Cold and Mountain Region Hydrological Systems Under Climate Change: Towards Improved Projections
Proceedings of H02, IAHS-IAPSO-IASPEI Assembly, Gothenburg, Sweden, July 2013 (IAHS Publ. 360, 2013).

Numerical modelling of snowpack seasonal evolution in various climatic conditions

ANDREY B. SHMAKIN & VASILIY S. SOKRATOV

Institute of Geography, Russian Academy of Sciences, Staromonetny St., 29, Moscow 119017 Russia
ashmakin@igras.ru

Abstract Results obtained by simulating snow characteristics with a numerical model of land surface heat and moisture exchange, SPONSOR, are presented. The numerical experiments are carried out for Franz Josef Land with strong winds and low temperatures, Dukant in the Tien Shan mountains with abundant relatively warm snow, and Valdai in western Russia with large interannual variability. The blizzard evaporation parameter is shown to have a great influence on snow depth at territories with high wind speed. At locations with regular warm events during winter, one should pay special attention to their modelling in terms of snow water equivalent and depth.

Key words snow modelling; blizzard evaporation; snow melt; interannual variability

INTRODUCTION

Modelling of heat/water exchange on land is a fast-developing technology for solving various tasks related to land–atmosphere interaction. Various types of models are used for such studies, and land surface models (LSMs) are among those which allow one to find reasonable compromise between relatively detailed description of the processes and not so significant computer resources. Normally, LSMs include explicit solution of the heat transfer equation, often accompanied by the water transfer equation, albeit many of the coefficients are parameterised using an implicit approach based on empirical knowledge of the phenomena.

In this paper, numerical modelling of snow cover is carried out, taking into account its interaction with the underlying soil, vegetation and atmosphere on a timescale of about an hour (i.e. with explicit solution of heat/water transfer equations). The task is solved by a numerical model of surface heat and moisture exchange, SPONSOR, developed at the Laboratory of Climatology of the Institute of Geography, Russian Academy of Science (Shmakin *et al*., 2009). The model took part in several major international projects on intercomparison of LSMs, such as PILPS, SnowMIP, etc. (Slater *et al*., 2001), and demonstrated quite good results. However, the snow cover block of SPONSOR was not detailed enough for the snow-related studies, and this paper describes a new version of the snow scheme in the model, and its further modifications for several types of natural conditions.

MODEL

General features of the land surface model

The SPONSOR model calculates all components of the heat and water balance on land, as well as the state variables (soil temperature and liquid/solid water content, etc.). All of these parameters are calculated at each time step. A set of external meteorological variables, such as temperature, humidity, wind speed, precipitation, radiation fluxes, etc., are required for the model. To operate the scheme, values of some parameters of the landscape are required, and some of the latter may have seasonal variation. These parameters are associated with the type of vegetation or soil at each location. Furthermore, values of deep soil temperature, and depth of groundwater table must be prescribed as the lower boundary conditions for heat and moisture. The number and thickness of the calculation levels in the soil can vary; in this work we used a seven-level version of the model.

Snow cover block

The new snow cover block includes a formalised description of the processes important to describe the hydrothermodynamic interaction of snow with the atmosphere and soil. These are: the formation of a new layer of snow, changes in the density of snow layers due to viscous and wind-influenced processes, change in temperature of snow layers due to heat exchange with the atmosphere and soil, as well as due to absorption of solar radiation, phase transitions of water within the snow, evaporation, melting, the transfer of melt water and its re-freezing; soil moisture changes due to snow melting and the infiltration of melt water and liquid precipitation, changes of snow albedo, and changes in the properties of snow due to change in the layer types. To evaluate changes in snow/soil temperature, correct description of the changing thermal conductivity and specific heat of snow plays an important role.

The basic structural unit of the snow cover model is a snow layer. It is assumed that, in general, at each time step with solid precipitation, a separate layer of snow with some initial properties is formed. Later, under the influence of various processes, the layer properties can change. With the development of snow thickness, adjacent snow layers of similar properties may be combined. From a computational point of view, the snow cover is considered as a multi-layered medium; each layer is characterized by temperature, weight (water equivalent), thickness, density, humidity, phase state of the contained water, heat capacity, thermal conductivity and maximum water-holding capacity (e.g. Sturm *et al*., 1997).

Further modification of the model included a simplified description of blizzard evaporation, as it was found that for conditions with very strong winds the procedure could be necessary. To evaluate the blizzard evaporation (i.e. the process of lifting of snow particles by strong wind and their further evaporation without falling back on the surface), we used the following parameterisation approach based on studies by Dyunin (1983):

$$I_m = Q_n / L_d \tag{1}$$

where I_m is the intensity of the blizzard evaporation, Q_n is carrying capacity of the blizzard, L_d is the maximum transport distance of the snow. The carrying capacity of the blizzard is calculated by the following formula: $Q_n = 0.077(\nu - 5)^3$, where ν is the wind speed at 10 m above surface level. The L_d value depends on the local climate and cannot be a constant (Dyunin, 1963). According to Cherniavsky (1894), it can be estimated as 1750 m for the western territory of Russia, while in western Siberia it is about 2–3 km. In the Arctic, it varies from 5 to 10 km; in Antarctica, according to Kotlyakov (1961), it can be 10–20 km and more.

There are no publications specifying the wind speed from which equation (1) starts to work. We have assumed that the blizzard evaporation should be taken into account with the linear dependence for wind speeds above 5 m/s, with a wind speed of 7 m/s evaporating the snow with a density of 96 g/m^3, and with a wind speed of 26 m/s evaporating snow with a density of 400 g/m^3. It is assumed that under strong winds, the snow can be evaporated without limitation, including the snow crust.

MODELLING RESULTS

Franz Josef Land

For comparison of calculated and observed values of snow cover characteristics in a very windy environment, we used data from the Institute of Geography expedition which took place during the International Geophysical Year in the archipelago of Franz Josef Land, from October 1957 to July 1959. The observations were made on glaciers of Franz Josef Land, namely, on the islands of Hooker and Hayes. The main object of study was the Churlyanis ice dome on Hooker island.

During numerical experiments, we compared time series of snow depth, calculated by the SPONSOR model, with observed data for the same location. The resulting curves are shown in Fig. 1.

The figure clearly shows the difference between observed data and modelled results. The largest bias with the observational data was obtained by calculation of snow depth without blizzard evaporation. With inclusion of the blizzard evaporation by the Dyunin formula, the calculated amount of snow cover is much closer to the observed one (D1). The D2 curve demonstrates snow

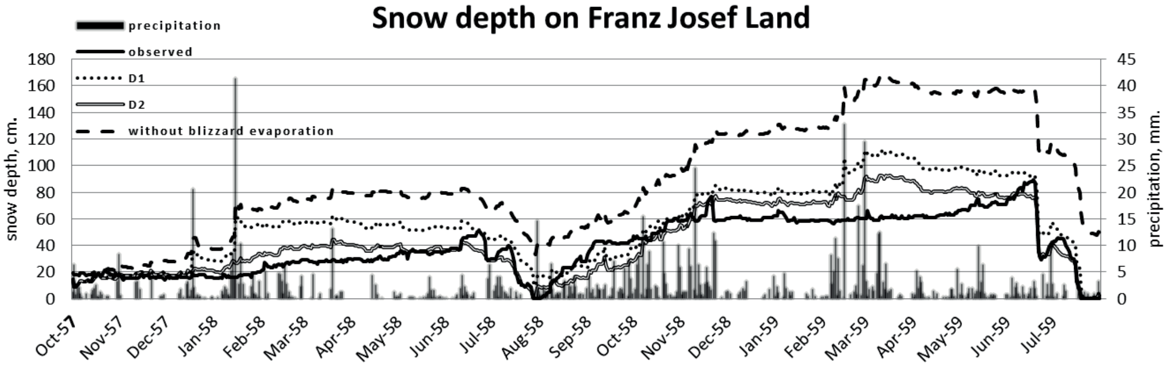

Fig. 1 Time series of observed and modelled snow depth at Churlyanis ice dome on Hooker Island, Franz Josef Land. Observed precipitation is shown with vertical columns.

depth calculated by the model without three cases of doubtful snowfalls: 18 December 1957, 15 January 1958 and 15 February 1959, when the observational data for each day demonstrate heavy snowfall, but it did not affect the snowpack depth during this period in any way. These snowfalls were marked as doubtful by the observers themselves. With a 10-fold reduction of precipitation on those days, the overall result has improved substantially (the D2 curve). While the model slightly overestimates the depth of snow cover, without blizzard evaporation consideration this bias would increase even more. In fact, one can see from Fig. 1 that changes in observed snow depth are not always related to observed precipitation events; probably, the very method of precipitation measurements with a gauge can provide questionable data.

The 22-month time series of observations is, perhaps, insufficient for accurate assessment of the model performance in this area, but some conclusions could be made if we consider the seasons separately (Fig. 2). The overall correlation coefficient between the observed and calculated snow depth equals 0.91. When analysing Fig. 2, we can conclude that in 1958 the correlation coefficients were generally better than in 1959, but in some seasons the bias can be variable, even by sign. In winter, for example, the correlation coefficients vary from year to year (in winter of 1957–1958 correlation coefficient between the series is 0.89, and in 1958–1959 it is 0.05). During the summer season, the correlation coefficients are high (0.87 for 1958 and 0.95 for 1959). Calculating the correlation coefficient between the time series of observed and modelled snow depth is useful for determining the accuracy of considering various processes in the model, because the observed snow depth is independent of any kind of external environment: it was measured separately, and observers were not comparing it with any other meteorological parameter. In turn, the model evaluates the snow depth on the basis of all other meteorological parameters (air temperature, wind, precipitation, radiation, etc.).

In spring (March–May), the correlation coefficients are negative, and Fig. 2 shows that the model gives decrease of snow depth, while in fact the latter is still accumulating. This means that the model does not take into account some processes/parameters affecting the intensity of snowmelt. However, an intensive process of melting begins in summer, which is clearly visible as estimated by the model and observational data, and the correlation coefficient for these periods is high; therefore, we can say that this period is well described by the model.

The blizzard evaporation formula, used in this work, shows a great influence on snow depth simulation in territories with high wind speed, such as Franz Josef Land. One could consider using equation (1) or similar ones for numerical modelling of the snow characteristics. Of course, this is not the only specific meteorological characteristic that should be taken into account for snow modelling, but this work demonstrates the importance of strong wind for snow depth simulation. With the blizzard evaporation parameterization included, snow cover models based on heat/water exchange description, could be successfully used for specific Arctic conditions.

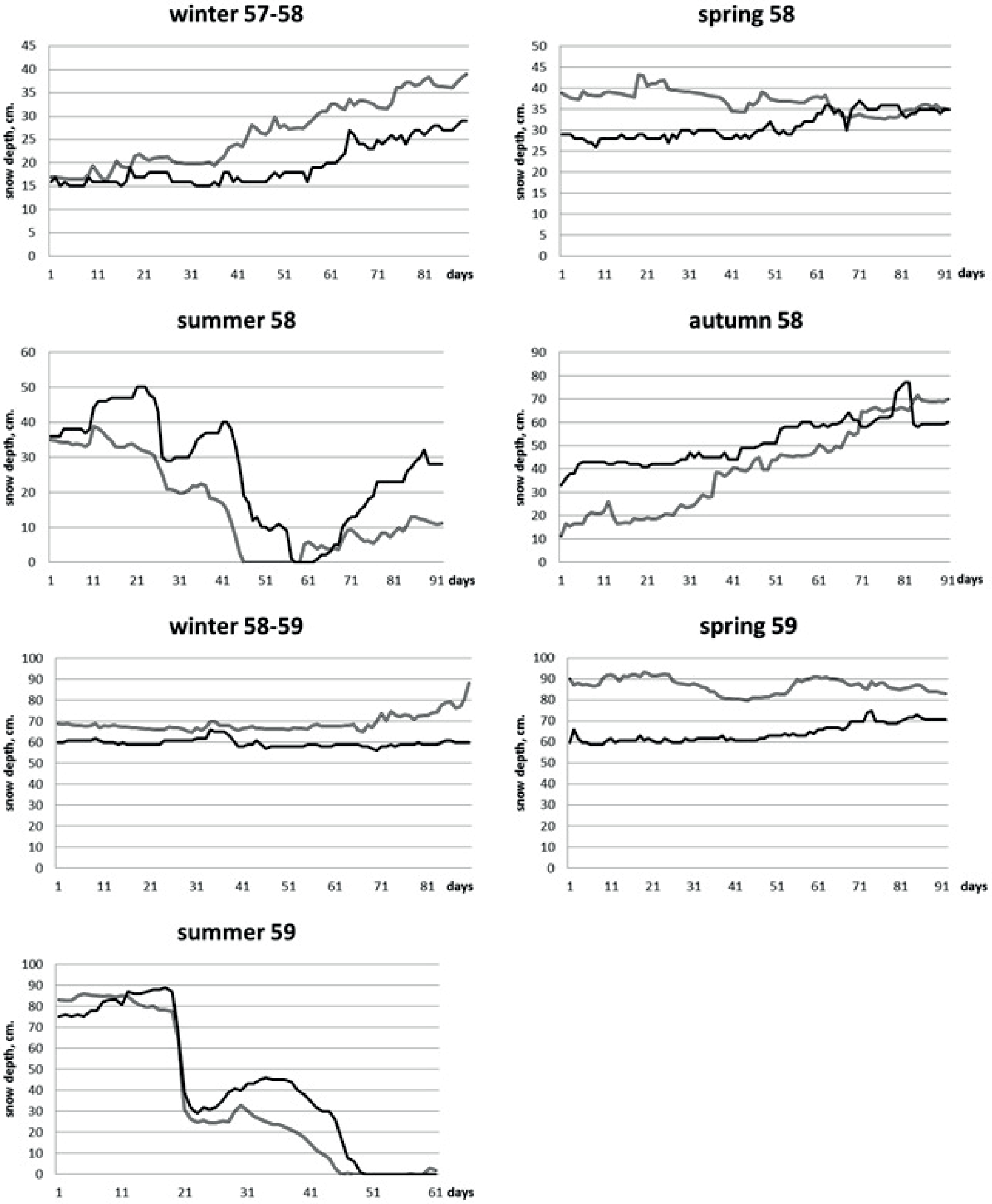

Fig. 2 Seasonal time series of observed and modelled snow depth at Churlyanis ice dome on Hooker Island, Franz Josef Land. Black curves represent observed values, grey ones – modelling results.

West Tien Shan

Figure 3 shows results of snow depth calculation at the avalanche-control station of Dukant (west of Tien Shan mountains, Uzbekistan) with and without blizzard evaporation, and observed snow depth data for four consecutive seasons. The region is located in subtropical zone, and only a short season in winter can generate snow cover there; however, the solid precipitation is usually abundant and relatively warm. Figure 3 shows that it is not necessary to use blizzard evaporation for calculating snow depth for this particular place; here the calculated snow depth, with blizzard evaporation and without it, are almost the same, and the difference is mostly less than 1 cm of snow. Although the Dukant station is situated at an altitude of 2000 m above sea level and the terrain is complicated, the wind speed data that we have do not show high values. The registered wind speed is mostly rather weak, probably because the station is shielded by mountainsides. The model needs to be adapted for better calculation of the snow depth in such an environment. However, the overall quality of snow modelling here is not too bad, albeit with some tendency for underestimation of the snow depth during the 1987–1988 and 1990–1991 seasons, while during 1988–1989 and 1991–1992 the model worked really well. A possible reason for different performance of the model in various years could be more frequent melting events in the 1987–1988 and 1990–1991 seasons, as during the melting snow pack usually compacts with increasing density, and even if the snow mass is described appropriately, the snow depth could be calculated with some errors.

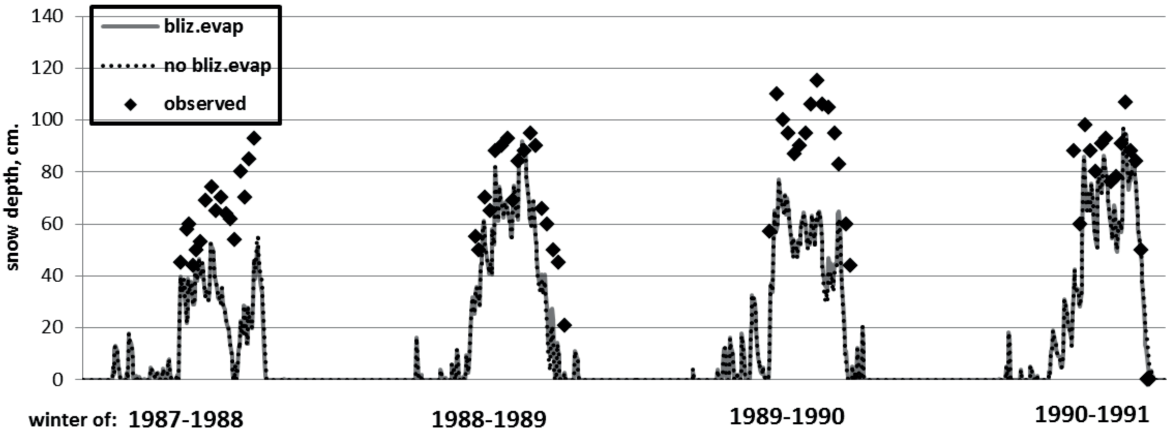

Fig. 3 Snow depth in Dukant (west of Tien Shan), observed (diamonds) and calculated with and without blizzard evaporation (curves).

Valdai, East European Plain

The model was also tested against the time series of data obtained in 1966–1983 at Valdai hydrological laboratory, located roughly between Moscow and St. Petersburg in a highland environment. The region is characterised by high interannual variability of the meteorological parameters, with some winters being relatively warm, and others similar to Arctic conditions. The particular location in Valdai area was Usadievsky catchment with prevailing grass vegetation and a shallow water table. The location is described in detail in Slater *et al.* (2001). Results of the modelling for some of those 18 years, both for snow water equivalent and snow depth, are shown in Fig. 4. For these experiments, we had several runs of the model with various values of external parameters which could only be specified with some uncertainty. Here the curves obtained with different values of the threshold air temperature determining the types of precipitation (rain or snow) are shown. This threshold air temperature cannot be prescribed with ideal accuracy for any precipitation event, and one can regard the situation when the observed values are located between the curves as good-quality results, while if the observed points are located outside the range between the curves, the model accuracy is definitely lower.

Experiments for Valdai demonstrate that warm (melting) events during winter can result in significant scatter of the modelled snow parameters. For example, in winter 1972–1973 all calculated curves are placed close to each other, and very close to the observed values; there were no significant warm events during that winter. In contrast, during the winter of 1974–1975, the observed values are located generally below all calculated curves, and the main reason for that are relatively frequent melting events during that winter. The snow water equivalent is modelled with a bias, but not very big one, while the snow depth is evaluated with poorer quality, probably due to compaction of the snow pack during the warm events. Thus, for modelling of the snow seasonal evolution, one should pay special attention to the warm events with melting and compaction of the snow cover.

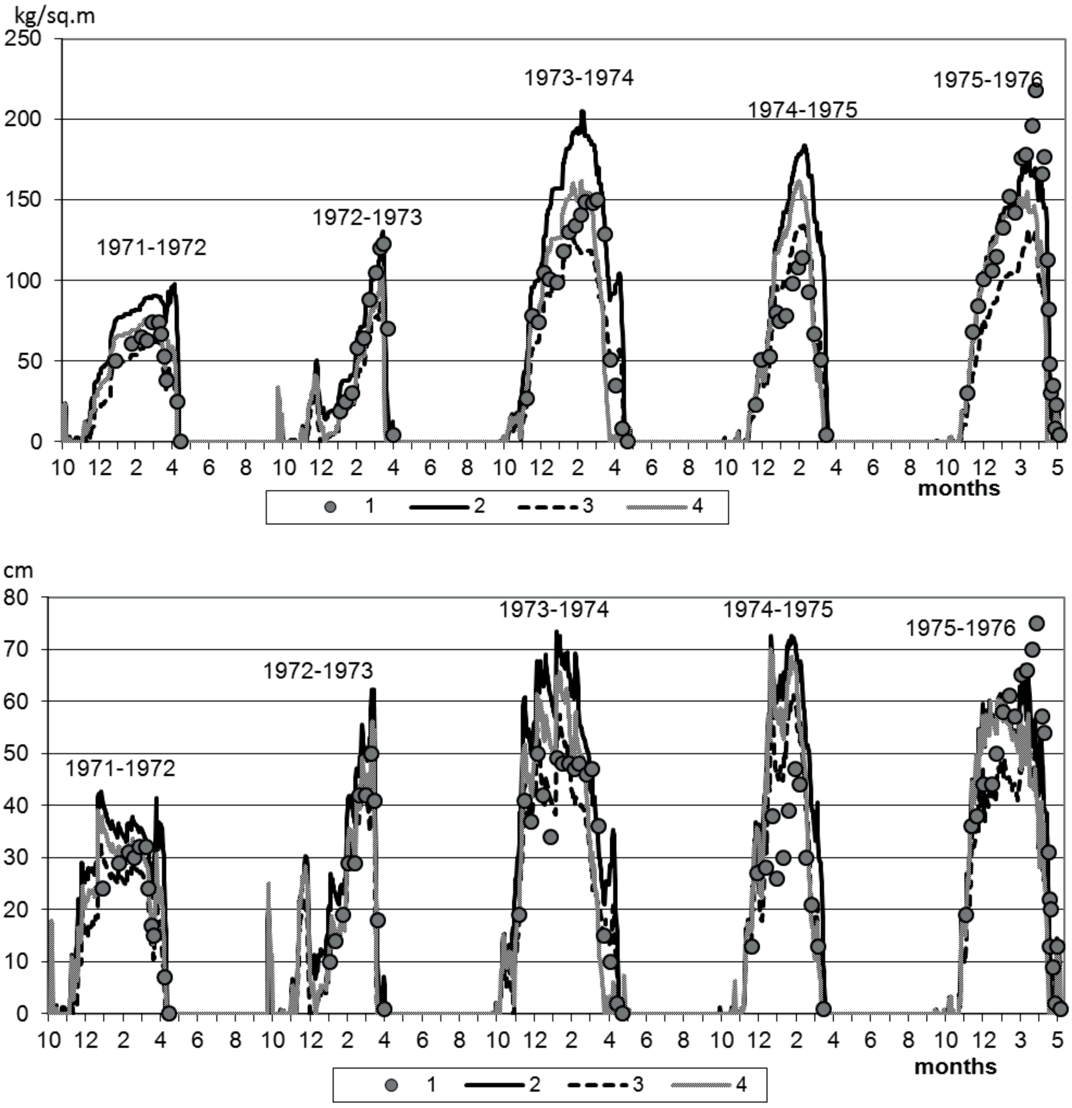

Fig. 4 Observed (1) and modelled (2–4) snow water equivalent (top) and depth (bottom) in Valdai in 1971–1976. Different curves are plotted with different values of the threshold air temperature determining type of precipitation: 0ºC (2), 1ºC (3) and 2ºC (4).

CONCLUSIONS

The new snow cover scheme, installed in the land surface model SPONSOR, allows evaluation of the snow cover characteristics with better accuracy; in most cases the model now reproduces the main snow cover features quite well. The model, tested against independently observed snow depth and/or water equivalent, demonstrates that in certain conditions, specific processes can be of great importance for the snow cover seasonal evolution. In the Arctic environment with very strong winds, the blizzard evaporation is so important that one cannot model seasonal snow variations without its parameterization. Among the methods for blizzard evaporation, the one developed by A. K. Dyunin demonstrated the best results, although in spring the model does not work very well, even with this parameterization. Moreover, it is very important to have precise measurements of precipitation, which can be problematic in the Arctic areas during certain conditions (e.g. high values of blowing snow). In the relatively warm conditions in Tien Shan mountains, melting events and heavy snowfalls are quite frequent so a proper description of melting during winter is important. For mid-latitude Russia with high interannual variability of weather conditions, warm events with melting and compaction of the snow cover can also be important, as well as proper description of frosty conditions during the entire winter. Due to existing uncertainty in several parameters, a good practice could be to run a set of experiments with different values of these parameters, and to analyse a whole family of modelled results.

Acknowledgements The study is supported by the Russian Foundation for Basic Research (grant no. 11-05-00573). The data for Dukant site were courteously provided by Maxim Petrov.

REFERENCES

Cherniavsky, A. S. (1894) Snow piles and their mitigation. *Railway Business* 25–27 (in Russian).

Dyunin, A. K. (1963) *Mechanics of Blizzards*. Novosibirsk, Siberian branch of the Russian Academy of Sciences (in Russian).

Dyunin, A. K. (1983) *In the Kingdom of Snow*. Novosibirsk, Nauka (in Russian).

Kotlyakov, V. M. (1961) The snow cover of Antarctica and its role in modern glaciation of the continent. In: *Research results of the International Geophysical Year. Part IX of the IGY, Glaciology, No. 7*. Moscow, Russian Academy of Sciences (in Russian).

Shmakin, A. B., Turkov, D. V. & Mikhailov, A. Y. (2009) Model of snow cover considering its layered structure and seasonal evolution. *Earth Cryosphere* XIII(4), 69–79 (in Russian).

Slater, A. G., Schlosser, C. A., Desborough, C. E., Pitman, A. J., Henderson-Sellers, A., Robock, A., Vinnikov, K. Ya., Mitchell, K., Boone, A., Braden, H., Chen, F., Cox, P. M., de Rosnay, P., Dickinson, R. E., Dai, Y.-J., Duan, Q., Entin, J., Etchevers, P., Gedney, N., Gusev, Ye. M., Habets, F., Kim, J., Koren, V., Kowalczyk, E. A., Nasonova, O. N., Noilhan, J., Schaake, S., Shmakin, A. B., Smirnova, T. G., Verseghy, D., Wetzel, P., Xue, Y., Yang, Z.-L. & Zeng, Q. (2001) The representation of snow in land surface schemes: results from PILPS 2(d). *J. Hydrometeorol.* 2(1), 7–25.

Sturm, M., Holmgren, J., Konig, M. & Morris, K. (1997) The thermal conductivity of seasonal snow. *J. Glaciol.* 43(143), 26–41.