

# **Development of the Pan-Arctic Snowfall Reconstruction: new land-based solid precipitation estimates for 1940-1999**

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## ABSTRACT

A new product, the Pan-Arctic Snowfall Reconstruction (PASR) is developed to address the problem of cold season precipitation gauge biases for the 1940-1999 period. The method used to create the PASR is different from methods used in other large-scale precipitation data products and has not previously been employed for estimating pan-arctic snowfall. The NASA Interannual-to-Seasonal Prediction Project Catchment Land Surface Model is used to reconstruct solid precipitation from observed snow depth and surface air temperatures. The method is tested at four stations in the United States and Canada where results are examined in depth. Reconstructed snowfall at Dease Lake, British Columbia and Barrow, Alaska is higher than gauge observations. Reconstructed snowfall at Regina, Saskatchewan and Minot, North Dakota is lower than gauge observations, probably because snow is transported by wind out of the Prairie region and enters the hydrometeorological cycle elsewhere. These results are similar to gauge biases estimated by a water budget approach. Reconstructed snowfall is consistently higher than snowfall from ECMWF Reanalysis-40 but does not have a consistent relationship with snowfall derived from the WMO Solid Precipitation Intercomparison Project correction algorithms. Advantages of the PASR approach include (1) the assimilation of snow depth observations captures blowing snow where it is deposited and (2) the modeling approach takes into account physical snowpack evolution. These advantages suggest the PASR product could be a valuable alternative to statistical gauge corrections and that arctic ground-based solid precipitation observing networks might emphasize snow depth measurements over gauges.

## 1. Introduction

The arctic water cycle is a key component of the global climate system and one which is changing quickly (Vörösmarty et al., 2001). Within this cycle, snowfall is a dominant feature which modulates radiative balance (Wang and Key, 2003), air temperature (Groisman et al., 1994), ground temperature (Stieglitz et al., 2003), geomorphology (Mackay, 1987; Grab, 2005), biogeography (Sturm et al., 2001; Chapin et al., 2005), large scale climate dynamics (Barnett et al. 1989; Cohen and Entekhabi 2001; Gong et al. 2002, 2003a, 2003b, 2004) and runoff (Lammers et al. 2001, Peterson et al. 2002, McClelland et al. 2004). Observations show long-term trends in all of these features (Chapman and Walsh, 1993; Serreze et al., 2000; IPCC, 2001; ACIA, 2004). The amount and distribution of snowfall throughout the Arctic is impacted by each of these climatic factors in turn through complex feedbacks.

Of all these factors, snowfall may be the most difficult to measure. The simplest bucket gauge and the most sophisticated shielded gauge share similar systematic biases: they disrupt the normal flow of snow in the atmospheric boundary layer. Snow preferentially falls away from the gauge. In addition, there are many unsystematic biases like unrepresentative gauge siting, mechanical failures, snow freezing around the gauge, etc. (Goodison et al., 1998). Improving estimates of frozen precipitation is a priority for several multi-agency programs including the Global Energy and Water Cycle Experiment (GEWEX, <http://www.gewex.org/>), Study of Environmental Arctic Change (SEARCH) (Morison et al., 2001), the Northern Eurasia Earth Science Partnership Initiative (NEESPI) (<http://neespi.gsfc.nasa.gov/science/science.html>), and the Arctic Community-

wide Hydrologic Analysis and Monitoring Program (<http://arcticchamp.sr.unh.edu/>). Finally, estimates of snowfall are crucial for predicting water resources in snow-dominated hydrologic regimes (Cherry et al., 2005a; Barnett et al., 2005).

Previous efforts to estimate “true” precipitation, liquid and frozen, include correcting gauges according to the amount of precipitation they catch relative to a well-tested reference gauge or a pit. This correction is typically a regression function based on meteorological parameters such as wind speed and temperature and has shown gauge errors of up to 80-120% for snowfall (Yang et al., 1998, 2000, 2005). Biases for liquid precipitation are smaller by an order of magnitude (Sevruk and Hamon, 1984) and will not be addressed here. The World Meteorological Organization (WMO) supported an extensive effort by Goodison et al. (1998) and contributors to compare solid precipitation between different gauges in this way. Corrections published in this report have led to a series of new solid precipitation estimates (Yang et al. 1998, 2000, 2005). Adam and Lettenmaier (2003) published an excellent summary of gauge corrections, including those by Legates and Willmott (1990), Groisman (1998), Groisman et al. (1991), Mekis and Hogg (1999), and Willmott and Matsuura (2001), and show how they can be used to adjust gridded precipitation datasets. Problems associated with the statistical correction approach may be that it does not account for gauge siting and other unsystematic biases (Goodison et al. 1998, Cherry et al. 2005b).

Other efforts to estimate solid and liquid precipitation include radar measurements (Marshall and Gunn, 1952), merged gauge-satellite analysis products (Xie and Arkin, 1997; Huffman et al. 1997), budgetary approaches (Fekete et al., 1999, 2004; Berezovskaya et al., 2004), and global climate reanalyses (Kalnay et al., 1996; Uppala et

al., 2005). Disadvantages to these approaches include a relatively short satellite era, errors associated with the indirect measurement of precipitation from satellites (Arkin and Ardanuy, 1989), inconsistent levels of assimilated observations in the reanalyses (Hagemann et al., 2005), and errors associated with interpolating or assimilating sparse observations of a heterogeneous substance to large grid cells (Hulme and New, 1997).

Our approach to estimating precipitation is different from the above-mentioned methods used in other large-scale precipitation data products and has not previously been employed for estimating pan-arctic snowfall. We focus on solid precipitation over the arctic hydrologic domain, because of the large uncertainties there. We use historical observations of daily snow depth, the NASA Seasonal-to-Interannual Prediction Project Catchment Land Surface Model, observed meteorological forcing, and a simple data assimilation technique to reconstruct the amount of snow which must have fallen to produce the observed snow depth, taking into account compaction, sublimation, evaporation, and other physics of the snowpack. The resulting product will be the Pan-Arctic Snowfall Reconstruction (PASR). In the following pages we will briefly describe the proof-of-concept study published in Cherry et al. (2005b); describe the model and the datasets used for forcing and restoring; compare reconstructions from four different stations in the arctic catchment to gauge, corrected gauge and gridded snowfall products; and discuss the results. Comparison of the full domain of the PASR product (2988 stations) to other arctic solid precipitation products will be reserved for a separate paper so that the four test stations can be studied in depth here.

## **2. Methods**

*a. Proof of concept*

The snowfall reconstruction method used to create the PASR was developed and tested using data from Reynolds Creek Experimental Watershed (RCEW) in Idaho, U.S.A. and is thoroughly described in Cherry et al. (2005b). RCEW is not in the arctic hydrologic catchment, but was used for the proof-of-concept because it receives a large amount of solid precipitation and has had a well-documented, hourly climate and snow water equivalent record for nearly 20 years. Because geography and meteorological conditions vary widely across the pan-arctic domain, the present paper demonstrates the method at four arctic sites with differing conditions. Throughout this paper the terms 'arctic' and 'the Arctic' will refer to the domain of the arctic hydrological catchment and the overlying atmosphere. We reconstruct snowfall by forcing a snow model with observed meteorological data. In this case, we use gauge precipitation as an initial guess of the actual precipitation ( $\mathbf{P}_{ini}$ ). As snow accumulates, the snowpack compacts and is subjected to evaporation, condensation, sublimation, and other snow physics represented in the model. At the end of each time step, the model snow depth is compared to the observed snow depth and the difference is considered "excess" ( $\mathbf{S}$ ), which can be positive or negative. Assuming a density of freshly fallen snow of  $150 \text{ kg}\cdot\text{m}^{-3}$ , this difference is then added to the initial guess of solid precipitation to reconstruct how much actual snowfall must have occurred ( $\mathbf{P}_{rec}$ ). This can be written as follows:

$$\mathbf{P}_{rec} = \mathbf{P}_{ini} + \mathbf{S} + e \quad (1)$$

where  $e$  represents all physical and numerical errors in the reconstruction. Physical errors are associated with model snow physics that do not precisely match the actual snow physics at any given time. An example of a physical error is when the modeled compaction of the snowpack does not match the observed compaction. Numerical errors are those introduced by the restoring procedure and are shown to be small (Cherry et al., 2005b), even for the large snowfall amount that occurs at RCEW. While excess ( $S$ ) may be negative (i.e. the initial guess of snowfall is an overestimate of the actual amount), net reconstructed solid precipitation can only be zero or positive.

This method was shown to reduce errors of snowfall estimates by 30% over corrected gauge observations at RCEW, when a number of meteorological parameters were used to drive the evolution of the snowpack, including surface air temperature (SAT), precipitation, incoming shortwave radiation, downwelling longwave radiation, relative humidity, and wind speed (Cherry et al., 2005b). However, SAT and gauge precipitation, along with snow depth, are the only abundant long-term meteorological observations available in the Arctic. This data constraint was tested at RCEW by forcing the land surface model with only SAT and assuming that the snow or soil surface temperature is equal to SAT, which resulted in up to 3% additional error per snow month. While errors introduced by lowering the number of snowpack-forcing parameters are significant (an average of 24% per year for an eight month snow season at RCEW), they are much smaller than the typical values of gauge undercatch in the Arctic of 80-120% (Yang et al., 1998, 2000, 2005). The error estimates at RCEW were made possible by the

high level of consistency between snow water equivalent from a snow course and a snow pillow, which provided a good estimate of “true” precipitation.

*b. Arctic reconstruction*

1. MODEL

The model used to reconstruct solid precipitation is the NASA Seasonal-to-Interannual Prediction Project (NSIPP) Catchment Land Surface Model (CLSM). This is a state-of-the-art land surface model based conceptually on TOPMODEL (Bevin and Kirkby, 1979), which uses the catchment as the fundamental unit of hydrological response. A topographic index in the model is used to estimate subgrid-scale heterogeneity in soil saturation, snow cover, etc. This topographic index is calculated from a 30-arc sec resolution (approximately 1 km) digital elevation model from the U.S. Geological Survey. The snow model in CLSM is that of Lynch-Stieglitz et al. (1994) and has been used successfully in a number of applications (Lynch-Stieglitz, 1994; Rind et al., 1997; Stieglitz et al., 1999, 2000, 2001, 2003; Déry et al., 2004, 2005a, 2005b, Cherry et al., 2005b). This model has three snow layers and includes growth and ablation processes such as melting, refreezing, compaction, sublimation, and heat exchange with a six-layer thermodynamic soil model; details are described in the 1994 paper. As in Cherry et al. (2005b), the major modification to the model of Lynch-Stieglitz (1994) is that albedo is modeled explicitly. The model was also modified to employ a specified boundary condition for the snow/soil surface temperature, which was SAT. The equations

for snow albedo are based on Hansen et al. (1983), as described by Stieglitz et al. (2001). When snow is freshly fallen (density =  $150 \text{ kg}\cdot\text{m}^{-3}$ ), albedo is 0.82 and when snow has aged and compacted it has a minimum albedo of 0.50 at 50 days. This parameterization is based on observations by the U.S. Army Corps of Engineers (1956). Further information about the physics and parameterizations of CLSM is given by Koster et al. (2000) and Ducharne et al. (2000).

## 2. DATA

Eight datasets are used to create the full PASR; a subset of these are used for the four test stations shown here in depth. Daily snow depth data are obtained from the Daily Snow Depth dataset (CSD; Environment Canada, 2000), Historical Soviet Daily Snow Depth version 2.0 (HSDSDv2; NSIDC, 1999), Global Surface Summary of the Day (GLOBALSOD; NCDC, 2005), and First Order Summary of the Day (FSOD; NCDC, 1998). Daily gauge precipitation and temperature minimum and maximum data were obtained from GLOBALSOD, FSOD, the Global Daily Climatology Network, v.1 (GDCN; NCDC, 2002), the Global Historical Climatology Network, v. 1 (GHCN; NCDC, 2004), and the Canadian Daily Climate Data (CDCD; Environment Canada, 2002). Elevation for the snow depth stations was obtained from ETOPO5 (NGDC, 1993). Duplicate stations were identified by national or WMO station identification number and/or location to a tenth of a minute. The location of all pan-arctic snow depth stations are shown in Figure 1, as well as the delineation of the arctic hydrologic catchment. Figure 1 also shows the locations of the four sample stations where results are analyzed

in depth (identified by stars). The catchment mask is from R-ArcticNet v3.0 (Lammers and Shiklamanov, 2000, updated). Figure 2 shows the number of snow depth, precipitation, and surface air temperature stations as a function of time in seven of the major arctic drainage basins and the total for all pan-arctic river basins. The total number of reconstructed snowfall time series is 2988, which is approximately the number of independent (i.e., non-duplicate) stations in the different snow depth datasets that fall inside the pan-arctic hydrologic catchment. Finally, comparisons are made between the four PASR test stations and 1) solid precipitation in the corrected gauge dataset by Yang et al. (2005) available at <http://www.uaf.edu/water/faculty/yang/yang.html>; 2) the snowfall fields from the European Centre for Medium-Range Weather Forecasts (ECMWF) Reanalysis-40, hereafter “ERA-40” (ECMWF, 2003; Uppala et al., 2005); and 3) gauge precipitation biases developed through a water balance approach by Fekete et al. (1999, 2004).

### 3. RECONSTRUCTION

The reconstructions are done by running the NSIPP CLSM at points where daily snow depth and SAT time series exist. During initial development of the PASR the authors did not wish to extrapolate snowfall over entire catchments or spatially interpolate between stations. For this reason, the NSIPP CLSM is run as a column (1-D) which has three sub-partitions to statistically represent sub-grid scale heterogeneity in snow and soil properties such as water saturation. Future use of the distributed (3-D)

capabilities of the NSIPP CLSM with the Déry and Yau (2002) blowing snow model is remarked upon in the Discussion and Conclusions section below.

At each snow depth station, we use daily precipitation and SAT forcing for the reconstruction from the nearest available station. The location of forcing data is kept the same for the entire period; if two stations are equidistant from the snow depth station the station with more existing data is used. In most locations, precipitation and minimum and maximum SAT were observed at approximately the same location where the snow depth was measured. Figures 3 and 4 show the distance of the precipitation and SAT forcing observations from the snow depth station, according to the location metadata associated with each dataset. Distances between stations were calculated along geographic Great Circles, but elevation differences were not considered.

While precipitation is referred to as a forcing variable, it should be noted that this method is not sensitive to the initial guess of precipitation. An alternative initial guess for precipitation could easily be the change in observed snow depth multiplied by the density of freshly fallen snow. Precipitation forcing was interpolated from daily observations down to a 20-minute time-step for the model runs. Precipitation was assumed to be distributed evenly throughout the day. SAT was interpolated linearly from daily temperature minimum and maximum. Daily temperature maximum was assumed to occur two hours after local solar noon and minimum was assumed to occur one hour before local sunrise. These interpolation algorithms are kept extremely simple because there are few hourly measurements from the Arctic to support the use of more complex schemes. Also, strong diurnal cycles of temperature and precipitation are associated with atmospheric convection (Dai et al., 1999). Such convection is suppressed in the Arctic by

the strong atmospheric inversion, which is caused by high surface albedo and low incoming radiation, and typically exists for much of the cold season (Serreze and Barry, 2005). Units for precipitation, SAT, and snow depth were converted to millimeters, Celsius, and meters, respectively.

It should be emphasized that while the model is run with a 20-minute time step, the intrinsic scale of the reconstruction is limited by the daily snow depth measurement. The availability of SAT minima and maxima provide some sub-daily information for forcing the snow model, but the assumption of precipitation occurring evenly throughout the day is unrealistic. Instead, the sub-daily time step should be thought of as Newtonian relaxation or nudging toward the observed snow depth value. A 20-minute to one-hour relaxation time scale has been used successfully in a number of studies employing assimilation of synoptic observations (Kalnay, 2003).

Missing daily data between 1 January, 1940 and 31 December, 1999 were infilled in the following way. For daily temperature or precipitation data, if three days in a row or fewer are missing, they are linearly interpolated from the temporally neighboring values. If more than three days are missing, the missing days are replaced by monthly climatological values. The threshold of three days was chosen because it is approximately the time of a synoptic event. For snow depth, the threshold was extended to 16 days because of the long residence time of snow on the ground, relative to precipitation.

The model runs are done by specifying the snow surface temperature or soil surface temperature, when there is no snow, with SAT. The initial guess of precipitation (solid and liquid) is made from gauge observations. Solid precipitation is reconstructed by restoring to observed snow depth and adding the excess (positive or negative), though

reconstructed snowfall is constrained to be positive. Finally, precipitation is only reconstructed between 1 October and 31 May, to avoid mixed-phase precipitation summer events. While there are still such events during the reconstruction period, especially during the shoulder seasons, they impact the snow pack and cannot be discounted.

Comparisons are made between the reconstructions and Yang et al. (2005) corrected gauge data. Monthly precipitation totals were used, multiplied by the percent of precipitation that occurred as snow (these percentages are provided with the Yang et al. dataset). Comparisons were also made between the reconstructions and ERA-40 snowfall data. ERA-40 and its predecessor ERA-15 have been analyzed and compared with other products specifically for the arctic domain (Walsh et al., 1998; Cullather et al., 2000; Serreze and Hurst, 2000; Bromwich et al., 2002; Genthon, 2002; Serreze and Etringer, 2002; Serreze et al., 2003) and have been used in polar climate studies (Betts et al., 2003; Déry and Wood, 2004). Finally, the differences between the reconstructed snowfall and the gauge snowfall (gauge biases) are compared to gauge precipitation biases from Fekete et al. (1999), which are calculated from observed river discharge and a water balance model (WBM).

### **3. Results at Four Arctic Stations**

Four stations in the arctic hydrological catchment were chosen for examining the PASR approach in depth. These locations were chosen because their records extend throughout all or most of the 60-year period and they were also at or near test sites which

were used to develop the WMO gauge corrections. These stations, Dease Lake (British Columbia) and Regina (Saskatchewan), Canada, and Barrow (Alaska) and Minot (North Dakota), United States, vary considerably in their geography and climate. Dease Lake ( $58.42^{\circ}$  N,  $130.00^{\circ}$  W) is in Northern British Columbia, east of Juneau, Alaska on the Pacific-Arctic hydrological Divide. The elevation of the snow station is 625 meters above sea level. The station is situated in the Tanzilla Valley, in the Cassiar Mountains. Regina ( $50.42^{\circ}$  N,  $104.62^{\circ}$  W, 526 m ASL) and Minot ( $48.25^{\circ}$  N  $101.28^{\circ}$  W, 511 m ASL) are both located on the flat, open North American Prairie. These areas are subject to high winds for much of the winter, which transport significant amounts of blowing snow. Barrow ( $71.30^{\circ}$  N,  $156.78^{\circ}$  W, 3 m ASL) has a coastal marine climate, being located on the Beaufort Sea, Arctic Ocean. These stations typify several of the important snowfall regimes in the arctic hydrologic catchment.

Figures 5a and 5b show cumulative monthly reconstructed snowfall at Dease Lake, plotted against the gauge record and the corrected gauge record. During the early part of the 1950s, significant portions of the gauge records (and therefore reconstruction forcing) are infilled with climatology. There are also months when significant portions of the temperature data or snow depth are also missing, such as in the early 1940s. This may account for the large differences between the reconstruction and the gauge record, particularly in the early part of the snow season. Starting in the late 1950s, the early season reconstructed snowfall is nearly the same as gauge estimates. During the later part of the season, the reconstructed snowfall at Dease Lake is far greater than that which was measured by the snow gauge. In other words, the observed snowpack is much deeper than a model snowpack built from observed meteorological conditions. Potential reasons for

this will be discussed further in the following section. The corrected snow gauge shows even less than the raw gauge data at times, particularly during water years 1995-1996, 1996-1997, and 1998-1999. The ERA-40 product was chosen for additional comparison because of its wide usage in the climate science community. At Dease Lake, ERA-40 snowfall is significantly less than either the gauge or the reconstructed snowfall during all years.

Figures 6a and 6b show results for Regina. This record shows significantly less reconstructed snowfall, particularly in the early part of the snow season. The corrected gauge record is generally slightly higher than the gauge record, though in 1982, it is nearly double. At Regina, there is no consistent relationship between ERA-40 and reconstructed snowfall, though it is nearly always less than the gauge snowfall.

Figures 7a and 7b show the reconstruction results for Barrow. The Barrow reconstruction shows more frozen precipitation than gauges some years and less in other years, until the 1990s, when reconstructed snowfall is generally much greater than gauge snowfall. The corrected gauge record shows a consistently higher amount of snowfall than the raw data record shows, however there are several years (1975-76, 1976-77, 1985-86, 1999) during which extreme snowfall rates were calculated by Yang et al. (2005). These are probably unrealistic and it is unclear what the source of these errors might be as they are not associated with high monthly values of wind speed. Yang has suggested that raw daily data quality is likely to be the problem (pers. comm.) and more reliable estimates from Barrow can be made from the detailed studies done separately at this station (Yang et al. 1998; Benning and Yang 2005). At Barrow, ERA-40 is

sometimes less than and sometimes greater than either the reconstructed snowfall or the gauge snowfall.

Figures 8a and 8b show the results for Minot. In general, the reconstruction is lower than the gauge precipitation, with the exception of the mid-to-late 1990s. The gauge correction is accounting for overcatch some years and undercatch in others. Two years in which the gauge correction calculates 50% or more undercatch (1974-75, 1983-84), the reconstruction is nearly identical to the gauge record. In a third year of high corrected gauge snowfall (1996-1997), the total reconstructed snowfall is about 180 mm less than the corrected value, but is still more than twice the gauge observation. The infilling climatology here produces no reconstructed precipitation. This means that the observed monthly mean climatological snow depth is much shallower than the one produced by modeling the depth with the observed SAT and precipitation. As in Regina, there is no consistent relationship between ERA-40 and reconstructed snowfall at Minot, though it is nearly always less than the gauge snowfall.

To elucidate some of the physical reasons behind the differences in reconstructed solid precipitation and solid precipitation from gauges, the non-zero daily differences (reconstructed minus gauge precipitation) were plotted against daily wind speed and mean temperature for each station in Figure 9. The daily wind speed data is from GLOBALSOD dataset and is only available from 1994-1999. The daily mean temperatures were used for the entire period 1940-1999. There is no clear relationship between these differences and wind speed or magnitude of temperatures below about -2 °C. Above -2 °C, however, there appears to be a relationship between positive snowfall errors and temperature, particularly at the Dease Lake station. Evidence at the other

stations for this relationship is much weaker. While the reconstruction method only calculates frozen precipitation, there appear to be a large number of rain-on-snow events at Dease Lake. As discussed in Cherry et al. 2005b, the model snowpack may not be metamorphosing accurately during rain-on-snow events; this might introduce errors into the reconstruction and is discussed further below.

There is also no significant correlation between the percentage of missing data that is infilled with climatology and the absolute difference between the reconstruction and the gauge snowfall. The percentage of infilling varies for each forcing/restoring parameter and between the stations. This is shown in Table 1. Regina has the least missing data of all the stations and Minot has the most, for the period 1940-1999. Whether or not reconstructions made with infilled data should be used for any particular application is discussed further in the following section.

In Figure 10, cumulative snowfall climatologies are shown for each product at each test station. Months with climatological infilling were not used to calculate these climatologies. At Dease Lake and Barrow the annual total reconstructed snowfall is greater than the gauge snowfall and ERA-40 snowfall. However, the corrected gauge snowfall is much greater at Barrow, due in part to the extreme years mentioned above. A second corrected gauge climatology, ignoring these extreme errors, is plotted for Barrow. This climatology is still much higher than the other solid precipitation estimates. At Regina and Minot, the reconstructed snowfall is less than both the gauge and the corrected gauge records, but higher than ERA-40. The potential influence of blowing snow at these sites will be discussed below.

#### 4. Discussion

The reconstruction approach is an independent way to estimate snowfall. Snow depth measurements have different systematic and unsystematic biases than do gauges (Cherry et al., 2005b). Differences between the reconstructed snowfall and the gauge snowfall are not correlated with wind speed, or low ( $< -2$  °C) temperatures, but more positive errors seem to appear during warm temperatures ( $> -2$  °C), at least at Dease Lake, when mixed precipitation and rain-on-snow events may be occurring. This method does not consistently reconstruct more or less snowfall than is captured by the gauges at each of the four test stations, but is consistent with gauge precipitation biases calculated using water balance methods (Fekete et al., 1999) discussed below.

Both the corrected gauge and the reconstructed snowfall show significantly more snowfall than ERA-40 at all four stations. Serreze et al. (2005) state that "comparing (precipitation) output from reanalysis ... directly with station data is unwarranted because of large differences in spatial scale of the data." While the creation of a gridded version of PASR may show stronger correlations with the ERA-40 product, the gridding procedure can obscure certain physical processes which are apparent when examining station data. Whether or not the snow depth station data are representative of that at the scale of the grid cell will certainly vary from station to station and is not examined in Serreze et al. (2005).

Differences between the reconstruction and the gauge precipitation do not correlate with wind speed because both neglect snow transported away by wind. For gauges, however, that snow is never measured (high wind causes high undercatch) and this is one reason why corrections have been developed. For reconstructed snowfall, the

snowfall is counted, but not where it fell. It gets counted where it ended up, because it is captured in the snow depth measurement in the deposition area. However, this may be the quantity which is most relevant for studies of large-scale hydroclimatology. Blown snow contributes to runoff in the location where it is deposited. Snow which does not sublimate during the transport will sublimate or melt in the deposition region. In the case of the North American Prairies, some blown snow is probably transported out of the region.

The reconstruction climatologies at four test stations (Figure 10) were qualitatively compared to estimated gauge biases from Fekete et al. (1999). Figure 11, reproduced with permission from Fekete et al. (1999), shows the global gauge biases determined by the water balance approach in that study. The correction coefficient is the amount by which runoff output from their Water Balance Model (WBM) forced with gauge precipitation must be multiplied to get observed runoff. This can be written as

$$\mathbf{R}_{\text{obs}} = \xi \cdot \mathbf{R}_{\text{wbm}} \quad (2)$$

where  $\mathbf{R}_{\text{obs}}$  is the “observed” runoff,  $\xi$  is the correction coefficient and  $\mathbf{R}_{\text{wbm}}$  is the runoff produced by the model forced with gauge precipitation. This is a simplified explanation of what was done in Fekete et al. (1999). What they call  $\mathbf{R}_{\text{obs}}$  has actually been interpolated from the observations thought to have the lowest errors and topographic information, modeled river routing, etc. While comparisons for exact station locations are avoided because of the low spatial resolution of the WBM approach, this figure shows gauge biases with sizable undercatch ( $\xi > 1$ ) in the Arctic in general, modest

undercatch along the Canadian and American Rocky Mountains, and apparent overcatch on the North American Prairies ( $\xi < 1$ ). This is consistent with the PASR test results at the four test stations. Reasons for the apparent overcatch biases on the Prairies most likely relate to transport and divergence of blowing snow.

Figure 12 shows histograms of daily maximum sustained wind speed for the four test sites from 1994-1999. Dease Lake is considerably less windy than the other three sites and this is where the PASR method reconstructs the most snow over the corrected gauge. At approximately  $10 \text{ m}\cdot\text{s}^{-1}$  small amounts of snow can be transported by wind, while speeds of  $20 \text{ m}\cdot\text{s}^{-1}$  are associated with considerable amounts of blowing snow (Arya, 2001). Dease Lake has few days above  $10 \text{ m}\cdot\text{s}^{-1}$  while Barrow, Regina, and Minot all have a significant number of these events. This is consistent with blowing snow climatologies from Déry and Yau (1999). The likelihood of blowing snow divergence or convergence at a station clearly affects how the snowfall reconstruction should be interpreted. The reconstructions capture snow that was deposited in a region because it appears as a positive snow depth change. Negative depth changes are used to readjust the initial guess of daily precipitation, but limited by the constraint that daily precipitation is zero or positive.

The present development study has made it clear that to interpret the PASR product one needs to include information about daily wind speed, where and when it is available, as well as wind convergence and divergence. Because of the high winds, both Barrow and the Prairie stations are expected to show strong gauge undercatch biases. Statistical gauge corrections (based on windspeed) show significantly more snowfall at Barrow and Minot than the raw gauge data. However, the reconstruction climatologies in

Figure 10 show modest gauge undercatch for Barrow but significant overcatch for the Prairies--both under high wind conditions--suggesting blowing snow convergence in Barrow and divergence on the Prairies. A user must have some knowledge of large-scale patterns of snow transport if he or she is concerned about the source region of snow. Déry and Yau (1999, 2002) show climatological estimates of likely regions of blowing snow and its divergence.

Treatment of missing data is another issue which affects all climate datasets, including the PASR. Because the PASR approach is model-based, forcing and restoring data must be infilled to keep the model running continuously over the snow season. The simple correlation tests between error and missing data percentage described above do not fully capture the impact of data infilling on the snowfall reconstruction. For example, several days of consecutive infilling with climatology for snowdepth is likely to have a much larger impact than occasional infilling of temperature forcing.

Finally, trace events ( $\leq 0.2$  cm) are another challenge to estimating arctic precipitation. While the model has difficulty simulating trace snowfall events on bare ground (especially if snow melts before depth is observed), it does capture trace events well once the snow is more than 13 mm deep. This is, of course, provided the trace amount is measureable as an increase in snow depth. The method cannot account for trace amounts smaller than the precision of the depth measurement device unless multiple trace events aggregate to a measureable amount of snow.

There are few arctic stations where the kind of proof-of-concept done at RCEW could be repeated. It is possible to estimate the error of the RCEW reconstruction using hourly snow water equivalent from a snow pillow, which does not have the same

systematic undercatch biases as a gauge. This proof-of-concept was meant to inform the larger pan-arctic reconstruction, not to claim that the unsystematic biases at this site are the same as those at every arctic site. However, snow depth measurements are taken along a transect and averaged, taken from a permanent graduated stake, or simply made in the same general area each day. There are objective reasons to think that snow depth measurements may be less systematically biased than gauge observations, simply because the gauge is a large, disruptive device with design shortcomings and subject to mechanical errors. Regardless, snow depth varies significantly on all scales (Neumann et al., 2006), particularly with topography, and both measurements have unsystematic biases associated with siting. These potential siting biases are neither estimated nor overcome by the method used to develop the PASR dataset.

## **5. Conclusions**

The present development study describes a new approach for estimating arctic solid precipitation and explores the product's physical underpinnings, interpretation, and potential robustness. Advantages of this approach include (1) the assimilation of snow depth observations captures blowing snow where it is deposited and (2) the modeling approach takes into account physical snowpack evolution. These advantages suggest the PASR product could be a valuable alternative to statistical gauge corrections.

The dynamical snow model, within the NSIPP CLSM, is the strongest asset of the reconstruction approach. The meteorological forcing with temperature helps determine the evolution of the snowpack. Forcing with gauge precipitation creates an initial

estimate of the precipitation input and the resulting impact on the growth, compaction, ripening, or ablation of the snowpack. The final restoring to observed snow depth at the end of the time step corrects the initial estimate of snowfall. This is quite different than a statistical correction of snowfall from a standard gauge, based on its past catch relationship with a “true” gauge, as a function of meteorological parameters.

Future work by the investigators includes comparison of the full pan-arctic domain of the PASR with other arctic precipitation estimates (and implications of our new estimate for the arctic freshwater budget), distribution of the full PASR product, and further development of the PASR product using additional meteorological observations as forcings. The development of the PASR described in the present study has provided insight into the physical processes which may impact the robustness of the reconstruction approach. Providing information about the potential for mixed-phase precipitation, blowing snow, and missing forcing data with the PASR will allow the user to choose reconstruction data which are appropriate for his or her application.

Finally, results from the PASR may influence the design of future observation systems. As shown in Figure 2, the number of precipitation gauges in the Arctic has dropped dramatically since the 1980s. The same is true of temperature, and especially of snow depth, and runoff. This apparent trend is due in part to the fact that observations can take many years to become available to researchers. On the other hand, the number of stations operating has also declined dramatically in Russia and Canada, and to a lesser extent in the United States (Vörösmarty et al., 2001). The results here suggest that using depth to reconstruct snowfall is a viable alternative to corrected gauges. The implication for future station design is that unmanned precipitation gauges might be replaced by

automatic snow depth sensors. Some automated systems like the SNOTEL network in the western United States have already begun to add ultrasonic depth sensors to the existing station instrumentation (<http://www.wcc.nrcs.usda.gov/snow/snotel-depth.html>).

The Arctic is changing quickly. Most arctic research is related to snowfall in some way or another. The arctic freshwater budget is not balanced by direct observations of precipitation (Vörösmarty et al. 2001); we need better estimates of snowfall. The reconstruction approach is an alternative to corrected gauge records, is not subject to the same biases, and has a flexible, model-based framework in which to develop further.

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## Figure Captions

**FIG. 1.** Locations of the daily snow depth stations used in the Pan-Arctic Snowfall Reconstruction (PASR) dataset. Projection is polar stereographic. Scale denotes elevation in meters. The black line delineates the arctic hydrological catchment. Stations from the Global Summary of the Day dataset (NCDC, 2005) are shown as crosses because they are only available for the period 1994-1999. Stations shown as circles (sometimes beneath crosses) typically have long-term observations. Stars denote stations where the reconstruction results are examined in depth.

**FIG. 2.** Number of stations per year where daily observations are made, for major arctic subcatchments and the entire pan-arctic catchment. The thin line denotes precipitation gauges, the thick line denotes temperature stations, and the dashed line denotes snow depth stations. Note that the scale on the y-axis changes from basin to basin. The sharp decline of observations at the start of the 21st century is a combined effect of station shutdown and lags in data reporting.

**FIG. 3.** Distance of the precipitation “forcing” station from the snow depth “restoring” station in kilometers.

**FIG. 4.** Distance of the temperature “forcing” station from the snow depth “restoring” station in kilometers.

**FIG. 5a.** Cumulative monthly reconstructed snowfall, corrected gauge snowfall (Yang et al. 2005), gauge snowfall, and ERA-40 for Dease Lake in British Columbia, Canada. The Yang et al. record starts in 1977 at this station. The ERA-40 data start in 1957. During some years the monthly value of the corrected gauge and gauge snowfall are nearly equal.

**FIG. 5b.** Cumulative monthly reconstructed snowfall, corrected gauge snowfall (Yang et al. 2005), gauge snowfall, and ERA-40 for Dease Lake in British Columbia, Canada. The Yang et al. record starts in 1977 at this station. The ERA-40 data start in 1957. During some years the monthly value of the corrected gauge and gauge snowfall are nearly equal.

**FIG. 6a.** Cumulative monthly reconstructed snowfall, corrected gauge snowfall (Yang et al. 2005), gauge snowfall, and ERA-40 for Regina in Saskatchewan, Canada. The Yang et al. record starts in 1977 at this station. The ERA-40 data start in 1957. During some years the monthly value of the corrected gauge and gauge snowfall are nearly equal.

**FIG. 6b.** Cumulative monthly reconstructed snowfall, corrected gauge snowfall (Yang et al. 2005), gauge snowfall, and ERA-40 for Regina in Saskatchewan, Canada. The Yang et al. record starts in 1977 at this station. The ERA-40 data start in 1957. During some years the monthly value of the corrected gauge and gauge snowfall are nearly equal.

**FIG. 7a. Cumulative monthly reconstructed snowfall, corrected gauge snowfall (Yang et al. 2005), gauge snowfall, and ERA-40 for Barrow, Alaska, USA. The Yang et al. record starts in 1973 at this station. The ERA-40 data start in 1957. During some years the monthly value of the corrected gauge and gauge snowfall are nearly equal.**

**FIG. 7b. Cumulative monthly reconstructed snowfall, corrected gauge snowfall (Yang et al. 2005), gauge snowfall, and ERA-40 for Barrow, Alaska, USA. The Yang et al. record starts in 1973 at this station. The ERA-40 data start in 1957. During some years the monthly value of the corrected gauge and gauge snowfall are nearly equal.**

**FIG. 8a. Cumulative monthly reconstructed snowfall, corrected gauge snowfall (Yang et al. 2005), gauge snowfall, and ERA-40 for Minot, North Dakota, USA. The Yang et al. record starts in 1973 at this station. The ERA-40 data start in 1957. During some years the monthly value of the corrected gauge and gauge snowfall are nearly equal.**

**FIG. 8b. Cumulative monthly reconstructed snowfall, corrected gauge snowfall (Yang et al. 2005), gauge snowfall, and ERA-40 for Minot, North Dakota, USA. The Yang et al. record starts in 1973 at this station. The ERA-40 data start in 1957. During some years the monthly value of the corrected gauge and gauge snowfall are nearly equal.**

**FIG. 9. Difference between daily reconstructed snowfall and (uncorrected) gauge snowfall versus wind speed and temperature for four test stations. Plots on the left are for wind speed and plots on the right are for temperature. Stations, from top to bottom are Dease Lake, Regina, Barrow, and Minot.**

**FIG. 10. Cumulative annual snowfall climatologies for four test stations using four different datasets: the reconstruction, raw gauge data, corrected gauged data from Yang et al. (2005), and ERA-40. For Barrow two different climatologies are plotted for the corrected gauge dataset: one calculated without the extreme errors of 1975-1977 and 1999 (the lower curve) and one calculated with the errors (the upper curve).**

**FIG. 11. Runoff Correction Coefficients from the Water Balance Model, reproduced from Fekete et al. 1999. This shows the bias in gauge precipitation compared to measured runoff and a balanced water budget. Coefficients greater than one (blue to black) represent gauge undercatch while coefficients less than one (pink to red) represent gauge overcatch.**

**FIG. 12. Histograms of maximum sustained wind speed at four PASR test sites. These daily data are from the GLOBALSOD dataset, available for the period 1994-1999.**

	Dease Lake	Regina	Barrow	Minot
precipitation	9.58	0.01	17.5	19.49
snow depth	12.98	1.84	17.64	19.48
temp max	9.65	0.07	18.96	19.48
temp min	9.66	0.16	20.26	19.48

Table 1: percentage of missing daily data from 1 January, 1940-31 December, 1999 for the four test stations. These missing data have been infilled according to the procedure described in the text.



FIG. 1. Locations of the daily snow depth stations used in the Pan-Arctic Snowfall Reconstruction (PASR) dataset. Projection is polar stereographic. Scale denotes elevation in meters. The black line delineates the arctic hydrological catchment. Stations from the Global Summary of the Day dataset (NCDC, 2005) are shown as crosses because they are only available for the period 1994-1999. Stations shown as circles (sometimes beneath crosses) typically have long-term observations. Stars denote stations where the reconstruction results are examined in depth.

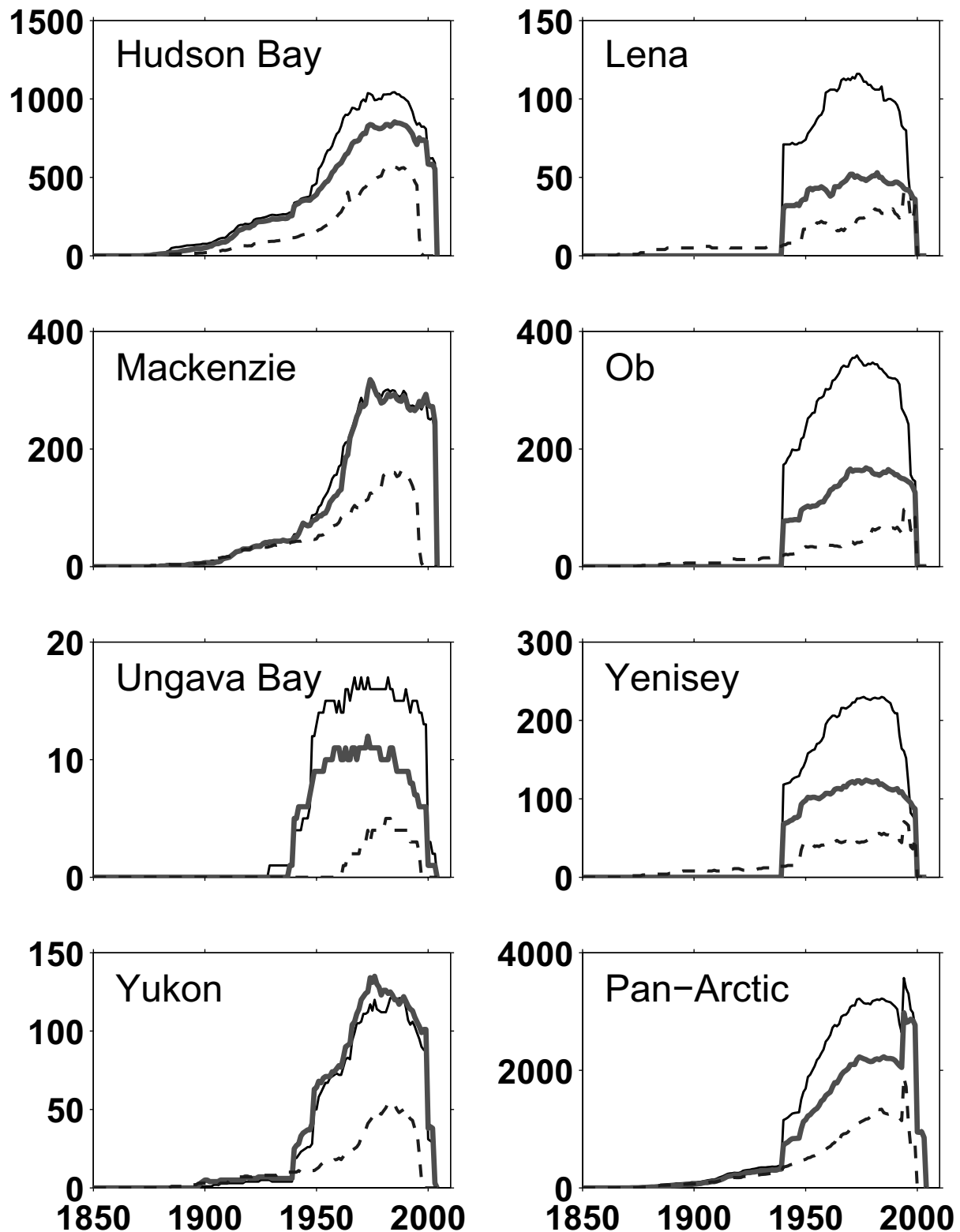


FIG. 2. Number of stations per year where daily observations are made, for major arctic subcatchments and the entire pan-arctic catchment. The thin line denotes precipitation gauges, the thick line denotes temperature stations, and the dashed line denotes snow depth stations. Note that the scale on the y-axis changes from basin to basin. The sharp decline of observations at the start of the 21st century is a combined effect of station shut-down and lags in data reporting.

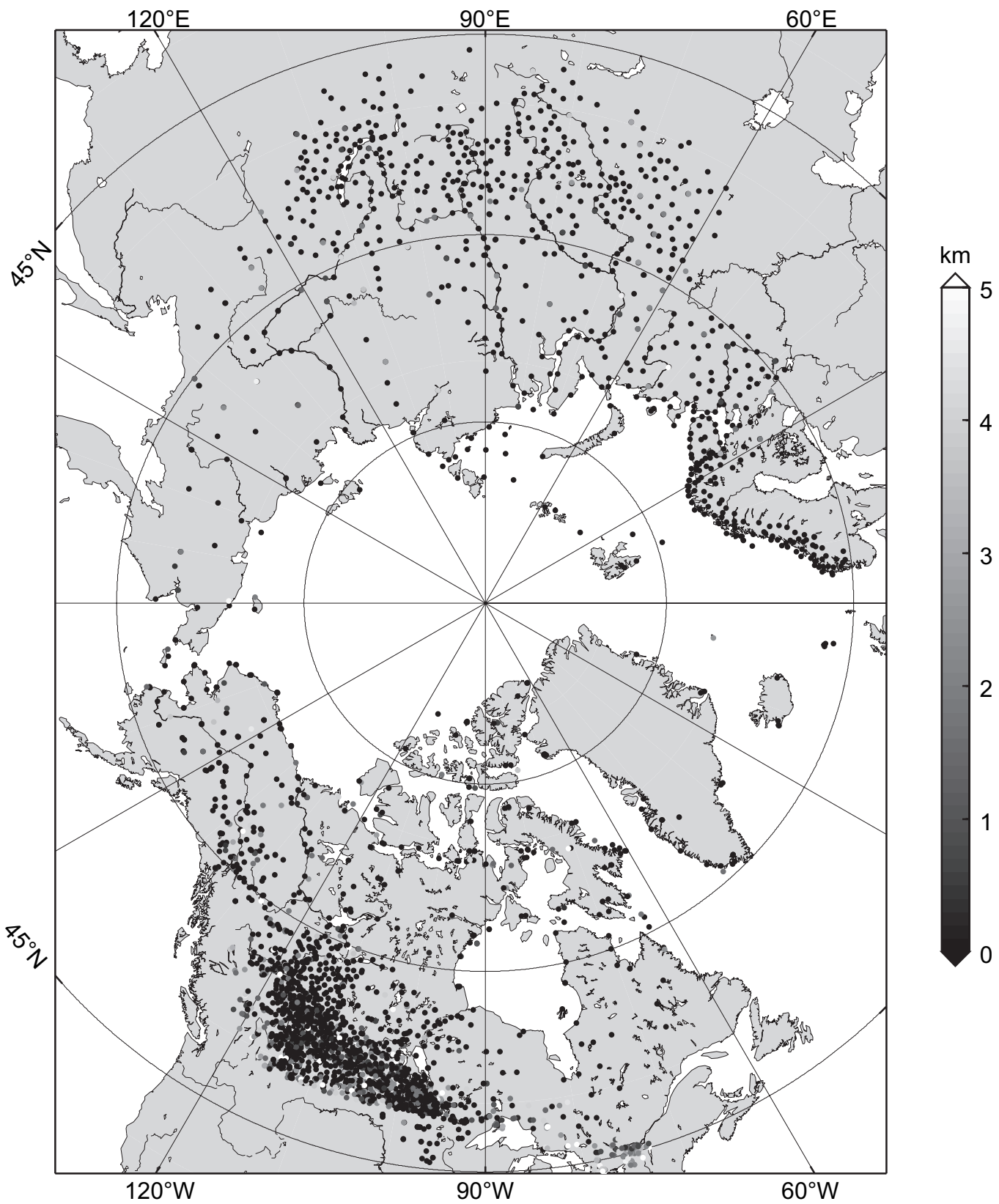


FIG. 3. Distance of the precipitation "forcing" station from the snow depth "restoring" station in kilometers.

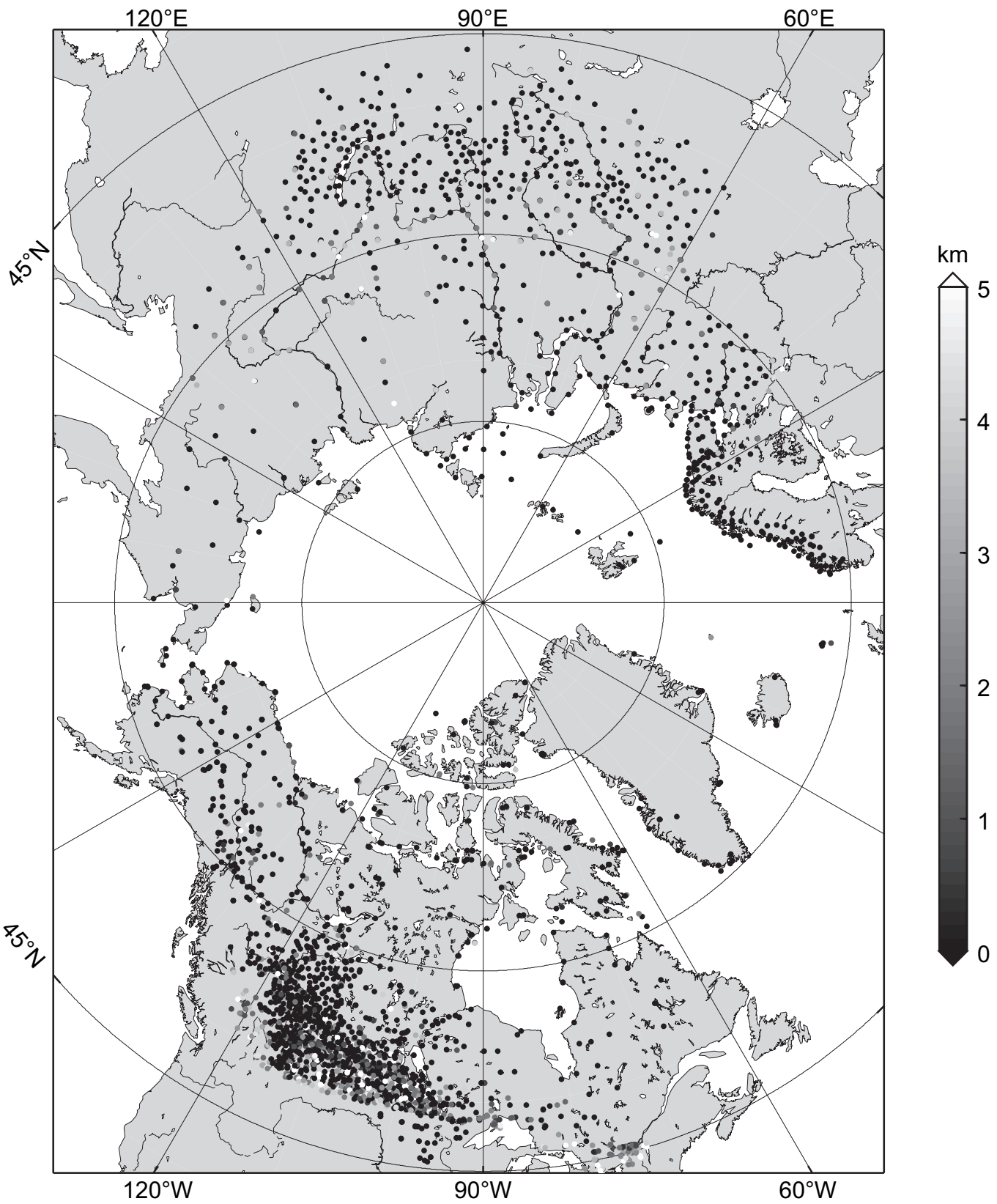


FIG.4. Distance of the temperature "forcing" station from the snow depth "restoring" station in kilometers.

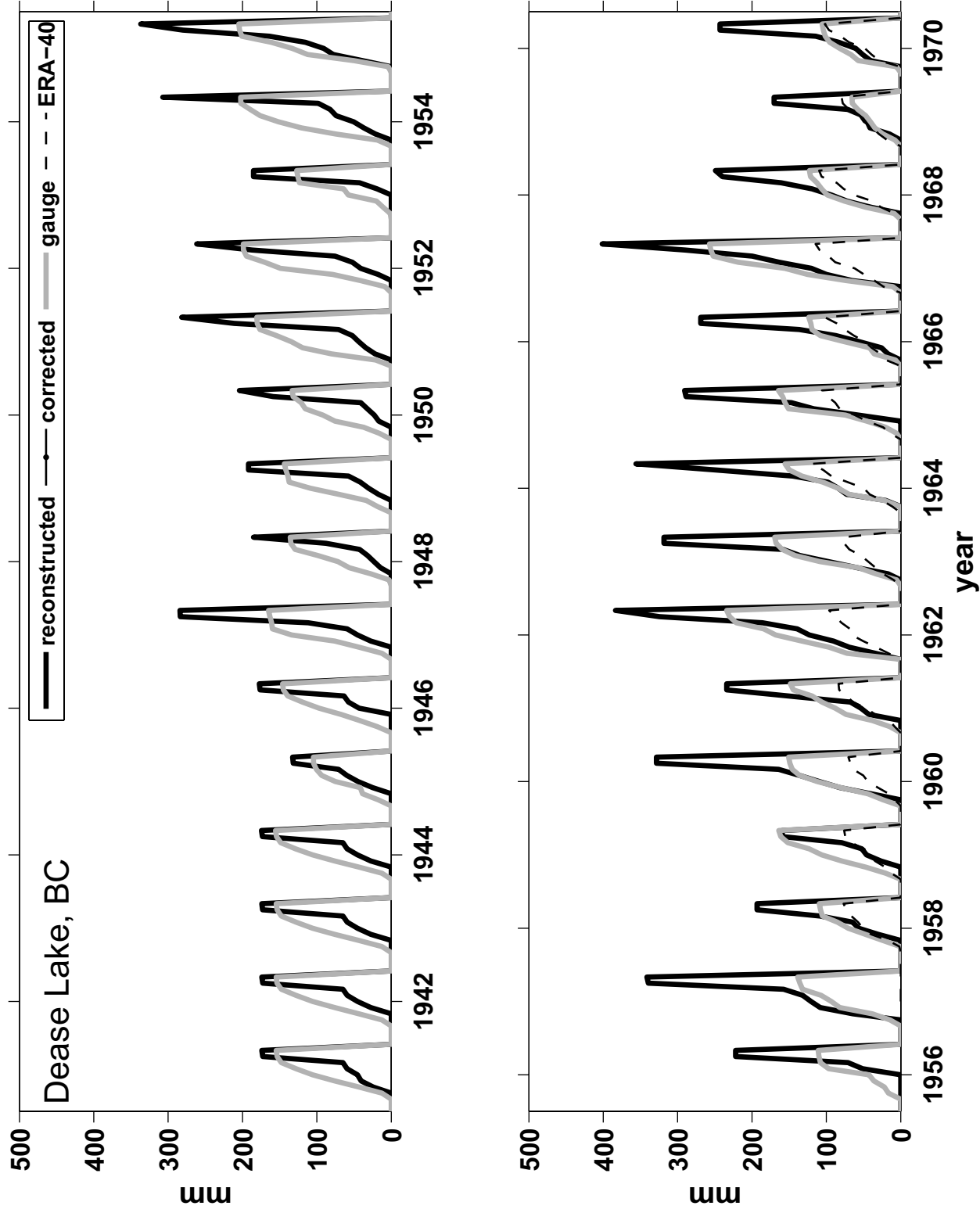


FIG. 5a. Cumulative monthly reconstructed snowfall (Yang et al. 2005), gauge snowfall, and ERA-40 for Dease Lake in British Columbia, Canada. The Yang et al. record starts in 1977 at this station. The ERA-40 data start in 1957. During some years the monthly value of the corrected gauge and gauge snowfall are nearly equal.

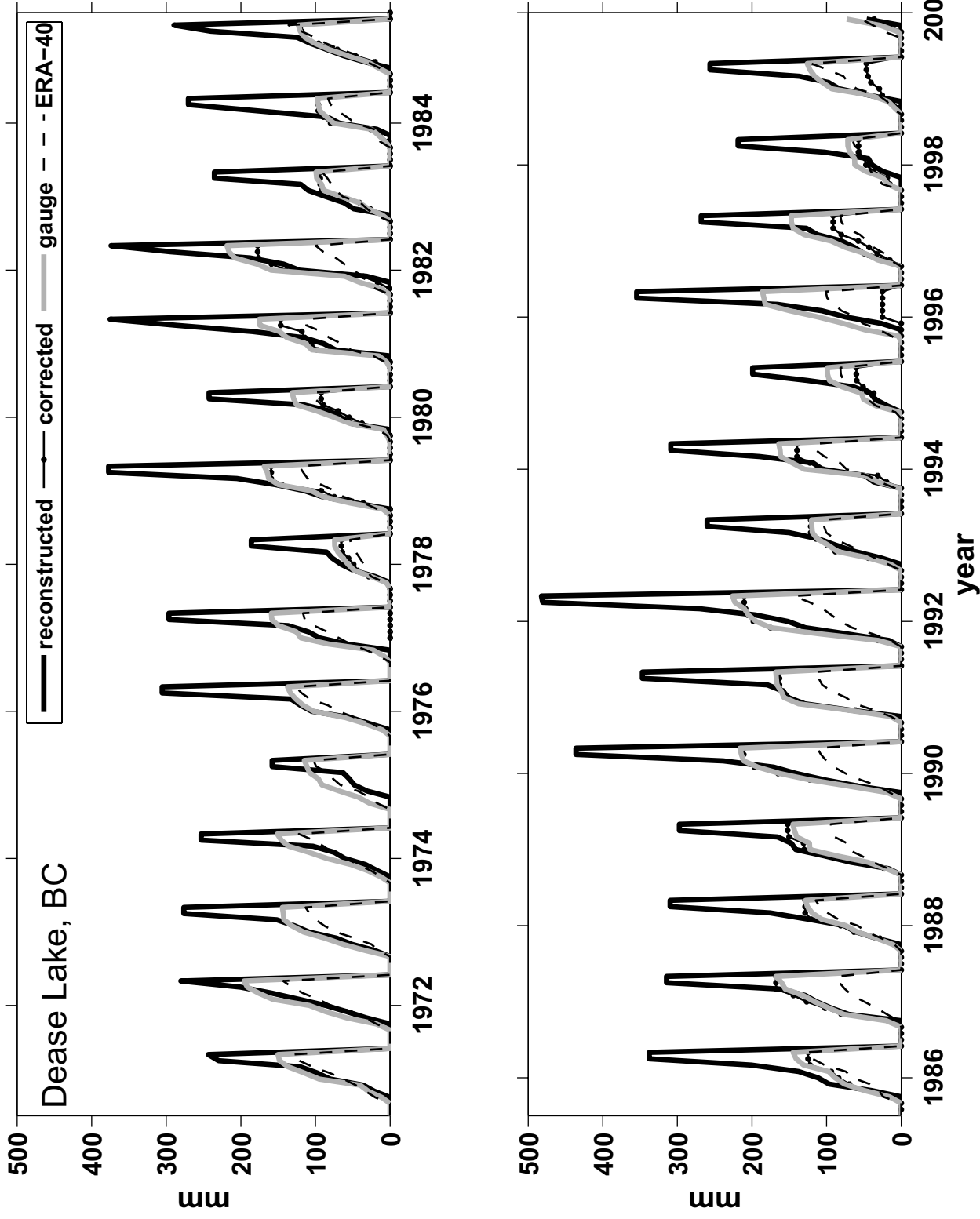


FIG. 5b. Cumulative monthly reconstructed snowfall, corrected gauge snowfall (Yang et al. 2005), gauge snowfall, and ERA-40 for Dease Lake in British Columbia, Canada. The Yang et al. record starts in 1977 at this station. The ERA-40 data start in 1957. During some years the monthly value of the corrected gauge and gauge snowfall are nearly equal.

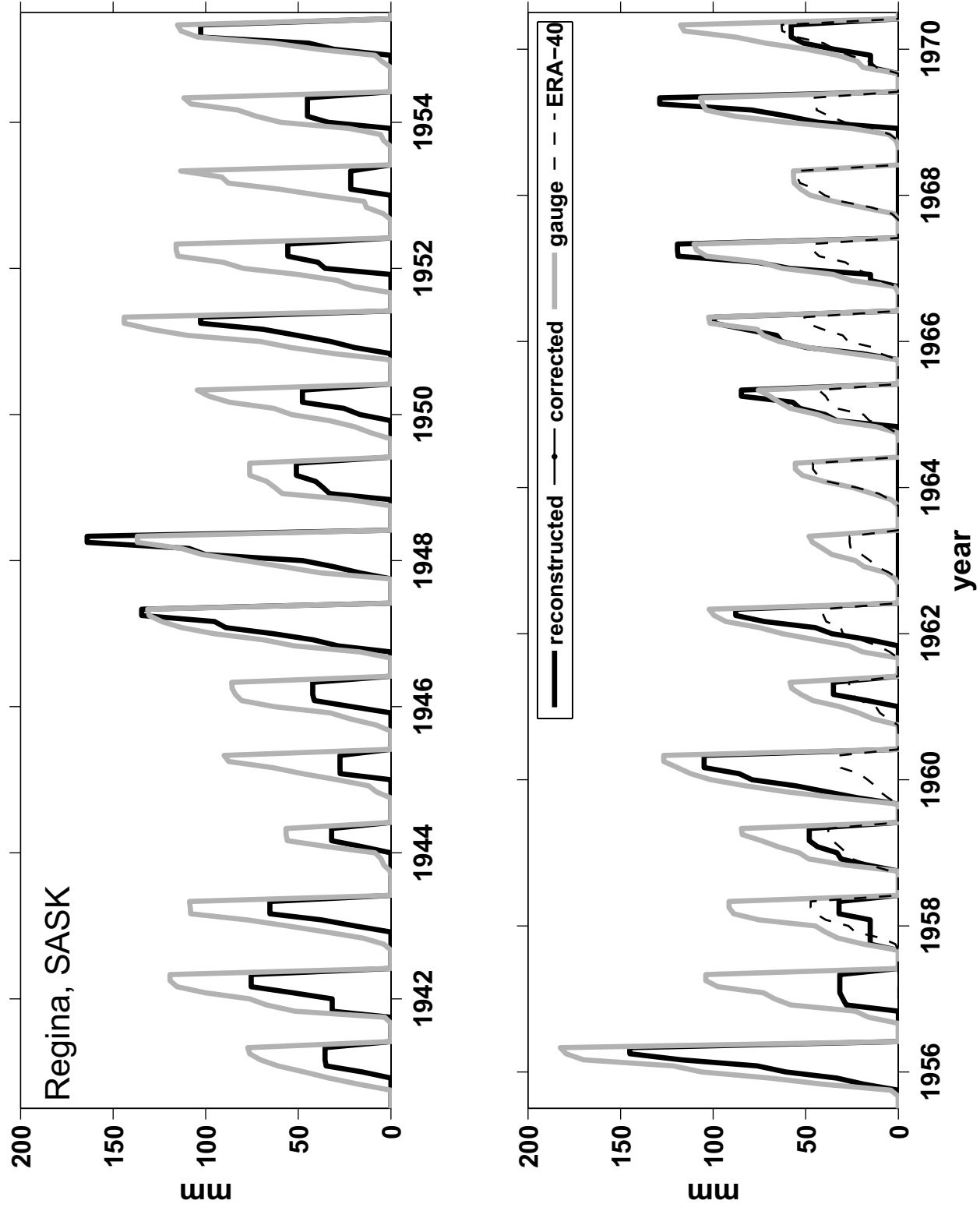


FIG. 6a. Cumulative monthly reconstructed snowfall, corrected gauge snowfall (Yang et al. 2005), gauge snowfall, and ERA-40 for Regina in Saskatchewan, Canada. The Yang et al. record starts in 1977 at this station. The ERA-40 data start in 1957. During some years the monthly value of the corrected gauge and gauge snowfall are nearly equal.

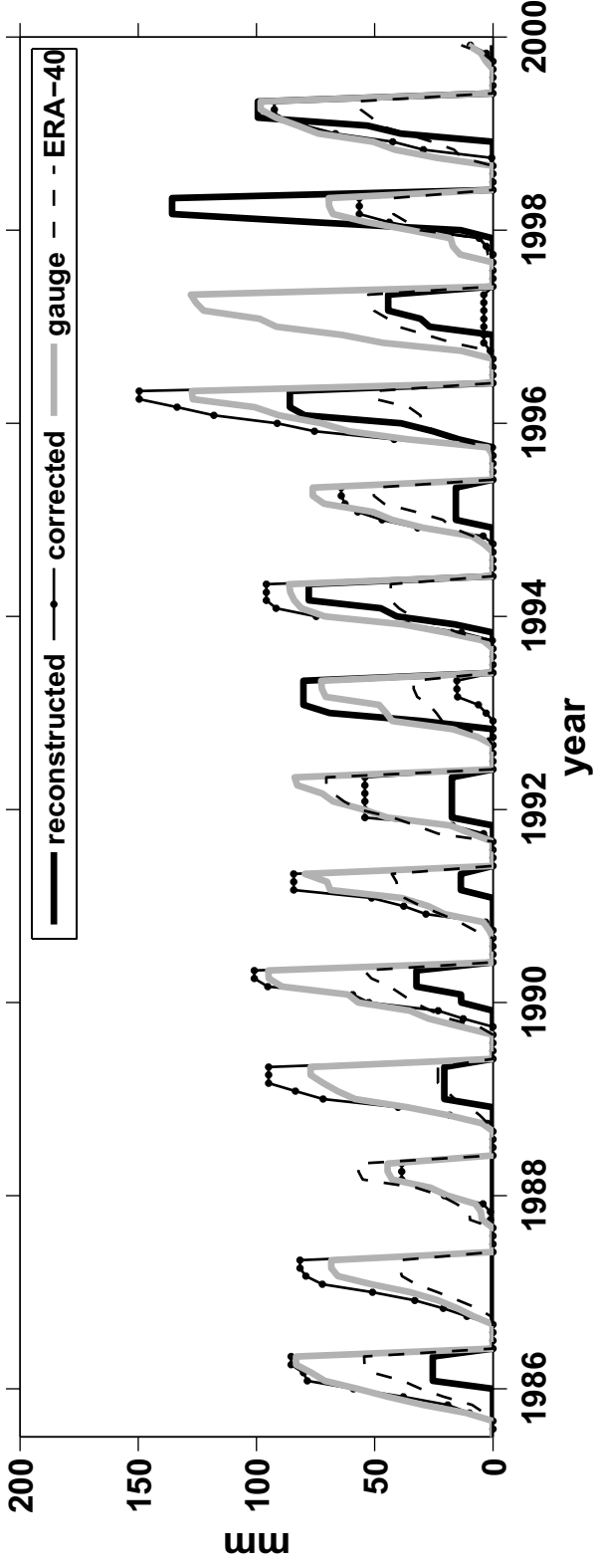
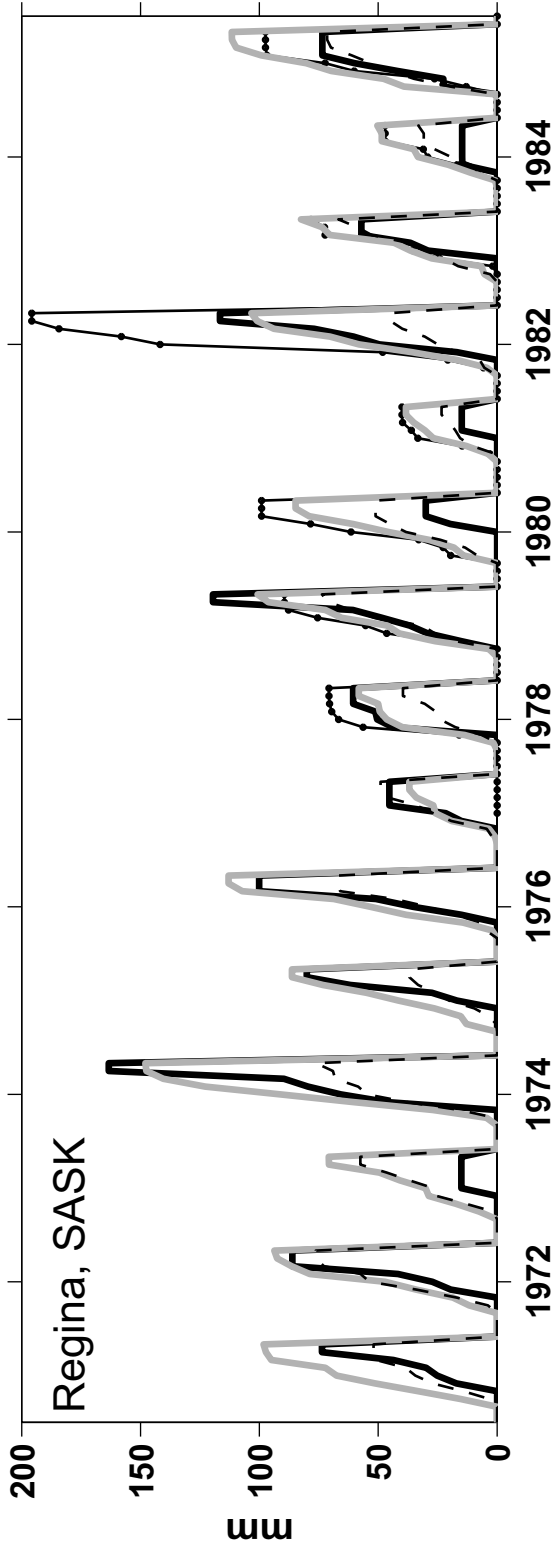


FIG.6b. Cumulative monthly reconstructed snowfall, corrected gauge snowfall (Yang et al. 2005), gauge snowfall, and ERA-40 for Regina in Saskatchewan, Canada. The Yang et al. record starts in 1977 at this station. The ERA-40 data start in 1957. During some years the monthly value of the corrected gauge and gauge snowfall are nearly equal.

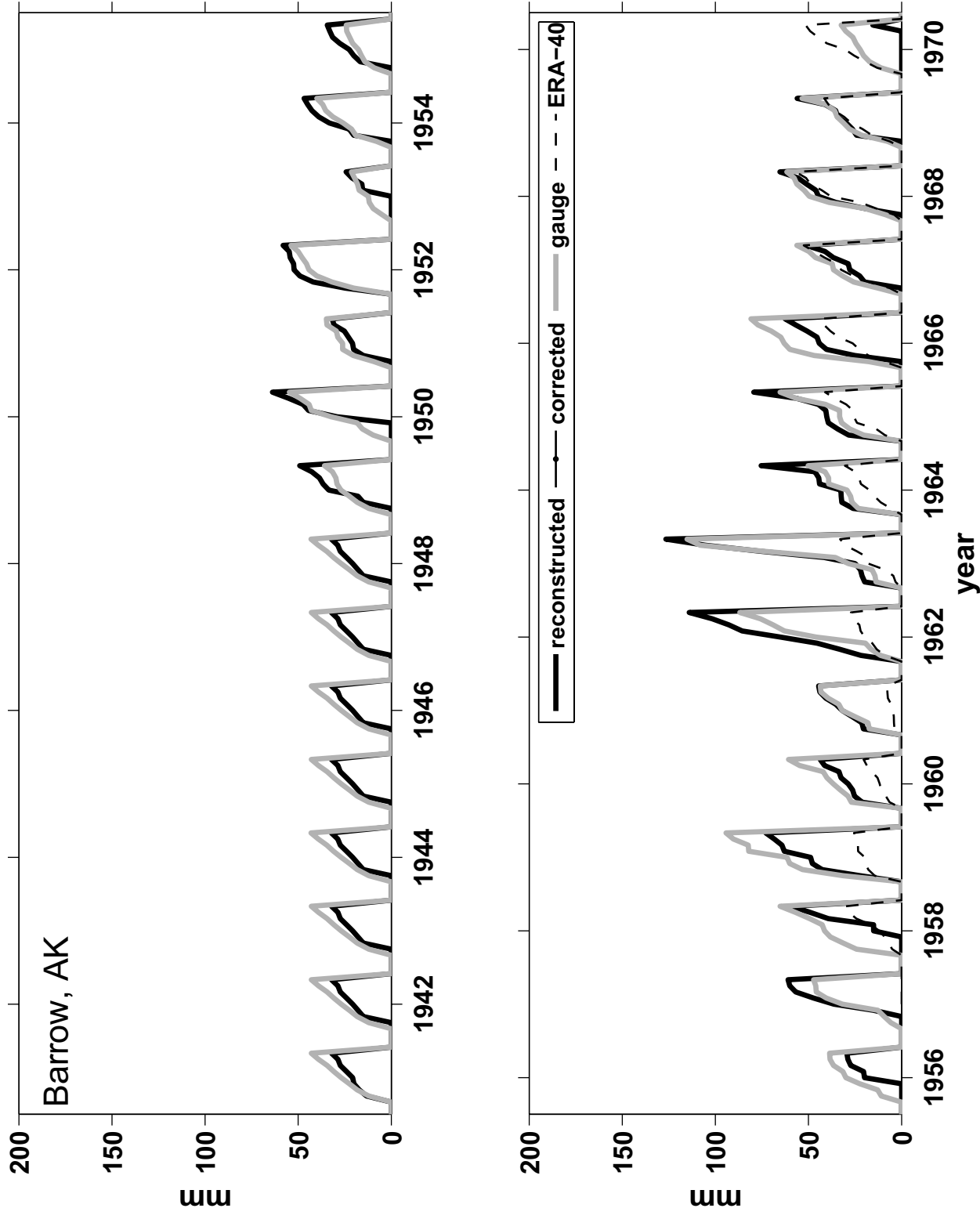


FIG. 7a. Cumulative monthly reconstructed snowfall, corrected gauge snowfall (Yang et al. 2005), gauge snowfall, and ERA-40 for Barrow, Alaska, USA. The Yang et al. record starts in 1973 at this station. The ERA-40 data start in 1957. During some years the monthly value of the corrected gauge and gauge snowfall are nearly equal.

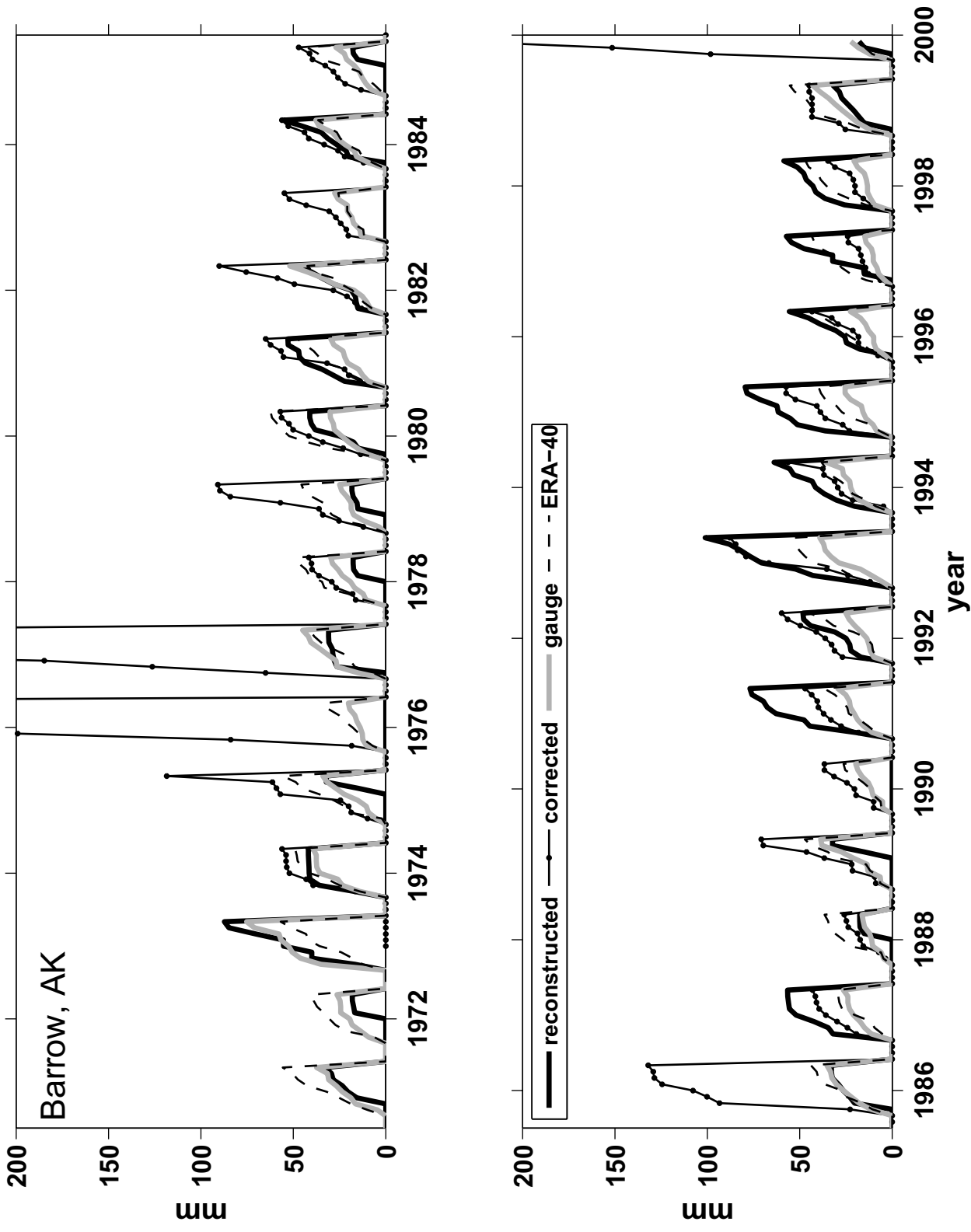


FIG. 7b. Cumulative monthly reconstructed snowfall, corrected gauge snowfall (Yang et al. 2005), gauge snowfall, and ERA-40 for Barrow, Alaska, USA. The Yang et al. record starts in 1973 at this station. The ERA-40 data start in 1957. During some years the monthly value of the corrected gauge and gauge snowfall are nearly equal.

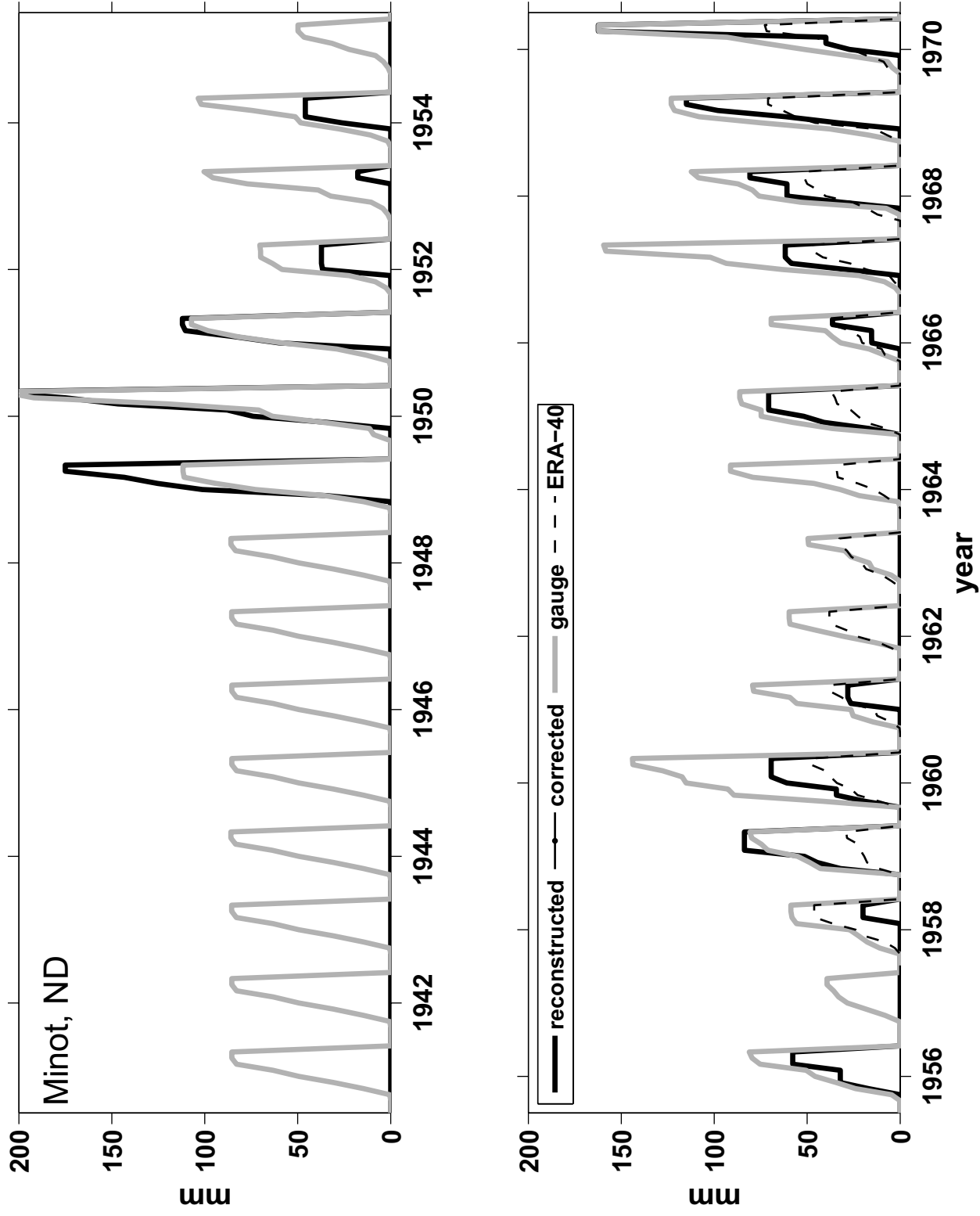


FIG. 8a. Cumulative monthly reconstructed snowfall, corrected gauge snowfall (Yang et al. 2005), gauge snowfall, and ERA-40 for Minot, North Dakota, USA. The Yang et al. record starts in 1973 at this station. The ERA-40 data start in 1957. During some years the monthly value of the corrected gauge and gauge snowfall are nearly equal.

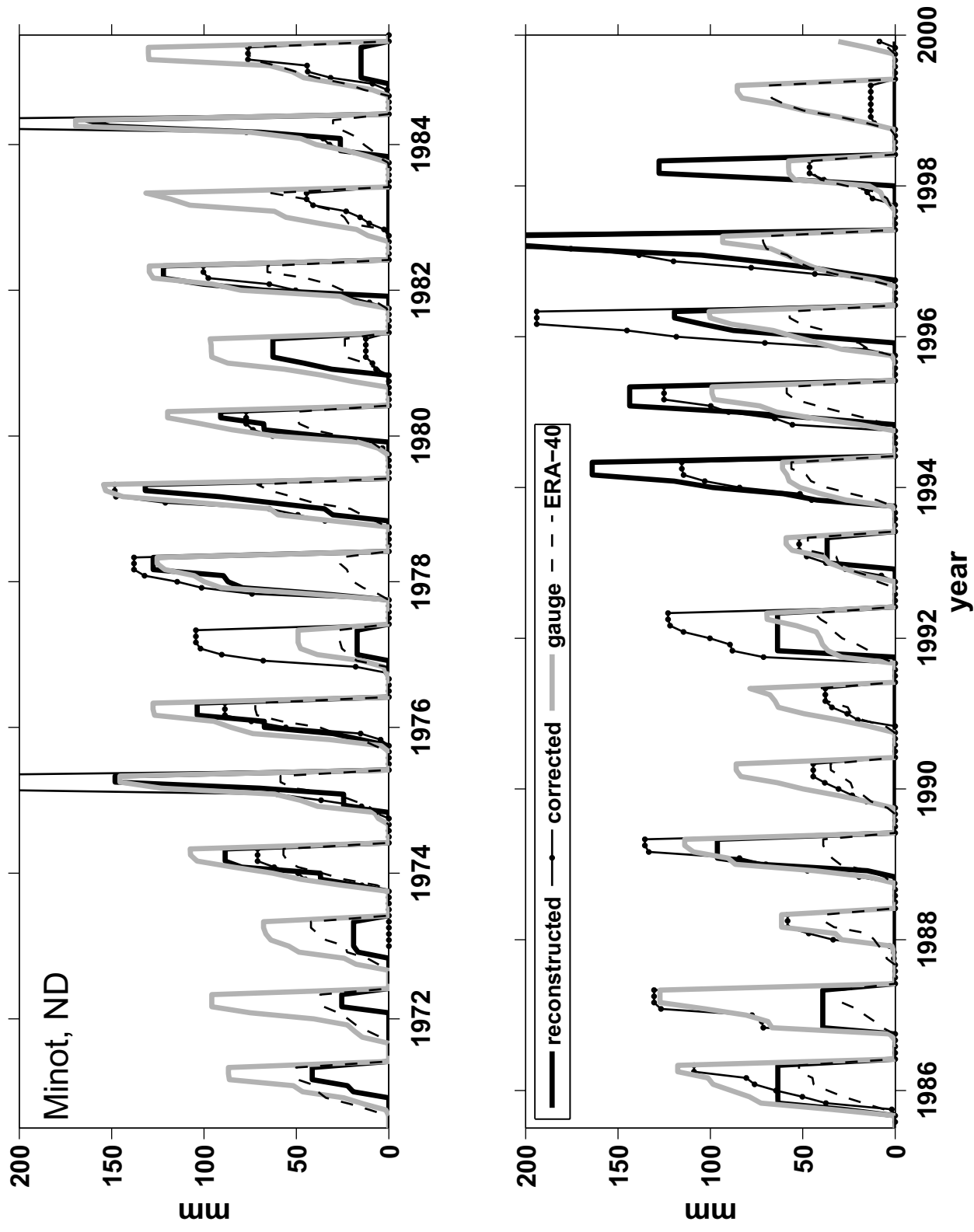


FIG. 8b. Cumulative monthly reconstructed snowfall, corrected gauge snowfall (Yang et al. 2005), gauge snowfall, and ERA-40 for Minot, North Dakota, USA. The Yang et al. record starts in 1973 at this station. The ERA-40 data start in 1957. During some years the monthly value of the corrected gauge and gauge snowfall are nearly equal.

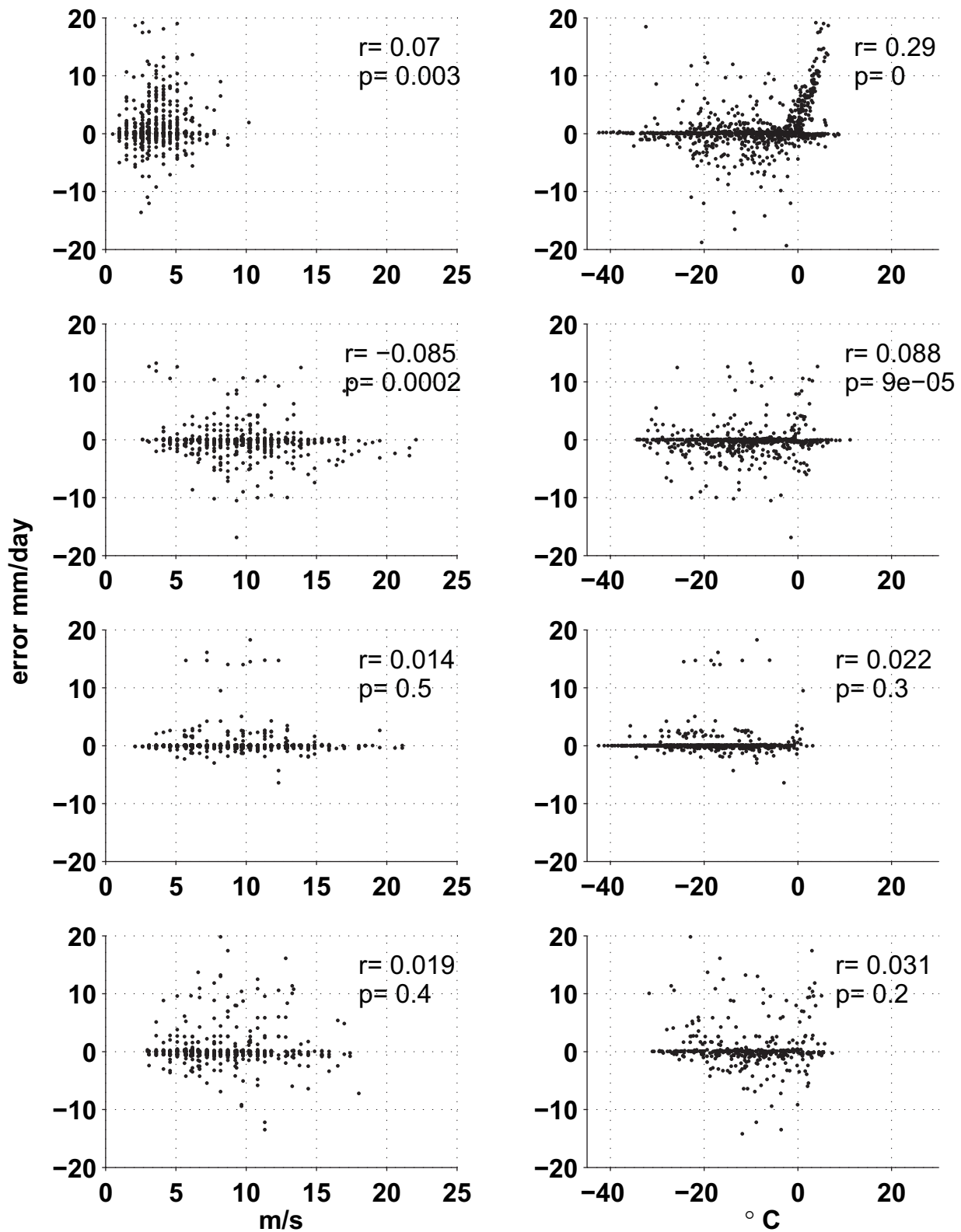


FIG. 9. Difference between daily reconstructed snowfall and (uncorrected) gauge snowfall versus wind speed and temperature for four test stations. Plots on the left are for wind speed and plots on the right are for temperature. Stations, from top to bottom are Dease Lake, Regina, Barrow, and Minot.

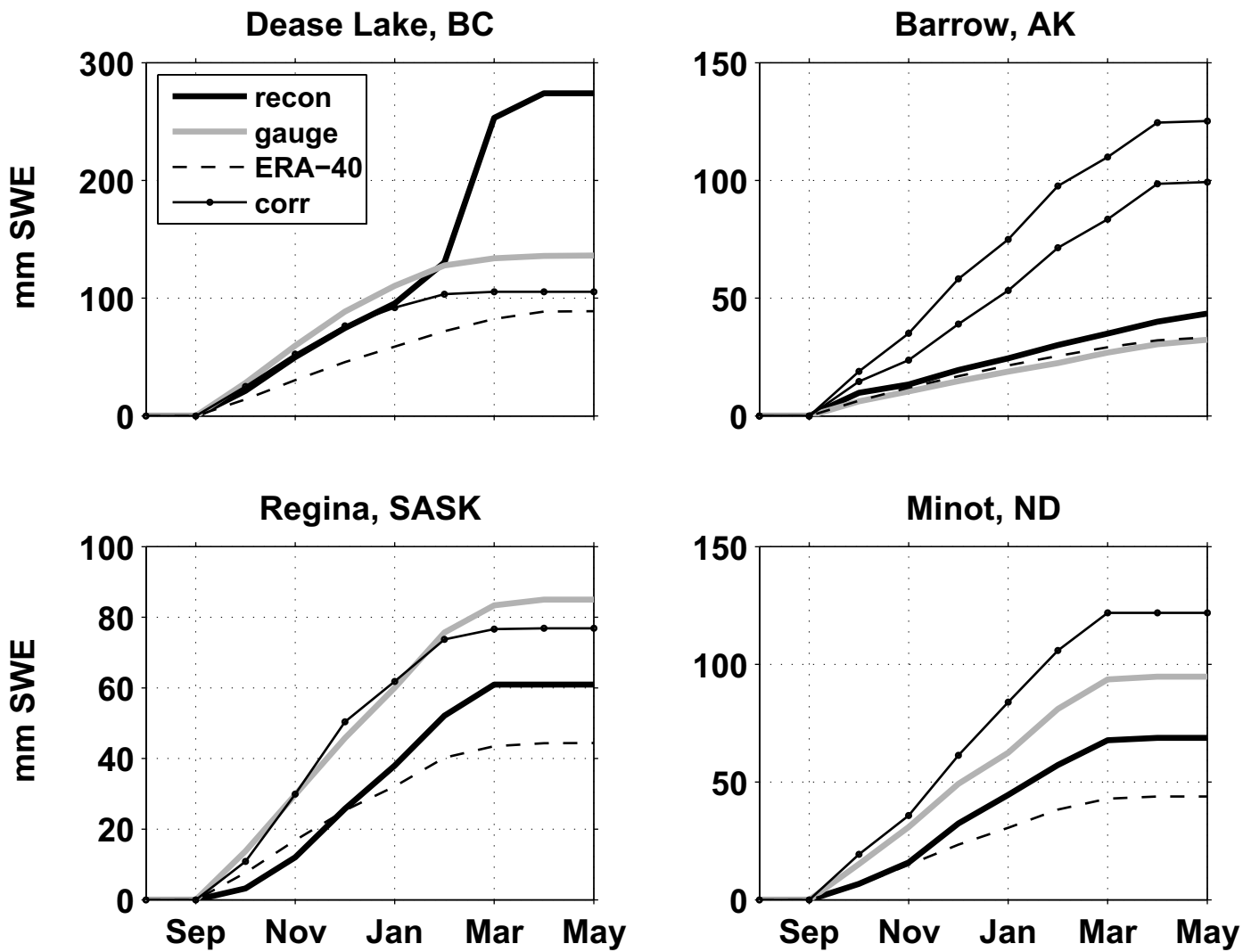


FIG. 10. Cumulative annual snowfall climatologies for four test stations using four different datasets: the reconstruction, raw gauge data, corrected gauged data from Yang et al. (2005), and ERA-40. For Barrow two different climatologies are plotted for the corrected gauge dataset: one calculated without the extreme errors of 1975-1977 and 1999 (the lower curve) and one calculated with the errors (the upper curve).

# WBM Runoff Correction Coefficients

30-minute spatial resolution

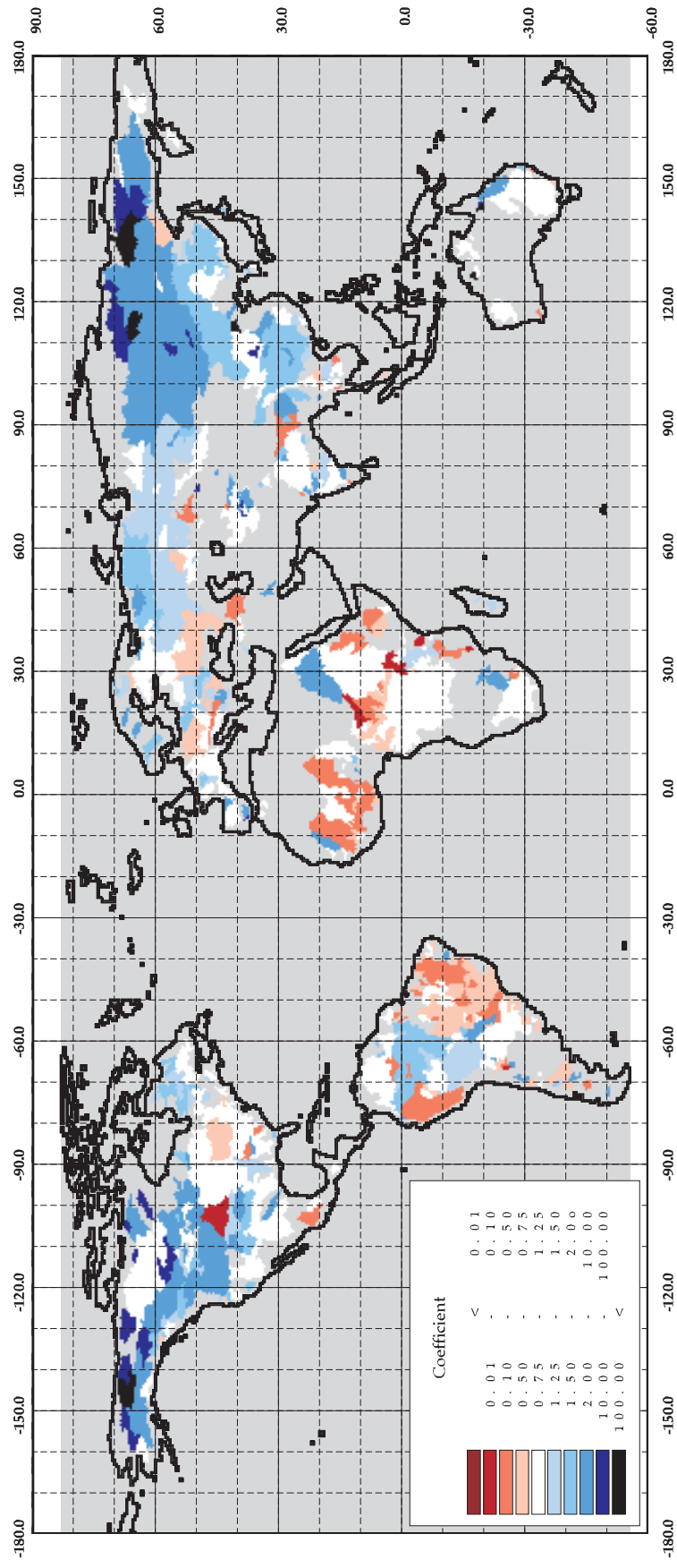


FIG. 11 . Runoff Correction Coefficients from the Water Balance Model, reproduced from Fekete et al. 1999. This shows the bias in gauge precipitation compared to measured runoff and a balanced water budget. Coefficients greater than one (blue to black) represent gauge undercatch while coefficients less than one (pink to red) represent gauge overcatch.

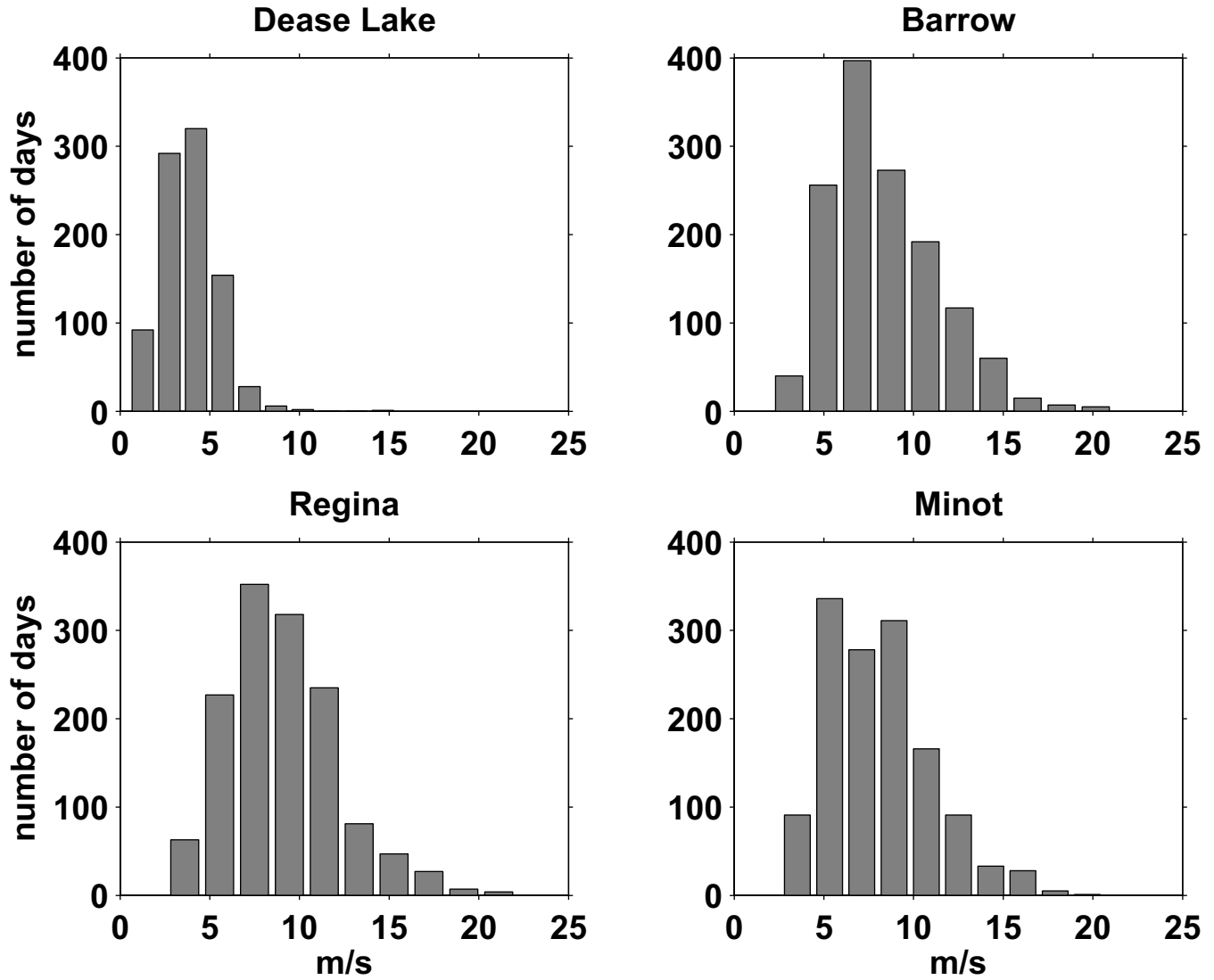


FIG. 12. Histograms of maximum sustained wind speed at four PASR test sites. These daily data are from the GLOBALSOD dataset, available for the period 1994-1999.