Interpolating sparse surface measurements for calibration and validation of satellite-derived snow water equivalent in Russian Siberia

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Abstract Geostatistical methods are used to interpolate and average point snow water equivalent (SWE) measurements to a spatial resolution appropriate for comparison with estimates from Special Sensor Microwave Imager (SSM/I) images. Block kriging, a form of optimal interpolation, is applied to station snow course measurements in and around the Ob' River basin in Russian Siberia, giving spatially averaged SWE estimates at the resolution of the sensors, 25 x 25 km cells. For the two dates studied, interpolated station data and SSM/I estimates agree well in flat, low-lying regions. In the Ural Mountains, the SSM/I estimates significantly underestimate SWE, with respect to the station data. In the Altai Mountains, the SSM/I algorithm indicates higher SWE than elsewhere, but comparison is difficult, due to the sparsity of stations there. The results will be used to adjust the SSM/I SWE retrieval algorithm for use in complex terrain, where remotely sensed snow data are particularly needed.

Key words kriging; snow water equivalent; SSM/I; Siberia; Russia; Ob’ River

INTRODUCTION

The mismatch of scale between surface observations and the satellite footprint represents a major challenge in validating, calibrating, and applying algorithms based on remotely sensed data. The present study is part of an ongoing effort to merge multi-source, multi-scale data sets in a hydrological model of the Ob’ River basin. The Ob’ rises in the Altai Mountains of Central Asia (latitude 50°N) and flows through Siberian taiga and tundra to the Kara Sea (72°N). Figure 1 shows the basin topography in the polar azimuthal projection used for the Equal-Area Scalable Earth Grid (EASE-Grid) format (Armstrong & Brodzik, 1995), the standard format for data from the polar orbiting satellites.

Interpolating station snow water equivalent

The station snow course data, “Former Soviet Union (FSU) Hydrological Snow Surveys”, were obtained from the National Snow and Ice Data Center (Haggerty &
Fig. 1 Topography of the Ob' basin and surroundings shown in the EASE-Grid polar azimuthal projection. Also shown are the locations of Former Soviet Union hydrological snow survey stations, the region over which gridded SWE is estimated (square), and the cross-sections selected for detailed analysis (dashed lines).

Armstrong, 1996). The FSU dataset gives snow water equivalent (SWE) as measured along a clear field (SWE$_C$) and forest (SWE$_F$) transect at each survey station, as well as the percent forest cover (100$f$) surrounding each station. For the selected dates, 20 January and 20 February 1989, a weighted average station SWE$_W$ was calculated:

$$SWE_W = fSWE_F + (1 - f)SWE_C.$$  

For stations or dates where only one type of transect was reported, SWE$_W$ was set to the available value, either SWE$_F$ or SWE$_C$. Locations are shown in Fig. 1; not every station reported on every date.

Optimal interpolation (kriging) requires a model of spatial covariance, the semi-variogram (SV). The SV of SWE$_W$ was constructed for each date as follows. First, the raw SV was built by plotting half the squared difference, $(SWE_{W,i} - SWE_{W,j})^2/2$, against separation distance ($d$) for all possible pairs of stations: $i, j, i \neq j$ (Fig. 2, scattered small dots). Next, the experimental SV was computed by averaging the squared differences into 50-km bins of $d$ (large circles). Finally, an analytical model was fitted to the experimental
SV (lines). A spherical SV model was assigned in each case. The non-zero intercept, known as the nugget effect, represents variability on scales smaller than the measurement scale. Here, the stations and their surroundings have been treated as points; the nugget effect of 1000 mm$^2$ reflects variation in SWE between clear field and forest transects, as well as along transects at a given station. The range (the distance beyond which station values are not spatially correlated) was estimated at 200 km for both dates, based on a flattening of the SVs at that distance. Beyond about 500 km, both SVs increase, indicating change in the statistical moments of SWE across the large expanse of the study region (spatial non-stationarity). The analysis assumed that the non-stationarity was due to a spatial trend in SWE, but that variability around such a trend, and co-variability as a function of distance, were consistent in space (stationary covariance). Ordinary kriging (OK) over a moving data neighbourhood was applied, avoiding detailed analysis of non-stationarity in the mean (Deutsch & Journel, 1998). The remaining parameter of the SV model, the positive variance contribution (sill), was set at 1000 mm$^2$ for 20 January and 2500 mm$^2$ for 20 February. The Geostatistical Software Library (GSLIB) routine $kb2d$ (Deutsch & Journel, 1998) was used for OK with isotropic spherical SVs and nugget as specified above to estimate average SWE for the $25 \times 25$ km EASE-Grid cells.

**Satellite Estimates of snow water equivalent**

The Scanning Multichannel Microwave Radiometer (SMMR) instrument on board the Nimbus satellite started acquiring passive microwave data in 1978 and stopped in 1987. The SSM/I on board the Defense Meteorological Satellite Program, launched in June 1987, has been continuing the snow measurements made by SMMR (Foster & Chang, 1993). Chang et al. (1987) developed an expression to derive $SWE$ [cm] for a uniform snowfield using two SMMR channels:

$$SWE = 4.8(T18_H - T37_H)$$

(2)
where $T_{18}$ and $T_{37}$ are brightness temperatures ($T_B$) [K] as measured in the 18- and 37-GHz channels, respectively, and the subscript $H$ indicates horizontal polarization. If $T_{18}$ is less than $T_{37}$, $SWE$ is assumed to be zero. For SSM/I $SWE$ retrieval, the expression is adjusted to:

$$SWE = 4.8 [(T_{19_H} - 5) - T_{37_H}]$$

(3)

where the 18-GHz $T_B$ in equation (2) is replaced with the 19-GHz $T_B (T_{19_H})$; subtracting 5 K to correct for bias with respect to $T_{18_H}$. The SSM/I estimates used in this study were obtained by applying equation (3) to pentad averages of the 19- and 37-GHz $T_B$.

RESULTS AND CONCLUSIONS

Maps of $SWE$ as estimated by both methods are shown in Fig. 3. On the left (Fig. 3(a and b)) are the block-kriged station values. On the right (Fig. 3(c and d)) are the SSM/I derived values for corresponding water pentads. The pairs of images are similar in magnitude at their centres, and both show an increase in $SWE$ from south to north.

Fig. 3 Maps of $SWE$ as determined by block kriging of the weighted-average FSU snow course station data (left), and from SSM/I brightness temperatures (right), for two dates in 1989.
Both methods show an overall increase in \textit{SWE} from 20 January to 20 February. However, the \textit{SWE} maxima around the mountainous rim of the basin in the block-kriged maps (Fig. 3(a and b)) do not appear in the SSM/I map; rather these areas appear as minima in Fig. 3(c and d).

Cross-sections of kriged and SSM/I-derived \textit{SWE} along sections A and B (Fig. 1) are plotted for the two dates (Fig. 4). The SSM/I values are plotted as stars, and the block-kriged values as open circles; the dashed lines indicate 2-sigma error estimates based on the kriging variance. Also included are topographic profiles of ground elevation along sections A and B, extracted from the digital elevation data (Fig. 1).
Section B shows a large increase in SWE from January to February in both the kriged and SSM/I estimates. In section A, a temporal increase in SWE is more apparent in the kriged than in the SSM/I estimates, particularly between 2500 and 3500 km. If the SSM/I value lies within the error bars on the kriged value, one cannot conclude that the two estimates differ at a 95% confidence level (assuming a normal distribution of errors). Under this interpretation, the estimates agree all along cross-section A for both dates; the bias increases from January to February, however. Where cross-section B crosses the Ural Mountains (2000–3000 km) there is a marked difference in the two estimates. Between 3000 and 4500 km, the section B estimates agree quite well in February, less well in January. Beyond 4500 km, the estimates diverge on both dates, although the SSM/I values still lie within the 2-sigma error bars of the kriged estimates.

The error bars in Fig. 4 can only be used to judge a difference or lack of difference between the two estimates, not the accuracy of either. Nonetheless, the kriged estimates should be fairly accurate in regions of high station density. The block-kriged 25-km SWE and the independent estimates based on SSM/I brightness temperatures agree fairly well in the flat, low-lying portions of the Ob' basin, notably between 3500 and 4200 km on Section A, where stations are dense.

Estimating SWE in remote mountainous regions remains problematic. Section B cuts through a region of fairly high station density; the kriged values are probably more accurate than the SSM/I values where they differ significantly between 2000 and 3000 km (Ural Mountains). Given that the SMMR and SSM/I snow depth and SWE retrieval algorithms were developed assuming a uniform snow field (Chang et al., 1987), it is not surprising to conclude that the SSM/I estimates are less accurate in regions of complex terrain. Snow in the high Altai Mountains, a major source of the Ob’s spring and summer flow, is not well represented by station surveys; the SSM/I retrieval algorithm does indicate local SWE maxima in this region, but comparison or validation is difficult due to the lack of surface stations. It is uncertain what station data will be available in real-time and in the future for hydrological modelling in this region. Further study will analyse the differences between kriged and SSM/I SWE estimates, over a range of elevations with good station density, to develop and adjust the SSM/I algorithm for SWE estimates in complex terrain.

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REFERENCES


