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Impact of BOREAS on the ECMWF forecast model

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Abstract. This paper reviews the impact of the Boreal Ecosystem-Atmosphere Study (BOREAS) on the development of a new land-surface parameterization in the European Centre for Medium-Range Weather Forecasts (ECMWF) global forecast system and discusses briefly the improvements that resulted in the model climate at high latitudes. We show how a long time series at a representative site was used to validate the model parameterizations of observed physical processes on both diurnal and seasonal timescales. As a direct result of BOREAS, the representation of several surface processes was greatly improved over the boreal forest. The introduction of separate tiles for tall and short vegetation meant that the boreal forest could be represented with some realism. In winter the albedo with snow under the trees was greatly reduced, and the introduction of a prognostic snow model with its own energy balance under the canopy meant that evaporation of snow in winter and spring was similarly reduced. An improved separate handling of liquid and frozen soil water meant that evaporation of frozen soil water is shut off, and surface runoff occurs when snow melts on frozen ground. Trees in the model now have larger more realistic unstressed vegetative resistance, as well as a stress factor for vapor pressure deficit, in addition to one for low light levels; both of which reduce summer transpiration to the lower levels observed over the boreal forest. The model does not yet have a global soil distribution, which means that it does not represent the organic soils that are characteristic of the black spruce sites, which have a larger water storage.

1. Introduction

This is a review of the impact of the Boreal Ecosystem-Atmosphere Study (BOREAS) on the land-surface parameterization and model climate in the European Centre for Medium-Range Weather Forecasts (ECMWF) global forecast system. It is both a review of the science accomplishments and a case study, which illustrates the benefits of a close integration between a field program and an operational forecast center. Previously, during the research phase of FIFE (First International Satellite Land Surface Climatology Project (ISLSCP) Field Experiment), the ECMWF research department had realized how valuable field data could be in the development and testing of new land-surface parameterizations. The FIFE data from 1987 was first used to identify errors in the land-surface model [*Betts et al.*, 1993], then to help develop (along with data from several field programs) a new surface scheme with four soil layers [*Viterbo and Beljaars*, 1995] (hereinafter referred to as VB95), and a revised boundary layer parameterization [*Beljaars and Betts*, 1993]. Subsequently, the new model showed improvements in forecast skill for summer precipitation and drew attention to the feedback between soil water and precipitation [*Beljaars et al.*, 1996; *Betts et al.*, 1996]. Consequently, during the planning phase of BOREAS, which was an international experiment involving the close collaboration between the National Aeronautics and Space Administration (NASA) and the Atmospheric Environment Service (AES) of Canada, the European Centre research department gratefully accepted the opportunity to collaborate as coinvestigators. The center provided daily forecasts during the field phase, and model surface fields for interdisciplinary research at the land surface, including the interaction of the carbon and water cycles on the 1000 x1000 km scale. In return the analysis of the field data provided direct insight into model errors in the surface energy balance related to errors in the model albedo with snow [*Viterbo and Betts*, 1999], surface evaporation [*Betts et al*, 1998a], and precipitation and runoff at high latitudes [*Betts and Viterbo*, 2000]. This work speeded the development of a new tiled land-surface model [*Van den Hurk et al.*, 2000], with improvements specifically designed to improve the representation of forests, and the physics of high-latitude cold processes. It should be noted that the importance of many of these physical processes, such as the effect of canopy shading on the albedo of forests in winter and its climatological significance has been known for some time [*McFadden and Ragotzkie*, 1967; *Robinson and Kukla*, 1984, 1985; *Otterman et al.*, 1984; *Cohen and Rind,* 1991]. A few papers have specifically addressed the impact of the low boreal forest albedo on winter climate [*Thomas and Rowntree*, 1992; *Bonan et al.*, 1992, 1995]. There has also been considerable development of improved models for high latitudes, such as the Canadian land surface scheme [*Verseghy et al.*,1993; *Verseghy*, 2000] and of detailed snow cover models for climate simulations [e.g., *Loth et al.*, 1993]. It was the BOREAS

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experiment, however, that catalyzed high-latitude improvements in the land-surface model at ECMWF.

Lemone [1983] has noted the considerable lag, as long as 10 years, between field programs and the publication of their results, which necessarily delays the development of improved parameterizations in forecast and climate models. In the case of BOREAS the tighter integration of the operational European Centre into the field program shortened this development time of improved parameterizations to about 4 years, between the field phases of 1994 and 1996, and the new tiled land-surface model for the 40-year ECMWF reanalysis (ERA-40), for which we shall show results here.

1.1. Significance of Land-Surface Interaction

The land-surface interaction is as important to the climate of a global model as the sea surface boundary condition but presents quite a different modeling problem. Over land, there is no measured field analogous to the sea surface temperature, which controls the surface sensible and latent heat fluxes over the ocean (together with surface wind, air temperature, and humidity). The fluxes over land in contrast are driven on diurnal timescales by the net radiation, and the partition into latent and sensible heat, which depends on the availability of water for evaporation (either from the soil or from the surface reservoirs). The accuracy of the downwelling fluxes depends on the model radiation physics and the determination or specification of clouds and aerosols. The outgoing fluxes depend on the calculated surface skin (radiation) temperature, and the albedo, which in turn depends on snow cover, vegetation type, and season (for example, leaf-out). Thus even the calculation of net radiation at the surface involves many physical parameterizations. The availability of water for evaporation depends on both a realistic hydrological model (partitioning model precipitation into surface and soil water reservoirs, drainage, and runoff) and a realistic vegetation model (to extract soil water for transpiration as a function of photosynthetic processes). The freezing and thawing of the soil plays an important role in the climate at high latitudes. It moderates winter temperatures, because during the freeze process the effective heat capacity of the soil is increased by a factor of 20 [*Viterbo et al.*, 1999], and it introduces a significant lag into the system. In spring a significant part of the net radiation goes into melting the ground and lakes (and in warming them). This energy becomes available in the fall and early winter, when the surface refreezes. Since water is not available for transpiration in spring until the ground melts and for evaporation until the ground, wetlands, and lakes warm with respect to the atmosphere, the soil thermal balances and the timing of snowmelt (since snow insulates the ground) have a large impact on the seasonal cycle of transpiration. In spring the boreal forest is characterized by very low evaporation and very large sensible heat fluxes off the forest canopy [*Betts et al.*, 1996, 1998a, this issue], which in turn produce deep dry boundary layers. In fall, when the lakes and ground are warm relative to the cooling atmosphere, the situation reverses. Evaporative fraction is high from the conifers and lakes (but not from the deciduous species after leaf-fall). However, net radiation is much lower by the time the surface freezes, sensible heat fluxes are very low, and boundary layers in the fall become very shallow, often capped by stratocumulus.

Since global measurements of soil water and soil temperature are not at present available for analysis, a model must derive them from its own physical parameterizations, using near-surface atmospheric measurements as constraints [e.g., *Douville et al.*, 2000]. However, the coupling between surface fluxes and the convective boundary layer is so tight, that errors in any of the physical parameterizations, whether for subsurface hydrology or thermal transfers, vegetation, cloud, or radiation physics, or for the stable/unstable boundary layer can all interact to give an erroneous diurnal cycle of the mixed layer and, in turn, an erroneous diurnal cycle of convective precipitation in a model. This paper will address the use of data from the BOREAS field program to improve the model formulation of the land-surface interaction for the high-latitude northern forests.

1.2. Validation Tools And Criteria

Several methods of validating land-surface models are available, including (1) off-line comparisons with field data of model landsurface physics, driven by observed near-surface meteorological forcing: these are useful tests of the ability of the model land-surface physics to reproduce the diurnal and seasonal cycle of the surface fluxes, given known meteorological forcing (this is the approach used in section 4); (2) comparison of model output time series at a single grid-point with field data time series: these test the model's ability to reproduce the surface diurnal and seasonal cycles with a fully coupled surface and boundary layer. This has been used primarily to identify existing model errors, as in *Betts et al.* [1998a]; (3) comparison of basin-averaged model surface fluxes with basin-averaged precipitation and runoff (an example of this approach will be shown in section 2): these give insights into key aspects of the model hydrology on regional scales, such as the accuracy of the precipitation in the analysis cycle and in short-term forecasts, and over the diurnal cycle, and the realism of the runoff parameterization and of spatially averaged evaporation. Fields for model parameters like the nudging of soil water, introduced to control model drift, may indicate the structure of errors in the model physics [*Betts et al.*, 1998b,1999b].

Several general issues are worth discussing. One perennial question is field data representivity. Current high-resolution global models have effective grid resolutions of 50 km or greater. Can a grid point time-series from a global model be usefully compared with a time series of field data, which are representative of only a few square kilometers? Field sites are usually located over carefully chosen vegetation types with adequate fetch for eddy correlation measurements to give representative fluxes on the hourly timescale. Globally, there are now 50-100 towers measuring continually the surface radiation, energy, CO_2, and hydrological balances. Most are on towers over forests typical of a region, while some are over grassland and crops. A model represents a grid box average, not a single vegetation type, but if a model has a tiled structure, then a single representative vegetation tile can be compared with the tower flux site data. Even so, exact agreement between measurements and model cannot be expected as flux sites are representative of much smaller areas than a global model grid. Where we have spatially distributed data, as from flux aircraft, these can be used to validate the grid average over the model tiles, as in section 4.3. However, the comparison of point time series measurements with a forecast model analysis (or short-term forecast) is much simpler than, say, with a climate model output, because the forecast model may represent quite well the time rate of change associated with synoptic advective processes.

Given the difficulties of intercomparing observed and model time series, the type of questions we ask are as follows:

1. How well does the model represent the observed diurnal cycle of the soil temperatures, surface temperature, humidity and the surface fluxes? Many model errors can often be identified by looking at the monthly mean diurnal cycle, partitioned if needed into broad subsets, such as disturbed rainy days and undisturbed sunny days.

2. How well does the model represent the corresponding observed seasonal cycles, and in addition the seasonal cycle of soil moisture and temperature, snow depth, snow melt and soil melt, albedo, green-up and leaf-fall?

3. How well does the model represent the transitory response to precipitation events, such as runoff and soil dry downs? This is a

key test of the soil hydrological model and adequate surface water reservoirs.

For forecast model studies we can use the same validation dates (for example, for item 1), while for climate model comparisons, longer-term averages must be compared, such as the monthly (or seasonal) mean diurnal cycles, bearing in mind any differences in the observed monthly precipitation and the climate model precipitation. A study of the diurnal cycle is essential, since day-night differences between the stable and the unstable boundary layer (BL) play a major role. The stable BL, in particular, is often poorly modeled, and nighttime measurements are also less reliable, particularly under low wind conditions. If only daily or monthly averaged model output are analyzed, compensating nighttime and daytime errors can be missed. In addition, the diurnal temperature and humidity range are key climate parameters (although unfortunately the humidity range is not always measured at routine climate stations). Fortunately, as mentioned above, a global network of flux measurements, primarily at forest sites, is coming into being. This is a rich resource of site validation data for global models, including the coupling to the diurnal cycle of CO_2 and the net CO_2 flux. Many sites were installed primarily to study the global carbon balance for climate purposes and not short-term meteorological issues, for which the sensible and latent heat fluxes are the primary surface drivers. What is needed is a major international cooperative effort to process and analyze these data for forecast and climate model validation and development.

2. Initial Insights From BOREAS

The availability of daily forecasts from ECMWF for the northern and southern study areas of BOREAS in 1994 and 1996 meant that scientists became aware in the field of systematic biases in the model. The most vivid example was the underprediction of surface maximum temperature by about 10 K on sunny days in spring, when there was snow on the ground, which suggested the model albedo was too large. There were also suggestions that model cloud cover was too high, suggesting perhaps a positive model bias in evaporation, in comparison with the rather low evaporation rates observed at the forest towers [*Sellers et al.*, 1995,1997]. More detailed comparisons between measurements at the northern study area old black spruce site and global model analyses [*Betts et al.*, 1998a] confirmed the high bias in spring of the surface albedo with snow, and of surface evaporation after snowmelt. The bias in albedo observed in the ECMWF model was so large in spring 1996 (the model had values as high as 80%, when the observed albedos with snow under conifers were generally <20% [*Betts and Ball*, 1997]) that it was decided to change the albedo of the boreal forest with snow before the 1996-1997 winter. The change was implemented in December 1996. *Viterbo and Betts* [1999] show that reducing the boreal forest albedo with snow from typical values near 80% to about 20% had a very large beneficial impact on spring forecasts at high latitudes and reduced the cold model bias in 5-day 850hPa temperature forecasts over eastern Russia by as much as 5 K. However, although a more accurate albedo improved the surface energy balance, in spring it made the model positive evaporation bias worse. In the operational one-layer canopy model, the higher net radiation (given a better but lower albedo) evaporated more snow (which was replaced by the snow analysis), and in addition, the model canopy transpiration remained unaware that the ground had not yet melted.

Positive errors in evaporation feedback and give positive errors in precipitation. This became clear from a study of the surface energy balance and hydrology of the largely forested Mackenzie river basin, which is west of the BOREAS region. *Betts and Viterbo* [2000] showed that the model had a positive evaporation bias on an annual basis of about 60 %. Figure 1 compares precipitation from the 11- to 35-hour operational forecasts with corrected monthly observations for the Mackenzie basin [see *Hogg et al.*, 1996] from September 1996 to December 1997, which spans the winter albedo reduction in the operational model, marked in December 1996. The top panel shows that after this date, the model has considerably more precipitation than the Mackenzie basin observations. The ratio of model to observed precipitation (shown below) peaks at 2.5 in April 1997 (when the snow evaporation peaks in the model) and averages about 1.4 for the rest of 1997. The overestimate of precipitation is consistent with the model having too much evaporation in spring and summer [*Betts et al.*, 1998a] over the boreal forest and with there being significant evaporation-precipitation feedback. During 1996, before the reduction in the model forest albedo with snow in December (which affected the surface energy balance), there is some indication that the model precipitation bias was smaller, although the surface radiation balance was much worse in spring 1996, because of the albedo error. It was clear that the model surface energy balance was still in error in spring 1997, with too much evaporation, leading to a positive precipitation bias, and too little surface sensible heat flux. Although the cold temperature bias at 850 hPa was greatly reduced [*Viterbo and Betts*, 1999], the transfer of heat to the atmosphere was coming through an incorrect mechanism, high evaporation followed by latent heat release, rather than direct sensible heating of a deep BL as observed. The high evaporation also led to an overestimate of cloud cover and a reduction of the amplitude of the diurnal temperature cycle.

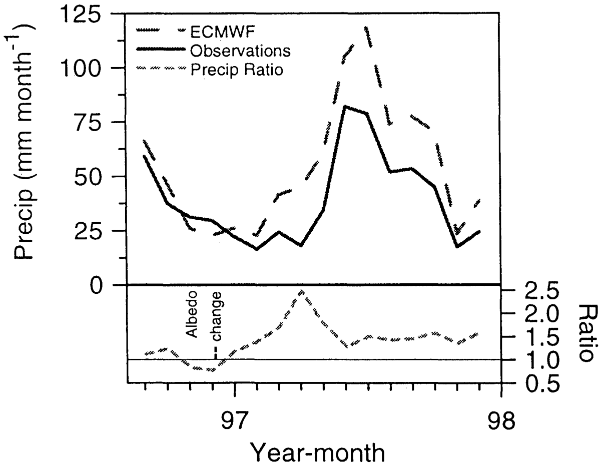

Figure 1. Comparison of ECMWF precipitation with Mackenzie observations and model/observed ratio [from *Betts and Viterbo*, 2000].

Consequently, an effort was undertaken to develop a tiled model with snow beneath "tall vegetation" in order to properly represent the spring energy balance of the system, when there is a very large sensible heat flux off a warm canopy, and very little evaporation, because the water in the ecosystem is still frozen, either in snow under the trees or in the soil [*Sellers et al.*, 1995; *Betts et al.*, 1996]. We will now briefly outline the main features of this new tiled landsurface model.

3. ERA-40 Land-Surface Model

At an operational forecast center, a major constraint on development is that global validation of a land-surface model is difficult. Only a limited number of vegetation types can be represented, and validation data are not available for many regions of the globe. Although the new tiled model is more complex than its predecessor (VB95), it has many fewer parameters than land-surface models such as Biosphere-Atmosphere Transfer Scheme (BATS) [*Dickinson et al.*, 1986], or even the revised simplified biosphere (SiB2) model [*Sellers et al.*, 1996]. Indeed, we attempted, if possible, to represent each important physical process by a single parameter, which could be determined from our data time series or from published process studies. Even so, some parameters for some ecosystems are poorly determined and will be updated as new studies become available, or their values can be derived indirectly from model error fields. The details of the model are given by *Van den Hurk et al.* [2000], so here we only summarize the important revisions. This model is now known by the acronym TESSEL (tiled ECMWF scheme for surface exchanges over land): it became part of the operational forecast system on June 27, 2000.

3.1. Soil Layers

The four-layer soil model, developed by VB95, was revised in August 1996 [*Viterbo et al.*, 1999], when soil water freezing and its thermal impact was added to the operational model. For the ERA-40 model, the soil hydrology and root extraction were made explicitly dependent on the liquid soil water fraction. As a result, there is no drainage through frozen soil layers, so surface runoff now occurs over frozen soil, and vegetation cannot extract water from layers where the liquid soil water is below the permanent wilting point, which removes transpiration when the ground is frozen. All the changes resulting from a better "frozen" physics are, of course, critical at high latitudes.

3.2. Tiles for Vegetation Type and Presence of Snow

The basic differences from the original VB95 scheme are that grid box surface fluxes are calculated separately for the different subgrid surface fractions (or "tiles"), leading to a separate solution of the surface energy balance equation and skin temperature for each of these tiles. This is an analog of the "mosaic" approach of *Koster and Suarez* [1992]. The four tiles in VB95 (bare soil, vegetation, snow, interception layer) are replaced by eight new tiles (bare soil, high vegetation, low vegetation, high vegetation with snow beneath, snow on low vegetation, interception layer, sea ice, and open water). For the moment, the distinction between water and land is still controlled by a land-sea mask, which implies that only six tiles are used over land and two over sea. The global uniform vegetation of VB95 is also replaced by a coverage map of vegetation types, divided in 17 broad categories. The land-surface parameters, such as minimal stomatal resistance, leaf area index, and rooting depth now vary with vegetation type. An exponential root distribution was used following *Zeng* [2001]. Two new environmental controls on canopy transpiration are introduced, no water extraction from frozen soils (mentioned above) and increased vegetative resistance in response to air humidity deficit for tall vegetation (meaning for all forests). For the boreal forests the impact is to eliminate transpiration when the ground is frozen in spring, and reduce summer transpiration. The model for the liquid interception layer is largely unchanged from *Viterbo and Beljaars* [1995], except that the contributions for the leaf area indices of both high and low vegetation types are added to determine the maximum reservoir. As intercepted water evaporates, the interception reservoir occupies an identical fraction of all snow-free tiles.

On top of the soil, a new single snow layer is introduced. The snow scheme includes prognostic equations for temperature, albedo and density and different energy balance equations for high and low vegetation tiles with snow. Evaporation of snow under tall vegetation is reduced by two processes: an additional aerodynamic resistance is included to limit the snow evaporation and a transfer coefficient between snow layer and canopy above controls the vertical energy exchange. Only a small fraction (3%) of the incoming short-wave radiation is allowed to penetrate the canopy and directly warm the snow layer. (This 3% figure for fractional penetration of solar radiation to the surface may be low for subarctic forests and deciduous forests in winter, but at present the model has only a single value for forests at all latitudes). For the boreal forests these changes greatly reduce the snow evaporation and improve the timing of snowmelt, which is largely controlled by the thermal coupling between the canopy and the underlying snow layer. More details are given by *Van den Hurk et al.* [2000].

It is worth commenting on the limitations of the ERA-40 model. While some aspects of the seasonal cycle which are thermally controlled are modeled, and a seasonal cycle of monthly albedo is now included, a seasonal cycle of vegetation has not yet been implemented. This is of greatest importance for crops and deciduous species in midlatitudes. The model at present retains the single soil type and the free drainage model of VB95, which for unfrozen soils does not properly represent surface runoff. Future developments to better represent different soils and hydrological processes are planned.

4. Off-Line Comparisons of ERA-40 Model With Field Data

The off-line development and validation of this model was based on nine field data sets [*Van den Hurk et al.*, 2000]. Here we will illustrate the high-latitude validation of the model using the detailed data sets from the northern study area (NSA) of BOREAS just west of Thompson, Manitoba, in Canada for the time period 1994-1996.

A 30-min. driver data set for nearly 3 years (January 1994 to November 1996) was assembled primarily from two mesonet stations [*Shewchuk*, 1997]. The two sites were typical of the region. One was an old jack pine stand (at 55.93°N, 98.64°W, elevation 282 m) on sandy soil, covered with lichen. The second site was a mixed forest of spruce and poplar (55.8°N, 97.92 °W, elevation 221m) on a clay and peat layer soil with a thick layer of moss. In both cases, tree heights reached 13 m and the measurements were near 19 m, about 6 m above the forest canopy. The measured driver data set variables were above canopy values of wind speed, pressure, air temperature, and specific humidity, incoming short-wave and long-wave radiation and precipitation. Precipitation

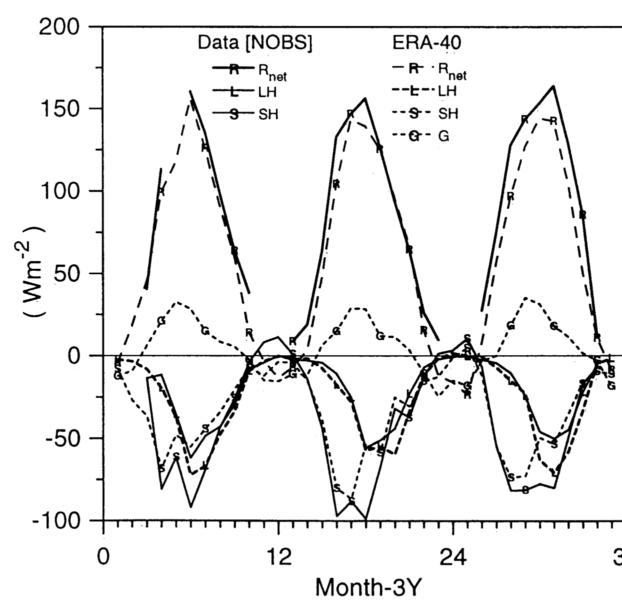

Figure 2. Three-year monthly mean fluxes from data and ERA-40.

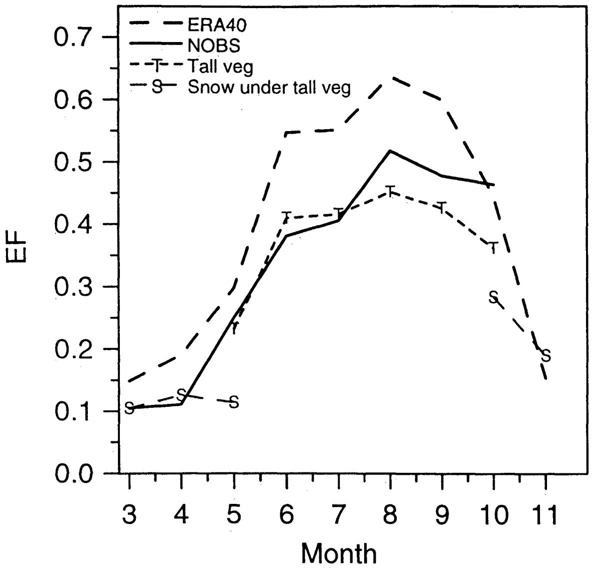

Figure 3. Mean annual cycle of EF from March to November.

required both additional data and special processing. In winter, two weighing Belfort precipitation gauges were used to measure snowfall, and snow depth was measured by an ultrasonic snow gauge. In summer an additional array of 10 precipitation gauges (installed to study the hydrology of a small basin in the NSA) were included, together with rain gauges at four flux tower sites to get a better representative average.

For validation data, we have primarily a nearly continuous set of flux measurements over a black spruce site [*Goulden et al.*, 1997: *Betts et al.*, 1999a]. There are gaps in the data, as the site was unmanned; which is the reason we did not use this site to provide the driver data set (in addition, incoming solar radiation was not measured here, only net radiation). Black spruce is the dominant land cover, although other conifers are present (such as jack pine) and deciduous species such as aspen; in addition to fens and lakes. For the summers of 1994 and 1996, flux measurements are also available over fen and jack pine sites in this study area. Aircraft flew grid patterns to assess the relation between spatially averaged fluxes and those measured at the flux tower sites. [*Barr et al.*, 1997; *Barr and Betts*, 1997; *Mahrt et al.*, 1998], and we will use some of these data here.

4.1. Comparison of Model and Old Black Spruce Site Data

Figure 2 compares the 3-year cycle (from January 1994 to November, 1996) of monthly mean data from the ERA-40 model and the NSA old black spruce site (NOBS). Net radiation (R_{net}) is shown positive (there are a few gaps in the data) and the sensible heat (SH) and latent heat (LH) fluxes are shown negative for clarity. The model ground heat flux (G) is shown positive in summer when it is downward. There was no measured ground heat flux. The monthly mean net radiation in the model is generally biased slightly

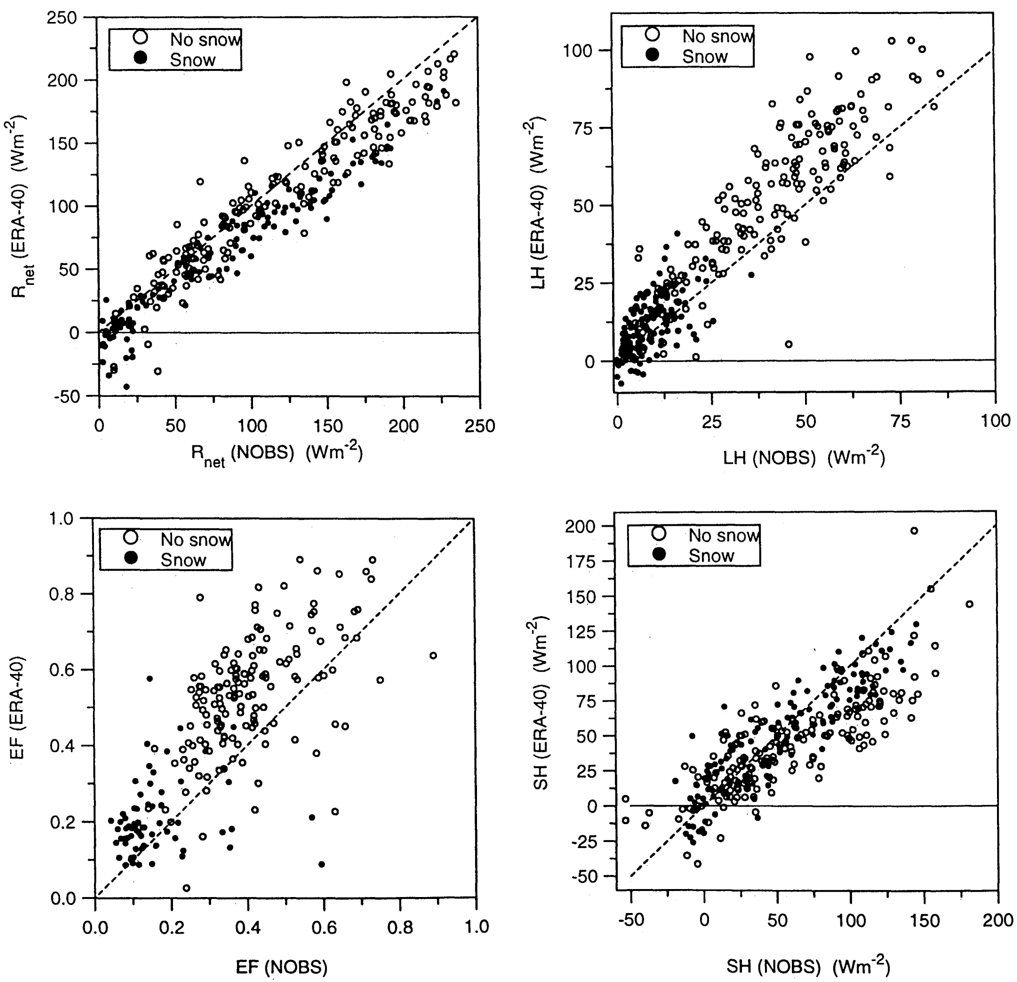

Figure 4. Comparison of R_{net}, LH, SH and EF (clockwise from top left).

low relative to the data. There are two reasons for this. One is albedo: the model albedo in winter with snow under the trees is about 20% for the grid average, while the spruce site albedo probably rarely reaches 0.15 [*Betts and Ball*, 1997]. In summer, the model climate albedo field at this grid point is 0.12, while the spruce site has a very low albedo (< 0.1). Errors in the skin temperature of the model also introduce small biases in the outgoing long-wave radiation. The characteristic feature of the observed SH and LH fluxes (solid lines) at these latitudes is the phase lag in spring of the LH flux. Typically, there is a double maximum in SH in both April and June; while evaporation is very low in April (because the ground is still frozen), and it reaches its summer maximum in June (July in 1996). The model captures this feature well, although the model has generally a lower SH and higher LH than the data at this site.

Figure 3 shows the mean annual cycle of evaporative fraction (EF) from March to November from the model (dashed) and the NOBS flux site (solid), as well as two of the model tiles. EF was derived by first averaging the SH and LH fluxes and then calculating from the means

$$EF = LH/(LH+SH). \tag{1}$$

The "snow under tall vegetation" tile is shown in spring and fall, for periods when the ground is snow covered, and the "tall vegetation" tile is shown from May to October. In the months of partial snow cover (May and October), the tile averages are for the days with and without snow (using a threshhold of 1 mm in snow water equivalent). The model reproduces well the mean rise of EF from spring to fall which is observed, and the two model tiles for the dominant vegetation type agree well with the black spruce observations. Note that the average over all the model tiles exceeds the EF for the NOBS data in almost all months (see later).

Figure 4 compares daily mean fluxes and evaporative fraction (EF), partitioning the data into days with snow on the ground (generally from November to mid-May) and those with no snow. The ttop left panel is the R_{net} comparison: it is clear that the model low bias of R_{net} is larger when there is snow on the ground. The panels on the right compare LH and SH fluxes. The clear separation in evaporation between warm and cold seasons is reproduced very well by the model, with low evaporation when there is snow on the ground, and the ground is frozen. The high bias of the model

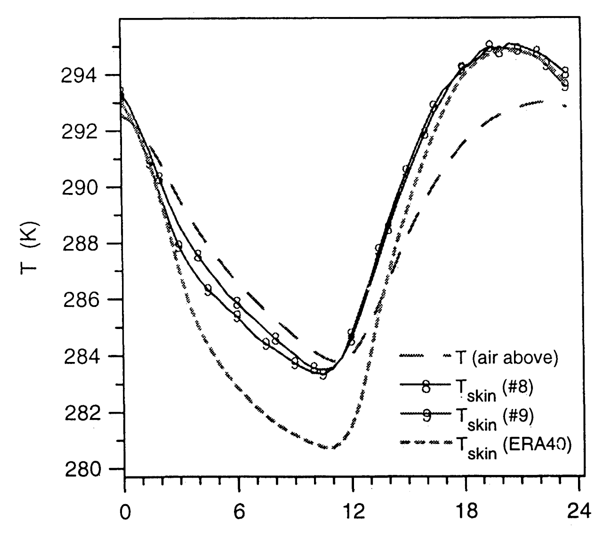

Figure 5. Comparison of mean summer diurnal cycle of air and skin temperature.

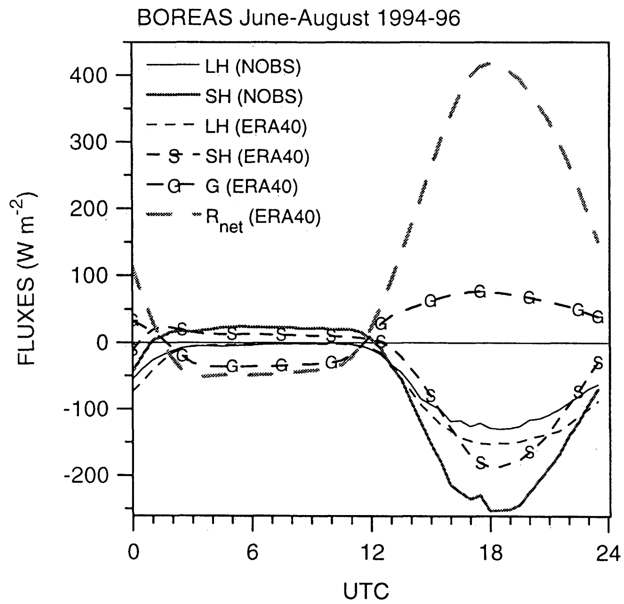

Figure 6. Comparison of mean summer diurnal cycle of surface fluxes.

evaporation (which is a grid average; see section 4.3) in comparison with the black spruce site is mainly in summer. The correlation between model and observed LH fluxes is high. The SH comparison shows a low model bias in the warm season, with less bias in the cool season. The final panel of EF is a useful summary, showing that although the model properly distinguishes between the warm and cold seasons, EF in the model, though well correlated with the black spruce site data, is generally higher. The few outliers in winter (when the data show high EF and the model low EF) are primarily days of snowfall: perhaps because the model has no model for the evaporation of snow on the canopy (the interception reservoir is only a liquid reservoir).

4.2. Verification of Diurnal Cycle

In these off-line comparisons the radiative and atmospheric forcing data are specified. However, the diurnal cycle of the surface heat fluxes and skin temperature of the model are independent checks on the model physics. We shall show two illustrations of this. Figure 5 shows the mean summer diurnal cycle of air temperature (specified) above the canopy, the radiometric skin temperature from two nearby BOREAS NSA mesonet sites (labeled 8 for the site at Thompson and 9 for the northern old jack pine site), and the ERA-40 model skin temperature. A characteristic of these forest sites is that the canopy skin temperature is quite closely coupled to the air temperature, even at night. Although the model skin temperature has a tighter coupling to the ground temperature at night than during the day [*Van den Hurk et al.*, 2000], the model skin temperature drops 2 K farther than the observed skin temperatures at night. In daytime the skin temperatures agree well, although there is a small lag after sunrise, while the model skin temperature rises from its low value. What physics is poorly represented in the model that in nature prevents the drop of skin temperature at night? Figure 6, which shows the corresponding summer diurnal cycle of the model and observed heat fluxes, gives a possible clue. The SH and LH fluxes are plotted negative in the daytime (local noon is about 1800 UTC). First, note that the primary balance in the model at night is between outgoing net radiation and ground heat flux. Unfortunately, we do not have ground heat flux measurements at night, and percent errors in the measurement of R_{net}

at night are larger than in the daytime, so we have no good observational check on this. At night the LH in both model and data are small, as expected. However, the observed SH flux at night, while small, is greater than in the model. This is significant as the flux estimates in the stable boundary layer at night are underestimates except at higher wind speeds [*Goulden et al.*, 1997]. It seems likely, as implied by the smaller drop in observed skin temperature at night in Figure 5, that the actual thermal coupling between the forest canopy and the atmosphere is much larger than in the model. This suggests that the downward SH flux in the model under stable conditions at night is probably too small, perhaps by a factor of 2.

4.3. Comparison of Model, Tower and Aircraft Data

In this section we show a comparison for the BOREAS NSA between the ERA-40 off-line simulation, the NOBS site, and spatially averaged data from flights of the Canadian Twin Otter aircraft. Two types of aircraft patterns are included. The primary one was a grid pattern, consisting of 9 legs 16 km in length, which mapped a square of size 16 by 16 km, located over the tower flux sites of the NSA. This pattern was flown twice, the second time with a reverse heading. This grid gives a representative average for a 16 by 16 km area [*Desjardins et al.*, 1997; *Ogunjemiyo et al.*, 1997, 1999]. This area is smaller than an ECMWF grid square, but it is still much larger than the flux footprint of a typical forest tower and includes a much wider spectrum of vegetation cover types. On some days, instead of the grid pattern or in addition to it, repeated flights were made past the tower flux (TF) sites over patches of relatively homogeneous forest. These runs were typically 10 km in length but were repeated 6 by 8 times to give a representative flux. These TF runs were flown past the NOBS tower, the NSA old jack pine tower and the young jack pine tower. No run was made past the northern fen site, because of limited fetch, but instead, a site was chosen for repeated runs, where the vegetation was recovering from a recent burn (with characteristically more deciduous vegetation). These aircraft patterns took 2-3 hours to fly and were generally in the 1600-1900 time period (local noon is near 1800 UTC). We generated averages from the model off-line run and the NOBS tower data for time periods corresponding to the aircraft pattern times. For days when the aircraft did not fly we show the 1600-1900 UTC average of the NOBS data.

Figure 7 shows the model-data comparison of EF for parts of three intensive field campaigns (IFCs) in 1994, while the Twin Otter (TW) was in the NSA. The first " IFC-1," starting on day 152 (June 1), is shown on the left, and the following IFC-2 and IFC-3 are later in the summer. For all averages, we first average the fluxes before computing an average EF, as in equation (1). The fractional variation of net radiation between different vegetation types is much smaller than the variation in EF [*Betts et al.*, this issue], so that EF gives a useful picture of the partition of the available energy into the separate sensible and latent heat fluxes. The model and the NOBS data track quite well, with the model (dashed) usually higher than the old black spruce site (as shown in Figure 3). The peaks in EF are days with low net radiation. The letter A marks the averages for the Twin Otter aircraft flights for just the days when either the grid pattern was flown or a representative set of the TF and burn site tracks. While the aircraft spatial average tracks the NOBS tower site reasonably well, it is higher by about 0.13, and the aircraft is also generally higher than the model, especially in midsummer. In addition, we show the $EF_{tallveg}$ of the ERA-40 tile representing tall vegetation for this grid square, excluding days when the interception reservoir fraction exceeds 20%. This is lower than the ERA-40 tile average but generally closer to the NOBS data. Why is the aircraft spatial average of EF higher than the NOBS site? One reason is that the EF from the NOBS site is the lowest of all the flux tower sites.

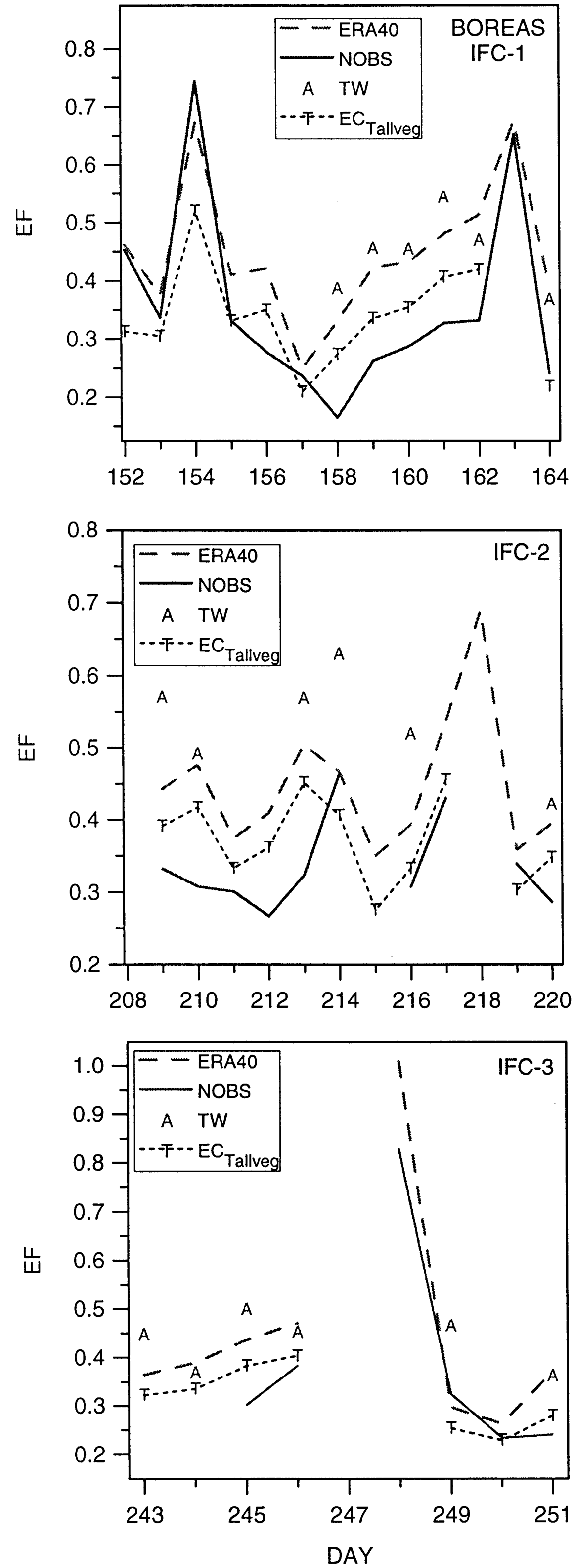

Figure 7. Near-noon comparison of model with NOBS and aircraft data

In addition, the aircraft spatial averages include a significant fraction of deciduous forest species, which have a much vegetative conductance and evaporative fraction in midsummer than all the conifer species (see comparison by *Barr et al.*, this issue). *Desjardins et al.* [1997] also speculated from the study of the flight

Table 1. Comparison of BOREAS NSA Evaporative Fraction Estimates (Figure 7)

Source	EF (Mean)
ERA-40	0.42
Twin Otter	0.47
Barr et al. [1997]	0.45
$EC_{tall\ veg}$	0.34
NOBS	0.33 (0.31)

tracks past individual towers (which also show higher aircraft EF than at the towers) that the tower sites might have been chosen in drier locations.

Table 1 summarizes these comparisons. The ERA-40 grid-point average (an average over all tiles) is a little lower than the average from the Twin Otter flights. *Barr et al.* [1997] made a separate estimate of spatially averaged EF from boundary layer rawinsonde budgets for the BOREAS study areas. Their estimate, shown in line 3 of Table 1, for undisturbed days (when sequential rawinsondes were launched) over all IFCs, was a little less than that from the Twin Otter flights. The last two values are averages for the tall vegetation tile (for the days shown in Figure 7) and the black spruce flux site for all days when there were data. The black spruce value in parentheses excludes the four days for which the tall vegetation tile was not shown in Figure 7. Our conclusion from Table 1 is that given the spread in EF between spatial averages and the NOBS site (the dominant vegetation type), which is a measure of the uncertainty in area average flux, the ERA-40 model and its tall vegetation tile are reasonably representative of the BOREAS northern study area.

4.4. Stratification by Wet Surface Index

It can be seen from Figure 7 that the ERA-40 land-surface scheme has some skill in reproducing the day-to-day variability in EF near local noon. A characteristic of the boreal forest (particularly the spruce forest) is a widespread surface moss layer, which is able to store and evaporate water for several days after rain events [*Price et al.*, 1997]. *Betts at al.* [1999a] used a wet surface index to stratify days based on rain amounts on preceding days. This wet surface index (WS) gives a qualitative indication of the surface water storage. Zero means that for at least 5 days, less than 1mm of rain fell, and the moss layer (as well as the canopy) is probably quite dry. Index 5 means that at least 5 mm fell the preceding day, and the surface, including the moss layer, is likely to be quite wet. Indices 1-4 represent intermediate conditions of either less rainfall or dry downs from rain events. They found that the diurnal cycle of relative humidity (RH) and the closely related pressure height to the lifting condensation level (P_{LCL}) varied with this index WS. Figure 8 shows the mean diurnal cycle for the driver data, averaged for the warm season months of June to September 1994-1996 and stratified by WS. The data are plotted against P_{LCL}, and the RH scale shown is a slight approximation [see *Betts et al.*, this issue]. As the surface index gets wetter, P_{LCL} decreases in the afternoon (that is, the mean cloud base gets lower) and RH increases. On the very wet days (with WS=5), cloud base is low and afternoon humidity is large. How do the surface fluxes respond to these precipitation differences and to this diurnal cycle of humidity contained in our driver data? The two panels in Figure 9 show the stratification of mean daytime EF using the same WS index for the NSA OBS site (top panel) and the ERA-40 run (bottom) for the same period, June to September 1994-1996. Both show a similar increase in mean EF as RH

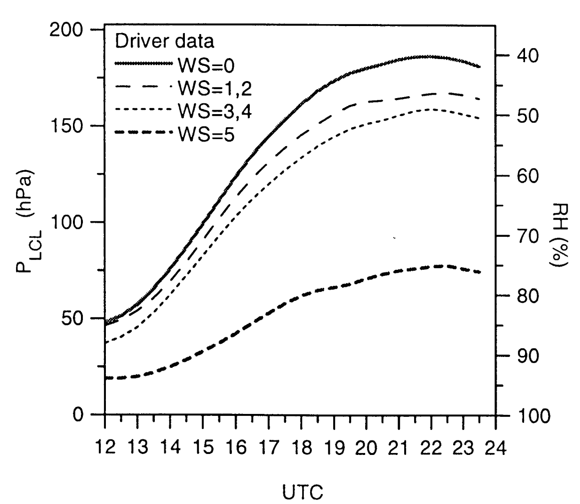

Figure 8. Diurnal cycle of driver data as a function of wet surface (WS) index.

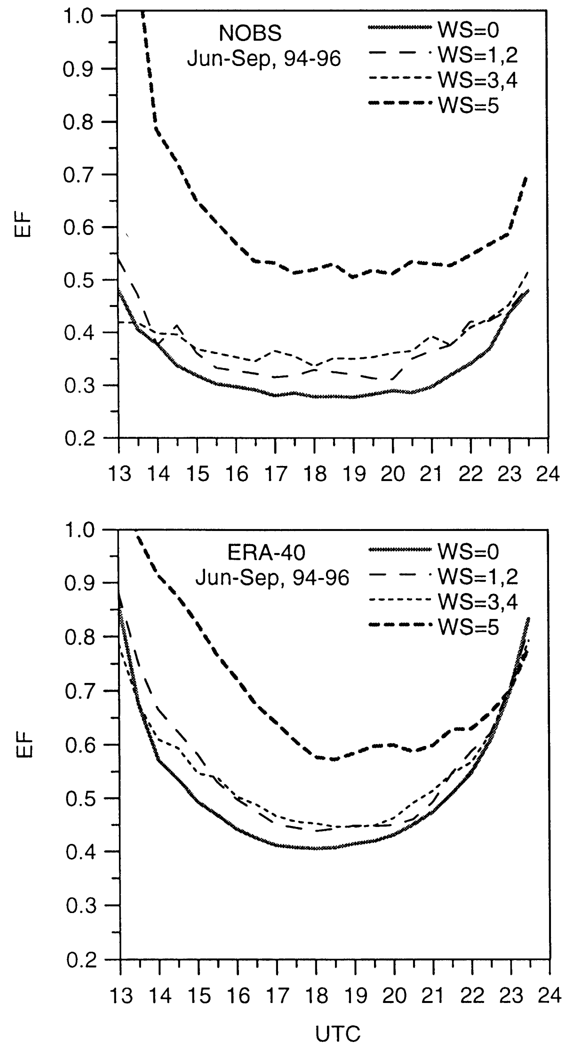

Figure 9. Daytime diurnal cycle of EF, stratified by wet surface index for NOBS and ERA-40.

Table 2. Hydrological Balance of BOREAS Off-line Runs (Units in millimeters)

Year	Precipitation	Evaporation ERA-40 (OPS)	Surface Runoff ERA-40 (OPS)	Total Runoff ERA-40 (OPS)	NSA NW1 Streamflow
1994	409	318 (483)	113 (0)	156 (10)	113
1995	467	288 (415)	74 (0)	144 (8)	97
1996	459	292 (478)	93 (0)	160 (10)	169

increases (P_{LCL} decreases), but all the ERA-40 curves are shifted to higher EF than at the NOBS site, which is consistent with earlier figures. However, the ERA-40 model has no moss layer, and it is reproducing the trend with WS through a dependence of vegetative resistance on soil water, and for large WS, evaporation from the interception reservoir. On the landscape scale where several forest species are involved with different soils (as well as wetlands) this approximate model treatment may be satisfactory, until more detailed validation suggests otherwise. The model at present has a single global soil type (a loam), which is not very representative of the NOBS site, where there is a 30 cm largely organic soil layer over clay.

4.5. Hydrological Balance

Some of the major terms in the mean annual hydrological balance of the ERA-40 off-line runs for BOREAS are shown in Table 2. The figures in parentheses are from parallel runs using the operational land-surface model, with the same driver data (so both runs have the same precipitation). Using the ERA-40 model, total evaporation has been reduced by 35 %, and there is now significant surface runoff. For comparison with the total model runoff, column 6 shows the observed streamflow for the NW1 streamgage on the Sapochi River in the NSA at 55.91°N, 98.49°W. We have used the basin drainage area of 433 km^2 to convert the streamflow to millimeters per unit area. This small basin includes the old black spruce site and is representative of the NSA. The ERA-40 total runoff is now much closer to the observed streamflow, and correspondingly, since these off-line runs are being driven by observed precipitation, we can conclude that the much reduced model evaporation is now much more realistic on an annual basis. However the ERA-40 surface runoff in spring is several weeks earlier than observed, as shown in Figure 10. Part of this error is probably due to the timing of snowmelt [see *Van den Hurk et al.*, 2000], which coincides in the model to snowmelt at the old jack pine mesonet site, where melt occurred about a week earlier than at the spruce and poplar site. The streamflow in spring is also delayed about two weeks by the freezing and thawing of meltwater as it percolates first through the snowpack and then though the soil [*Marsh and Woo*, 1984; *Quinton and Marsh*, 1999], and these processes are not represented in the model. In addition, the midsummer peak in streamflow in 1995, following heavy rain in early August, is not properly represented in the ERA-40 model, because with the present scheme, surface runoff only occurs over frozen ground.

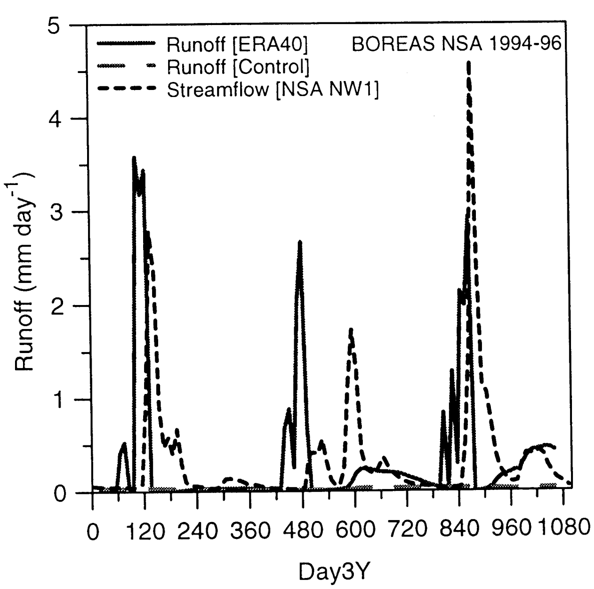

Figure 10. Comparison of runoff from two land-surface models with observed NSA streamflow.

5. Global Impact on High-Latitude ECMWF Model Climate in Spring Shown by Long Integrations

Following the single-column offline tests of the new land-surface model, year-long global integrations of the ECMWF model were completed with both the operational and the ERA-40 tiled land surface schemes, initialized from the operational model on October 1, 1998. In these long integrations the wind and temperature fields above 500 m were relaxed toward the operational analysis with a 6-hour timescale to constrain long-term drifts. This is a computationally inexpensive alternative to full data assimilation, which shows the main features of the land-surface impact in midlatitudes: it was used by *Viterbo et al.* [1999]. We have no global validation data, but we can compare the different response of the two land-surface models. Plate 1 shows the large impact of the new land-surface scheme over the northern forests in spring in these off-line global runs. The panels are March, April, May 1999 average difference fields between the ERA-40 tiled land-surface scheme and the operational scheme. The top panel shows the large drop of the surface latent heat flux in spring of 20-30 W m^{-2} over the boreal forest (with similar increases of surface sensible heat flux, not shown). This corresponds to a reduction in the evaporation in spring of nearly 1 mm d^{-1}. The middle panel shows the resulting reduction in low-level cloud fraction of the order of 25%, resulting from the reduced evaporation. This suggests that the subjective impression, during the 1996 BOREAS field campaign that the ECMWF operational model was giving too large a cloud fraction may well have been correct. The bottom panel shows the consequent large increase in the mean difference between maximum and minimum temperature of the order of 6 K, which results from the reduced evaporation and cloud cover.

6. Conclusions

This paper has reviewed how the data from BOREAS impacted the development of the land-surface parameterization in the

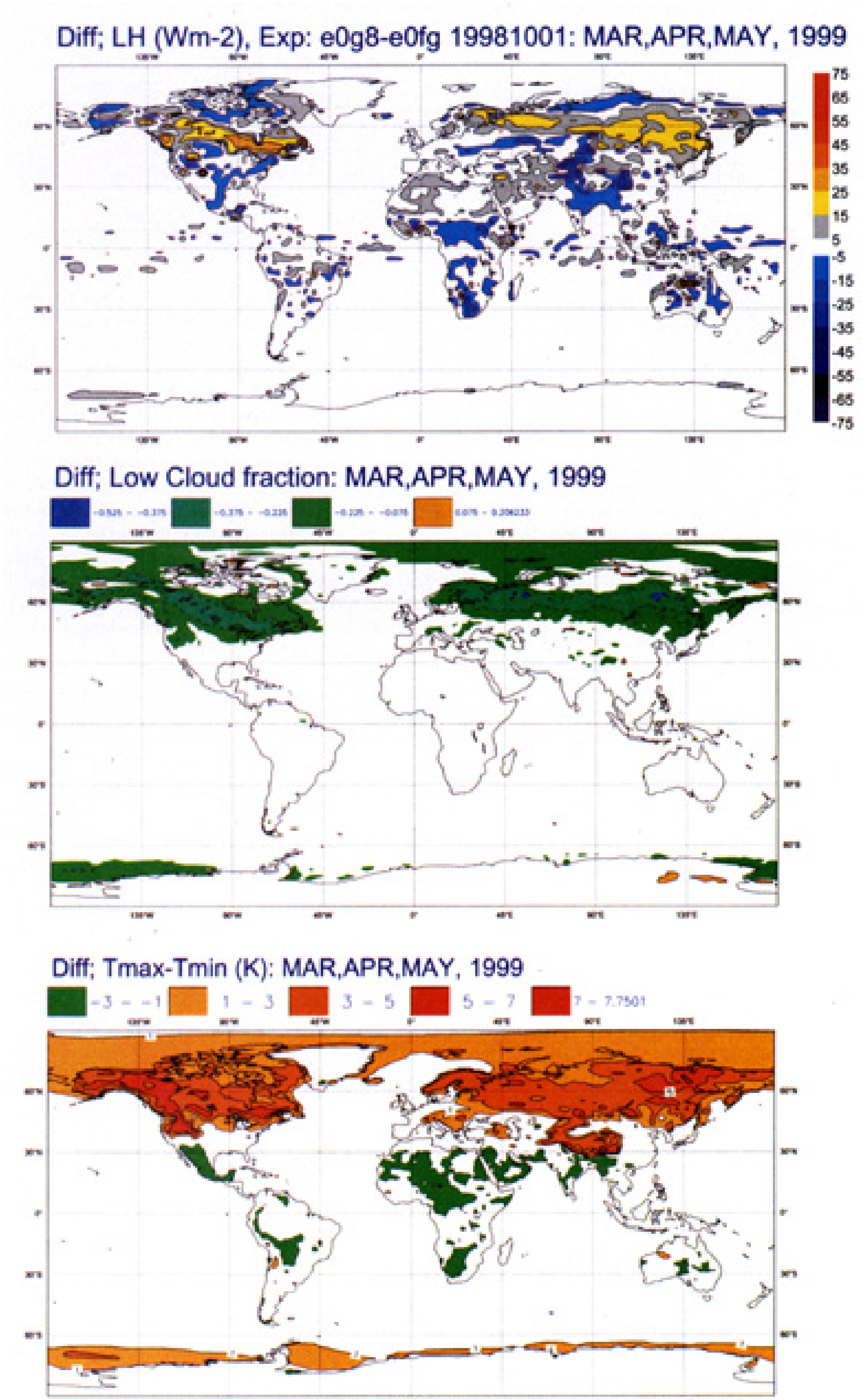

Plate 1. March, April, May 1999 average differences between ERA-40 and operational land-surface model for integrations starting October 1, 1998, for surface latent heat flux (top), low cloud cover (middle), and T_{max}-T_{min} (bottom).

ECMWF global forecast system and improved the model climate at high latitudes. We have shown how point-scale data from a flux tower can be used to validate the model parameterizations of observed physical processes on both diurnal and seasonal timescales. Some aspects of the model surface hydrology and its interactive with the precipitation physics and model spin-up can be assessed only on larger scales, but much model development can be done with long time series at representative flux sites. Data do not exist on all spatial scales, and all data have error signatures, some well known and some not, so care must be taken to distinguish model biases from uncertainties in the data or its representivity. Given, however, the still relatively primitive representation of land surface physical processes in our numerical forecast models, the first test is to ask whether the processes that can be seen in the data are represented in the model on the diurnal and seasonal timescales. The corollary of this is that we tried not to add complexity, if it could not be validated. As a direct result of BOREAS, the representation of several surface processes was greatly improved over the boreal forest.

The introduction of separate tiles for tall and short vegetation meant that the forest could be represented with more realism. In winter the albedo with snow under the trees was greatly reduced, and the introduction of a prognostic snow model, with its own energy balance under the canopy, meant that evaporation of snow in winter and spring was also greatly reduced. An improved handling of liquid and frozen soil water meant the evaporation from frozen soil water is not allowed, and surface runoff occurs with snowmelt over frozen ground. All these improvements to the handling of the freeze-thaw processes significantly improve the seasonal cycle at high latitudes. This is particularly marked in late spring, when the model now has large sensible heat fluxes off the forest canopy and reduced evapotranspiration. Trees in the model now have larger more realistic unstressed vegetative resistance and also have a stress factor for vapor pressure deficit, as well as one for low light levels; both of which reduce summer transpiration to the lower levels observed over the boreal forest. Some processes that we observed have not yet been implemented, such as lower vegetative resistance under diffuse light, and the increased evaporation when the surface moss reservoir is wet [*Betts et al.*, 1999a]. The model does not yet have a global soil distribution, which means that it does not represent the organic soils characteristic of the black spruce sites, which have a larger water storage.

It is clear that while there remains more work to be done to improve the land-surface physics in our Earth system models, significant incremental progress is being made. A land-surface data set will be prepared from ERA-40 as part of the second International Satellite Land Surface Climatology Project (ISLSCP-II). Global surface data sets from such reanalyses can then be used to drive and intercompare land-surface models off line, which will lead to further model development.

We have given a scientific perspective on BOREAS, which clearly illustrates the benefits of a tighter integration between a field program and an operational forecast center. In addition, it is worth commenting that the close collaboration and free exchange of data among NASA, AES Canada, and ECMWF was essential to the success of BOREAS. Traditionally, there have been barriers to the free exchange of data and scientific products, even though much lip service is paid to the idea of free exchange. For the parties involved in the BOREAS-ECMWF collaboration, the example of FIFE provided a powerful motivation to work together to overcome any difficulties. As a result, there were real-time benefits to the field experiment, and for the subsequent analysis by cross-fertilization with the global modeling perspective on the land-surface-atmosphere interaction. In exchange, there have been considerable benefits for the ECMWF forecasts, both medium range and seasonal, and a much improved ERA-40 data set for the ISLSCP community, which closes the loop in the science, and prepares the way for further progress. Such tight international collaboration is of critical importance, if we are to understand and model the Earth and its biosphere as a fully interacting system.

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