

THE VALIDATION OF A SNOW PARAMETERIZATION DESIGNED FOR USE IN GENERAL CIRCULATION MODELS

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ABSTRACT

A snow model designed for use in general circulation models is tested using atmospheric forcing derived from observations. Using data for four sites from the former Soviet Union, simulations of snow mass, snow density, snow depth, albedo and net radiation are validated. While we find the model works well in capturing the general patterns of the seasonal cycles and interannual variability in these quantities, there are systematic problems in the modelling of snow density and snow albedo towards the end of each snow season. An analysis of the snow densification parameterization shows that the model is very sensitive to the value of a poorly known parameter describing the temperature dependence of the snow compactive viscosity. Many problems in the simulation of snow stem from the simplification of complex processes occurring over time, such as layering and ripening, by a single bulk snow layer which prevents the model responding dynamically to changes in the forcing. In the simulation of the four sites, we used various methods for downward longwave radiation, and we find that the simulation of snow by the model is very sensitive to this quantity. Climate models do not provide this flux accurately in transitional seasons at high latitudes which suggests that it is probably premature to include very complex snow schemes into climate models. In this paper, we were able to obtain far better simulations by choosing the 'right' method for downward longwave radiation than by tuning parameters or increasing the complexity of the snow parameterization. However, the ability of the model to capture seasonal and interannual variability irrespective of which downward longwave radiation was used is important, and provides some support for the inclusion of complex snow schemes into climate models. © 1998 Royal Meteorological Society.

KEY WORDS: snow; snow processes; land surface models; BASE (Best Approximation of Surface Exchange); Russia; longwave radiation; model sensitivity; snow density

1. INTRODUCTION

Over 50% of Eurasia and North America can be seasonally covered by snow (Robinson *et al.*, 1993) leading to significant spatial and temporal fluctuations in surface conditions. The properties of snow (e.g. high albedo, low roughness length and low thermal conductivity) lead to impacts ranging from the micrometeorological scale (Kukla, 1981) to the global scale (Vernekar *et al.*, 1995). Foster *et al.* (1982) showed that over the Eurasian continent between 18 and 52% of variance in winter temperature could be explained by the extent of snow cover present in autumn and that extensive snow cover over continental areas leads to the development of anticyclonic conditions. More recently, Foster *et al.* (1996) performed an intercomparison of numerous general circulation models (GCMs) and found that while some general patterns in the global snow regime are reproduced, large differences exist in mass, duration and extent of snow cover. These differences emphasise the need to examine the simulation of snow in a variety of land surface schemes used in GCMs in order to explain the disparity between models.

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Other modelling experiments have also highlighted the importance of snow. Gallimore *et al.* (1986) showed that modest model improvements could lead to a significant sensitivity of regional climate to snowcover. Marshall *et al.* (1994) and Marshall and Oglesby (1994) showed that improved snow hydrology caused substantial shifts in the global water storage regime. Yeh *et al.* (1983) found that in addition to increased solar absorption and evaporation, rapid removal of snow cover could decrease the meridional temperature gradient and high latitude zonal wind. Accurate prediction of snow mass, duration and extent is necessary if we wish to predict future climates, especially given the reported high latitude sensitivity to increasing greenhouse gases.

The parameterizations of snow processes used in GCMs range greatly in terms of complexity. At the simplest extreme, the model used by Manabe (1969) represents snow as a single layer with a prescribed albedo. At the other extreme, multi-layer schemes such as those of Loth *et al.* (1993) and Lynch-Steiglitz (1994) include advanced albedo calculations based on snow age or temperature and explicitly model the metamorphism of snow as well as liquid water storage and transmission in the snowpack. Even greater complexity is available in the models of Anderson (1976) and Jordan (1991), but these are too numerically intensive for use in current GCMs. Models of intermediate complexity do not include all snow processes. Validation of these models' ability to simulate snow as well as validation of their ability to simulate components of a snow regime such as albedo or snow density is clearly important in attempts to develop improved models, particularly in the context that snow schemes used in GCMs are known to simulate different accumulation and seasonal coverage of snow even if the atmospheric forcing is prescribed (Pitman *et al.*, 1993a).

Recent work by Loth *et al.* (1993), Douville *et al.* (1995), Robock *et al.* (1995), and Yang *et al.* (1997) all show the ability of specific land surface schemes to simulate snow cover. They also demonstrate how differing levels of simplification can affect the simulation of snow. In this paper, a land surface scheme BASE (Best Approximation of Surface Exchange, Desborough, 1997, Appendix B) is validated using the same data used by Douville *et al.* (1995), Robock *et al.* (1995), and Yang *et al.* (1997). This paper therefore provides evidence on the ability of one land surface scheme to simulate snow, supporting work performed with other land surface schemes. However, this paper also highlights the role of snow densification as an important area deserving of further attention in both modelling and observational efforts.

In Section 2 we briefly describe the experimental design and the data used. Section 3 describes the validation of simulations by BASE against available observed data. Finally, Section 4 discusses the implications of the results and concludes.

2. DATA AND EXPERIMENTAL DESIGN

2.1. Forcing data

The data used in this study have previously been used by Douville *et al.* (1995), Robock *et al.* (1995), and Yang *et al.* (1997). The data are described by Robock *et al.* (1995) and are not reported here in detail. For the years 1978–1983 measurements at several hydrometeorological stations within the mid-latitude zone of the former Soviet Union were taken over a grass covered plot. As noted by Robock *et al.* (1995), these plots were flat pieces of land with an area greater than or equal to 0.10 ha; the soil type was representative of the main soil type and landscape of the region and did not differ significantly from the prevailing soil type and landscape of the climatic zone; and the mean depth of the water table and its seasonal variations were typical for a large area.

Four stations are used in this paper: Yershov (51.4°N, 48.3°E), Uralsk (51.3°N, 51.4°E), Khabarovsk (48.5°N, 135.2°E) and Ogurtsovo (54.9°N, 83.0°E). All sites are within 10° of latitude, but are widely spaced longitudinally. In each case, vegetation cover was a grass (naturally occurring or artificially grown) although surrounding vegetation may have been forest, agricultural land or grass (Robock *et al.*, 1995). The grass covering was allowed to grow uninhibited and was rarely cut (C.A. Schlosser, personal

communication, 1995). We also ran for two additional sites (Kostroma, 57.8°N, 41.0°E) and Tulun (54.6°N, 100.6°E) but we have not included results from these stations because they do not affect our conclusions.

In order to be used in off-line experiments, BASE requires the input of atmospheric forcing data every 30 min and the specification of various parameters which characterise the land surface. BASE requires the input of precipitation, incoming shortwave radiation, air temperature, windspeed, atmospheric pressure and specific humidity all of which were available from the meteorological records, recorded at three hourly intervals starting at midnight Moscow time. Actinometric data were recorded at 00:30, 06:30, 09:30, 12:30, 15:30 and 18:30 h local time.

BASE also requires downward longwave radiation ($L\downarrow$) but the original meteorological data set did not include this variable which had to be calculated separately. Several methods were employed (Appendix A), resulting in differences of up to 60 W m^{-2} in $L\downarrow$. The semi-empirical methods used were proposed by Monteith (1973) (MONT), Satterlund (1979) (SATT) and Idso (1981) (IDSO). Net radiation (NET), available from the observed data, was also used to derive an estimate of $L\downarrow$ given:

$$L\downarrow = R_{\text{net}} - S\downarrow(1 - \alpha) - \epsilon\sigma T_{\text{g}}^4 \quad (1)$$

where R_{net} is the observed net radiation, $S\downarrow$ is the observed downward shortwave radiation, α is the observed surface albedo, ϵ is the surface emissivity (not measured), σ is the Stefan-Boltzmann constant and T_{g} is the observed surface temperature. Observations of T_{g} were taken using an unshielded thermometer placed on a grass surface (i.e. swept clear of snow during winter) and are therefore unlikely to be an accurate representation of the radiating surface temperature. Finally, R_{net} was measured using an unreliable pyranometer (Vinnikov, personal communication). These problems indicate that $L\downarrow$ derived using net radiation cannot be considered more reliable than using semi-empirical methods.

Figure 1 shows the range in $L\downarrow$ derived using the four methods used in this paper. The differences between the four methods vary through the annual cycle with the four methods giving different $L\downarrow$ during winter, but IDSO and NET converge during summer to consistently higher values than MONT and SATT. Since we do not know which is the most accurate estimation of $L\downarrow$ we have conducted all the subsequent simulations with all four $L\downarrow$ methods. The sensitivity of BASE's simulation of snow to differences in $L\downarrow$ is discussed later.

2.2. Validation data

A variety of observational data describing snow characteristics are available for the sites used in this paper. Available snow based measurements, taken every 10 days, include snow water equivalent, snow density, fractional snow cover and snow depth near the station. Barry *et al.* (1994) note that snow depth measurements are taken along transects close to the station so providing a consistent average value. Fractional snow cover observations correlated poorly with both observed depth and observed albedo. Whether this is due to problems with the scale of observation or related to the definition of the quantity is unknown.

Overall, these data provide relatively high resolution observations across a wide longitudinal range of the former Soviet Union. The observations cover a large number of years and are probably as consistent in quality as any available long term snow data set. The ability of BASE to simulate the individual seasonal cycles as well as interannual variability represented in the 6 years of data therefore presents a rigorous test of the model.

2.3. Experimental design

In each simulation, BASE was initialised using estimates derived from observations. The model then used the first year of observed data to recursively spin-up. Having reached equilibrium (as defined by Chen *et al.*, 1997), BASE was then forced with the full 6 years of observed meteorological data with a 30 min timestep (using a cubic spline interpolation for the forcing data following Chen *et al.*, 1997). This paper describes the results from these 6 years. A recursive spin-up was used to negate the possibility of

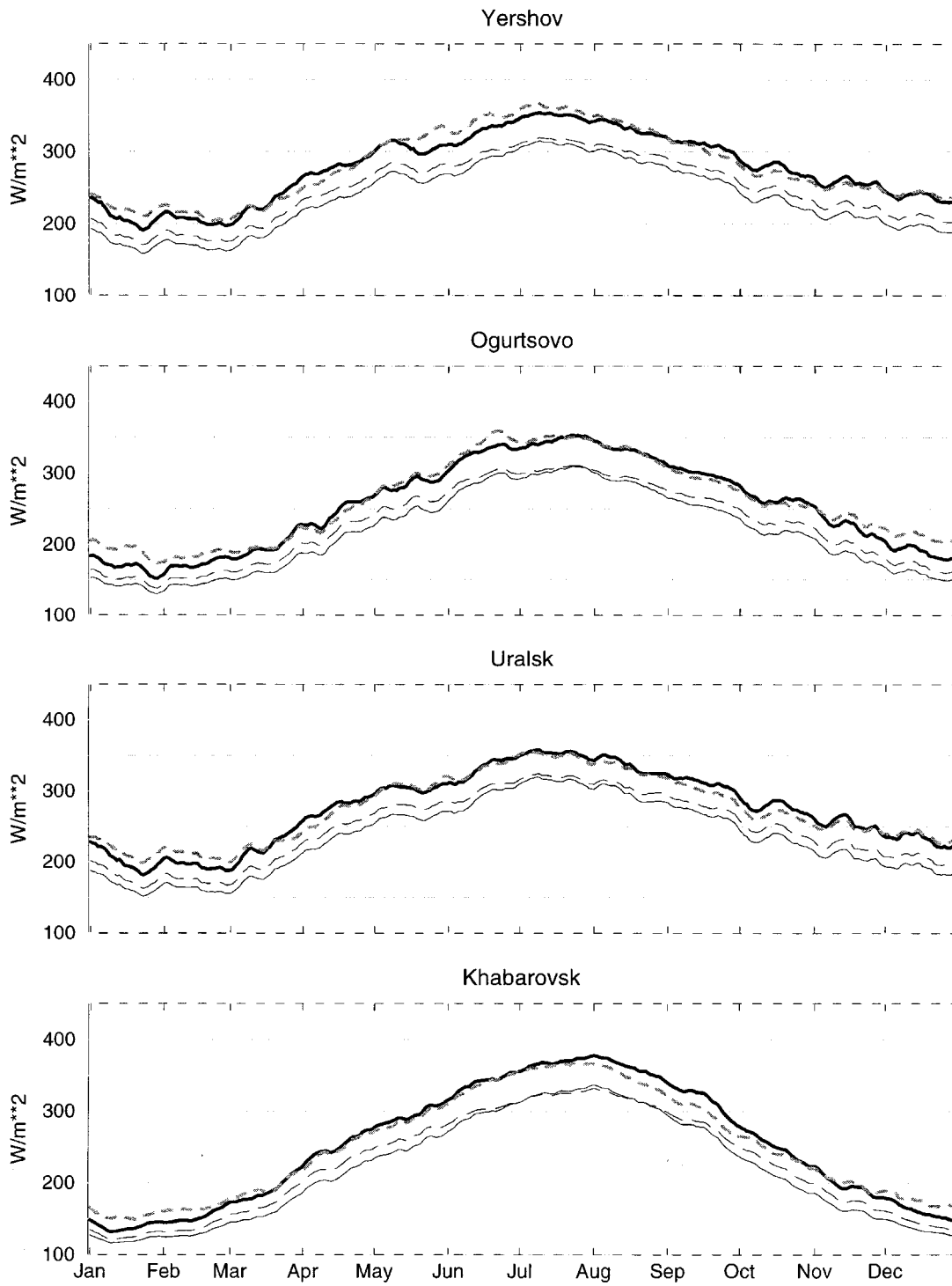


Figure 1. Average annual cycle in downwelling infrared radiation (W m^{-2}) derived using the methods described in Appendix A for the four stations used in this paper. The thick solid line is for the IDSO method, the thin solid line is for the MONT method, the thick dashed line is for the NET method and the thin dashed line is for the SATT method (see Section 2 and Appendix A)

analysing model output which was unduly influenced by initialisation and to ensure that all quantities within the model were internally consistent. In terms of the analysis of the simulation of snow by BASE, the decision over whether to recursively spin-up or to initialize and just simulate the 6 years is not important. However, these simulations were conducted using an identical methodology to Slater *et al.* (1998) where soil moisture was investigated and the spin-up method was important.

Since the observed data apply to a site (effectively a point in a GCM) the spatial heterogeneity parameterization, dealing with fractional snow cover included in BASE to describe the relationship between grid square average snow depth in metres d_n and grid square fractional snow cover (A_n) on spatial scales of 300 km \times 300 km, was modified from the Equations (B4)–(B7) to:

$$A_n = \text{minimum of } [5d_n, 1] \quad (2)$$

This relationship was estimated from the observed data by plotting observed snow depth against observed snow cover. This relationship was derived specifically for these simulations and is not intended for use in GCMs. Due to our limited confidence in this relationship and in the observation of snow cover we decided not to attempt to validate snow cover in the simulations we performed.

Finally, we chose to switch between rainfall-snowfall using an air temperature criteria of 273.16 K. This is the default value used in BASE when used off-line. We chose this value independently of Yang *et al.* (1997) but we note that this was the value they recommended for these stations.

3. VALIDATION OF SNOW VARIABLES

3.1. Snow mass or snow water equivalent

The ability of a land surface model to simulate snow mass or snow water equivalent (SWE) is important in determining the amount of precipitation which will be lagged in the hydrological cycle while the simulation of excess snow mass will tend to cause an overestimation of snow cover and thereby affect seasonal albedo and the amount of energy available to turbulent and radiant energy exchange.

Overall, the general patterns of SWE are simulated well by BASE, at least to the temporal accuracy of the observations (10 days). Figure 2 shows that the inception of the season is simulated very well every year for all stations although this is due in part to the insensitivity of the model to the specification of criteria for precipitation occurring as snow rather than rain at this time of year. Figure 2 does, however, indicate that BASE is able to accumulate new snow onto a snow-free surface accurately.

Figure 2 shows that large differences in SWE occur as a result of using different methods for $L\downarrow$. At Yershov and Uralsk, up to a factor of two difference is exhibited in the SWE for different $L\downarrow$ forcing (e.g. years 1982/1983). Since the same amount of precipitation occurred as snow under each $L\downarrow$ forcing, these results indicate that ablation is very sensitive to the method used for $L\downarrow$. Using the MONT and SATT forcing, which produces less $L\downarrow$, BASE overpredicts the observed SWE by almost a factor of two. In contrast, if the IDSO method is used, the model simulates excessive mid-season ablation while if the NET method is used, BASE produces a near perfect simulation. While the simulation for Yershov represents the extreme sensitivity to $L\downarrow$, the sensitivities simulated by other stations are considerable. However, at times when the different $L\downarrow$ methods predict similar SWE, BASE tends to simulate SWE accurately (e.g. early in the winter of 1979 at Yershov, during the winter of 1979/1980 at Uralsk and during the 1980/1981 winter at Khabarovsk). These generally represent rapid accumulation periods where differences in $L\downarrow$ are not able to dominate the simulation.

The inception of the snow season is captured well by BASE, even to the extent of resolving early season fluctuations, e.g. Yershov (1979/1980 and 1980/1981) and Ogurtsovo (1982/1983). There is evidence from Figure 2 that snowpack ablation is time lagged in BASE and the reduction in SWE is simulated with less accuracy than the early winter accumulation. Unfortunately, the resolution of the observed data for determining the end of the snow season is not ideal, but where observations do exist, BASE tends to overpredict SWE (e.g. the spring of 1978 and 1983 at Yershov, 1979 and 1981 at Ogurtsovo, 1981 at

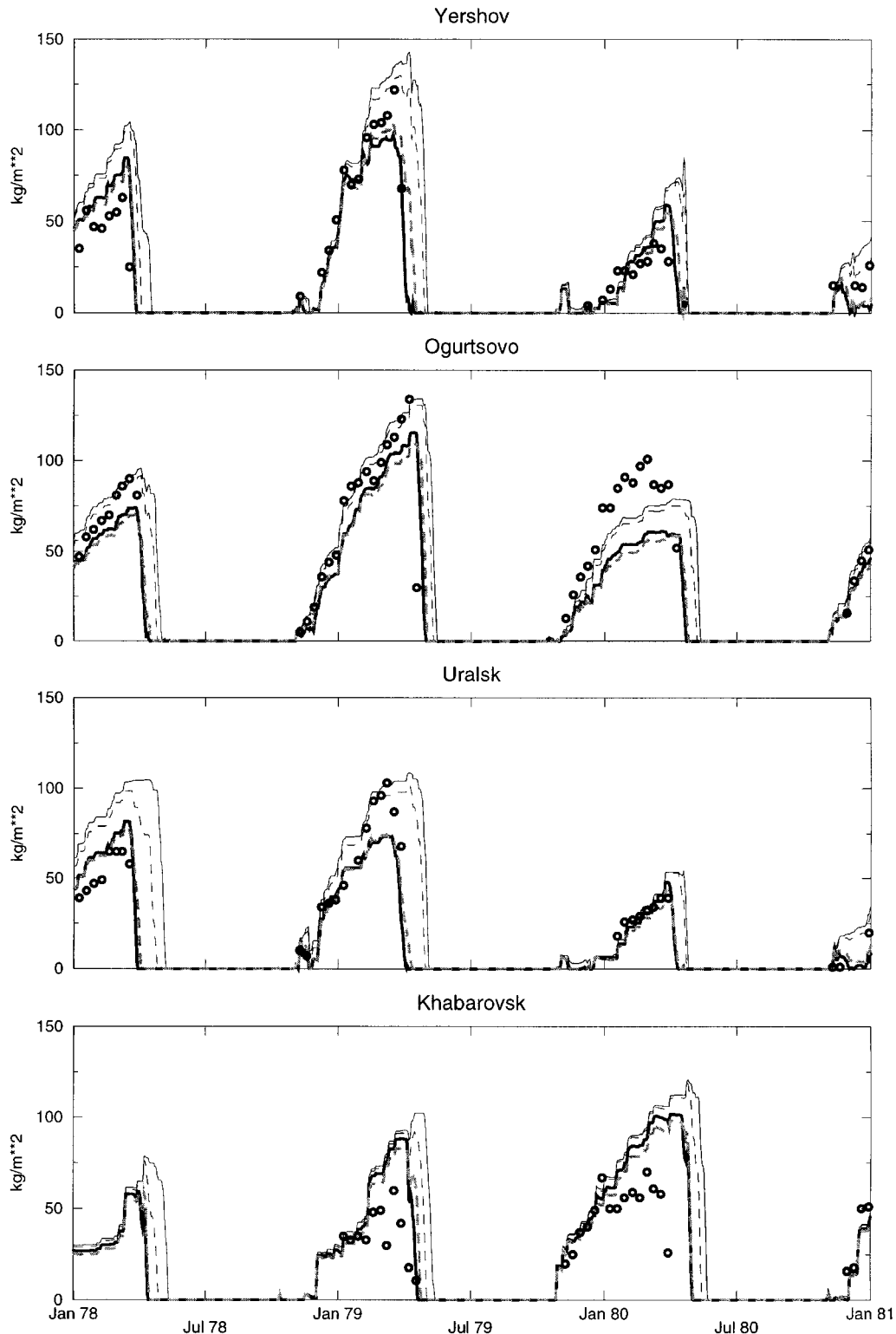


Figure 2. Snow water equivalent (kg m^{-2}) simulated by BASE for the 6 years of observations. Where available, the observed values are shown by open circles. The four curves are for the four methods of providing downwelling infrared radiation using the methods described in Appendix A. The thick solid line is for the IDSO method, the thin solid line is for the MONT method, the thick dashed line is for the NET method and the thin dashed line is for the SATT method (see Section 2 and Appendix A)

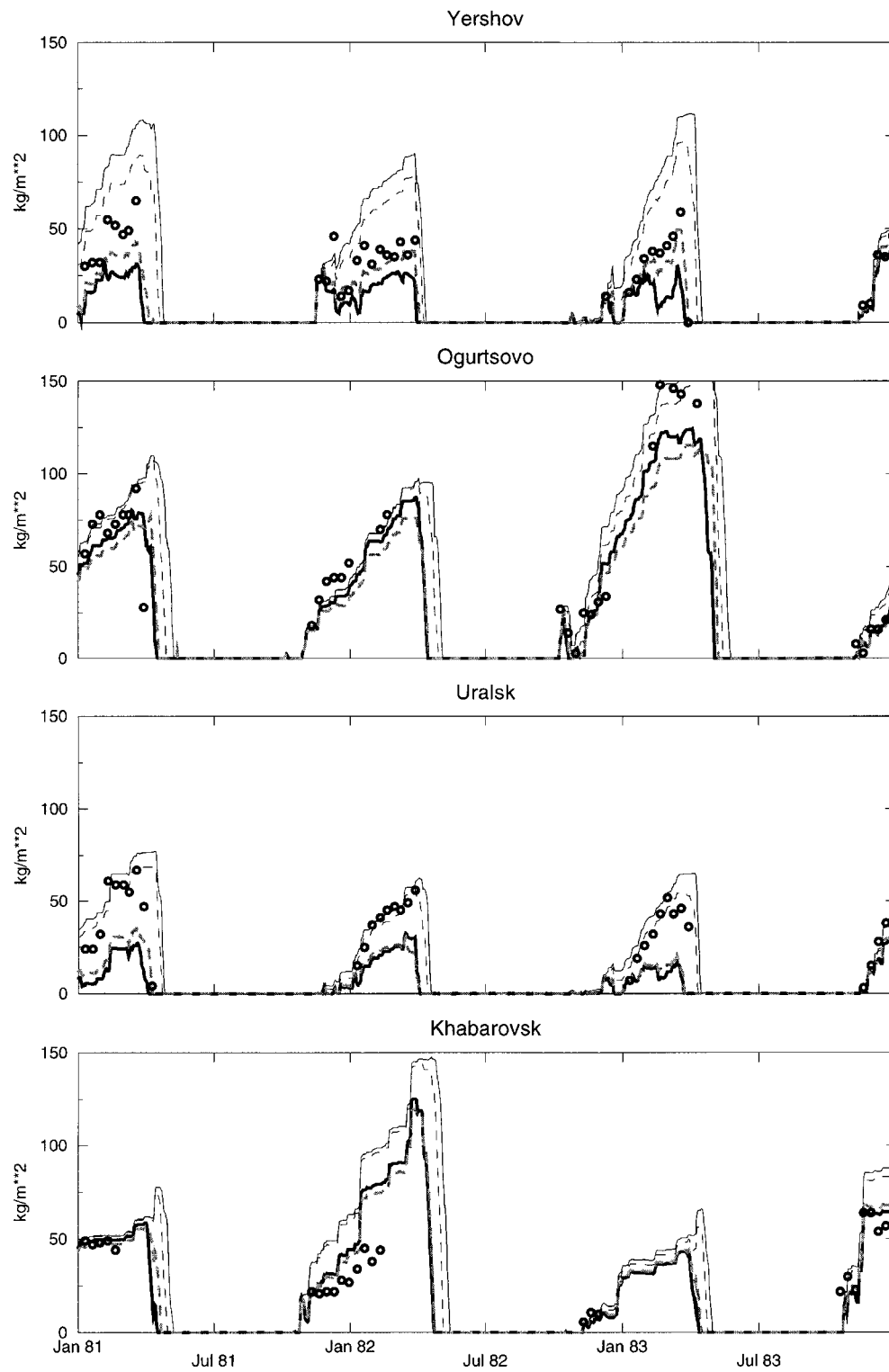


Figure 2 (Continued)

Uralsk and 1979 and 1980 at Khabarovsk). These all show instances where mid-winter SWE was observed to decrease, but this was not captured by BASE. The rapid end-of-season snow ablation is modelled well, although the exact timing of the event is not captured and this error can be up to 30 days depending on which $L\downarrow$ method was used.

Simulations at Khabarovsk overpredict SWE in most years and there is a characteristic peak in SWE prior to the end of each season which is not observed. Investigation of the forcing data at these times showed that air temperatures were sufficiently below freezing when the precipitation occurred to rule out snow/rain criteria as a possible cause of error, which points to an 'on the ground' ablation underestimation. During the 1980/1981 season at Khabarovsk, this overprediction, when viewed with observed albedo, is persistent and serious although similar findings were reported by Yang *et al.* (1997) suggesting that perhaps local climate anomalies due to topographic complications may be responsible for the difference between modeled and observed SWE (C.A. Schlosser, personal communication, 1995). The end-of-season peak is largely associated with the MONT and SATT forcing rather than the IDSO and NET simulations. For the end-of-winter 1980/1981 peak, precipitation certainly fell as snow, but the difference in the snowpack and ground temperature between the MONT/SATT and IDSO/NET forcing was sufficient to ensure that under the less intense (MONT/SATT) $L\downarrow$ forcing, considerably more snow accumulated, while under the higher $L\downarrow$ forcing, ablation occurred within a day. Overall, this suggests that near the end of the snow season, BASE's snow pack may be too cold. This shows that the lower/higher $L\downarrow$ forcing leads to lower/higher ground or snow temperatures which leads to significant differences in whether additional snow accumulates or melts (Figure 2). The link between snow and albedo provides a positive feedback enhancing the initial difference. Apart from the lagged snowpack ablation, there does not appear to be any other systematic error in the SWE simulations and shallow snowpacks are simulated with equal skill to deeper packs.

A few observations can be made from these results. No $L\downarrow$ method leads to a consistently higher quality simulation at all stations by BASE but the IDSO and NET methods remove the snow closer to the correct time more often. However, the quality of BASE's simulation of SWE depends more on which $L\downarrow$ method is used than on minor changes to how the model parameterizes snow. Clearly a complex snow scheme can only improve a simulation if the host model can provide $L\downarrow$ accurately. Finally, the sensitivity of BASE to relatively small variations in $L\downarrow$ has important implications for the model when linked to GCMs since the accuracy of $L\downarrow$ at high latitudes is probably poor (Garratt and Prata, 1996).

3.2. Snow density

Early season calculations of density have some general similarities with the patterns of observations suggesting that the parameterization is of some value for new snow (Figure 3). BASE produces good simulations at Ogurtsovo and Khabarovsk most years with the model rarely more than 50 kg m^{-3} from observations. In contrast, the density parameterization in BASE (see Appendix B) performs poorly at Yershov every year and relatively poorly at Uralsk most years. While early winter density is often simulated well, larger differences develop later in the snow season when BASE fails to resolve the end-of-season snowpack ripening (i.e. the phenomena which occur as a snowpack matures to become isothermal and subsequently melts) for some sites. There is a tendency for BASE to fail to increase the density of snow quickly enough although this is not systematic (e.g. Uralsk, 1978/1979 and 1979/1980, Figure 3). Despite the deficiency in the simulation of density at Uralsk and Yershov, BASE still simulates snow water equivalent reasonably well (Figure 2).

While BASE does simulate high resolution fluctuations in density, the amplitude of the fluctuation is usually damped and temporally lagged (e.g. Yershov 1981/1982 and Uralsk 1980/1981, Figure 3) because of the simplicity of the parameterization. Additional problems in density simulations occur where the early season snowpack fully ablates and snow density is reset to a value of 100 kg m^{-3} rather than continue to increase. This is evident at Yershov and Uralsk in the winter of 1980/1981 and 1982/1983 where the different $L\downarrow$ regimes allow for density to be either reset or to continue to increase, resulting in density differences of up to 100 kg m^{-3} . Again, this points to the need for a host model to provide $L\downarrow$ accurately before a complex snow parameterization becomes warranted.

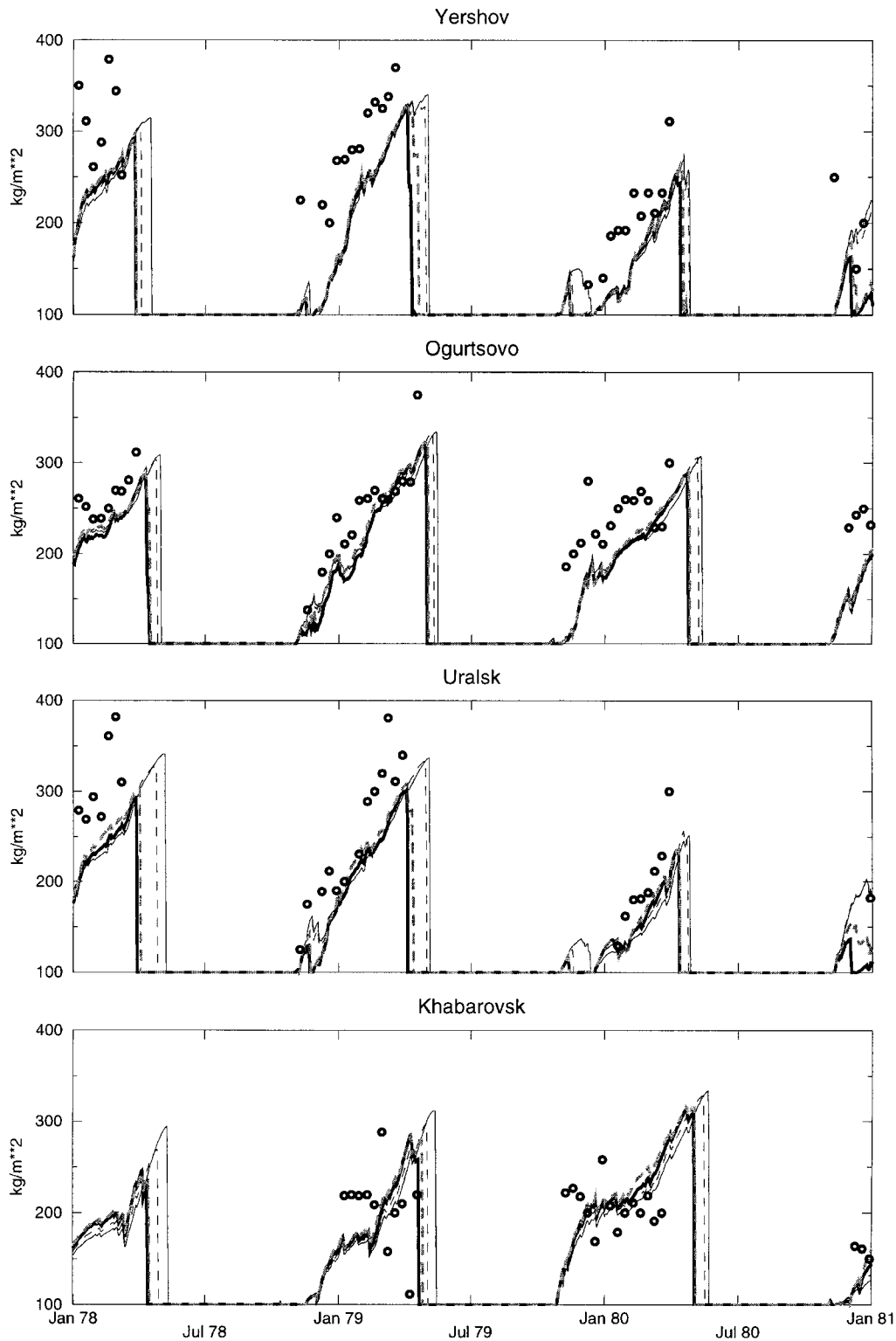


Figure 3. As for Figure 2 but for snow density (kg m^{-3})

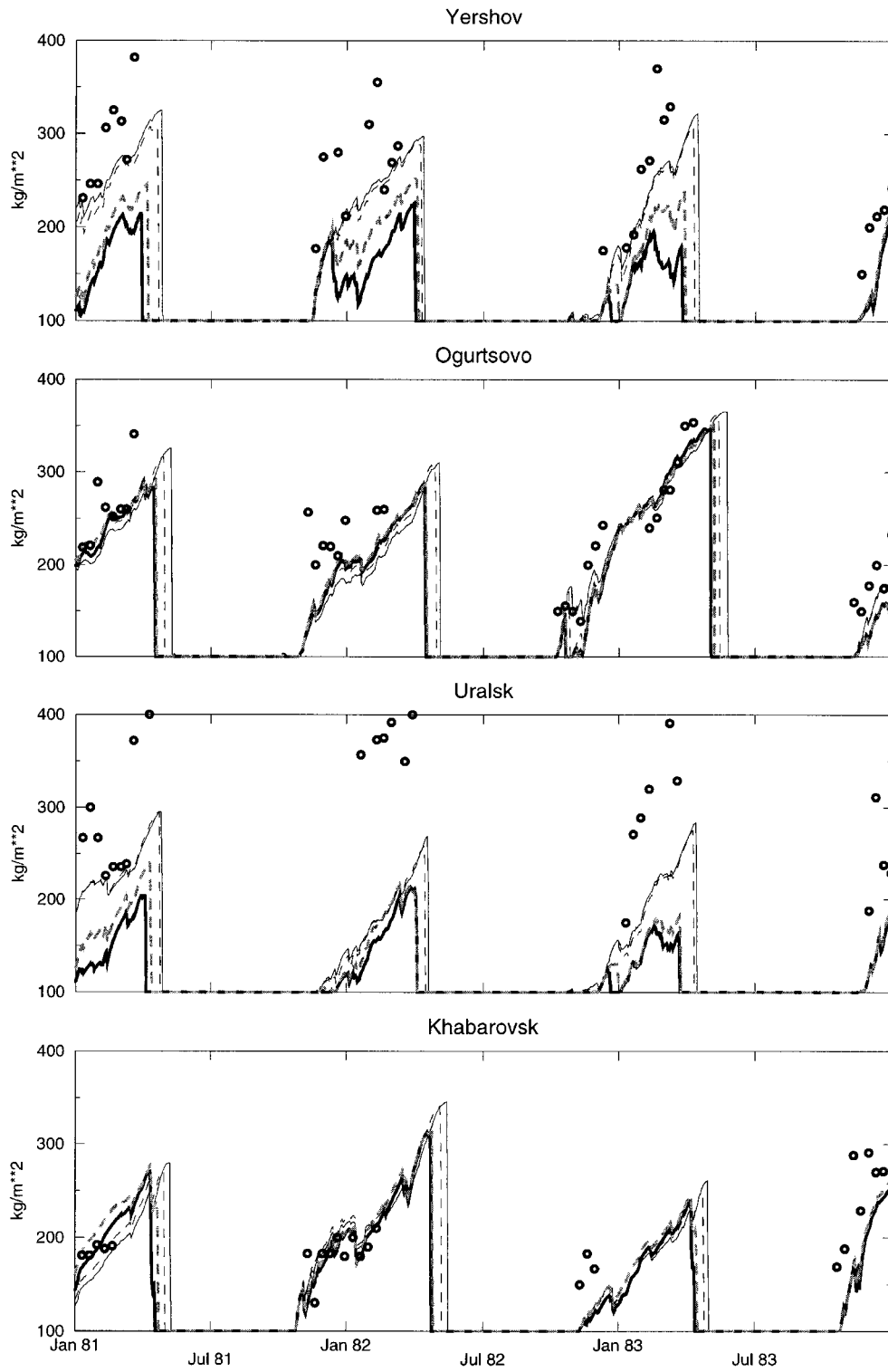


Figure 3 (Continued)

In real snowpacks, density increases towards the end of snow season from a combination of mechanical compaction and inter-snowpack phase changes leading to granularisation (Colbeck, 1974). Compaction is included in BASE (Equation (B8)), but granularisation is not. Since BASE only permits a single snow layer, snowpack stratigraphy and metamorphism due to temperature gradients are ignored and a complex system of snow of different age and characteristics is simplified into a single bulk layer. Additionally, densification due to diurnal freeze/thaw cycles at the surface of the snowpack cannot be accounted for, nor can factors such as ice layers (Woo and Heron, 1981). BASE, in common with all other land surface schemes, also ignores wind action which can pulverise grains, allowing them to form into a more dense snowpack (Sturm *et al.*, 1995).

BASE's snow density parameterization (Cogley *et al.*, 1990) follows Kojima (1967) who used data obtained over a snow temperature range of 268–273 K. Sturm *et al.* (1995) argue that regimes of snow exist each with different characteristics. This, combined with the lower temperatures simulated by BASE (as low as 233 K), suggests that the constants used in BASE for snow densification are probably not appropriate at these stations. The parameterization of snow density in BASE relates the compactive forces of snow self loading to the mass and temperature of the snowpack despite the temperature dependence of compactive viscosity being an area of considerable uncertainty (Pitman *et al.*, 1991). In BASE, snow densification depends on self-loading and temperature with the parameterization containing a parameter (k_n see Equation (B8)) which, according to Kojima (1967) lies in the range 2600–4600.

Figure 4 illustrates the sensitivity of BASE to the assumptions inherent in the densification parameterization. Varying the parameter (k_n) used in Equation (B8) across the range given by Kojima (1967) leads to very considerable differences in the density of snow such that after only 1 week, density variations from 125 to 500 kg m⁻³ develop (Figure 4(a)). Varying the snow mass, but holding k_n constant, also leads to differences in BASE such that the additional snow mass leads, through self-loading, to higher snow densities, but in this case (Figure 4(b)), the difference in density after 8 weeks is only about 125 kg m⁻³. Finally, Equation (B8) contains a temperature term and Figure 4(c–f) show the sensitivity of density across the range of temperatures from 253 K to 273 K. After 8 weeks the density differences due to temperature are small (100 kg m⁻³), but the differences between the four panels is substantial again indicating that the value of k_n is paramount. Figure 4 shows that colder temperatures result in slower metamorphism rates (*cf.* Colbeck, 1983) and that a unit temperature change for a cold snowpack will not have the same impact as it would at warmer temperatures. There is, therefore, merit in adjusting the value of k_n to match the criteria of the specific stations rather than assign a single global value. Determining the correct assignment of this value is not possible without further observed data, however BASE can match the observed density over time at all stations by tuning k_n . Decreasing the initial value of k_n from 4000 to 3500 on the basis that the climate of the stations used in this paper are colder than the maritime climate of Hokkaido (the area from which measurements were taken by Kojima, 1967) led to a significantly improved simulation of density and snow depth at Yershov and Uralsk for instance.

Of further interest in the simulation with a lower k_n is that for all stations the influence of different $L\downarrow$ forcing on density for deeper snow was not as pronounced as in the control experiment (with $k_n = 4000$). For lower density snowpacks, the density differences were greater between the different $L\downarrow$ forcing. Those instances of low density were typically associated with a very thin snow cover and are in keeping with the hypothesis that as temperature gradients increase, so does snow metamorphism. A low density snowpack (density is inversely proportional to a snow's insulating ability) which is thin can develop a large temperature gradient (Colbeck, 1991). These will lead to more rapid redistribution of water molecules causing crystals to lose their original form, becoming more dense and compacting at a faster rate thereby resulting in an overall increase in density (Colbeck, 1983).

The lack of observed data to develop parameterizations of snow density is a significant problem in improving the simulation of snow in land surface schemes. Experiments have shown that BASE's simulations can be improved by varying a single parameter (k_n) but choosing parameter values based on observed data, rather than effectively tuning these parameters, is presently impossible. While density is not the most important problem in land surface modelling, systematic measurements of density at a series of sites over time would help generalise Kojima's (1967) methodology.

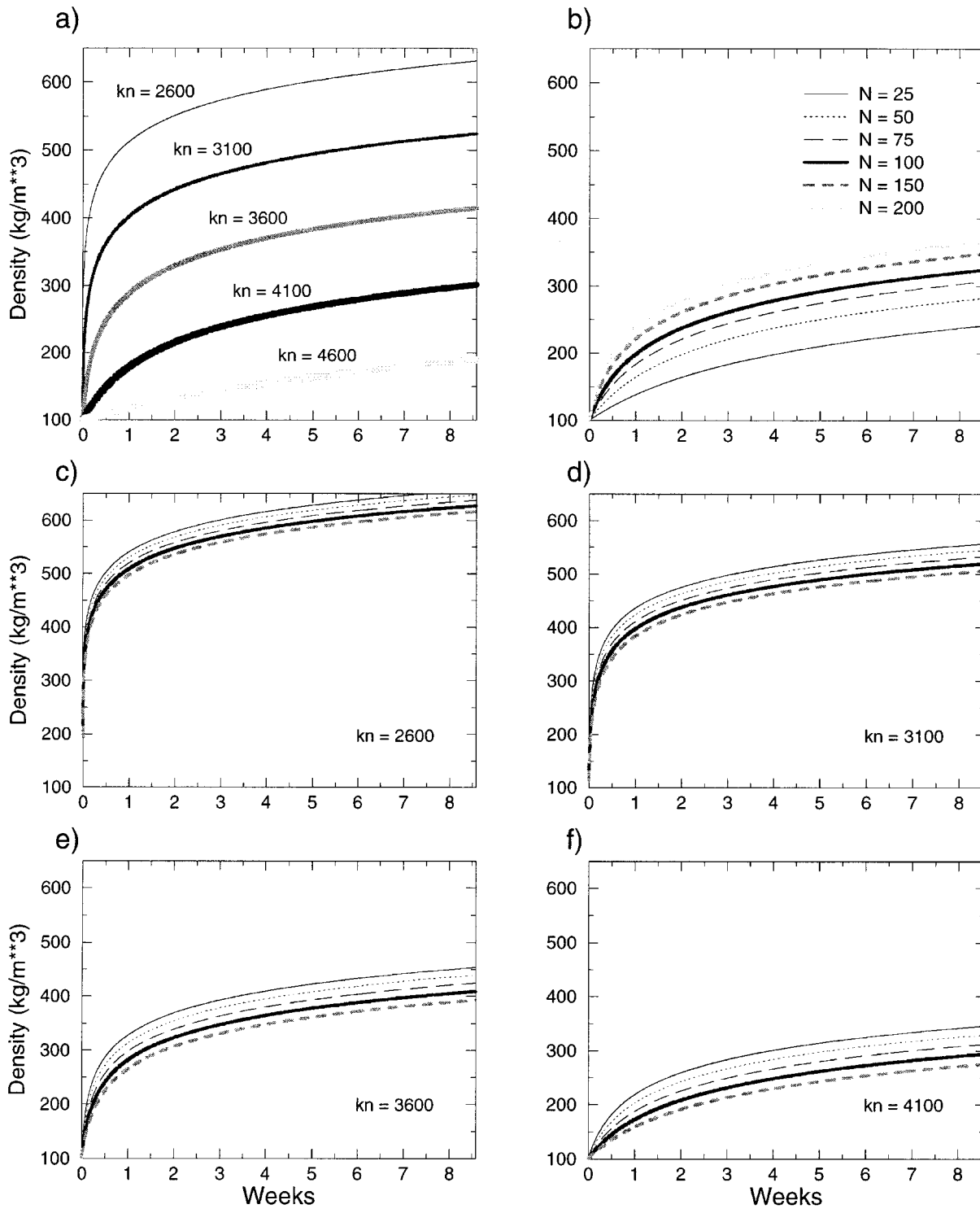


Figure 4. Sensitivity of the densification parameterization incorporated in BASE (Equation B8) to changes in the underlying assumptions. In each case, density is shown plotted against time (in weeks). (a) Varying k_n from 4600 (bottom curve) to 2600 (top curve); (b) varying snow mass from 25 (bottom curve) to 200 kg m⁻² (top curve); (c) varying T_n from 253 K (bottom curve) to 273 K (top curve) with k_n set to 2600; (d) as for (c) but with k_n set to 3100; (e) as for (c) but with k_n set to 3600; (f) as for (c) but with k_n set to 4100

3.3. Terrestrial albedo

The initial snow albedo appears to be specified well and BASE does a very good job of simulating the early snow season fluctuations in albedo, i.e. picking up the sharp peaks and troughs such as can be seen at Uralsk 1978/1979, Yershov and Ogurtsovo 1982/1983 (Figure 5). However, on average, BASE calculates too high an albedo for snow covered periods.

The extended end-of-season cover contributes to the overestimations of albedo, especially at Ogurtsovo and Khabarovsk, while cases of low fractional snow cover, such as simulated using the IDSO or NET $L\downarrow$ forcing in the later years at Uralsk, produces an underestimate. Indeed much of the deviation in terrestrial albedo is due to fractional coverage changes rather than to the actual snow albedo. This is indicated by the simulated rapid decline in albedo as ablation takes place, rather than the observed steady decrease in albedo as the snowpack ages and becomes contaminated. This suggests the effects of snow metamorphism and/or aging are not always resolved well, particularly for the colder stations of Ogurtsovo and Khabarovsk. This can be seen in each year in Figure 5 where observed albedo follows an asymmetric pattern, falling more rapidly from a peak in early to mid-winter than is simulated by BASE. Additionally, once BASE establishes a snowpack, the model has difficulty in capturing mid-season decreases in albedo.

An albedo scheme with a more dynamic response may be preferable. The problem in the BASE scheme stems from the fact that albedo is snow temperature dependent, but snow temperature is calculated for the integrated snowpack rather than just a thin surface snow layer. Differentiating layers within the snowpack and calculating the albedo using just the top few centimetres or using a calculated snow age would provide more skill in simulating the end of season albedo, but these methods would be computationally expensive.

3.4. Net radiation

BASE's simulated net radiation is a function of albedo and surface temperature since all other components of the radiation balance are supplied. Despite some inaccuracies in the observed net radiation data (see Section 2), the modelled net radiation appears to be consistently underpredicted when BASE is forced by the MONT/SATT data for all stations (Figure 6). This is true even for Uralsk in which the MONT and SATT forcing produced a SWE that was much closer to observations than when using other methods for $L\downarrow$. Figure 6 shows that the differences between IDSO/NET and observations only occurs at particular times of the year (e.g. around April each year the large difference is caused by the extended duration of modelled snow cover under a particular $L\downarrow$ forcing). This is most evident at Ogurtsovo and Khabarovsk where observations and simulation often differ by more than 40 W m^{-2} during this period. Outside of this period, fluctuations in net radiation are modelled very accurately, to the extent of resolving variations in net radiation of the order of 5 W m^{-2} . Clearly, over the winter period, there is a consistent underprediction in net radiation due to the poor albedo simulation described above (the difference is small because of the low solar radiation flux in winter). Part of the problem in simulating net radiation may also be due to surface temperature. During periods of snow cover, the surface radiating temperature is the snow temperature. The use of a single layer snow parameterization does not permit rapid change in snow temperature which may bias the radiating temperature. Similarly, surface snow melt during the day, or simply warming the first few centimetres of snow, cannot occur which may generate errors in the simulation of the surface radiating temperature (this is analogous to the need to simulate a skin soil temperature). This problem might also explain the lagged ablation in BASE apparent in some years. Unfortunately, the lack of observational data for snow temperature prevents further analysis of this problem.

4. DISCUSSION AND CONCLUSIONS

This paper has shown that a model containing a snow parameterization of intermediate complexity can simulate the observed seasonal and interannual changes in snow water equivalent, snow density and

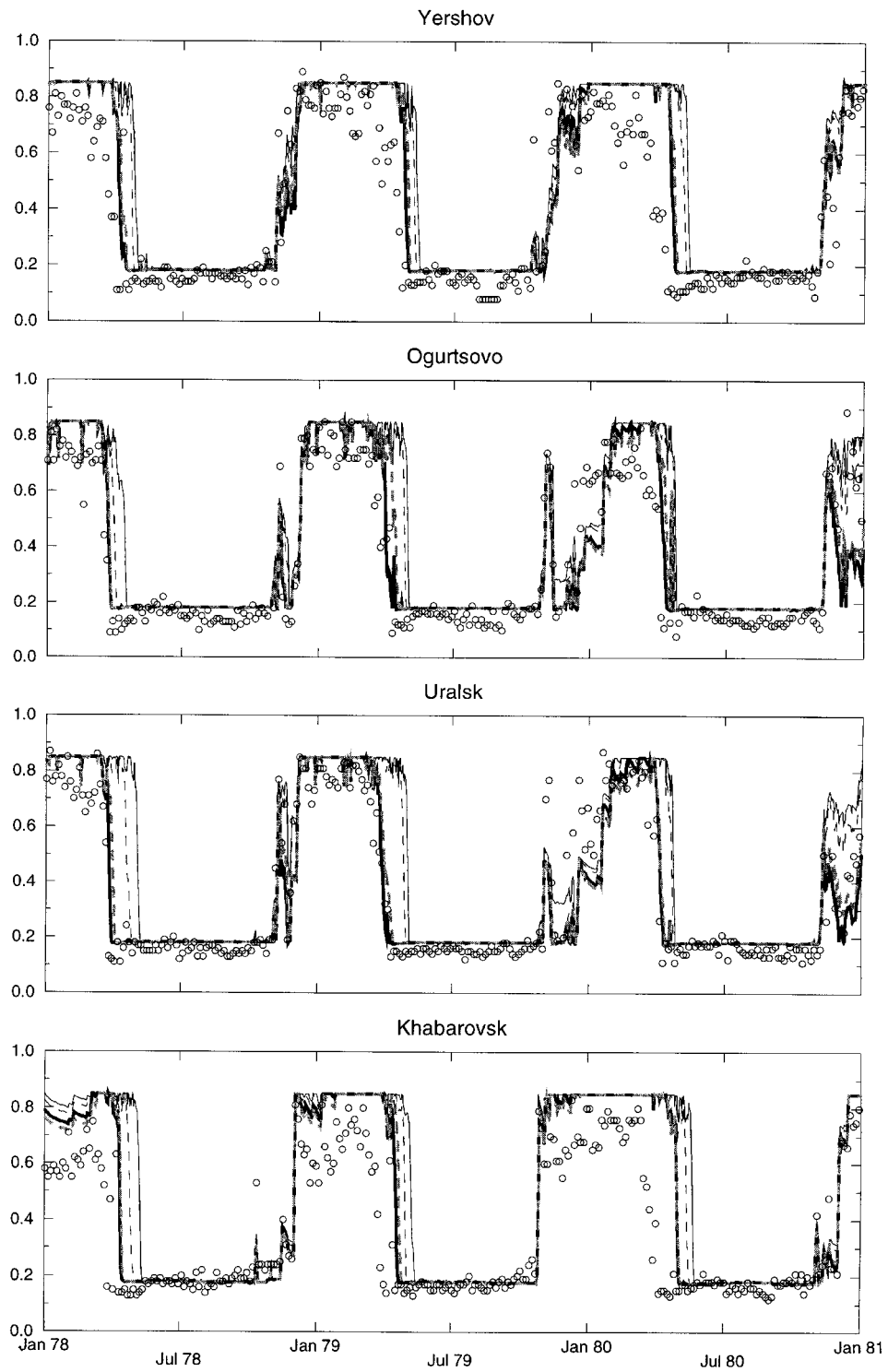


Figure 5. As for Figure 2 but for surface albedo. Observations are shown by open circles

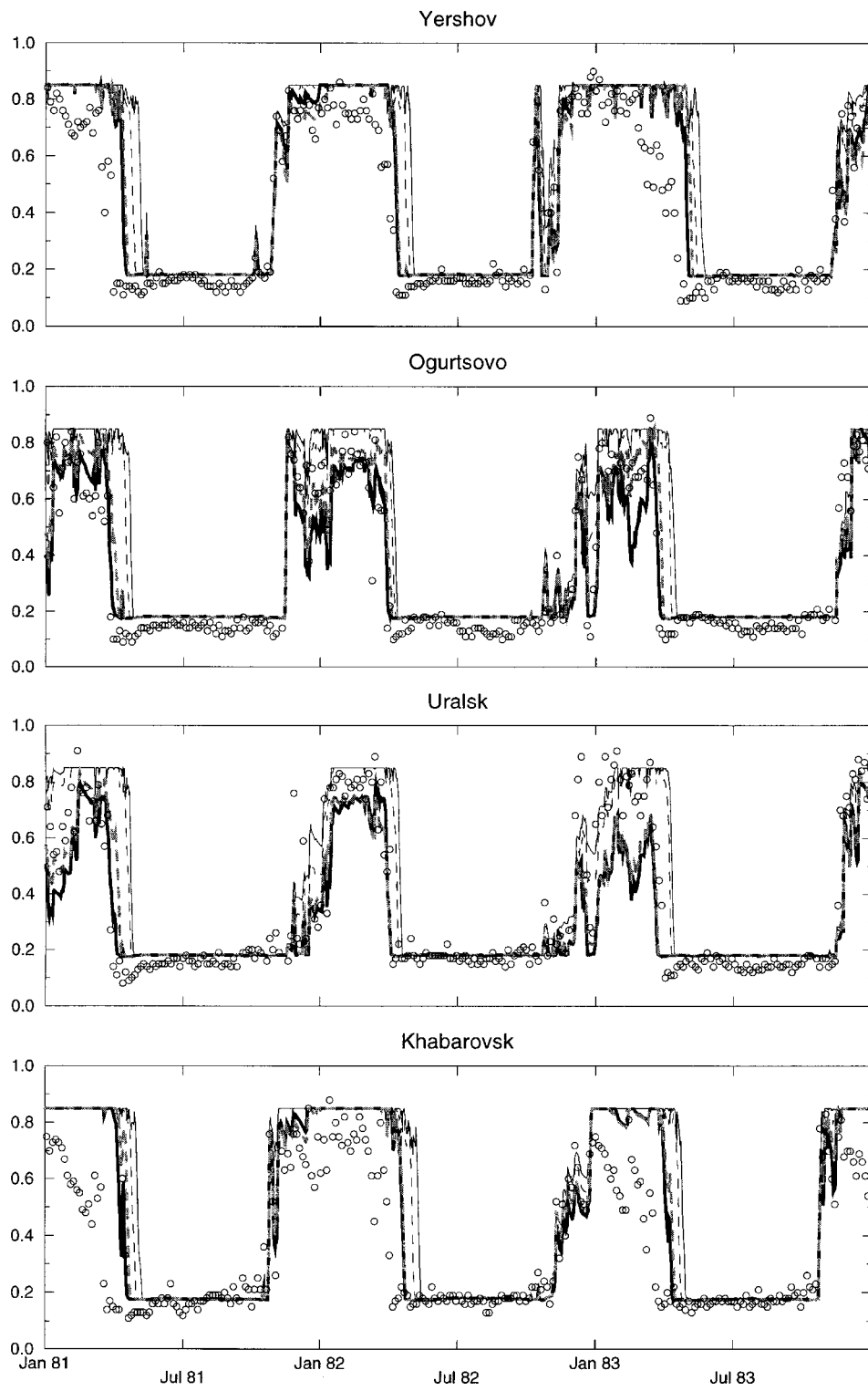


Figure 5 (Continued)

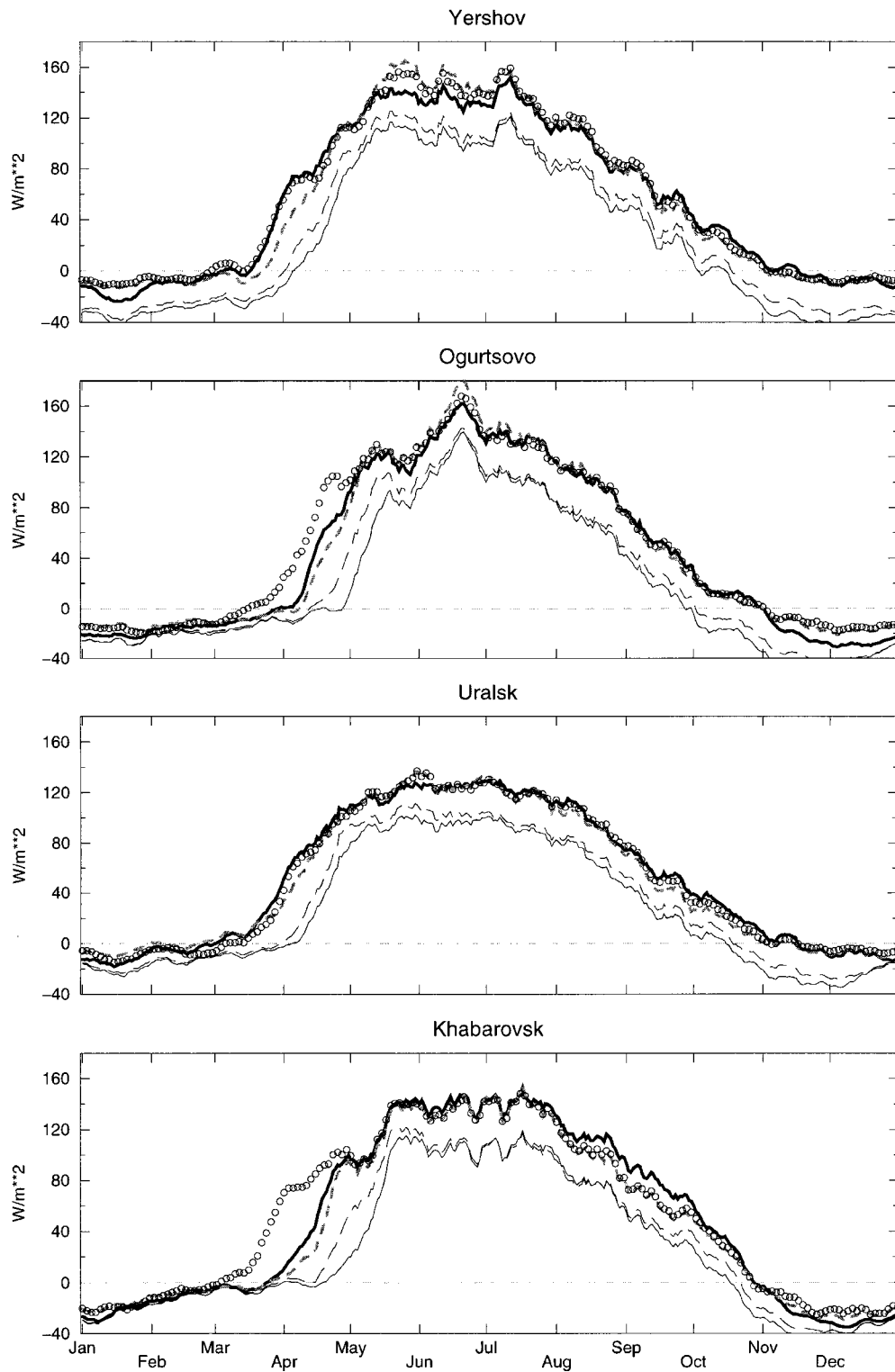


Figure 6. As for Figure 1 but for net radiation. Observations are shown by open circles

albedo with reasonable skill. The ability of the model to simulate the year-to-year differences in the observed quantities should give us some confidence that these types of model can simulate the changes in the same quantities resulting from climate change. Comparing the performance of BASE to other models such as BATS (Yang *et al.*, 1997) or SSiB and the Bucket (Robock *et al.*, 1995) is difficult because each model was run with slightly different assumptions. It is clear, however, that BASE maintains a snow cover longer into spring than these other models which may be linked to the different fractional coverage parameterizations used in the simulations. A more detailed intercomparison of snow simulated by a suite of models is being performed within Phase 2(d) of the Project for the Intercomparison of Landsurface Parameterization Schemes (Henderson-Sellers *et al.*, 1995).

While the overall simulations are reasonable, we have shown that some problems exist. The simulated temporal extent of snow cover tends to be overestimated, but this could be related to the provision of $L\downarrow$ to which BASE is sensitive. Problems also exist in the simulation of snow density, but by varying a single parameter, BASE could simulate seasonal density variations accurately. This is worrying since it implies that we may need to know the value of this parameter globally and we may need to know how this parameter may change with climate.

We note that some of BASE's weaknesses probably relate to the representation of the complexity of a snowpack with a single bulk layer approximation. Some of the difficulties in resolving densification, albedo and net radiation may all be related to the inability of a deep layer to evolve as quickly as a skin layer in terms of the characteristics of the surface actually interacting with the atmosphere. The need for a skin temperature in soil modelling is well known and we speculate that there may be a need to mirror this in snow modelling at GCM scales.

In principal, we might conclude that the level of complexity included in BASE is appropriate and that the model performs well enough to be considered satisfactory. However, notwithstanding the issue of whether the model works well enough to warrant the additional complexity, there are some difficult issues which need to be addressed. The fact that BASE is very sensitive to $L\downarrow$ is worrisome since the provision of this variable by GCMs to within $10\text{--}20\text{ W m}^{-2}$ of observations in transitional seasons at high latitudes is problematic. If GCMs cannot provide this variable accurately then including an overly complex sub-model may lead to worse simulations than a simpler scheme because of the inclusion of processes which permit the overestimation of $L\downarrow$ to be amplified. Similarly, BASE has been shown to be very sensitive to the parameterization of snow densification. Including land surface models into GCMs when they can be shown to be sensitive to a generally unknown global parameter requires care, although this is routinely done in land surface modelling with other parameters which are required globally (root distribution, leaf angle orientation etc). When the resulting variable (snow cover) is known to be a strong positive feedback on regional warming or cooling, even greater care should be exercised.

We should also be careful in interpreting the results shown in this paper. BASE performed well in offline tests, but showed considerable sensitivity to $L\downarrow$ and k_n . Pitman *et al.* (1990) showed that a similar land surface scheme (BATS, Dickinson *et al.*, 1986) was sensitive to a parameter value in offline tests, but when coupled to a host model which resolved surface-atmospheric feedbacks, BATS was much less sensitive (Pitman *et al.*, 1993b).

Overall, this paper indicates that a land surface scheme of intermediate complexity can simulate these cold environments well. There is skill in the simulation of snow water equivalent and other associated variables and BASE can simulate the seasonal cycle and the year-to-year variability of snow. However, there is a lack of knowledge concerning how to model snowpacks at the scales of GCMs and how to develop parameterizations at these scales rather than just implement site specific methods and assume they work at GCM scales. In this paper we have only dealt with the validation of BASE at the point scale because the observational data we have do not contain information concerning horizontal heterogeneity. Within a GCM, the parameterization of horizontal heterogeneity is likely to be very important. Liston (1995) discussed this and indicated that for the macroscale, the division of the surface into two homogeneous areas, one for snow and one for vegetation is generally adequate for the calculation of surface fluxes. Problems develop, however, if the snow cover is highly heterogeneous since many small patches of snow interact with the atmosphere and soil differently to one larger single patch. Accounting

for this horizontal heterogeneity in a realistic fashion is a matter for future research and is beyond the scope of this paper. However, we note that in order to answer these types of questions and improve land surface models in general, the collection of long term observations, across a range of climate conditions of snow variables, is one of the higher priorities in climate research.

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APPENDIX A. METHODS FOR PARAMETERIZING DOWNWELLING INFRARED RADIATION

$L\downarrow$ can be calculated via an atmospheric emittance method or using the surface radiation balance. This section explains those methods used in the simulations described in this paper.

A.1. Clear sky atmospheric emittance methods

To calculate $L\downarrow$ accurately requires the air temperature, humidity, CO₂ and O₃ content of the atmospheric column. Since these are rarely available in full, methods have been evolved for calculating $L\downarrow$ with limited meteorological data. All methods were derived for use in clear sky conditions and generally follow:

$$L\downarrow = \epsilon_a \sigma T_a^4 \quad (\text{A1})$$

where ϵ_a is the atmospheric emissivity and σ is the Stefan-Boltzmann constant. The difficulty in Equation (A1) is in calculating ϵ_a which largely depends on the amount of water vapour in the atmosphere (Idso, 1981). The contribution of CO₂ and O₃ is included implicitly on the basis that at atmospheric temperatures their effect is restricted to a narrow spectral band (Swinbank, 1963). Monteith (1973) suggested:

$$\epsilon_a = 0.53 + 0.06e^{1/2} \quad (\text{A2})$$

where e is vapour pressure near the surface. Aase and Idso (1978) determined that such empirical equations perform well in air temperatures above 0°C but in cold regions the equations should be modified hence Satterlund (1979) proposed:

$$\epsilon_a = 1.08[1 - \exp(-e^{T_a/2016})] \quad (\text{A3})$$

and Idso (1981) suggested:

$$\epsilon_a = 0.179e^{1/7} \exp\left(\frac{350}{T_a}\right) \quad (\text{A4})$$

These methods represented improved methods for calculating $L\downarrow$, but both assume clear sky conditions and require ϵ_a to be adjusted for cloudy conditions.

A.2. Cloud effects on ϵ_a

Monteith (1973) suggested that when the sky is covered with a fraction c of cloud, the apparent emissivity is ($\epsilon_{a,c}$) given by:

$$\epsilon_{a,c} = \epsilon_{a,0}(1 + mc^2) \quad (\text{A5})$$

where $\epsilon_{a,0}$ is the clear sky emissivity and m is the Boltz parameter which accounts for decreasing temperature with increasing height. m ranges from 0.2 for stratus or cumulus to 0.04 for cirrus.

Kimball *et al.* (1982) proposed a more sophisticated method which requires cloud height and cloud base temperature. While Yang *et al.* (1997) used this method, we did not believe the observed data for cloud height and missing data for cloud base made this method suitable. Alados-Arboledas *et al.* (1995) compared a Boltz parameter equation with methods based on Kimball *et al.* (1982) and Martin and Berdahl (1984) and found that the Boltz type equation performed best overall, although both the Boltz and Kimball *et al.* (1982) method showed discrepancies close to experimental error.

On the basis of the literature cited above and the limitations of the data set, the parameterization used for the influence of cloud cover was a Boltz/Monteith type used by Robock *et al.* (1995):

$$L\downarrow = \epsilon_a C \sigma T_a^4 \quad (\text{A6})$$

where C describes the increase in clear sky emissivity caused by clouds and is:

$$C = 1 + 0.2(C_L + C_M)^2 + 0.04C_H^2 \quad (\text{A7})$$

where C_L is the lower cloud cover fraction as provided in the observed data set. C_M and C_H are the middle and high level cloud cover fractions determined by:

$$C_M = C_H = (C_T - C_L)/2 \quad (\text{A8})$$

where C_T is the observed total cloud cover fraction. The methods described thus far for determining $L\downarrow$ are all semi-empirical approximations, the accuracy of which can be variable (Hatfield *et al.*, 1983; Alados-Arboledas *et al.*, 1995). If all other components of the surface radiation balance are available or can be calculated, an alternative method for determining $L\downarrow$ is via the surface radiation balance.

A.3. Radiation balance method for $L\downarrow$

$L\downarrow$ can be calculated via closure of the surface radiation balance:

$$L\downarrow = R_{\text{net}} - S\downarrow(1 - \alpha) + \epsilon_s \sigma T_s^4 \quad (\text{A9})$$

where α is surface albedo, R_{net} is net radiation, $S\downarrow$ is incoming shortwave radiation, ϵ_s is the surface emissivity (a quantity poorly known for snow, but set to 0.98 here) and T_s is the surface temperature. The observed albedo data contained many missing values which were infilled using previous day's values. The observed data set supplies measurements of unshielded surface temperature (only representative of grass surface temperature, not the entire plot).

We used four methods of determining $L\downarrow$ to eliminate this variable as a source of unexplainable model behaviour (the four methods result in variations in $L\downarrow$ of up to 60 W m^{-2} in both summer and winter). The methods employed were Monteith (1973) (MONT), Satterlund (1979) (SATT), Idso (1981) (IDSO), and the radiation balance method (NET). Of these equations, MONT typically returned the lowest average value, IDSO returns the highest average value and SATT lies between the two, but has been tested against temperatures as low as -36°C (Satterlund, 1979). The SATT method had been used by Douville *et al.* (1995), Robock *et al.* (1995), Schlosser (1995), Yang *et al.* (1997) and the NET method by Yang *et al.* (1997).

APPENDIX B. MODEL DESCRIPTION OF SNOW IN *BASE*

B.1. Model structure

The land–surface scheme *BASE* (Best Approximation of Surface Exchange) (Desborough, 1997) has a lineage running from Deardorff (1978), through *BATS* (Dickinson *et al.*, 1986) and *BEST* (Pitman *et al.*, 1991). *BASE* is designed to simulate the surface and sub-surface climate of a GCM grid square and to simulate the energy and water flux exchange between the land surface at the lower atmosphere. The snow

sub-model of BASE is a single layer snowpack that sits upon the three layer soil model. It is described by the four variables of snowpack temperature, mass, density and fractional extent. The sections below give details of BASE's snow model.

B.2. Snow sub-model

B.2.1. Albedo. Following Cogley *et al.* (1990) the snow albedo parameterization in BASE is centred on a temperature dependent reduction of fresh snow albedo. Albedo is calculated in two areas of the spectrum, i.e. the visible $\alpha_{n,\text{vis}}$ and near infrared $\alpha_{n,\text{nir}}$ with the division being at $0.7 \mu\text{m}$. The final integral snow albedo (α_n) is an average of the two component albedos; based on the assumption that the incident solar flux is equal above and below the division (Dickinson *et al.*, 1986).

$$\alpha_n = (\alpha_{n,\text{vis}} + \alpha_{n,\text{nir}})/2 \quad (\text{B1})$$

The temperature dependence of albedo, following Cogley *et al.* (1990), is parameterized as:

$$\begin{aligned} \alpha_{n,\text{vis}} &= 0.90 - 0.20 \cdot T_m^3 \\ \alpha_{n,\text{nir}} &= 0.80 - 0.16 \cdot T_m^3 \end{aligned} \quad (\text{B2})$$

where T_m , the temperature dependent metamorphism factor, is given by:

$$T_m = \begin{cases} 0.0 & T_n \leq 263.16 \text{ K} \\ 0.1(T_n - 263.16) & 263.16 \text{ K} < T_n < 273.16 \text{ K} \\ 1.0 & T_n \geq 273.16 \text{ K} \end{cases} \quad (\text{B3})$$

and T_n represents snow temperature. This results in a maximum α_n of 0.85 and a minimum of 0.67. While this scheme does not account for the diurnal cycle or diurnal hysteresis of albedo, which has been noted by Harding *et al.* (1995) it does capture the broad changes in surface albedo (Pitman *et al.*, 1991; Yang *et al.*, 1995) and has purposely been kept simple in order to minimise computational resources.

B.2.2. Fractional snow cover and vegetation masking. Snow and rain fall uniformly over the grid square. The fraction of the ground considered covered by existing snow (A_n) is, following Cogley *et al.* (1990), parameterized as:

$$A_n = \sqrt{\frac{d_n}{2D_{n,\text{eff}}}} \quad \text{for } 0 < A_n < 1 \quad (\text{B4})$$

where d_n is the snow depth (m) given by:

$$d_n = N/\rho_n \quad (\text{B5})$$

where N is the snow mass or water equivalent (kg m^{-2}) and ρ_n is the snow density (kg m^{-3}) and $D_{n,\text{eff}}$ is the effective snow depth (m) given by an empirical relationship based on snow density:

$$D_{n,\text{eff}} = 0.038 + 0.000144\rho_n \quad (\text{B6})$$

Vegetation will exist above the snow cover unless snow depth increases enough to cover it. The snow masking of vegetation (fraction $A_{n,c}$) is parameterized, following Dickinson *et al.* (1986) as:

$$A_{n,c} = \frac{d_n}{(d_n + 5Z_{0F})} \quad (\text{B7})$$

where Z_{0F} is the foliage roughness length (m). Note that it is not possible to completely mask all vegetation. The fractions calculated here are then used in calculating quantities such as the albedo of the entire grid square. This parameterization was altered as noted in Section 3.

B.2.3. Snowpack metamorphism and densification. Metamorphism of snow is characterised by a prognostic variable for snow density (ρ_n). The change in snow density over time is caused by mechanical compaction from overlying snow (self-loading) and input of fresh snow to the snowpack. These processes are included in a highly simplified form in BASE. The snow density parameterization included in BASE was described in full by Cogley *et al.* (1990).

Thus, in BASE, the densification rate of the snowpack due to self-loading (ρ) is:

$$\begin{aligned}\frac{\partial \rho}{\partial t} &= \rho_n \sigma_v / \eta \\ &= \frac{1}{2} \rho_n g \frac{N}{\eta} \\ &= \frac{1}{2} \rho_n g N / [10^{-7} \exp(-0.02 \rho_n + k_n / T_n - 14.643)]\end{aligned}\quad (\text{B8})$$

following Kojima (1967) and Cogley *et al.* (1990) where η is the compactive viscosity of snow (MPa s) and T_n is snow temperature (K). k_n is used to characterise the temperature dependence of η . Kojima (1967) gives a range of 2600–4600 for k_n which was arbitrarily assigned a value of 4000 by Cogley *et al.* (1990). This value is used in the standard version of BASE.

Equation (B8) does not include the reduction of bulk density due to the addition of new snow (see Equation B9).

Following the compaction calculation, the density of the snowpack is the weighted average of the initial snowpack and any new snow fallen during the timestep (P_{sn}). New snow falls at a density of 100 kg m^{-3} ($\rho_{n,\text{new}}$) while the maximum density is 450 kg m^{-3} . Therefore, the change in snow density with time is given by Equation (B8) combined with the impact of fresh snow:

$$\frac{\partial \rho_n}{\partial t} = \frac{\partial \rho_1}{\partial t} + \frac{P_{sn}(\rho_{n,\text{new}} - \rho_n)}{N} \quad (\text{B9})$$

Note that with the constraints on the model given here, snow does not develop into firn.

B.2.4. Mass balance. When BASE is used off-line, precipitation falls as snow when the air temperature at the surface is at freezing or below. Snowmelt only takes place once the temperature of the snowpack goes above 273.16 K.

Snow mass is not intercepted by the canopy although it can cover the canopy from the ground up. Evaporation of snow can occur when sufficient energy for sublimation is available.

B.2.5. Energy balance. The snowpack exchanges energy with the canopy air space above and the soil below. The movement of energy through the snow is governed by its thermal conductivity ($K_{T,n}$):

$$K_{T,n} = 2.805 \times 10^{-6} \rho_n^2 \quad (\text{B10})$$

where the constant 2.805×10^{-6} has been both theoretically (Schwerdtfeger, 1963) and empirically (Mellor, 1977) derived. The heat flux (W m^{-2}) from the snow to the ground (G_{n1}) is:

$$G_{n1} = A_n K_{T,n1} (T_n - T_1) \quad (\text{B11})$$

where T_1 , is the temperature of the top soil layer (K) and ($K_{T,n1}$) is the thermal conductivity at the snow/soil interface given by:

$$K_{T,n1} = \frac{\rho_n A_n (K_{T,n} + K_{T,1})}{N + d_1 (\rho_n A_n)} \quad (12)$$

where N is the snow mass (kg m^{-2}), d_1 is the depth of the upper soil layer (m) and $K_{T,1}$ is the thermal conductivity of the upper soil layer. The temperature of the snowpack is updated using:

$$\frac{dT_n}{dt} = \frac{H_{cn} - G_{n1}}{d_n \cdot C_{vn}} \quad (13)$$

where H_{cn} is the heat flux from the canopy air space to the snow (W m^{-2}) and the volumetric heat capacity of snow (C_{vn}) is calculated as

$$C_{vn} = C_{pi} \rho_n \quad (B14)$$

where C_{pi} is the specific heat capacity of ice. As no liquid water is modelled within the snowpack it makes no contribution to the thermal properties.

The snowpack has a maximum temperature of 273.16 K and once reached, additional available energy is used to melt snow. Rain incident upon the snow does not carry any energy, though it is known that the percolation of water through a snowpack can release latent heat.

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