The Energy Balance of the Winter Boreal Landscape

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ABSTRACT

During the winter of 1993/94 a study to quantify the winter energy balance of the main cover types of the boreal landscape took place. The study was based on the southern edge of boreal forest in Canada. Measurements were made over a mature jack pine stand and a frozen lake. Shortwave albedos of 12% to 14% over the jack pine and 20% to 88% on the frozen lake (both depending on snow cover) were measured. There were correspondingly large contrasts in the total radiation inputs and the turbulent heat fluxes. The mean net all-wave radiation input was large and positive into the forest and negative over the lake. The sensible heat fluxes were of the same sign as the radiative inputs with positive values over the forest peaking at +200 W m^{-2} and falling to −100 W m^{-2} over the lake. The evaporation from the forest depended on whether the there was snow on the canopy. When the canopy was snow-free, the evaporation was low, about 50% of net radiation but, when there was snow on the canopy, the evaporation was large, 4 mm over a 36-hour period. The results of these experiments are being used to design much-improved descriptions of boreal forest within the next generation of climate models.

1. Introduction

The boreal forest is a complex mosaic of land surface types varying from closed coniferous canopies, through sparsely vegetated cleared areas, to open lakes. These surface cover types interact in very different ways with the incoming radiation and overlying atmosphere. The consequence is a substantial spatial variation in microclimate and surface fluxes of radiation, heat, and water vapor. This is particularly marked in the winter period when the vegetation interacts strongly with snow. Climate models simulate large changes of temperature in northern latitudes in response to increased atmospheric carbon dioxide. These changes are large because of the strong feedback between increased temperature and reduced snow cover. However, there will be further modification of climate arising from changes in landscape cover type. Forest cover will change both as a direct result of land clearing and as a result of the response of the vegetation community to climate warming.

The southern boreal forest of western Canada is known to be extremely sensitive to hydrological and climatic conditions. This forest exists on a climatic margin between the semiarid prairies, where low rainfall limits primary productivity, and the boreal/subarctic, where the limitation is low temperatures. In the summer low soil moisture results in drought stress on trees and the risk of large-scale fires. The sensitivity of forest survival to available water and climate is evident at locations where southern exposure on slopes gives slightly higher radiative heating and, hence, increased likelihood of water stress. Here, natural grasslands have developed rather than the mixed-wood aspen–spruce or jack pine stands that are characteristic of the region. In contrast, nearby wetlands display pockets of permafrost and cold summer soil temperatures, which restrict tree growth to scattered patches of stunted larch and black spruce. The southern boreal forest is therefore characterized by a natural mosaic of open areas (grassland, wetland, lake, burnt), deciduous forest (aspen dominated), and coniferous forest (pine and spruce), on which are superimposed human modifications such as clear-cuts and cleared farmland.

The long, cold winter characteristic of the boreal forest causes cessation of growth for five to seven months of the year and is the most distinguishing feature of this biome. The components of the energy balance, which determine the net energy supply to the canopy and the surface, are small during this winter period—with direct effects on biological activity. Most of the components of the energy balance: the radiation flux (solar and thermal), sensible heat flux (turbulent transfer of

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heat from the surface to the atmosphere), latent heat flux (evaporative cooling or warming upon condensation), and to a lesser degree the soil heat flux, are strongly influenced by the presence and depth of snow cover and its interaction with vegetation. Snow reflects most solar radiation incident upon it and hence can dramatically reduce the net radiative energy input. However, when the snow covers a forest canopy, the multiple reflections within the canopy scatter rather than reflect the majority of incident radiation and the albedo remains low (van de Hulst 1957; Oke 1978).

The exposure of snow to the atmosphere varies with landscape type: in open areas it forms a smooth surface but in coniferous forests it only partially covers the canopy. Sublimation of snow occurs where it is well exposed, such as from blowing snow particles in open areas and from intercepted snow clumps in coniferous canopies (Pomeroy and Gray 1995). Sublimation can result in substantial latent heat fluxes from the surface and loss of seasonal snow cover before melting. However, once the snow reaches the forest floor, it is shaded from solar radiation and sheltered from the vigorous turbulent exchanges above the canopy; hence snowmelt and sublimation are suppressed. Because winter climate is a distinguishing aspect of the boreal forest, it is important to document the winter energy balance of the boreal forest and the sensitivity of this balance to land surface and climate conditions.

General circulation models suggest that the strong interactions between the surface energy fluxes and snow cover cause important feedbacks on climate. In the double CO_2 simulations warming is considerably enhanced in high latitudes (IPCC 1992) because the effects of the snow cover on the radiation budget induce a positive feedback on the temperature. Similarly, interactions between snow cover and forest can cause large changes in global climate. Bonan et al. (1992) and Thomas and Rowntree (1992) have investigated the sensitivity of GCMs to the specification of the land surface parameters describing the boreal forests. Both studies show a marked decrease of temperature in the spring associated with the removal of the forest; this is primarily related to the increase in albedo. Interestingly, the simulation of Bonan et al. extended throughout the year and showed that the effect of forest removal continued into the summer months, after a maximum in April. The continued modification of climate through the snow-free months is attributed by the authors to be in part a result of a memory effect of the interactive ocean included within this model.

Current GCMs generally have a very simple representation of snow cover and an even simpler description of the complex interactions between snow and forest. Future GCMs will need to include a much more complete description of the boreal forest. It will be essential that these schemes correctly represent the physics of the snow in the forest and that they are tested against measurements, on all scales up to that of the GCM grid cell. It is against this background that two international experiments (BOREAS, GEWEX) have recently started in the boreal forest region of Canada (Sellers et al. 1995; Krauss 1994). These studies have the objective of quantifying the atmospheric and hydrologic interactions. This paper outlines some of the first measurements taken under these experiments.

2. Study site

Prince Albert National Park, Saskatchewan, Canada (53° 52′N, 106° 8′W), on the southern edge of the boreal forest, was selected for study because of its relatively natural mix of landscapes and also because it is contained within the southern study area of the BOREAS (Boreal Ecosystem–Atmosphere Study) experiment. The area experiences a subhumid, strongly continental climate, with cold dry winters, moist cool springs, and hot dry summers. The mean annual precipitation is 463 mm, of which 33% occurs as winter snowfall. The mean climatology of Prince Albert, 80 km to the south of the study sites, is shown in Fig. 1. The snow-covered period normally lasts six months, with seasonal snowmelt in late April. Midwinter melts are rare.

The mosaic of vegetation cover as derived from classification of a recent Landsat thermatic mapper image of the region is shown in Fig. 2, along with the two specific study sites. The first site is a mature, slightly open, jack pine stand (*Pinus banksia*) in the Beartrap Creek catchment. The fetch is level and reasonably uniform for at least 100 m. The pine canopy is 16–22 m in height with a sparse understorey of deciduous bushes. The winter "leaf" (branch and needle) area

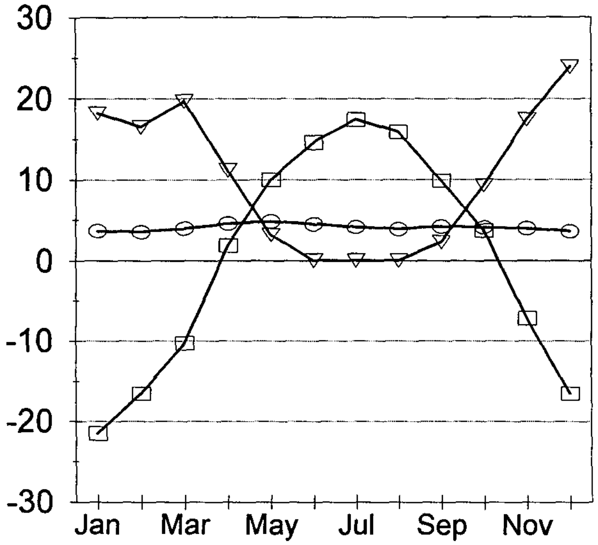

FIG. 1. Mean climatology for Prince Albert, Saskatchwan. Open squares: temperature (°C), open triangles: snowfall (mm), and open circles: wind speed (m s^{-1}).

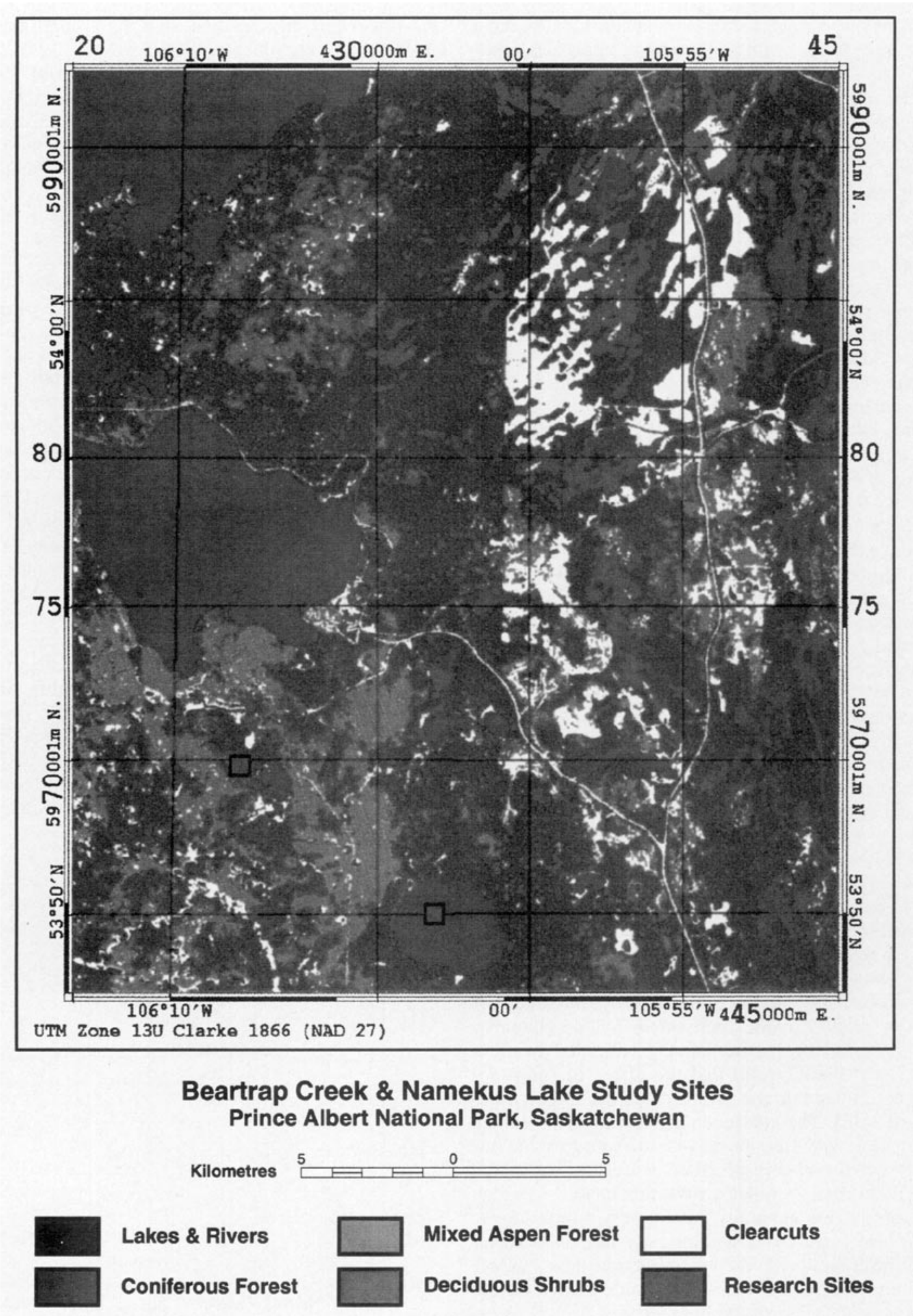

FIG. 2. Landsat TM image of the experimental area showing the location of the Bear Trap Forest and Namekus Lake sites.

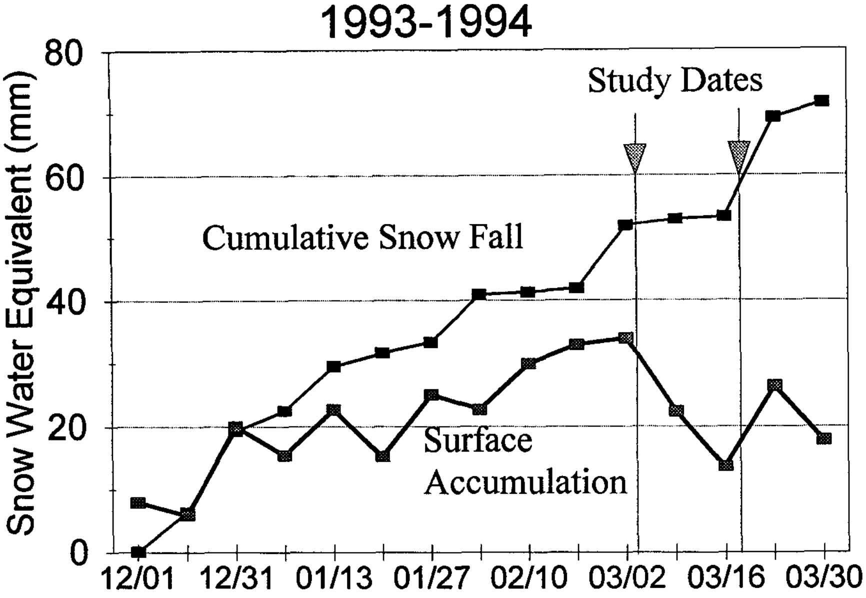

FIG. 3. Snow accumulation at the pine site for entire winter.

index of this stand is 2.2 and the canopy cover of sky is 82%, measured with a Li-Cor LAI-2000 Canopy Density Meter (Gower and Norman 1991). The second site is on the snow-covered frozen surface of Namekus Lake, 10 km southeast of the jack pine stand. The surface is quite smooth (snow over ice) and the fetch exceeds 700 m.

The pine site was equipped with a 26-m meteorological tower from which measurements were taken of shortwave radiation (downward and reflected) and net all-wave radiation above and below the canopy, along with vertical profiles of temperature, humidity, snow particle flux, and wind speed through the canopy. A "Hydra" eddy correlation system was also operated from the top of the tower for direct measurement of sensible and latent heat fluxes (Shuttleworth et al. 1988). A second eddy correlation system measuring only sensible heat flux was operated from the tower, and this provides a useful confirmation of the Hydra measurements. This consisted of a "Solent" sonic anemometer (Gill Instruments Ltd., Lymington, U.K.), a fine wire thermocouple, and a Campbell Scientific CR21 with software similar to that described by Shuttleworth et al. (1988).

A single weighed tree system was operated, which gave a continuous output of the weight of snow on the forest canopy (Pomeroy and Schmidt 1993). These data are a useful check on the eddy correlation evaporation measurements (although there are uncertainties about how representative a single tree is of the entire canopy).

The net all-wave radiation above the canopy is a particularly important quantity, being the overall energy input to the canopy available for evaporation and sensible heat. It was measured with two instruments, both on short (~2 m) booms at the top of the tower, a Middleton radiometer and a REBS (Radiation Energy Balance Systems, Seattle, Washington) Q*5. There have been reported problems with the calibration and design of these latter instruments, but in this case, good agreement was found between the two different designs. Below the canopy shortwave and net all-wave radiation were measured with "Delta 7" tube radiometers (Delta-T Devices Ltd., Cambridge, U.K.). These tubes (almost one meter in length) average the speckled light found beneath the canopy (Szeicz et al. 1964).

The lake site was equipped with a 3-m mast with instruments to measure the two shortwave components of the radiation balance, the all-wave net radiation, temperature, humidity, snow particle flux, and wind speed (at 1, 2, and 3 m) above the snow surface. A second Hydra was operated to provide fluxes of heat and water vapor.

Nipher shielded snow gauges under the forest canopy and in a nearby open area provided weekly snowfall and snow interception. Weekly snow surveys, both underneath the forest canopy and on the lake, were undertaken to give average snow depth and the water equivalent of the pack.

Snowfall and accumulation at the pine site are shown in Fig. 3 for the entire winter. There was much less snow on the ground than fell over the winter. The total cumulative winter snowfall at the beginning of March was 55 mm water equivalent. The pine snow accumulation was divided between surface accumulation and interception. Despite interception losses there was

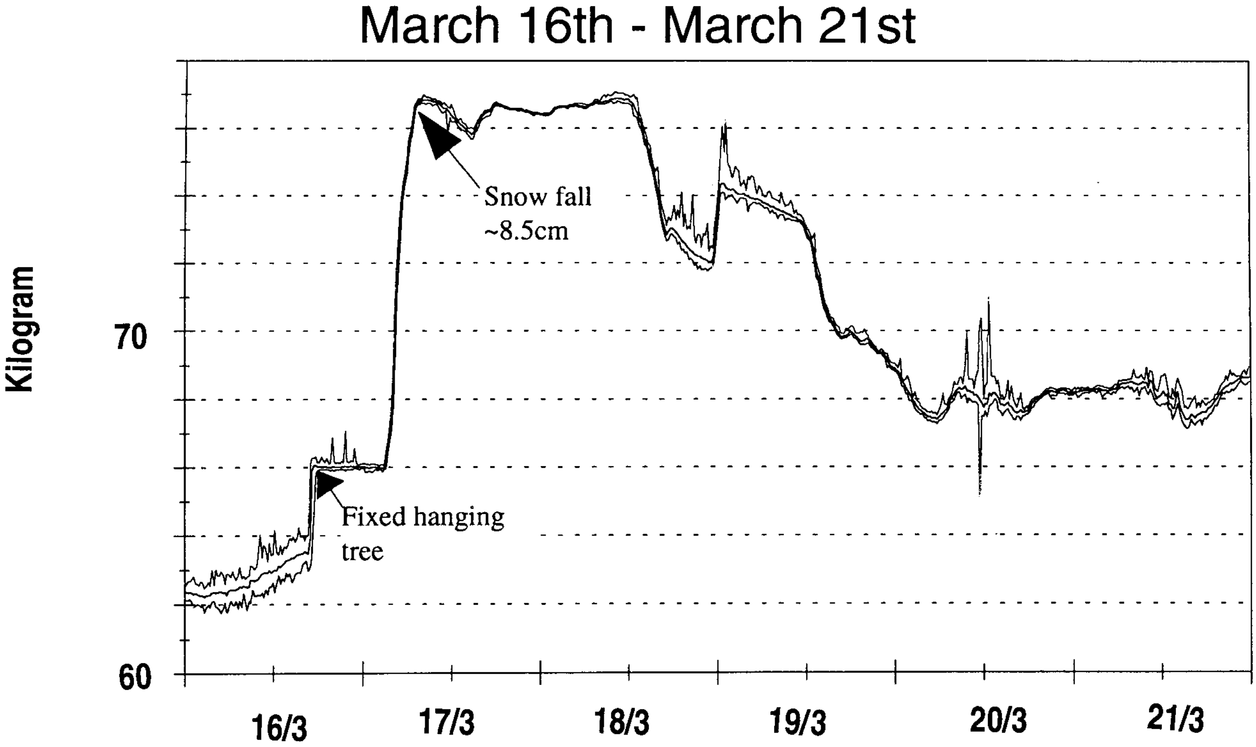

FIG. 4. Weight of the hanging tree, 16–21 March 1994, showing mean and extremes.

more snow under the pine canopy than on the lake surface. This is primarily because wind erosion removes snow from the lake surface and then transports and sublimates it. There was no intercepted snow in the pine canopy on 3–4 March; however about 10 mm snow water equivalent was lodged on to the forest canopy on 18–19 March, with an uncertain amount already sublimated.

3. Microclimate and comparative energy balances

Two periods were chosen for analysis, 3–4 March and 18–19 March. These periods are useful because a complete set of measurements was taken, the incoming and outgoing energy fluxes were relatively high, and contrasting snow conditions prevailed. On 3–4 March, the pine canopy was snow-free but there was 30-mm water equivalent (47 depth cm) of snow on the forest floor. The lake had a snow cover depth of approximately 10 cm. The radiation inputs were high, as were the wind speeds and air temperatures.

On 18–19 March the pine canopy had intercepted a substantial load of snow, which was undergoing sublimation (Fig. 4). Snow fell on the morning of 17 March, with an accumulation of 11 kg on the weighed tree. The weighed tree indicates a slight sublimation of the snow on the canopy around noon on 17 March followed by a further small snowfall. On the morning of 18 March there was a slight increase in the weight of the tree followed by rapid loss of 5 kg until midnight. At midnight there was a further rise in weight followed by a steady loss throughout 19 March, accelerating during the afternoon, with a total loss of 4 kg over the day. The flux and radiation measurements were disrupted on the morning of 18 March by the fall of snow, and the averages for this period, presented below, are for 1400–2400 (local time) 18 March and 0000–2400 19 March.

The lake on 18–19 March had a shallow snow cover (9.6 cm deep on 16 March). There is evidence from the blowing snow particle detectors and eddy correlation equipment that this experienced some wind redistribution and in-transit sublimation. Throughout the afternoons of 18 March and 19 March the wind speeds were moderate (3–6 m s^{-1}) and the air temperatures relatively high, rising to above zero on the afternoon of 19 March. The incoming solar radiation was high on the afternoon of 18 March (peak 452 W m^{-2}) but lower on 19 March (peak 256 W m^{-2}).

On 3–4 March the wind speeds had a strong diurnal pattern with a peak midafternoon decreasing almost to zero at night (Fig. 5a). The diurnal variation was most pronounced over the lake, with higher daytime and lower nighttime wind speeds compared to above the forest canopy. The wind speeds below the canopy were low throughout, almost zero at night but peaking to 0.5 m s^{-1} during the day. The air temperatures were similar

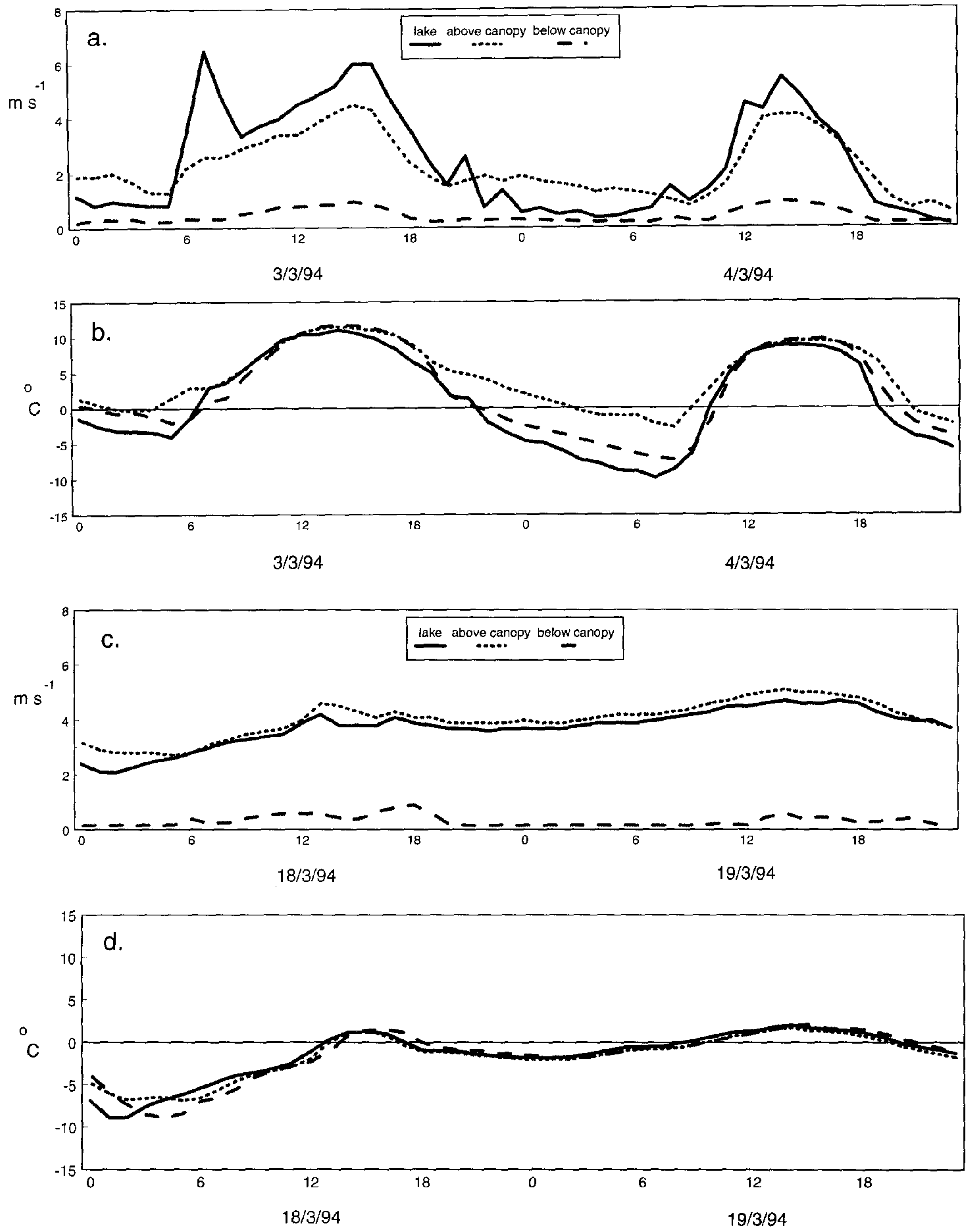

FIG. 5. Hourly air temperatures and wind speeds for the lake site (at 3 m) and above and below the forest canopy.

during the daytime, with the lake between 0.5° and 1°C cooler than the two forest temperatures, which were almost equal (Fig. 5b). At night the air temperatures diverged with the lake being the coolest and the above-canopy temperature markedly higher than that below the canopy or the lake.

On 18–19 March conditions were more dominated by the large-scale flows. There were rapid rises of temperature and wind speed at the time of the snowfall during the first half of 18 March (Figs. 5c, d). After this time the wind speeds over the lake were similar to those above the canopy, as were all air temperatures. It is evident that for both periods the air temperatures were similar, to within 1°C, when the wind speed was appreciable, but diverged when the wind speed dropped. The temperature comparisons demonstrate

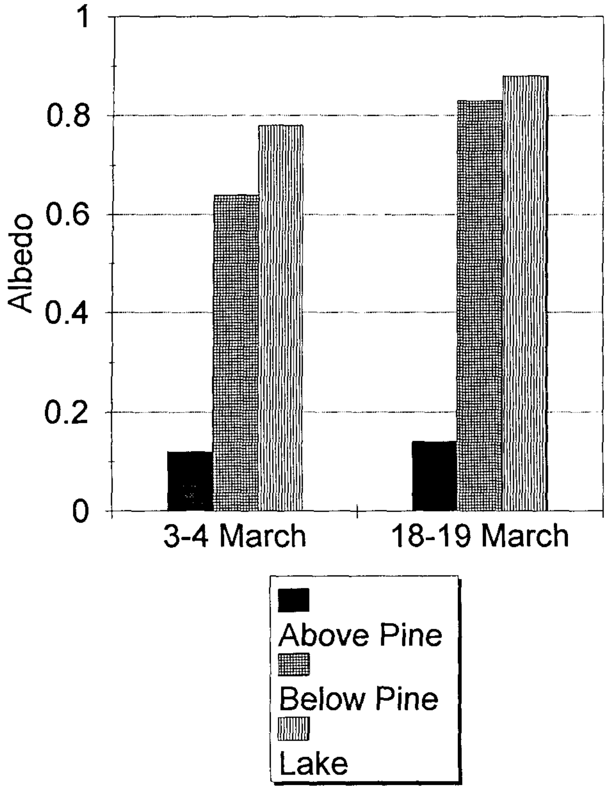

FIG. 6. Measured shortwave albedo values for the lake and above and below the forest site canopy.

the direct climate modification of the pine canopy, whose effect in winter was to increase the air temperatures and moderate the temperature fluctuations. The wind speed below the canopy was only above zero during daytime. Even when the above-canopy wind speed was high at night (i.e., during the night of 18–19) the below-canopy wind speed was still zero. This indicates that the coupling through the canopy is due to penetration of daytime convective turbulence.

Albedos at the sites contrast sharply, as shown in Fig. 6. The albedo at the top of the pine canopy on 3–4 March was about 0.12; that is, 88% of shortwave radiation is absorbed by the canopy and the ground surface. The albedo at the ground under the pine canopy was 0.64. Thus, despite the open canopy and the high albedo of the forest floor the overall forest albedo remains low, confirming the effectiveness of radiation trapping within the canopy. The albedo for the lake snow cover was high, at about 0.78, with only 22% of the shortwave radiation being absorbed by the lake snow cover. Albedos were slightly higher on 18–19 March because of fresh snow on the canopy and snow surfaces. The pine canopy albedo increased to 0.135 and the lake snow cover albedo to 0.88. However, it should be noted that even with snow on the canopy the albedo remains low.

The net longwave fluxes for the different surfaces are shown in Fig. 7. On 3–4 March the highest longwave loss was from the forest. The warmer temperatures over the forest, compared with those over the lake resulted in a higher loss. In contrast, there was very little net longwave loss from the subcanopy snow cover. The longwave emissions from this surface were intercepted and reemitted both upward and downward by the canopy, which was at a similar temperature. The net longwave exchanges were smaller on 18–19 March because of increased cloud cover (particularly on 19 March). The combined effects of incident shortwave radiation, albedo, and net longwave fluxes on the net radiation flux is shown in Fig. 8. There was substantial contrast between sites with net radiation an input to the pine canopy and a loss from the lake snow cover. Beneath the pine canopy the net radiation was small and downward on 3–4 March and essentially zero on 18–19 March.

The contrast between the turbulent fluxes from the snow-covered lake and the forest was considerable. There was also a strong dependence of the fluxes on the vertical distribution of the snow in the forest, (Table 1 and Fig. 9). This was marked for both the sensible and latent heat fluxes. Over the lake the flux of sensible heat was small and downward (see, e.g., Figs. 9b, d). This is a characteristic of open snow surfaces. The

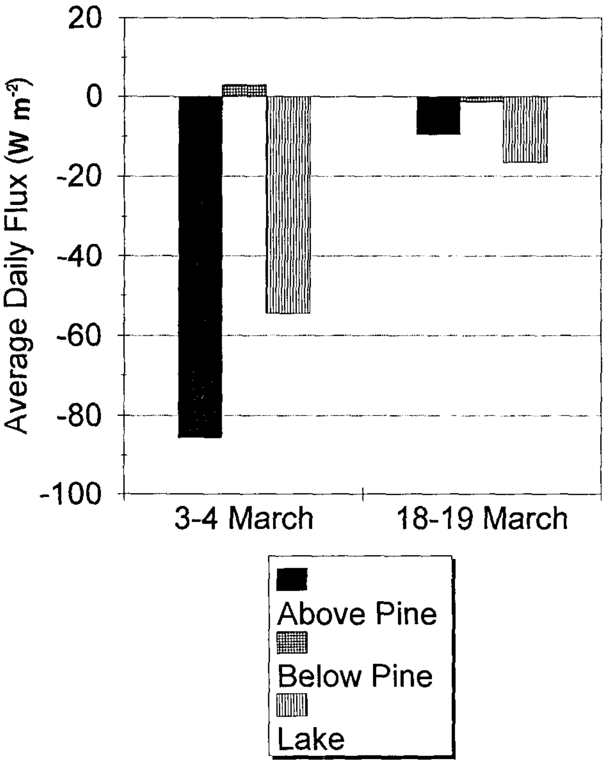

FIG. 7. Daily mean net longwave fluxes for the lake and forest (above and below canopy) sites.

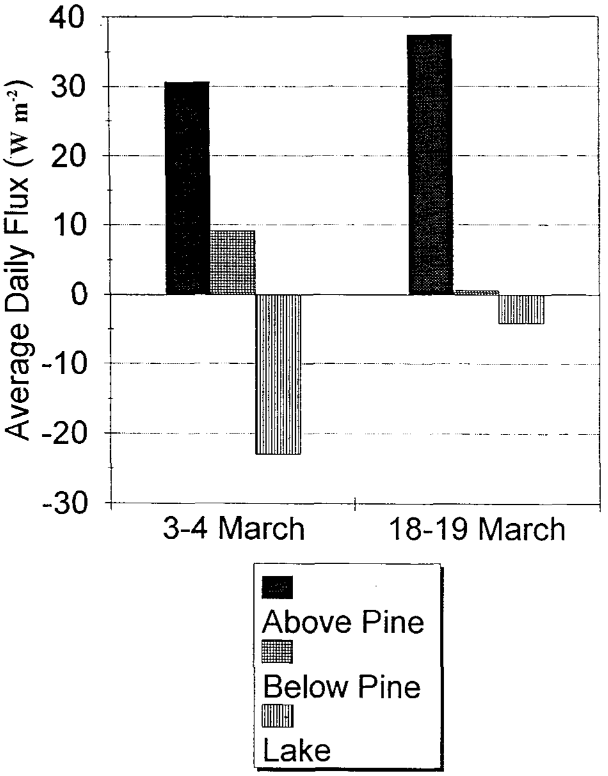

FIG. 8. Daily mean net all-wave radiation flux for the lake and forest sites.

smooth surface of snow makes the transport of heat inefficient and, hence, the fluxes small. The generally negative net radiation usually results in a snow surface that is cooler than the air (and, of course, if the air temperature rises above zero, the surface temperature cannot follow) (cf. Harding 1986). Thus, the lower atmosphere is stable, leading to negative heat fluxes and increased resistances. In contrast the snow-free forest canopy on 3–4 March showed an upward heat flux commensurate with the net radiation input. However, the snow-covered canopy found on 18–19 March exhibited a small downward heat flux to compensate (in part) for the large evaporation.

The latent heat flux was small over the lake on 3–4 March. Over the forest the evaporation rates are low: about one-third of the net radiation during the day but one-half over the whole 24-hour cycle with a mean daily total of 1 mm. At this time the only source of water was on the forest floor, and this is to some extent isolated from the vigorous turbulent exchange above the canopy (the low temperatures and frozen soil make transpiration impossible).

During 18–19 March evaporation from the lake was again small but slightly greater than during the earlier period. The wind speed for this period was between 3 and 6 m s^{-1}, and it is likely that the evaporation was enhanced by blowing snow (Pomeroy et al. 1993). Over the forest the evaporation rates were high—averaging over 80 W m^{-2} with a total of over 4 mm over the 36-hour period (from 1200 18 March to 2400 19 March). This was the result of the very significant effect of large roughness coupled with free availability of moisture on the canopy.

Over Namekus Lake the sum of the turbulent fluxes agrees reasonably with the input net radiation (particularly considering the small size of the fluxes). Unfortunately, neither snowmelt nor ground (ice) heat fluxes were measured at this site, but the energy closure neglecting these suggests they were small. However, for both periods there appears to be an imbalance between the sensible and latent heat fluxes and the radiation input in the forest; the energy from net radiation and downward sensible heat flux could not provide sufficient energy to supply the high observed evaporation. Measurements of ground heat flux are typically 2 W m^{-2} and so could not provide this energy. It seems unlikely that the flux measurements were seriously in error: the net radiation was measured by two completely separate systems, which agreed to within 10 W m^{-2}. The sensible heat fluxes were again measured by two systems (Table 1), and again the agreement was within a few watts per square meter. Finally, measurement of the latent heat flux using eddy correlation is strongly supported by measurements from the weighed tree. It is possible that during the latter period there was some strong horizontal advection at this site (with the localized radiation measurements being unrepresentative of the "footprint" of the measurements of the turbulent fluxes), or there is also the possibility of some large change of heat storage within the trunk space. This imbalance obviously requires further study.

Large evaporation rates have been documented for rain-covered canopies (see, e.g., Stewart 1977), but direct measurements for snow-covered canopies are rare. However, the few reported studies suggest that

TABLE 1. Daily averages of components of the energy balance for two sites (W m^{-2})

Date	Net radiation	Latent heat flux (HYDRA)	Sensible heat flux (HYDRA)	Sensible heat flux (Solent)
		(a) Beartrap Forest site		
3 March	30.6	16.6	27.5	32.5
4 March	30.9	18.8	46	48.3
18 March (1400–2400)	39.3	129.8	−16.3	−8.7
19 March	21.0	69.0	−11.2	−8.9
		(b) Namekus Lake		
3 March	−23.2	6.6	−29.5	
4 March	−22.7	7.9	−15.5	
18 March (1400–2400)	−9.3	20.2	−11.9	
19 March	1.1	7.7	−6.3	

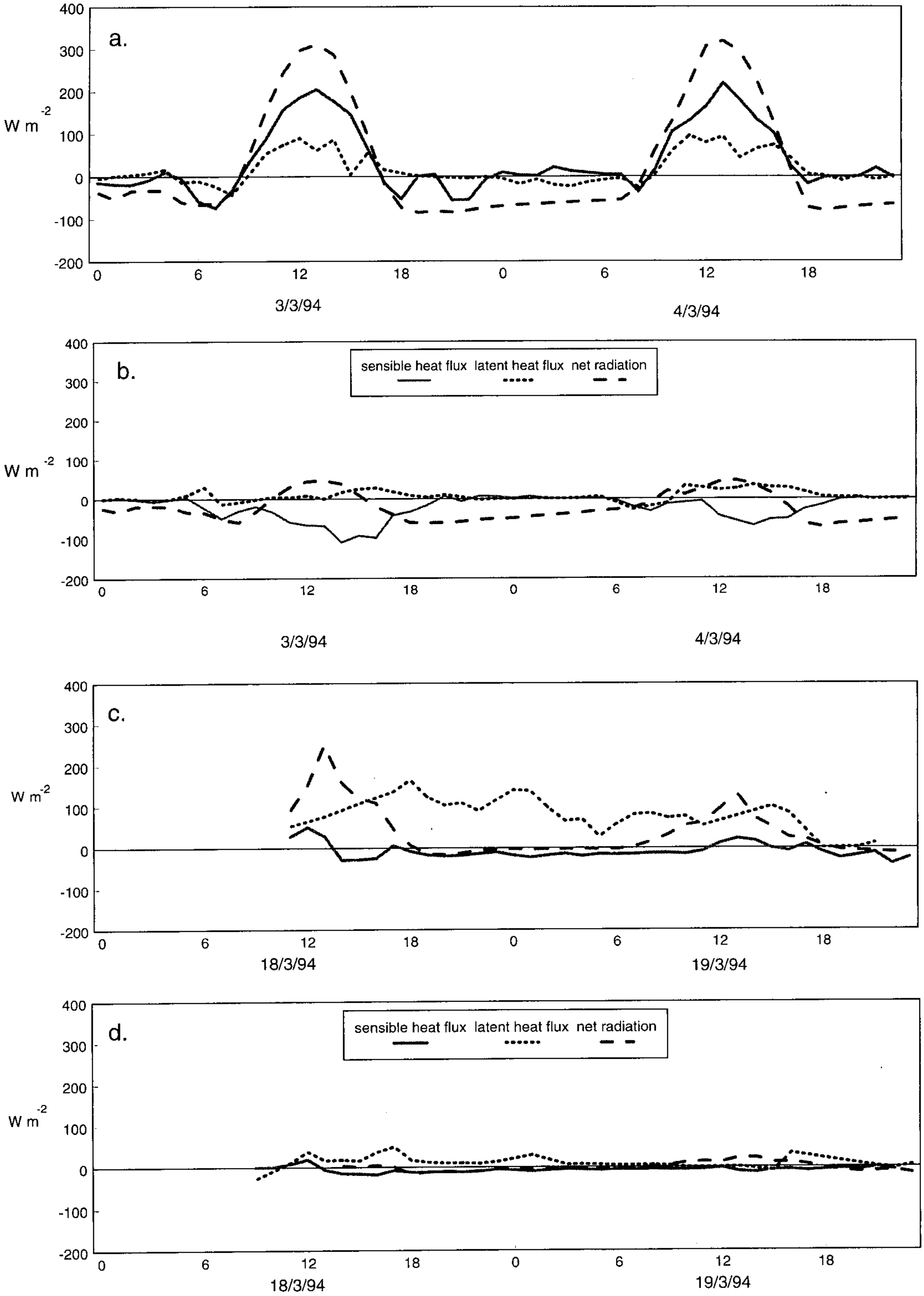

FIG. 9. Hourly net radiation, latent heat flux, and sensible heat flux for the lake and forest sites.

high evaporation rates are possible. Calder (1990) reports rates of up to 0.5 mm h^{-1} from wet snow on a coniferous canopy, and Lundberg and Halldin (1994) describe a mean evaporation of 0.3 mm h^{-1} from a snow-covered forest in northern Sweden, sustained for 7 hours. These large rates lead, typically, to a 30% reduction in the cumulative snowpacks observed under the boreal forest (Pomeroy and Schmidt 1993) and will have an important impact on regional water resources and on the availability of soil water for plant growth in the following summer.

4. Conclusions

This paper presents a comparison of the surface meteorology and energy balance of a snow-covered lake and an adjacent forest area (including some of the first direct measurements of turbulent fluxes from a snow-covered forest) for two contrasting periods during March 1994. There were small, but measurable, contrasts in the temperature and wind speed measured over the lake and forest (the lake was cooler and windier), but it is the comparison of the energy balance of the forest and the open snow surface that is dramatic: The mean net radiation flux was into the forest canopy but out of the snow-covered lake. Similarly, when the forest canopy was clear of snow, the sensible heat flux was of different sign (and magnitude) over the forest and lake. When the forest canopy was snow covered the partition of the sensible and latent heat fluxes was different again, exhibiting a large upward latent heat flux and a compensating downward sensible heat flux. In this area of the boreal forest of Canada, lakes cover 15% of the land area, with other deforested areas (swamps, clear-cuts, etc.) covering a further 18%. It is evident from these measurements that these unforested areas have exchanges with the overlying atmosphere that are radically different from the continuous forested areas. However, from the point of view of climate models these patches are of subgrid scale. Such patchiness is usually not represented in current GCM surface schemes; when it is, it is as a tile with no allowance for the interaction between tiles. This heterogeneity, the associated sharp contrasts in energy balance, and advection of energy from one cover to another should be represented in some way in the GCMs if the correct representation of boreal forest region is to be made.

The high evaporation rates observed over the snow-covered forest are important to the regional water balance. Snow surveys suggest that there is 30% less snow accumulating within these forests, compared to local clearings. This data provides confirmation of the physical mechanisms underlying the snow survey observations. The data presented for 18–19 March may be at the extreme end of possible climatic conditions in this region, as the air temperature at this time was about zero—higher than is usual at this time of year—leading to, perhaps, a higher evaporative demand. There is a strong need for a physically based model to allow the extrapolation of these data in both space and time.

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