Evapotranspiration in intermediate-aged and mature fens and upland black spruce boreal forests

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ABSTRACT

The Canadian boreal forest consists of a mosaic of landscapes of varying soil drainage and forest age driven by wildfire. The hydrological consequences are complicated by plant responses to soil moisture and forest age, both potentially influencing evapotranspiration. Evapotranspiration was measured using the energy balance residual technique in 2006 and 2007 at forested upland and fen sites that originated following fire in 1964, 1930 and about 1850, near Thompson, Manitoba, Canada. Both net radiation and sensible heat flux density were greater at the older sites than those at the younger sites. Evapotranspiration was also greater at the older sites by between 4 and 19% for the 1930–1964 comparison, and 15% for the 1930–1850 comparison. There was no difference in net radiation between upland and fen sites of the same age, although upland sites had a higher sensible heat flux density. Albedo was greater at the fen sites. Evapotranspiration was greater at the upland sites by 11–20%, likely driven by greater leaf area at the upland sites. These intermediate to mature boreal forest sites still show the persistence of the impact of fire, and it is clear that changes in drainage and local hydrology will also have an impact on local evapotranspiration. The implication is that even these small changes in evapotranspiration can have a great regional and global effect because of the large land area of the boreal forest. Copyright © 2009 John Wiley & Sons, Ltd.

KEY WORDS boreal forest; energy balance; eddy covariance; fire; drainage; wetlands

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INTRODUCTION

The circumpolar boreal forest has a large impact on global cycling of carbon and water (Dixon et al., 1994; Chapin et al., 2000). Within this large area, the forest is continuously being renewed by fire, insects, disease and human disturbance, such as harvesting. This creates a multi-aged forest mosaic that is determined by the current and past disturbance regime. Fire is one of the main disturbances in the North American and Siberian boreal forest, and any shift in the frequency or extent in area burned will change the age structure of the forest. For example, Flannigan et al. (2005) estimated that Canadian forests could experience a doubling of area burned with a warming climate, expected to occur near the end of the current century. These potential shifts in age structure could cause changes to the water balance of the forest if younger forests have different hydrological characteristics than older forests. The boreal forest also has topographic variation, resulting in differences in vegetation, soil moisture content and drainage. Across this landscape, large fires usually burn both the lowland and upland communities during the same event (Turetsky et al., 2004), creating a mosaic of various ages.

The water balance of these boreal ecosystems is likely determined by forest age, topographic position and soil drainage for a given climate regime. In particular, upland, relatively well-drained forests contrast with low-lying areas dominated by fens. However, it is not clear if evapotranspiration (ET) differs among these areas. For example, Mackay et al. (2007) found that the drivers of ET were fundamentally different within and across seasons between wetland and upland forests in northern Wisconsin. Current models of boreal ET predict smaller fluxes from poorly drained forests than well-drained ones; however, these results have not been tested against ET data from poorly drained sites (Bond-Lamberty et al., 2009). A comparison of previous studies of boreal forest ET showed that broad-leaved deciduous forests generally have greater ET than coniferous evergreen forests (Baldocchi et al., 2000). In addition, ET decreased immediately following a fire, increasing to a maximum at about 20–25 years of age, then decreasing and remaining approximately constant for many years (Amiro et al., 2006a). This general relationship was based on comparisons among geographically widespread forests from Alaska, USA, to Manitoba, Canada, and exhibited quite a bit of variability among sites.

The area near Thompson, Manitoba, Canada, has been one of the focal points for northern boreal forest research since the BOREAS experiment in the early 1990s (Sellers et al., 1997). Since the intensive campaigns during...
the BOREAS period, this area has supported long-term measurements of the carbon exchange of boreal forests (e.g. Dunn et al., 2007). There have also been extensive studies of fire chronosequences, including carbon exchange (Litvak et al., 2003; Goulden et al., 2006), forest inventory and dynamics (Bond-Lamberty et al., 2002, 2004, 2006; Wang et al., 2002, 2003), respiration (Czimczik et al., 2006), and tree transpiration (Ewers et al., 2005). In particular, the development of knowledge on the dynamics of fire chronosequences has been central to our understanding of boreal forest processes, especially tied to potential changes that could occur with a warming climate. In the present study, we investigated the effects of both forest stand age and soil drainage on ET using the chronosequences near Thompson. We focused on intermediate and mature stands (40–150 years) because previous studies have shown that very young stands are clearly different (Chambers and Chapin, 2002; Liu et al., 2005; Amiro et al., 2006a,b). We compared ET for forested fens and upland forests of the same age, as these are the basic forest types resulting from topographic variations and differences in soil drainage. Although these landscape types are very common, they do not cover the full breadth of boreal ecosystems.

METHODS

Site description

The chronosequence is located in the northern boreal forest near Thompson, Manitoba (Table I). The ages were chosen due to their similar topography (glaciated landscapes with local elevation changes of less than 10 m) and soil type (clay and silty clay, Cyr, 2005), and close proximity to each other (~15 km). Additional characteristics of this chronosequence have been reported by Bond-Lamberty et al. (2004). The sites also represented the key changes in canopy transpiration controls from changing species composition with succession and hydraulic changes due to tree age (Ewers et al., 2005). The oldest site, known as Northern Old Black Spruce, was a previously established research site that was included in BOREAS. This site was burned in approximately 1850, and the main tower has been in operation since 1994 (Dunn et al., 2007). In 2006, the forest was composed of black spruce (Picea mariana) with moss ground-cover. The dominant moss types in the upland areas were feather mosses (Pleurozium schreberi and Hylcomium splendens), with sphagnum mosses (Sphagnum spp.) in the lowland areas. The understory included Labrador tea (Ledum groenlandicum), wild rose (Rosa spp.) and green alder (Alnus crispa) (Bond-Lamberty et al., 2002). Upland forests dominated the site, although there were wetland areas within the site flux footprint.

The second age of the chronosequence was burned in 1930 (Table I). In 2006, the forested fen site was composed of stunted black spruce with a groundcover of mosses. The vegetation at the upland areas were feather mosses (Pleurozium schreberi and Hylcomium splendens), with sphagnum mosses (Sphagnum spp.) in the lowland areas. The understory included Labrador tea (Ledum groenlandicum), wild rose (Rosa spp.) and green alder (Alnus crispa) (Bond-Lamberty et al., 2002). Upland forests dominated the site, although there were wetland areas within the site flux footprint.

Table I. Site characteristics of five northern boreal forest sites located near Thompson, Manitoba, Canada.

<table>
<thead>
<tr>
<th>Site</th>
<th>1964 Fen</th>
<th>1964 Upland</th>
<th>1930 Fen</th>
<th>1930 Upland</th>
<th>1850 Upland</th>
</tr>
</thead>
<tbody>
<tr>
<td>Location</td>
<td>N55.91369°</td>
<td>N55.92033°</td>
<td>N55.90442°</td>
<td>N55.90795°</td>
<td>N55.87962°</td>
</tr>
<tr>
<td>W98.38176°</td>
<td>W98.38986°</td>
<td>W98.52043°</td>
<td>W98.38209°</td>
<td>W98.48081°</td>
<td></td>
</tr>
<tr>
<td>Year burned</td>
<td>1964</td>
<td>1964</td>
<td>1930</td>
<td>1930</td>
<td>~1850</td>
</tr>
<tr>
<td>Measurement height (m)</td>
<td>4-5</td>
<td>9</td>
<td>4.5</td>
<td>12</td>
<td>30</td>
</tr>
<tr>
<td>Soil classification</td>
<td>Orthic grey luvisol</td>
<td>Orthic grey luvisol</td>
<td>Orthic grey luvisol</td>
<td>Orthic grey luvisol</td>
<td>Orthic grey luvisol</td>
</tr>
<tr>
<td>Tree density (trees ha⁻¹)</td>
<td>3537</td>
<td>7215</td>
<td>4951</td>
<td>6225</td>
<td>5164</td>
</tr>
<tr>
<td>Black spruce</td>
<td>2.67</td>
<td>3.34</td>
<td>4.09</td>
<td>8.74</td>
<td>9.34</td>
</tr>
<tr>
<td>Aspen</td>
<td>0</td>
<td>2546</td>
<td>0</td>
<td>354</td>
<td>0</td>
</tr>
<tr>
<td>Jack pine</td>
<td>0</td>
<td>3183</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Birch</td>
<td>0</td>
<td>424</td>
<td>0</td>
<td>71</td>
<td>0</td>
</tr>
<tr>
<td>Total</td>
<td>3537</td>
<td>13 793</td>
<td>4951</td>
<td>6649</td>
<td>5164</td>
</tr>
<tr>
<td>Diameter at breast height (cm)</td>
<td>2.67</td>
<td>3.89</td>
<td>4.09</td>
<td>8.69</td>
<td>9.34</td>
</tr>
<tr>
<td>Basal area (m² ha⁻¹)</td>
<td>2.3</td>
<td>10.8</td>
<td>7.4</td>
<td>42.8</td>
<td>38.3</td>
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<tr>
<td>Black spruce</td>
<td>0</td>
<td>3.2</td>
<td>0</td>
<td>1.9</td>
<td>0</td>
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<tr>
<td>Aspen</td>
<td>0</td>
<td>5.8</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Jack pine</td>
<td>0</td>
<td>0.2</td>
<td>0</td>
<td>2.0</td>
<td>0</td>
</tr>
<tr>
<td>Birch</td>
<td>2.3</td>
<td>20.0</td>
<td>7.4</td>
<td>45.7</td>
<td>38.3</td>
</tr>
<tr>
<td>Total</td>
<td>0.4</td>
<td>2</td>
<td>3</td>
<td>7</td>
<td>5.5</td>
</tr>
</tbody>
</table>

The youngest age included in this chronosequence was burned in 1964 (Table I). The fen had black spruce trees, with a sphagnum moss groundcover. The upland site had a canopy of black spruce, jack pine (Pinus banksiana), trembling aspen and a small number of birch, with a groundcover of feather mosses (Bond-Lamberty et al., 2002).

**Calculation of evapotranspiration (ET)**

We determined ET through the energy balance residual technique (e.g. Amiro and Wuschke, 1987; Adams et al., 1991; Blanford and Gay, 1992; Mizutani et al., 1997; Gash et al., 1999; van der Tol et al., 2003; Loescher et al., 2005). This method calculates the latent energy flux density ($\lambda E$) as the difference between net radiation ($R_n$), ground heat flux density ($G$) and sensible heat flux density ($H$):

$$\lambda E = R_n - G - H$$

In Equation (1), we assume that storage is small, especially at time scales of more than 1 day. Following Amiro (2009), we set $\lambda E$ to zero at night, due to low turbulence and other complicating factors that occur at night, such as the accumulation of dew on the net radiometers. We recognize that $\lambda E$ is often finite at night in many ecosystems (Dawson et al., 2007; Fisher et al., 2007), especially when the vapour pressure deficit (VPD) is large. We evaluate this potential under-estimation in a subsequent section of the paper. Following Equation (1), we convert $\lambda E$ (energy units) to ET (water units) by dividing by the latent heat of vaporization (set at a constant 2450 J g$^{-1}$).

Differences between the energy balance residual method and the direct measurement of ET by eddy covariance were less than 5% when summed over the growing season for several boreal forest sites in Saskatchewan (Amiro, 2009). This relationship was especially close during the middle of the growing season period, although ET may be overestimated during the early spring due to energy stored by the ecosystem during snow melt. In the present study, our measurements started in mid-May at the earliest, and did not include the period of snowmelt. We recognize the strength of direct ET measurements using eddy covariance, constrained by energy balance, as the best measurement method. However, at remote sites with limited power, we have difficulty operating fast-response humidity sensors needed for the direct eddy covariance ET measurement. Coupled with the additional cost of these humidity sensors, the energy balance residual technique is a good method to investigate multiple sites in remote locations.

**Field measurements**

Instruments were mounted above the tree canopy on towers. The 1930 site used scaffold-type towers for both the upland and fen sites, which measured at approximately 12 and 4-5 m in height respectively. The 1964 fen site also used a scaffold-type tower, which was 4-5 m in height. The upland tower for 1964 was a 9-m triangular tower, while the 1850 site had a 30-m triangular tower.

At each of the towers, three-dimensional wind velocities and air temperature were measured at the top of the towers with a CSAT3 sonic anemometer--thermometer (Campbell Scientific, Edmonton, AB, Canada and Logan, UT, USA) at most sites, and a K-Style anemometer--thermometer at the 1850 site (Applied Technologies Inc., Longmont, CO, USA). The CSAT3s were measured at 10 Hz and the flux cross products were saved every 30 min. Data at the 1850 site were measured at 4 Hz. Virtual sensible heat flux was calculated using the eddy covariance method and adjusted for vapour pressure to calculate $H$. Turbulent flux data were coordinate-rotated following Tanner and Thurtell (1969).

Net radiation was measured above the canopy using a NR-lite net radiometer (Kipp and Zonen, Delft, The Netherlands), except at the 1850 site, where a Q7 net radiometer (REBS, Seattle, WA, USA) was used. Ground heat flux was also measured at the 1964 and 1930 sites using HFT3 heat flux transducers (Campbell Scientific). Four soil heat flux sensors were installed horizontally 2 cm below the active moss and litter layers, within the non-living organic layer at all sites. The results from the four sensors were averaged at each site. We did not measure energy storage in the shallow layer above the heat flux plates. Lafleur et al. (1997) have shown that storage in 10 cm of peat soil is typically a factor of 3 greater than the flux measured at the 10 cm depth, averaged over periods of 12 days or less. Storage in our 2 cm depth would be much less than this, especially when averaged over several days.

Supporting meteorological measurements included air temperature and relative humidity using HMP45C probes (Campbell Scientific) and a tipping-bucket-type precipitation gauge (TE525M, Texas Electronics, Inc., Dallas, Texas, USA). Soil temperature was measured at each site using four replicate 24-gauge copper-constantan thermocouples at 5 cm depth, as well as a single profile at 5, 10, 30 and 50 cm depths below the live moss layer. Theta probe soil moisture sensors (ML2x, Delta-T Devices, Ltd., Cambridge, UK) were used to measure soil moisture levels at the peat surface, as well as at 15 and 30 cm depths. Incoming and outgoing photosynthetically active radiation (PAR) was measured above the canopy at each site using LI-190 PAR sensors (LICOR Inc., Lincoln, NE, USA).

The data from the instruments were collected by data loggers (CR1000 or 10X, Campbell Scientific, Canada) and data were manually downloaded on a regular basis during the growing season. Data loggers were powered by 12-V batteries charged by solar panels. Data at the 1850 site were collected on a desktop computer because AC power was available. The high frequency data were recorded and then averaged by the data loggers or computer over 30-min periods. Data were collected from June 16 to October 31, 2006, and May 14 to October 15, 2007.
Data quality control and gap filling

The CSAT3 data were excluded when more than one sample was missed in a 30-min period. Turbulent flux terms are often underestimated during low wind conditions at night (Goulden et al., 1996). Hence, measurements of $H$ were not used below the friction velocity ($u_*$) cut-off of 0.25 m s$^{-1}$ because of this poor turbulent mixing. The selection of a $u_*$ threshold of 0.25 m s$^{-1}$ was based on our previous experience with fire-disturbed forests at other locations (Amiro et al., 2006b). Gap-filling of data followed the standard protocol of Fluxnet-Canada (Amiro et al., 2006b; Barr et al., 2006). Data gaps of less than 2 h were filled using linear interpolation. To fill gaps in sensible heat flux that were longer than 2 h, a 240-point moving window was used, with a linear regression between $H$ and $R_n - G$, and moved in 48-point increments. The data from the net radiometer for the 1964 upland site had a large gap from June 29 to July 19, 2007, which was filled using a regression relationship ($R^2 = 0.97$) between half-hourly $R_n$ data at the 1964 upland and fen sites. Note that ET at night was set to zero, so that the night-time $H$ measurements were only used for site comparisons of $H$, irrespective of ET.

Uncertainties

Our uncertainty is a combination of components that contribute to the final estimate of ET. Both the precision and accuracy of the measurements of $R_n$, $H$ and $G$ need to be evaluated. The uncertainty in the measurement of $R_n$ was estimated by placing a four-component net radiometer (CNR1, Kipp and Zonen, Delft, Netherlands) at each of the 1964 and 1930 sites from mid-July to mid-September in 2007. Similarly, we calculated $H$ using a second CSAT3 sonic anemometer–thermometer at these sites. In addition to the instrument comparisons, we further evaluated whether the energy residual technique and a direct eddy covariance measurement of ET would give the same conclusions. This was done by measuring ET directly using a LI7500 open-path infrared gas analyser (LiCor Inc., Lincoln, NE, USA) mounted at each of the 1964 sites from June 19 to July 24, 2008. ET fluxes were calculated for the same 30-min periods and were quality controlled for water on the sensors (using the AGC signal from the instrument) and for turbulence by only keeping night-time values during conditions when $u_* > 0.25$ m s$^{-1}$. We calculated daily-total ET, filling day-time gaps using our standard protocol. ET values during night-time gaps were set to zero.

Statistical analysis

We address whether differences are real through two methods. The first method was to estimate the uncertainty in instrument measurement to evaluate whether differences are outside of these uncertainties. Secondly, we compared the energy balance components and the estimate of ET using regression analyses between the sites plus an analysis of covariance to determine if the slopes of the regressions were different from unity (MATLAB, The MathWorks Inc., Natick, MA, USA, version 7). The reference dataset to test for a slope of unity was generated through a second dataset that was perfectly correlated, i.e. an identical dataset to that used on the abscissa. In all the cases, our comparisons used gap-filled daily values, e.g. ET in millimetres per day.

RESULTS

Climate conditions

Both 2006 and 2007 had slightly more growing-season precipitation and were slightly warmer than average. Data from Thompson airport, about 75 km away from the sites, recorded 406 mm of precipitation for the May-to-October period in each year, compared to the 30-year normal of 378 mm (Environment Canada, 2006). Mean air temperatures for the same period were 10.2°C (2006) and 9.7°C (2007) compared to the normal of 9.4°C. Soil temperature measurements at the sites showed thawing at the 5 cm depth before May 10, at the 30 cm depth by June 1, and at the 50 cm depth by July 1 at all upland sites. Typically, the fen sites thaw more slowly than the upland sites by about 10 days, and remain colder at the 50 cm depth by almost 3°C throughout the summer. Near-surface temperatures are more similar with fen sites being about 1°C colder than upland sites of the same age.

Instrument uncertainty

The comparisons of net radiation instruments at the 1930 and 1964 sites showed consistently higher readings from the CNR1 four-component radiometer compared to the NR-lite net radiometer during the day and lower (more negative) at night. We selected a subset of 30-min data when the relative humidity was <90% and wind speed at the sensor height was between 1 and 5 m s$^{-1}$ to evaluate conditions when moisture on the sensors should not be an issue. There is a known wind response of the NR-lite radiometer at wind speeds >5 m s$^{-1}$, so we avoided these conditions. For the population of four site comparisons, the regression was CNR1 = 1.128* NR-lite + 13.0 with units in W m$^{-2}$ ($R^2 = 0.97$, $n = 3278$). This gave a mean ratio of CNR1:NR-lite of 1.22 over the period. Selecting times between 1000 and 1400 h local time when the sun was closer to zenith improved the slope of the regression, differing by only 2% at a given site, but we still observed an offset in the regression of about 35 W m$^{-2}$. Note that all four of the NR-lite radiometers were new instruments with factory calibrations so they should have been in good agreement when installed.

Comparison of $H$ measurements with a second set of CSAT3 sonic anemometer–thermometers showed very close agreement differing by less than 3% at all sites, and by only 1% at some sites. In all the cases, for both $R_n$ and $H$, scatter was low, with $R^2$ values being 0.95–0.98.
We used the 2008 data set to compare ET measured directly using eddy covariance with the fast-response infrared gas analyser against the energy balance residual technique. We only had sufficient power to make continuous measurements through the night at the upland site for about 10 days, but had 33 days for the fen site. At the upland site, the energy balance residual technique gave a mean ET of 2.71 mm day\(^{-1}\) compared to 2.44 mm day\(^{-1}\) for the direct eddy covariance technique. The lack of closure of the energy balance at many sites by about 20% (Wilson et al., 2002) has resulted in many authors adjusting the turbulent flux measurements to close the balance (e.g. Twine et al., 2000). If we increase the eddy covariance measurement of ET by 20%, the 10-day mean would be 2.93 mm day\(^{-1}\). At the fen site, the 33-day mean ET was 2.20 mm day\(^{-1}\) from the energy balance residual and 1.76 mm day\(^{-1}\) from direct eddy covariance. A 20% adjustment would increase the eddy covariance estimate to 2.11 mm day\(^{-1}\). Therefore, the fractional difference between the energy balance residual and the adjusted eddy covariance technique is \(-8\%\) for the upland site and \(+4\%\) at the fen site. Using a lesser closure factor of 15% would decrease the upland difference to \(-3\%\) and increase the fen site difference to \(+9\%\). Note that we could also increase H by a percentage for energy balance closure in the residual technique, which would decrease the ET estimate.

Data capture of \(R_n\) and H was typically 97% at each site. Hence, only about 3% of the data were gap filled, which did not add substantially to the uncertainty.

**The effect of stand age**

\(R_n\), H and ET were compared between stands of different ages with similar soil drainage and topographical location. Over the growing seasons of 2006 and 2007, four experimental comparisons were possible for the age effect: 1930 fen versus 1964 fen in both 2006 and 2007, 1850 upland versus 1930 upland in 2006 and 1930 upland versus 1964 upland in 2007. All four age-effect experiments showed lower daily ET at younger sites than older sites, typically by 8–17% (Tables II, III). Regressions comparing pairs of sites of different ages had slopes significantly different from a 1 : 1 slope \((P < 0.001)\). Note that there is no true independent variable in these comparisons and the selection of the site for the abscissa is arbitrary. In all the cases, \(R^2\) values were high with the largest difference and lowest \(R^2\) was in the comparisons with the 1850 upland site. Note that, in Table III, \(R_n\), G and H are averages that include night-time data. \(\lambda E\) is a calculated value that is set to zero at night. Owing to this difference, the energy balance components in Table III do not exactly sum to \(R_n\) and can differ by up to 4%.

Daily H values were generally larger during the growing season at older sites than at younger sites (Table II). The smallest difference in H between sites of different ages was in 2007 between the 1964 and 1930 upland sites, where H was not significantly different between the two sites \((P = 0.24)\). The largest difference in H was between the 1850 and 1930 upland sites, where the 1930 site was 33% less. All other comparisons of H were significantly different \((P < 0.001)\).

### Table II. Regression coefficients for intersite comparisons as daily averages for \(R_n\) (W m\(^{-2}\)) and H (W m\(^{-2}\)) and daily totals for ET (mm day\(^{-1}\)).

<table>
<thead>
<tr>
<th>Component</th>
<th>x</th>
<th>y</th>
<th>Slope</th>
<th>Intercept</th>
<th>(R^2)</th>
<th>RMSE</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Age effect</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(R_n), 2007</td>
<td>1930 Upland</td>
<td>1964 Upland</td>
<td>0.93*</td>
<td>0.4</td>
<td>0.96</td>
<td>10.6</td>
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<tr>
<td>(R_n), 2006</td>
<td>1850 Upland</td>
<td>1930 Upland</td>
<td>0.83*</td>
<td>-2.3</td>
<td>0.89</td>
<td>19.7</td>
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<tr>
<td>(R_n), 2006</td>
<td>1930 Fen</td>
<td>1964 Fen</td>
<td>0.92*</td>
<td>0.5</td>
<td>0.97</td>
<td>9.2</td>
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<tr>
<td>(R_n), 2006</td>
<td>1930 Fen</td>
<td>1964 Fen</td>
<td>0.92*</td>
<td>0.3</td>
<td>0.99</td>
<td>6.8</td>
</tr>
<tr>
<td>H, 2007</td>
<td>1930 Upland</td>
<td>1964 Upland</td>
<td>0.96</td>
<td>5.2</td>
<td>0.82</td>
<td>15.4</td>
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<td>H, 2006</td>
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<td>0.66*</td>
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<td>1964 Fen</td>
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<td>0.94</td>
<td>6.4</td>
</tr>
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<td>H, 2007</td>
<td>1930 Fen</td>
<td>1964 Fen</td>
<td>0.90*</td>
<td>1.0</td>
<td>0.95</td>
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</tr>
<tr>
<td>ET, 2007</td>
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<td>0.05</td>
<td>0.82</td>
<td>0.4</td>
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<td>1930 Upland</td>
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<td>0.07</td>
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<td>ET, 2007</td>
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<td>1964 Fen</td>
<td>0.96*</td>
<td>-0.11</td>
<td>0.94</td>
<td>0.2</td>
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### Topographic effect

<table>
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<tr>
<th>Component</th>
<th>x</th>
<th>y</th>
<th>Slope</th>
<th>Intercept</th>
<th>(R^2)</th>
<th>RMSE</th>
</tr>
</thead>
<tbody>
<tr>
<td>(R_n), 2006</td>
<td>1930 Upland</td>
<td>1930 Fen</td>
<td>1.00</td>
<td>2.0</td>
<td>0.98</td>
<td>9.2</td>
</tr>
<tr>
<td>(R_n), 2007</td>
<td>1930 Upland</td>
<td>1930 Fen</td>
<td>0.98</td>
<td>0.6</td>
<td>0.99</td>
<td>7.1</td>
</tr>
<tr>
<td>(R_n), 2007</td>
<td>1964 Upland</td>
<td>1964 Fen</td>
<td>0.96*</td>
<td>0.9</td>
<td>0.97</td>
<td>9.2</td>
</tr>
<tr>
<td>H, 2006</td>
<td>1930 Upland</td>
<td>1930 Fen</td>
<td>0.88*</td>
<td>8.4</td>
<td>0.90</td>
<td>8.8</td>
</tr>
<tr>
<td>H, 2007</td>
<td>1930 Upland</td>
<td>1930 Fen</td>
<td>0.88*</td>
<td>7.1</td>
<td>0.92</td>
<td>9.1</td>
</tr>
<tr>
<td>H, 2007</td>
<td>1964 Upland</td>
<td>1964 Fen</td>
<td>0.72*</td>
<td>7.4</td>
<td>0.81</td>
<td>13.1</td>
</tr>
<tr>
<td>ET, 2006</td>
<td>1930 Upland</td>
<td>1930 Fen</td>
<td>0.89*</td>
<td>-0.04</td>
<td>0.90</td>
<td>0.3</td>
</tr>
<tr>
<td>ET, 2007</td>
<td>1930 Upland</td>
<td>1930 Fen</td>
<td>0.80*</td>
<td>0.08</td>
<td>0.82</td>
<td>0.4</td>
</tr>
<tr>
<td>ET, 2007</td>
<td>1964 Upland</td>
<td>1964 Fen</td>
<td>0.86*</td>
<td>0.07</td>
<td>0.76</td>
<td>0.4</td>
</tr>
</tbody>
</table>

Approximately 130–150 days are included in each comparison.

* Denotes slopes significantly different from unity at the \(P = 0.05\) level.

RMSE = root mean square error.

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Daily ET was also significantly less at younger sites (Table II). The largest difference in ET was between the 1964 and 1930 upland sites in 2007, where there was a 19% difference. We investigated possible driving factors by regressing the difference in daily ET between each pair of sites against the daily mean air temperature, daily mean net radiation, day of year (DOY), daily mean soil volumetric moisture content at 15-cm depth and VPD at noon. The regressions were not significant ($P < 0.05$), and there were no days on which these environmental variables were useful in pinpointing factors driving differences.

**Comparison of upland and fen forests**

Three comparisons of upland and fen forests were possible; 1930 upland and fen for both 2006 and 2007, and 1964 upland and fen for 2007. For both 2006 and 2007, $R_n$ was not found to be significantly different between the two 1930 sites (Table II; $P = 0.76$ for 2006, $P = 0.10$ for 2007). At the 1964 sites, $R_n$ was 4% lower at the fen site, which was significantly different ($P < 0.001$).

A pair of four-component CNR1 net radiometers was used to compare the two 1930 sites in July, and then the two 1964 sites in August 2007. This allowed a comparison of shortwave and longwave radiation fluxes between the upland and fen sites using linear regression. The difference in reflected shortwave radiation was not significantly different between upland and fen sites at either the 1930 or 1964 sites, differing by only 1% ($P = 0.58$ for 1930, $P = 0.56$ for 1964). However, albedo calculated at noon was 8.0 ± 0.32% at the 1930 upland site compared to 8.8 ± 0.1% at the 1930 fen site as a mean of three days (±1 S.D.) when incoming solar radiation exceeded 600 W m$^{-2}$. Under similar conditions, the 1964 upland albedo was 8.6 ± 0.36% and the 1964 fen albedo was 9.7 ± 0.35%. It is clear that albedo was greater at the fen sites, and that the 1964 site had higher albedo than the 1930 site when comparing similar forest types. Daily mean outgoing longwave radiation was not statistically different between upland and fen sites ($P = 0.28$ for 1964, 0.11 for 1930).

Each comparison showed $H$ to be less at the fen sites than upland sites (Tables II, III; $P < 0.001$). During 2006 and 2007, $H$ at the 1930 fen was 12% less than at the upland site. At the 1964 site in 2007, $H$ was 28% less at the fen site than the upland site. The largest difference was caused by a period in late July 2007 when the upland site had much higher $H$ values. This was during a drying period when volumetric soil moisture at the upland site decreased to 0.32 m$^3$ m$^{-3}$ at 15 cm depth, and 0.1 m$^3$ m$^{-3}$ at 2 cm depth. This contributed to the lower $R^2$ values in the regression (Table II).

ET was consistently less at fen sites than upland sites, typically by 10–20% (Tables II, III). The ET regressions tend to have more scatter (lower $R^2$ values) than for either $R_n$ or $H$, but the regression slopes are statistically different from unity ($P < 0.0001$).

For 2006, cumulative ET was calculated for the growing season period of DOY 166–304 (Figure 1a). As this was the year in which research sites were established, the data were not available in the early part of the growing season and trees had already begun transpiring at the beginning of the data set. The 1850 upland site had the largest cumulative ET for this year (304 mm), followed by 1930 upland (258 mm), 1930 fen (227 mm) and 1964 fen (195 mm). The same general patterns were seen for cumulative ET during 2007, even though the measurements started on DOY 134 (Figure 1b). The fen sites consistently had lower cumulative ET than the upland sites of the same age. Cumulative ET was 284 mm at 1930 fen, 239 mm at 1930 fen, 238 mm at 1964 upland and 215 mm at 1964 fen. In both years, the greatest increases for cumulative ET occurred before DOY 258.

We also analysed the difference in daily ET between each pair of upland and fen sites against daily mean air temperature, daily mean net radiation, DOY, daily mean soil volumetric moisture content at 15 cm depth and VPD at noon. Most regressions were not significant ($P < 0.05$). However, on a few days in 2007, the 1964 fen site had greater ET than that at the 1964 upland site when the VPD at noon exceeded 2 kPa. The regression between VPD at noon and the ET difference between these sites for the full season was significant ($P = 0.002$).
Figure 1. Cumulative evapotranspiration (ET) in (a) 2006 and (b) 2007, comparing upland and fen forests of various ages.

but had a low $R^2$ of 0.06 ($n = 143$). Hence, it was not valuable as a predictive tool.

DISCUSSION

Uncertainty

The largest instrument uncertainty in the energy residual technique is our ability to measure $R_n$. Brotzge and Duchon (2000) compared the CNR1, NR-lite and REBS Q7 net radiometers, and found inherent differences among their measurements. Daily averages of net radiation were found to vary by as much as 20% among model types. When seven different NR-lite instruments were compared with each other, daily estimates varied by up to 5%. Cobos and Baker (2003) found very close agreement of the NR-lite radiometer against Eppley radiometers measuring all four components of the radiation budget. Our experience at other forest sites is that the CNR1 estimates slightly more net radiation than the NR-lite radiometer by about 8% when averaged over about 6 months of 30-min data (unpublished data). However, in the present study, using new radiometers with factory calibrations, we obtained an average ratio of CNR1:NR-lite of 1.22 for 3278 30-min averages in summer. This tended to be caused more by an offset than by a slope difference when comparing 30-min or daily averages. There is no independent standard for the net radiation measurement, so this disagreement adds to the uncertainty. However, we do obtain better energy balance closure using the NR-lite radiometer, clearly because the CNR1 radiometer gives a greater reading. This difference contributes to the bias, as opposed to a random measurement uncertainty.

The measurement of $H$ had an error of less than 3% at all sites. The error in $G$ is not well known, although its importance diminishes with longer averaging times. The amount of energy in $G$ is typically less than 5% over the growing season, with the younger sites having greater values than the older sites (Table III). This is caused by thicker canopies in the older sites for a given topographic location as indicated by the leaf area index (Table I). This low value of $G$ also indicates that the error in measurement of the storage and ground heat flux terms is likely a small contributor to the error in the estimate of ET, especially in the older sites.

We estimated the overall random instrument error as the sum of the squares of the individual error components, based on the relative magnitudes of the terms (Bevington, 1969). From Table III, this was 5% of 90 W m$^{-2}$ as an average for $R_n$ and 3% of 36 W m$^{-2}$ as an average for $H$, yielding an overall measurement error of 4.6 W m$^{-2}$ for $\lambda E$ on an average of 50 W m$^{-2}$ or 9%. Hence, the uncertainty in the measurement of $R_n$ created a greater uncertainty in $\lambda E$. The apparent bias among net radiometer types added additional uncertainty in comparisons among sites, but this mostly caused an offset on the actual magnitude of the ET estimate, with a very small effect on the site inter-comparisons. The conclusions are that differences in ET among sites that are greater than 9% are likely outside the bounds of measurement error and represent true differences. Differences less than 9%, although having regression slopes that are statistically different from unity, could be caused by measurement uncertainty.

The age effect

Net radiation was greater at older sites than that at younger sites. In a boreal forest, chronosequence study that included data from the sites near Thompson, Alaska and Saskatchewan, Amiro et al. (2006a) found that summertime albedo decreased and net radiation increased with age. They attributed the differences to species composition where early-succession broad-leaved forests were replaced by black spruce. In summertime, coniferous forests had an albedo of about 0.08, whereas broad-leaved forests had an albedo of 0.16 for stands of the same age (Amiro et al., 2006a). For mid-day incoming shortwave radiation of about 700 W m$^{-2}$, this difference would be 56 W m$^{-2}$ in $R_n$. Typical mid-day values for $R_n$ are about 400–500 W m$^{-2}$; therefore, the albedo differences cause $R_n$ to be 11–14% greater at pure coniferous sites compared to that at broad-leaved sites. In the current study, mid-day albedo was also greater for the 1964 sites than that for the 1930 sites for both upland and fen forests.
by about a factor of 1-1. This reduced net radiation at the younger sites compared to that at the older sites.

\[ H \] was lower at younger sites than that at older sites. The largest difference in \( H \) was between the 1930 and 1850 upland sites in 2006 at 33%. We are unsure as to why \( H \) differs so greatly at these sites because these are the most similar of all the comparisons in terms of tree density, diameter at breast height (DBH) and total basal area. The 1850 upland site was composed completely of black spruce trees, while the 1930 upland had a broad-leaved component of only 5%. The 1850 upland site had undergone some self-thinning as the canopy aged, and therefore the tree density and leaf-area density were slightly lower than those at the 1930 upland. Another noticeable difference between these two sites was that the bryophytes at 1850 upland were better established. The bryophyte hummocks were more pronounced, and the organic peat layer was thicker.

All sites showed a larger proportion of energy partitioned into \( \lambda E \) than \( H \), therefore more energy was used to evaporate water than was dissipated as heat (Table III). \( ET \) was less at the younger sites, typically by 5–20%, depending on the comparison. Much of this can be explained for the 1964 and 1930 upland site comparison by the difference in tree species. For example, the upland sites have a broad-leaved component of 21% at the 1964, 5% at 1930 and 0% at the 1850 sites (Table I). Coniferous evergreen trees are able to begin photosynthesis almost immediately following above-zero nocturnal temperatures (Goulden et al., 2006), whereas deciduous broad-leaved trees are not able to transpire until they experience leaf-out (Pejam et al., 2006). Despite this, forests with a greater broad-leaved tree component often have higher \( ET \) due to higher rates of transpiration per unit leaf area (Ewers et al., 2002). In a previous chronosequence comparison, summertime \( ET \) was greater at ages less than about 25 years compared to that at older forests, which was explained as a function of a greater broad-leaved component in the youngest forests (Amiro et al., 2006a). Hence, our expectation was that \( ET \) at the 1964 upland site would have been greater than that at either of the older upland sites. It is likely that the much greater basal area and leaf-area index at the older sites (Table I) increased transpiration, irrespective of the species mix. However, transpiration differences at these sites cannot be explained by leaf-area index or basal area alone and the species mix does matter (Ewers et al., 2005).

At the fen sites, the difference in \( ET \) can be explained through the amount of vegetation (although all coniferous) contributing to transpiration. At each of the 1964 and 1930 fen sites, there is very little open water, even in spring, so \( ET \) is largely a function of tree transpiration and moss evaporation. The greater basal area and leaf area at the 1930 site (Table I) supports more tree transpiration.

**Uplands and fens**

There was little difference in \( \lambda E \) for any given site age when comparing uplands and fens. Although the 4% difference between the 1964 upland and fen sites was statistically different from a slope of unity (Table II), our uncertainty in the measurement of \( \lambda E \) is of the same order. Further, our comparisons using a four-component radiometer showed no difference in either of the outgoing components, indicating that the topography did not affect the landscape radiation exchange. Although the fens have a more open forest canopy (Table II), this difference does not seem to have much effect on net radiation or its components. This also means that differences in \( ET \) are largely caused by measured differences in \( H \).

For the 1964 site comparison, \( H \) was lower at the fen during a period in early July, where we suspect that the upland site may have experienced a brief period of water stress, which increased \( H \) and lowered \( ET \). For the 1930 site comparison, \( H \) at the fen is higher than that at the upland during days with relatively low \( H \) values. This pattern seems to shift once the daily \( H \) values exceed approximately 80 W m\(^{-2}\), and \( H \) becomes greater at the upland site. During days with high \( H \), such as early spring before water becomes available, more energy is partitioned into heating the air at the upland sites than at the fen sites. We propose that, during these periods, more energy is needed to warm the upper soil organic layer at the fen sites than at the upland sites because soil moisture is greater at the poorly drained fen sites. The ground temperatures demonstrate a distinct lag behind air temperatures, and this lag is more pronounced at the fens. Soils at the fens also do not reach temperatures as high as at the upland sites at depth, most notably at 30 and 50 cm depths. Therefore, the better drainage at the upland sites allows them to warm up faster in the spring.

For all topographic-effect comparisons, \( ET \) was greater at upland sites by more than 10% (Tables II, III). This was likely driven by greater transpiration from the greater tree biomass and leaf area at the upland sites. The lesser tree transpiration in open fen forests results in an increased relative contribution by understorey and bryophyte plants, which can contribute as much as 50% of the total \( ET \) (Constantin et al., 1999).

**Implications and conclusions**

For boreal black spruce stands in Manitoba, real differences in \( ET \), \( \lambda E \) and \( H \) existed among stands of different ages and soil drainage. The data concentrated on water and energy budgets for forest stands between the ages of 40 and 150 years, as well as stands that have very different drainage capabilities. The number of forest fires has been predicted to increase by as much as twofold in a future climate with a tripling in atmospheric CO\(_2\) concentrations, which would renew stands at an increasing rate (Flannigan et al., 2005). If these predictions come true, the boreal forest as a whole would be younger, potentially changing the boreal forest’s contribution to the global water cycle. We would also see a change in the radiation balance and partitioning of these stands. It has been suggested that this may cause a cooling feedback, which would counteract the warming that would be encountered.
due to increased CO₂ release by forest fires (Randerson et al., 2006). Early successional boreal forests often have a large proportion of broad-leaved trees, and an overall younger boreal forest would see a change in water use, especially in more southern boreal stands (Chapin et al., 2000). The chronosequence pattern is complicated with very young stands having lower ET, stands with a high broad-leaved component less than 25 years of age having higher ET, and then a relatively unchanging situation with older forest ages, averaged over the broad boreal landscape (Amiro et al., 2006a). However, the present study demonstrates that boreal coniferous forests that are about 75 years old have roughly 10–20% greater ET than that at forests about 40 years of age. This indicates that the effect of fire persists for a long period after the event in both upland and fen forests.

The recommendation for modelling and landscape characterization is that fen areas can be expected to have ET rates that are 10–20% lower than adjacent upland areas during a growing season. This difference may only be about 0.2 mm day⁻¹, totally 30 mm over 150 days. For large drainage basins, on the order of 10,000 km², this difference represents 3 × 10⁶ m³ water annually that is lost to the atmosphere. The management implications for topography are mostly related to changes in drainage by humans for hydroelectric power or roadway construction. However, a changing climate will likely alter precipitation patterns and also change the rate of ET. Decreases in water availability will decrease the relative fraction of fen to upland forest areas. Although our data suggest that a greater relative upland forest area will increase ET, water stress in upland forests will limit this increase. The small differences over large scales need to be addressed in the development of regional and global climate models, along with the assessment of impacts caused by landscape changes.

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