; Figure 2c). Wind conditions at the study pond were moderate during the 2003 study season, with a mean wind speed of 1·0 m s^{-1}.

Surface energy balance

The seasonal trend in net radiation Q^* over the open water surface was similar to that from the adjacent peatland (Figure 2a and b). Over both surfaces the bulk of Q^* was partitioned to Q_{E} early in the season, whereas Q_{H} over the peatland accounted for the largest portion of the surface turbulent fluxes as the season progressed and drier conditions prevailed (Figure 2a and b). Substrate heat flux comparisons involve the storage and transfer of heat energy through the peat surface in the peatland, and the heat stored in the water column of the pond in addition to the flux out of the bottom of the pond through the sediments and underlying substrate. Much larger fluctuations in Q_{G} were observed in the pond, which likely reflected greater variance in storage and transfer of energy within the pond's shallow water column than those measured in the peatland (Figure 2a and c). Precipitation events were associated with decreases in temperatures over both the peatland and pond water surfaces (Figure 2c and d).

Spatial variability in actual evapotranspiration

The ratio of PET to AET over the peatland differed considerably from that over the pond water surfaces. Cumulative PET exceeded AET in the peatland, but was less than AET in the pond (Figure 3a and b). Furthermore, an increase in AET was observed during drier periods (Figure 3a and b). PET was approximately 10% greater during the first 30 days, but ∼25% greater near the end of the study. PET is largely controlled by Q^* and Q_{G}; however, it is also quite sensitive to temperature. Early in the season the available energy is slightly larger over the pond (∼6% greater), but the air temperature just above the water surface (used in the calculations) is much larger over the peat (∼17%). Further, much of the energy early in the season is going into heat storage in the pond, but by later in the season the available energy is ∼14% greater over the peat, where the temperature is also approximately double that of the pond. Thus, the primary difference between the two sites is due in large part to the ground/water heat flux (storage). As the substrate heat flux increases, so the available energy decreases. However, as the season progresses, the heat storage in the pond is increasing, and is released slowly at the end of the season as opposed to the more rapid release of stored heat in the peat. Thus, there is a larger difference in available energy and PET between the two sites, with the peat being greater.

Relating the AET to PET over both surfaces resulted in Priestley–Taylor coefficients of 0·69 and 1·11 for the peatland and pond respectively. The α values are reflected in the pond water levels remaining relatively stable over a season, ranging from 37 to 57 cm in depth (Figure 3c). Although the water table depth within the riparian peatland responded rapidly and with greater amplitude than the pond to precipitation events (indicative of the peat's high specific yield), water table depth in the riparian peatland ranged from 50 to 80 cm below the surface (Figure 3c). Low α values for peat are not commonly reported in humid climates, but can be expected when water tables drop below the capillary fringe and rooting depths (Ingram, 1983; Price and Waddington, 2000). Further, α values reported here are similar to those found in other sheltered, and impacted, peatlands with lower water tables in eastern Canada (Van Seters and Price, 2001; Petrone *et al*., 2004). Such, water levels were observed during this study and can be

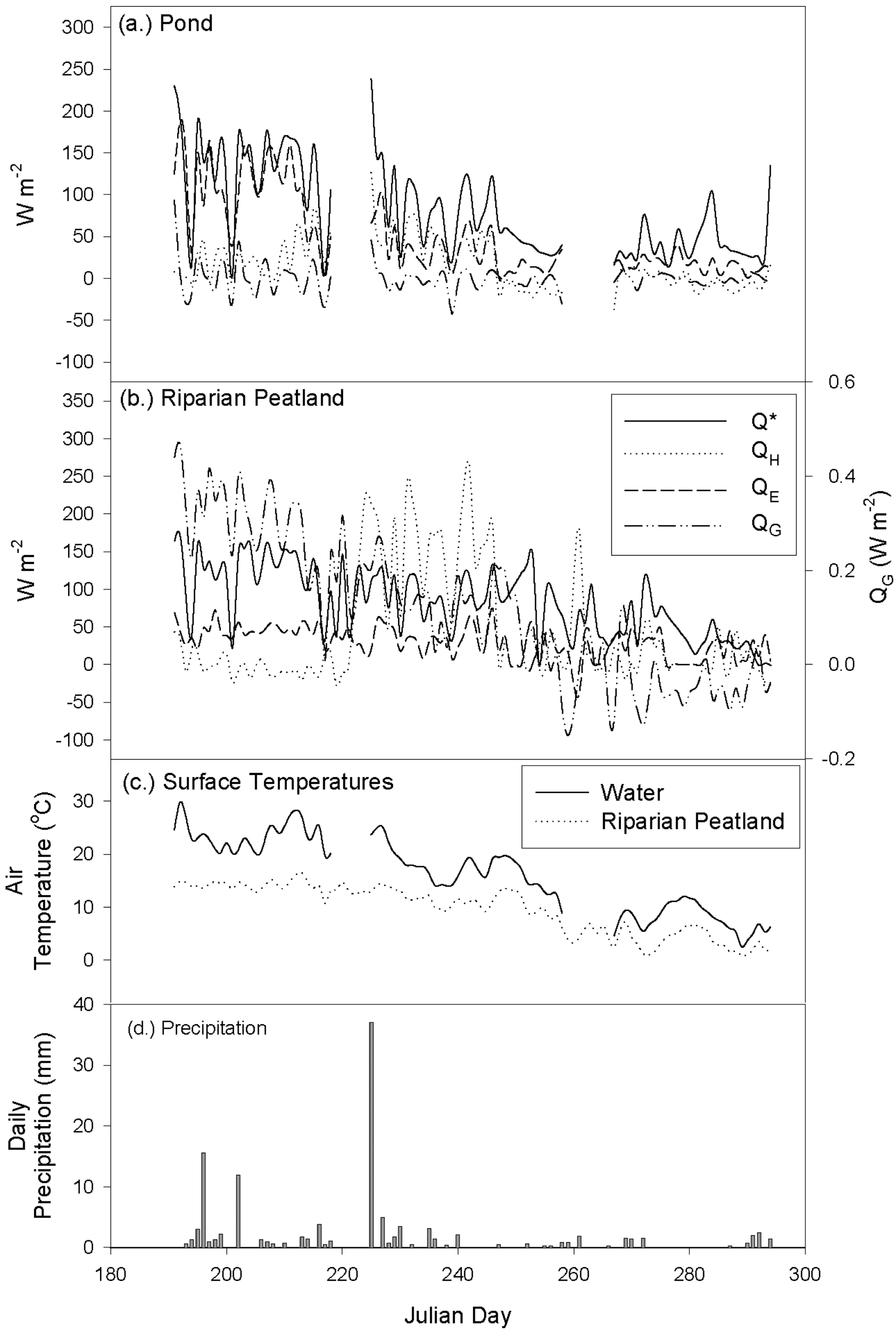

Figure 2. Energy balance of the (a) pond and (b) peatland surfaces at Pond 43, 2003; shown are the net radiation Q^*, the turbulent fluxes (sensible heat flux Q_H, latent heat flux Q_E), substrate heat flux Q_G. (c) Mean daily air temperatures (measured at the lowest levels at each Bowen ratio tower) for both the water and peat surfaces and (d) daily total precipitation P for the 2003 study period

expected frequently in the WBP where extended drought periods are common, and low α values may be indicative of the peatland's role in helping to conserve water and maintain pond levels.

Differences in the partitioning of the sensible and latent fluxes were reflected in variations in the micrometeorological conditions between the pond and riparian peatland sites (Figure 4). More frequent and pronounced lapse conditions were measured over the peatland compared with the more stable and near-neutral conditions evident over the pond during the study period (Figure 4a). In contrast, stronger lapse vapour pressure conditions were observed over the pond site. However, there were more near-neutral vapour pressure gradients over the pond than were evident over the peatland (Figure 4b). Thus, the observation of greater cumulative AET observed from the pond likely reflects differences not only in vapour pressure gradients, but also in mixing (mechanical turbulence)

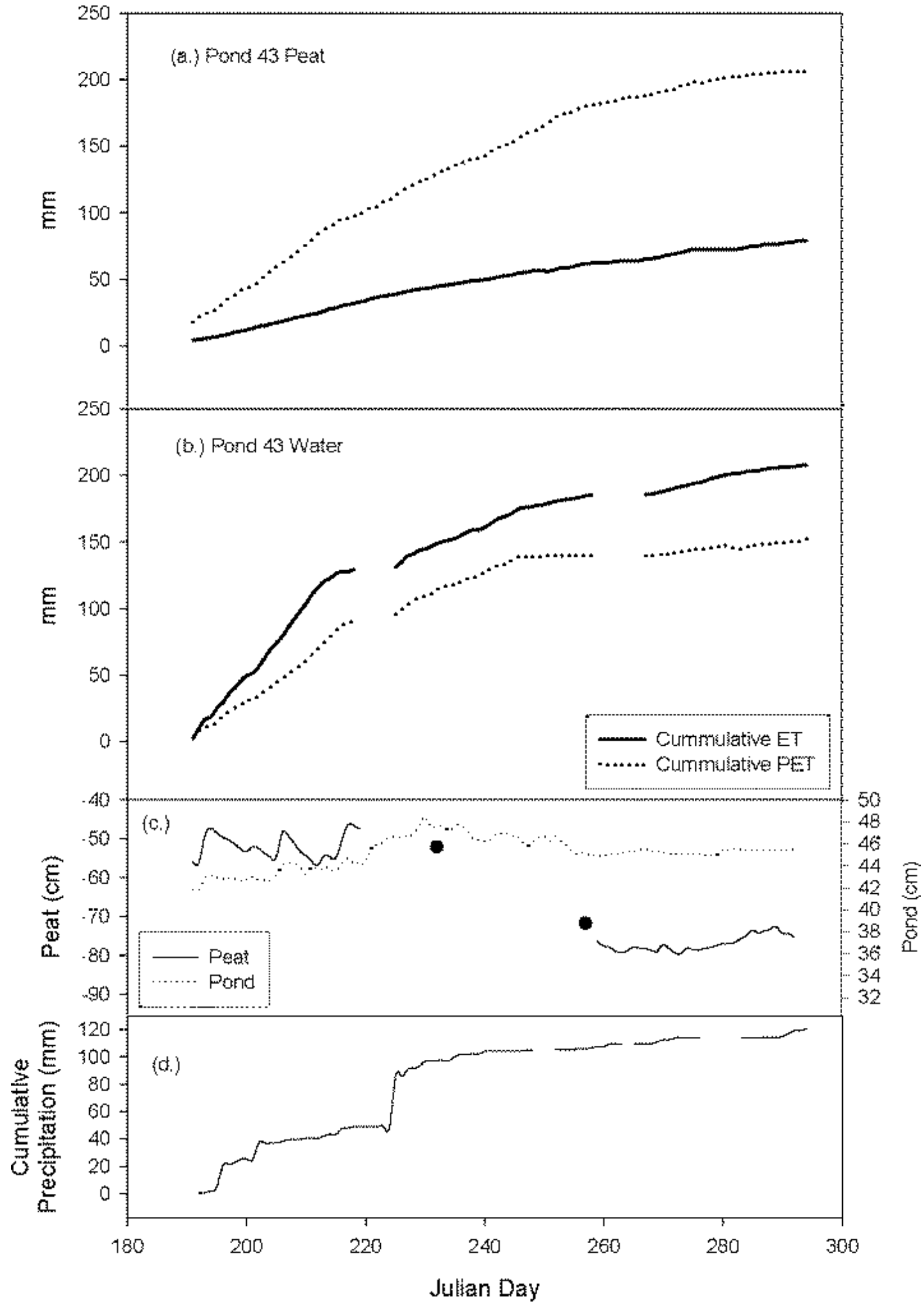

Figure 3. Cumulative PET and AET for the (a) pond and (b) peatland; (c) peatland water table depth (WT) and pond water level (WL) and (d) cumulative precipitation P for Pond 43, 2003. Solid circles (●) represent point measurements of water table depths on JD 232 and 257

over these two surfaces; this is discussed in the next section.

Micrometeorological controls on pond actual evapotranspiration temporal dynamics

To reconcile the correlation of less frequent lapse vapour pressure gradients over the pond with the larger AET observed there, the direction of wind flow within the catchment and how it may influence the stability conditions and, or, the amount of turbulent mixing over the two surfaces were examined using the frequency of wind directions and the Monin–Obukhov stability length over the pond surface.

Over Pond 43, greater instability was observed when air flow was from the north to southeast, and from the west, whereas more stable conditions were observed over the pond when the flow was from the south-southeast to the west-southwest (Figure 5). The frequencies of wind directions were 52% and 36% of the time for the unstable and stable conditions respectively. Thus, potential decreases in evaporation associated with weaker vapour pressure gradients were often offset by along-wind gradients in wind stress, even increasing evaporation rapidly until the wind equilibrated with the local z_o (Condie and Webster, 1997). That is, increasing fetch can increase turbulent mixing at the pond surface and, therefore, Q_E despite more humid conditions. This supports the notion of a fine balance between the competing effects of increasing humidity, surface temperature

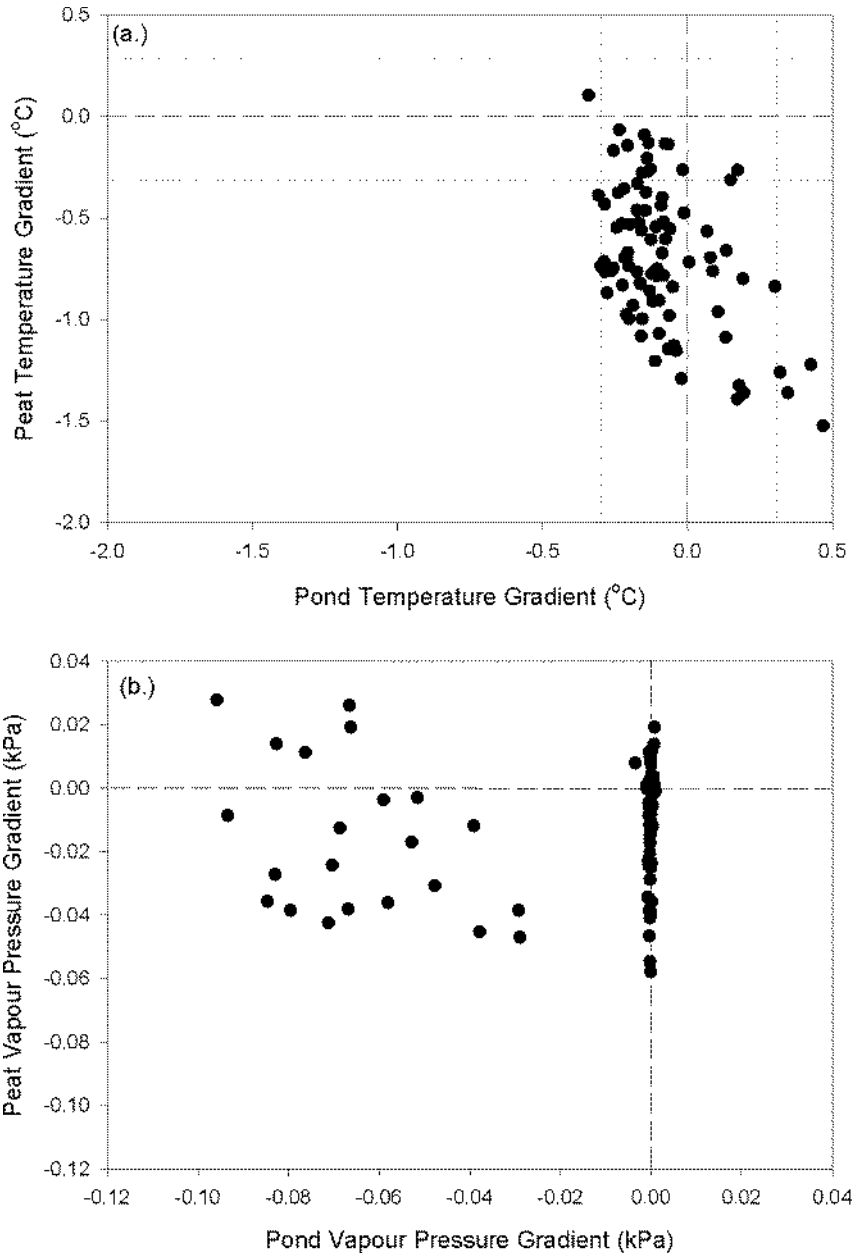

Figure 4. (a) Daytime temperature dT/dz and (b) vapour pressure de/dz gradients over the pond and peatland surfaces at Pond 43, 2003. Gradients were calculated based on surface and 3 m height measurements. Dotted lines in (a) indicate predefined limits for transitions from neutral conditions

and wind stress with downwind distance. If the adjacent land is very smooth ($z_o \approx 10^{-4}$ m) then evaporation will decrease with fetch, whereas with very rough surfaces ($z_o \approx 10^{-1}$ m), such as in this case, evaporation will increase with increasing fetch (Weisman and Brutsaert, 1973).

For Pond 43, when air flow was from the east-southeast and southeast, enough fetch existed that vapour gradients equilibrated. Furthermore, flow over Pond 43 from this direction also passes over an adjacent pond (Pond 48). Thus, under these conditions the much weaker gradients exert a stronger influence than the potentially higher wind velocities due to the longer fetch, and a decrease in the latent heat flux over Pond 43 was observed (Figure 6). Thus, stability conditions within variable landscapes such as this may be strongly driven by the variation

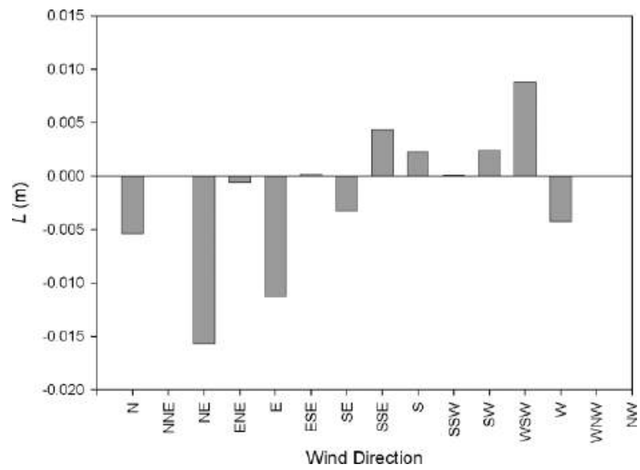

Figure 5. Monin–Obukhov stability length L over the water surface of Pond 43, 2003. Data are grouped for wind direction in the catchment

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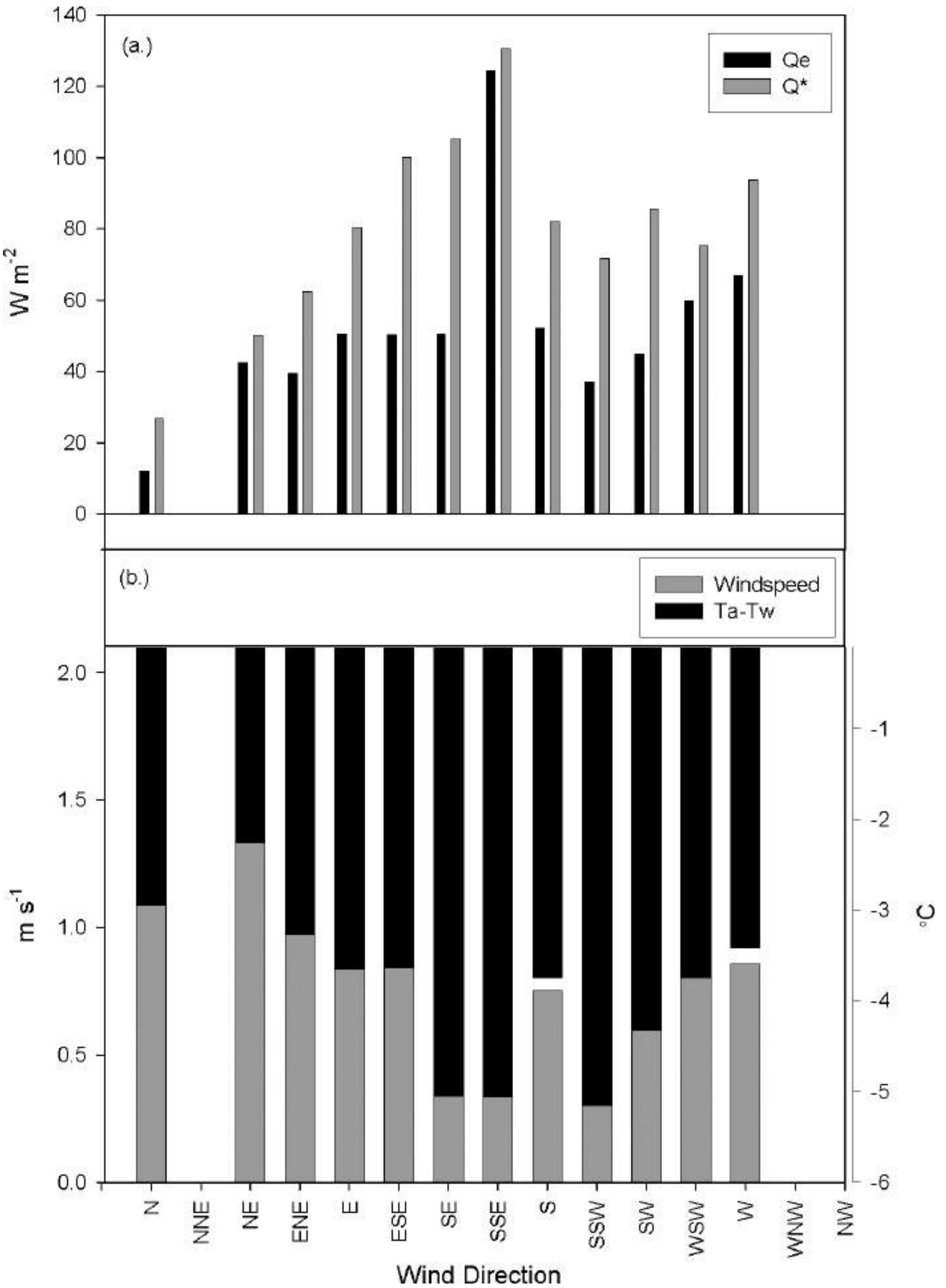

Figure 6. (a) Average latent heat flux Q_E and net radiation Q^* and (b) wind speed u and vertical temperature gradient dT/dz grouped for wind direction over the water surface at Pond 43, 2003. The vertical temperature gradient is defined as the difference between air temperature measured at 1·5 m above the pond surface and the temperature of the water surface ($T_a - T_w$)

in the sequence of surfaces over which air masses move. For example, more lapse conditions associated with greater sensible heat flux were observed over the peatland. In contrast, the pond was frequently under neutral conditions, suggesting the greater importance of turbulent transport in the pond's larger AET, and not gradients in vapour pressure. These observations strongly suggest that the influence of the edge effect of the surrounding forested areas on the amount of evaporation occurring from the open water surface of a pond should be considered.

The effect of direction of air mass movement on the dynamics of pond vapour fluxes was also evidenced by an increase in latent heat flux from the pond when the flow was from the south-southeast (Figure 6). An increase in the vertical temperature gradient and a decrease in wind speed was observed when the winds were from the south-southeast, and was associated with the maximum Q_E (Figure 6). However, there was also second peak in the latent heat flux from the pond when more fetch was present over the pond concurrent with less from the surrounding hillslopes (west-southwest and westerly flows). These trends are evidence of a sheltering effect of the surrounding forested uplands, which is a function of the lake radius or fetch. We observed a decrease in sheltering when the flow was from these directions and, therefore, an increase in Q_E, presumably resulting from greater fetch and turbulent mixing.

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Harvest implications and relation to other studies

Much of the previous research on forest harvesting effects on adjacent surface water levels has focused on a two-pronged effect that acts to increase pond, or stream, water levels through: (1) the removal of the 'transpirational pump', causing a (temporary) increase in the hillslope water table; and (2) increased soil evaporation, which will partially offset (1). Therefore, this lack of a canopy allows more water (from precipitation, etc.) to reach the soil, which becomes more saturated, so that further precipitation increases the duration and amount of surface and groundwater flow (Heatherington, 1987; Satterlund and Adams, 1992; Pike and Scherer, 2003). Studying the effects of logging on adjacent surface water levels in a humid system in northern California, Keppler and Ziemer (1990) observed increased levels after harvesting due to the removal of transpiring vegetation. However, this may not be the case in areas where soils have higher infiltration rates and/or receive much less precipitation, i.e. in the WBP (Devito *et al*., 2005).

Other studies in similar systems have shown that changes in vegetation, especially in the riparian zone, once regeneration has begun can act to decrease adjacent pond, or stream, levels (Hicks *et al*., 1991; Pike and Scherer, 2003). This is especially true in conifer-dominated catchments, common in much of the Boreal zone, where deciduous trees may succeed in riparian areas following harvesting (Hicks *et al*., 1991). These deciduous trees generally use more water than the conifers they have replaced for similar leaf areas, thereby removing water available to the adjacent ponds and streams (Berndt, 1971; Hicks *et al*., 1991).

Swanson and Rothwell (2001), working in an aspen-dominated catchment in west-central Alberta, observed only an average 16 mm (4%) decrease in evapotranspiration after cutting. Assuming that this increase in available moisture translates to a 4% increase in runoff to the pond, the rise in pond water level would be much smaller than the more than 100% increase in pond evaporation due to the loss of sheltering. This lower amount of available moisture from the harvested hillslope in an aspen-dominated catchment may also be due to the increased root uptake as the suckering of the regenerating aspen increases, mitigating transpirational decreases (Berndt, 1971; Hicks *et al*., 1991). Thus, the impacts of the forest cover on the turbulent regime of the pond will be much more significant to the catchment water balance of a sub-humid aspen-dominated catchment.

CONCLUSIONS

Knowledge of the processes occurring within, and the linkages between, the mosaic of terrain types in the WBP is required for the successful scaling of hydrologic process such as AET in order to assess the sensitivity of the WBP water balance to land-use and climatic change. However, previous studies (e.g. Metcalfe and Buttle, 1999) have suggested that good estimates of AET can be difficult in boreal landscapes because of the mosaic of wetlands, upland and areas of depression storage; as such, it is essential first to look at the relative contributions of the pond water body and the adjacent riparian peatland. Results from this study emphasize the importance of this partitioning of the catchment AET, but also the influence of the adjacent terrain and upland areas in the evaporative regime of the pond water body. If the sheltering and turbulent influences of the adjacent forested areas are not considered in the processes governing the permanence of these WBP ponds, then forestry practices may inadvertently enhance the evaporative losses from the ponds, over and above the controls exerted by the regional climate.

Furthermore, these findings provide insights as to how gross or synoptic-scale atmospheric conditions may fail to provide enough resolution to understand the spatial variation of evapotranspiration in a landscape as complex as the WBP.

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