# Reconstructing solid precipitation from snow depth measurements and a land surface model

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[1] The amount and distribution of snowfall in the Arctic has significant effects on global climate. However, measurements of snowfall from gauges are strongly biased. A new method is described for reconstructing snowfall from observed snow depth records, meteorological observations, and running the NASA Seasonal-to-Interannual Prediction Project Catchment Land Surface Model (NSIPP CLSM) in an inverse mode. This method is developed and tested with observations from Reynolds Creek Experimental Watershed. Results show snowfall can be accurately reconstructed on the basis of how much snow must have fallen to produce the observed snow depth. The mean cumulative error (bias) of the reconstructed precipitation for 11 snow seasons is 29 mm snow water equivalent (SWE) for the corrected gauge measurement compared to −77 mm SWE for the precipitation from the corrected snow gauges. This means the root-mean-square error of reconstructed solid precipitation is 30% less than that of gauge corrections. The intended application of this method is the pan-Arctic landmass, where estimates of snowfall are highly uncertain but where more than 60 years of historical snow depth and air temperature records exist.

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## 1. Introduction

[2] The Arctic is an obvious place to look for evidence of climate change because changes there are expected to be some of the largest in magnitude of anywhere on the planet [*Intergovernmental Panel on Climate Change* (*IPCC*), 2001]. Surface air temperatures at high latitudes are expected to rise significantly, and in fact, an increase has already been observed [*Chapman and Walsh*, 1993; *Serreze et al.*, 2000]. Changes in the Arctic associated with warming include increased river discharge [*Peterson et al.*, 2002], a longer growing season [*Foster*, 1989; *Foster et al.*, 1992; *Brown and Braaten*, 1998; *Stone et al.*, 2002], and a change in the distribution of plant species [*Sturm et al.*, 2001]. Subsurface warming has been observed in borehole measurements [*Lachenbruch and Marshall*, 1986; *Pavlov*, 1994; *Osterkamp and Romanovsky*, 1999; *Oberman and Mazhitova*, 2001; *Romanovsky et al.*, 2002; *Romanovsky and Osterkamp*, 2000]. Trends in Arctic cloudiness and shortwave radiation observed from satellites show increased cloudiness in spring and summer, decreased cloudiness in winter, and decreased surface albedo during all seasons, with the strongest decreases in fall and winter. Changes in cloudiness are associated with increased cooling in the summer, fall, and winter, suggesting that clouds in these seasons may be damping warming trends [*Wang and Key*, 2003].

[3] Trends in the magnitude, distribution, and timing of precipitation are less clear. *Dai et al.* [1997], *Groisman and Easterling* [1994], and *Ye et al.* [1998] show increases in precipitation in some, but not all regions. Sparse observations in the Arctic lead to considerable uncertainty about the amount and distribution of precipitation. The problem of sparse observations is further exacerbated when station records are interpolated to fit a grid [*Hulme and New*, 1997]. Much of the uncertainty regarding precipitation and other Arctic freshwater trends is tied to the difficulty in measuring solid precipitation. Furthermore, according to estimates by *Aagaard and Carmack* [1989], the pan-Arctic landmass is the single greatest contributor to the Arctic freshwater budget. The uncertainty associated with Arctic precipitation makes it difficult to interpret observed changes in surface runoff as being caused by changes in subsurface storage of water, land use, or precipitation distribution [*Berezovskaya et al.*, 2004; *Fekete et al.*, 2004; *McClelland et al.*, 2004]. Multiagency programs such as the Study of Environmental Arctic Change (SEARCH) [*Morison et al.*, 2001], the Northern Eurasia Earth Science Partnership Initiative (NEESPI) (NASA and Russian Academy of Science, NEESPI science plan, available at http://neespi.gsfc.nasa.gov/science/science.html), and the Arctic Freshwater Initiative (NSF/ARCSS Freshwater Initiative, All-hands meeting notes, 2004) have emphasized the need to get better estimates of the mean precipitation, as well as estimates of spatial and temporal variability to help answer some of these questions.

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[4] Three major climate feedbacks in the Arctic relate to the amount and distribution of snowfall and its role in the Arctic freshwater budget. First, there is concern that warming of the ground in permafrost regions will result in additional release of carbon to the atmosphere [*Oechel et al.*, 1993; *McKane et al.*, 1997a, 1997b]. Snow is an efficient insulator, which prevents the ground from exposure to surface air temperatures during the period of snow cover. *Stieglitz et al.* [2003] show that changes in below-ground temperatures can be impacted as much by temporal variations of snow cover as by changes in the surface air temperature. In a changing climate, distribution of snowfall will help determine soil temperature and therefore carbon fluxes to the atmosphere. Second, snowfall distribution on land may play an important role in the ice albedo feedback [*Groisman et al.*, 1994]. Third, freshwater input into the Arctic ocean and Nordic and Labrador seas from the landmass may ultimately affect the formation of deep water and therefore meridional overturning circulation in the North Atlantic [*Peterson et al.*, 2002]. For these reasons, recent emphasis has been placed on predicting the magnitude and spatial distribution of future warming and precipitation at high latitudes. Section 2 presents the background and motivation for conducting this study. The test site, the errors, the model, and the experiments are described in section 3. This is followed by the results and a discussion of the major findings at the test site and the implications for applying the method to the Arctic.

## 2. Motivation

[5] One of the biggest challenges in monitoring Arctic climate change is measuring frozen precipitation. In high-wind conditions, which are common across the Arctic [*Déry and Yau*, 2002], precipitation gauges disrupt the boundary layer atmospheric flow, causing frozen precipitation to preferentially blow over and around, rather than into, the gauge [*Goodison et al.*, 1998]. Liquid precipitation is less susceptible to this undercatch problem because it is denser and has a faster falling velocity. Though during high-wind events, gauges can also undercatch rain at the rate of 2–10% [*Sevruk*, 1982]. Estimates of snowfall undercatch for some types of gauges are as high as 70% or more [*Yang et al.*, 1998, 2000]. Because most places in the Arctic have below-freezing surface air temperatures for 9–10 months of the year, a large percentage of annual precipitation is frozen. However, because the undercatch problem is so severe, it is difficult to estimate the total amount, let alone the percentage of precipitation that falls in each phase or in which season.

[6] In addition to undercatch caused by disrupting the wind flow near the gauge, other systematic errors in measuring solid precipitation include evaporation/sublimation, wetting losses from water sticking inside the gauge, blowing snow, the tendency of observers to ignore trace events, and gauge location which is unrepresentative of the catchment [*Sevruk*, 1982; *Legates*, 1987; *Goodison et al.*, 1998]. All of these systematic biases lead to underestimation of precipitation, with the exception of biases associated with measurements in areas of blowing snow deposition. *Adam and Lettenmaier* [2003] provide a useful summary of the relative magnitude of biases.

[7] Early gauge corrections tended to be empirical comparisons between a gauge and another gauge, precipitation in a pit, or changes in snow water equivalent (SWE) on a snow course. However, *Goodison et al.* [1998, and references therein] have shown that these type of corrections are not very portable; they are site-specific because systematic biases in measurement are not separated from unsystematic biases at the site. Recent efforts have focused on identifying systematic biases at many sites and developing correction algorithms for gauge precipitation data sets. *Legates and Willmott* [1990] produced the first comprehensive global precipitation product that included gauge corrections. Then, from 1986 to 1993, the extensive WMO Solid Precipitation Measurement Intercomparison program focused on improving the estimates of systematic errors associated with many different types of solid precipitation gauges, in comparison to a double-fenced intercomparison reference (DFIR) gauge [*Goodison et al.*, 1998]. *Adam and Lettenmaier* [2003] went on to modify the correction method used by *Legates and Willmott* [1990] on the basis of the WMO Solid Precipitation Measurement Intercomparison results for wind-induced undercatch and applied it to the C. J. Willmott and K. Matsuura global monthly gridded precipitation data set (Terrestrial air temperature and precipitation: Monthly and annual time series (1950–1999) (version 1.02), available at http://climate.geog.udel.edu/∼climate/html_pages/download.html#ghcn_T_P2). Because of the high level of metadata available there, separate gauge corrections were made for Canada by *Groisman* [1998], *Mekis and Hogg* [1999], and *Adam and Lettenmaier* [2003]. For a detailed description and analysis of these different gauge corrections, see *Adam and Lettenmaier* [2003].

[8] These previous studies have shown that systematic biases in measurement, particularly for solid precipitation, are substantial. However, there are a number of disadvantages to this approach. Primarily, it ignores the unsystematic (i.e., site-specific) biases that may be as large as or larger than systematic ones. These include problems with ice or snow forming around the gauge, other types of instrumental malfunction, as well as gauge siting in an area of snow deposition that is atypical for that watershed. It is also necessary to obtain documentation about the type of gauge and baffles (if any) used at each station, or to make an educated guess if that information is unavailable. Similarly, the height of the gauge and anemometer and other aspects of the measurement site, including vegetation, must be known or assumed. The corrections developed by the WMO for solid precipitation were developed from a small number of sites. Many systematic errors are sensitive to the location of the gauge [*Goodison et al.*, 1981, 1998]. Some gauges were not tested in that study, so no corrections are available. Finally, in gridded precipitation products interpolation introduces further error, particularly when there are only a small number of stations in each grid cell. Currently, it is impossible to quantify the magnitude of these errors at any particular site, unless considerable testing has occurred there or at a similar site. Some members of the community believe, however, that corrected gauge measurements are still underestimating snowfall (B. E. Goodison, personal communication, 2004).

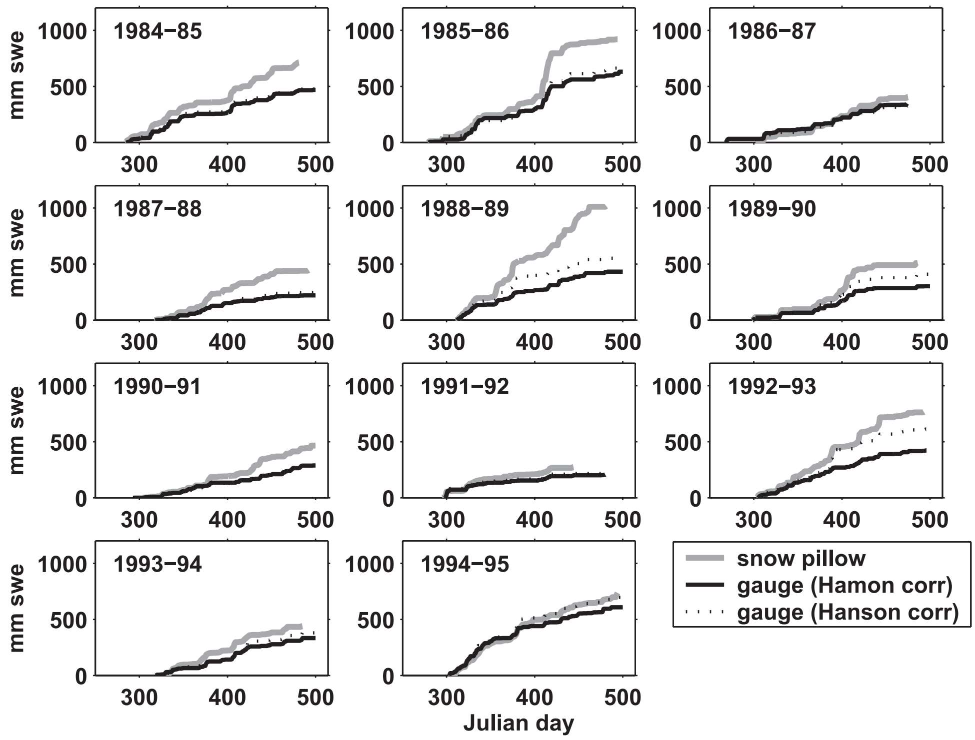


**Figure 1.** Cumulative snowfall records at RCEW from positive daily changes of snow water equivalent (SWE) on a snow pillow and from a single dual-gauge record that has two different corrections applied to it.

[9] There are several advantages to using snow depth measurements to solve an inverse problem for snowfall. Snow depth measurements are simple, inexpensive, and not susceptible to significant instrumental errors. The sampling technology (a ruler or graduated cylinder) has changed very little over time. Finally, it is important to recognize that the simplicity of depth measurements and the small marginal cost of additional measurements has meant that observers typically take multiple measurements and average them, recording what is actually an areal mean. Unfortunately, neither the spatial variance, nor the area over which the measurements were averaged is known for historical stations and may be assumed to range from 1 to 100 $m^2$ or more.

[10] This study is designed to avoid both the systematic and some unsystematic biases associated with snow gauges. Snow depth measurements will be used, along with the NASA Seasonal to Interannual Prediction Project (NSIPP) Catchment-based Land Surface Model (CLSM, details discussed in section 3) to solve an inverse problem for snowfall. This method calculates the snowfall that must have occurred to produce the observed snow depth, given the physics of the model (compaction, sublimation/condensation, snowmelt, etc). Because true precipitation in the Arctic is unknown, a test site must be used to evaluate the success of this method as a proof of concept. One of the WMO Solid Precipitation Intercomparison test sites was the Reynolds Creek Experimental Watershed (RCEW) in southwestern Idaho. Partly for this reason, and also because of the long time series and high quality of data and of documentation at this site [*Marks et al.*, 2001], RCEW is used to test the reconstruction method.

[11] Figure 1 shows cumulative frozen precipitation data from the Reynolds Mountain climate station at RCEW. These data were taken hourly starting on 1 January, 1984. Data from a snow pillow and two different corrections applied to a dual gauge system are plotted (instruments described in detail in section 3). The *Hanson et al.* [1999] correction is based on calibration with a Wyoming snow gauge, which has catch that is nearly identical to that of the WMO's DFIR at this site. The *Hamon* [1973] correction is based on an empirical relation between a covered and uncovered gauge at the RCEW site. While the *Hanson et al.* [1999] correction estimates much more solid precipitation than the earlier study, it still estimates significantly less than what was measured with the snow pillow. Underestimates of snowfall at RCEW from a corrected gauge record are consistent with the idea that correction algorithms are not accounting for all undercatch biases and therefore provide the motivation for this study.

[12] The objective here is to develop and test a method for reconstructing snowfall based on known principles of snow physics and which avoids the problems associated with gauges. We will reconstruct snowfall records using the snow depth record from RCEW, in addition to meteorological data, and a state-of-the-art land surface model. Our hypothesis is that because most measurement biases underestimate snowfall, this new method (which corrects for more biases than does gauge correction) will show significantly more snowfall than previous products and can be validated using nongauge based observations at a test site. Our intent following this study is to extend this method to the pan-Arctic landmass, where historical snow depth and surface air temperature records exist, in

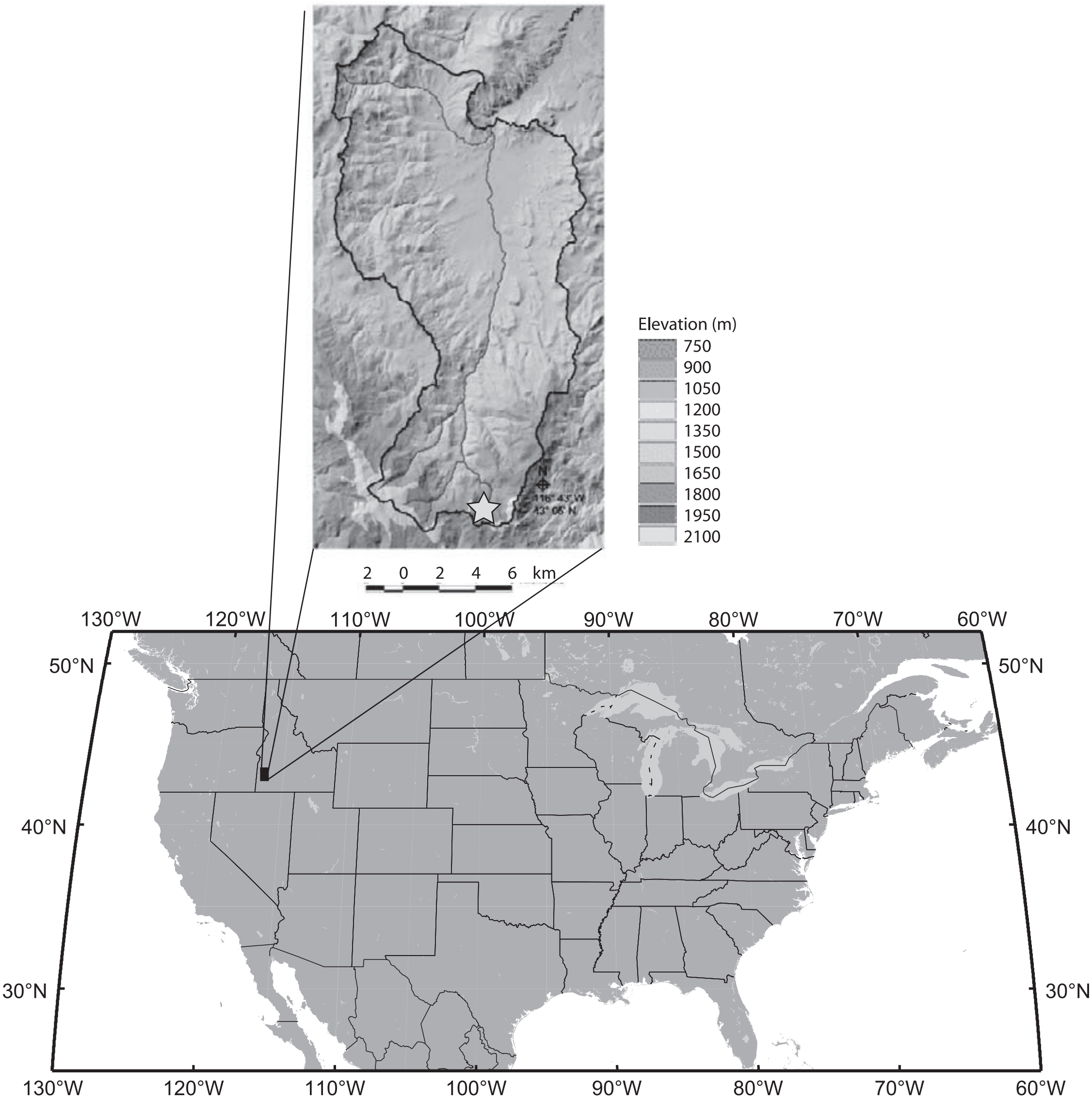


**Figure 2.** Reynolds Creek Experimental Watershed in southwestern Idaho. The Reynolds Mountain station is marked with a star. The catchment map is modified from *Seyfried et al.* [2000]. See color version of this figure at back of this issue.

order to reduce the uncertainties associated with the Arctic frozen precipitation budget and its role in climate.

## 3. Method

[13] First, a basic description of the test site will be given. Next, it is necessary to establish a benchmark of "true snowfall" to which the reconstructions can be compared. For reasons mentioned above, corrected gauge measurements can be used for comparison, but there should be an independent way to evaluate the reconstructions. In this study, data from a snow pillow will be used. After a benchmark truth for model validation is established, other meteorological observations used in the reconstruction will be described. The details of the model will then be described. Next, to show that the model physics can adequately produce observed snow depth and SWE, the model will be run in a control, forward, mode. In this run, observed liquid precipitation from the gauge (much less susceptible to undercatch) and solid precipitation from the snow pillow are used, along with temperature and other meteorological observations, to model the growth and ablation of a snowpack. Finally, three experiments will be conducted to show how snowfall can be reconstructed from historical snow depth records in the Arctic.

### 3.1. Proof of Concept at Reynolds Creek: Site Description

[14] Reynolds Creek Experimental Watershed (RCEW) is located in southwestern Idaho (43°12′N, 116°45′W, see Figure 2). The station used for this study is the Reynolds Mountain station, which is near the highest point in the watershed, approximately 2200 m above mean sea level. The high elevation and proximity of vegetation near this site mean that deposition of large-scale blowing snow is negligible, i.e., average snowfall in the catchment is not overestimated. The dual gauge precipitation measurement system has been operating at Reynolds Mountain station,

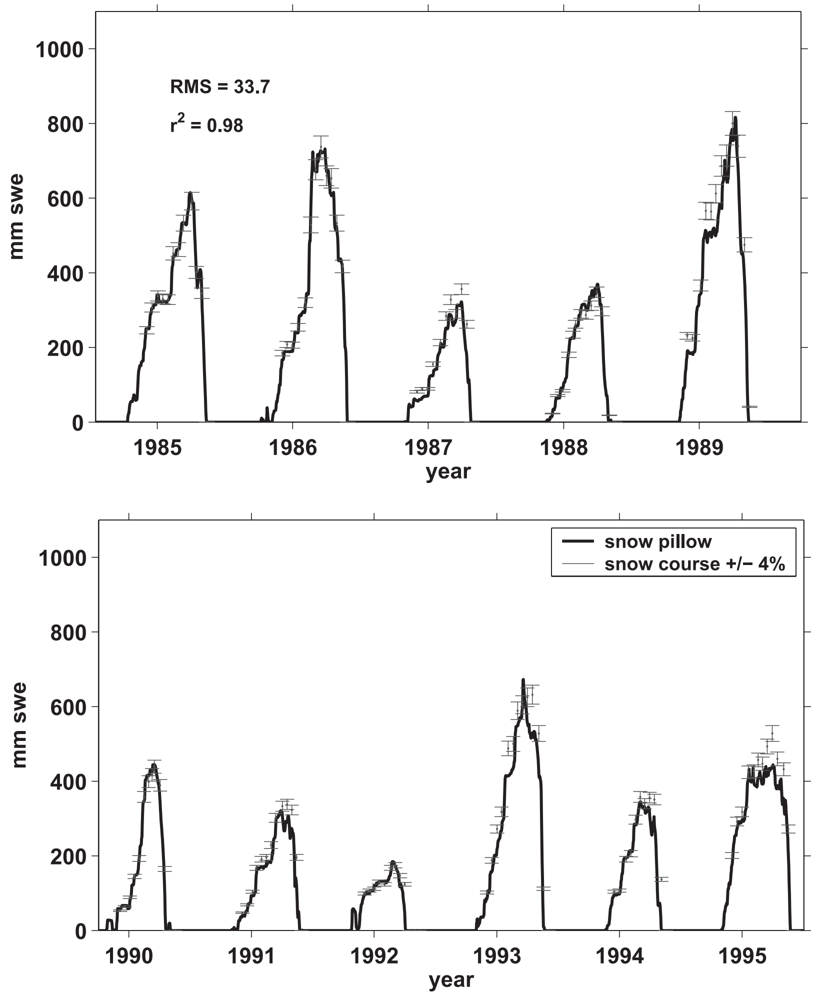


**Figure 3.** SWE observations from the snow pillow (shown daily) and snow course (semimonthly) at RCEW. Density measurements from the snow course are subject to an average error of ±4% caused by wetting of the snow-coring tube [*Johnson and Schaefer*, 2002]. Year ticks denote 1 January and statistics refer to all snow seasons from 1984–1985 to 1994–1995.

on an hourly basis, since January of 1968, while the snow pillow has operated hourly since October of 1982 and the meteorological station since March of 1983. The snow course has been consistently measured on a biweekly basis since 1969. While observations continue to be made, 1995 is the last complete year for which data has been quality checked and publicly distributed by the staff at RCEW. Precipitation at this site is fed by Pacific Ocean moisture brought with the Westerlies from the western or southwestern part of the catchment. The average annual precipitation is estimated at 1100 mm [*Hanson et al.*, 2000], 60–90% of which falls as snow.

### 3.2. Defining "True Precipitation": The Challenges of Snow Heterogeneity and Measurement Errors

[15] Two distinct problems are associated with estimating snowfall and an effort is made in this study to discuss them separately. The first (which we will call type A) is that snowfall is heterogeneous in space. Multiple measurements must be taken to estimate average snowfall over some area. In lieu of multiple measurement sites which may be costly, observers try to choose a site location that is representative of the area over which snowfall is being estimated. When the site location for the measurement is unrepresentative of the catchment, unsystematic biases can occur which are not accounted for by gauge corrections. The second type (type B) of problem for estimating snowfall is simply the instrumental error, which is the failure of the instrument to capture true snowfall at the site. A portion of this type of error may be systematic and therefore can, in part, be corrected for by the gauge intercomparison method. Three different solid precipitation measurement systems (a dual gauge system, a snow course, and a snow pillow) are described here and their associated errors (types A and B) are discussed. The purpose for discussing these errors is to make clear that the reconstruction method can correct for some, but not all of these errors. There is strong agreement between two different measurements of SWE at the Reynolds Mountain station, the snow pillow and snow course, as shown in Figure 3. The precipitation that will be defined as "truth" still contains both type A and type B errors, but these are shown to be small relative to the errors from the corrected gauge record.

[16] The gauges at Reynolds Mountain station are Universal Recording Gauges. The covered gauge employs an Alter-type shield with baffles individually constrained

30 degrees from the vertical. The covered gauge is 6 m from the uncovered gauge. Both have their orifices at 3.05 m above the ground and are 1.17 m in diameter [*Hanson et al.*, 2001]. The shielded gauge has been shown to catch up to 50% more precipitation than identical unshielded for the same conditions [*Larkin*, 1947; *Larson and Peck*, 1974; *Goodison et al.*, 1981; *Sturges*, 1984; *Hanson*, 1989; *Yang et al.*, 1998]. While correction functions have been developed for this dual gauge system, Figure 1 suggests that the corrected gauges still underestimate true snowfall at Reynolds Mountain (type B error). Catchment-scale variability of snowfall at RCEW (type A error), as measured by dual gauges is discussed by *Hanson et al.* [2001] and is complicated by elevation effects. Smaller-scale (<1 km) snowfall variability is not quantified by the dual gauge system.

[17] The snow courses consisted of two permanent poles, between which five samples of depth and SWE from a snow tube were taken according to the method described by *Goodison et al.* [1981]. These samples were taken every two weeks during the snow season (1 December to 1 June). *Johnson and Schaefer* [2002] report a snow core SWE measurement uncertainty of ±4% of true SWE based on an extensive comparison between snow pits and snow cores. Heterogeneity in snow depth is estimated over the snow course [*Marks et al.*, 2001]. The standard deviation and coefficient of variation (100% standard deviation/mean) was computed for the five samples from each biweekly snow course measurement date. *Dickinson and Whitely* [1972] showed that errors associated with density measurements (type B) can be separated from those associated with snow depth measurements. These authors showed that the standard error of the water equivalent increased much slower with a decreasing number of samples than did depth (as reported by *Goodison* [1981]). Results show that as the mean SWE increases, so does the standard deviation, but correlation between the two is weak. Additionally, as mean SWE increases, the coefficient of variation and its spread also increase. As the depth of snow on the snow course increases, within-course variability (heterogeneity) increases. On the catchment scale, density varies 10–15% from station to station within RCEW.

[18] In addition to heterogeneity and measurement errors in density of the snowpack, there are measurement (type B) errors associated with trace amounts of snow on the course. By convention, trace measurements of snow course depth are set to zero, as snow tube samplers perform poorly with small amounts of snow. Furthermore, data collection on the snow course does not begin until 1 December. There may be snow events prior to this date and a series of trace events could sum to a significant amount of precipitation.

[19] A snow pillow is a large rubber bladder full of air, which measures the overlying burden of snow by forcing antifreeze up a tube into a protected instrument house. Snow pillows are known to have several sources of error, none of which are well quantified and all are sensitive to the location of the snow pillow. These errors do not appear to be significant at RCEW, because SWE from the snow pillow and snow course are generally in good agreement. However, these errors may explain why there is an occasional discrepancy between the two SWE measurements. Typical errors include temperature and barometric pressure effects on the liquid in the standpipe. The temperature effect depends on the coefficient of expansion of the fluid. There is also the problem of the antifreeze solution staying mixed. Some snow pillows can show a delay in response time to overburden of a day or so [*Penton and Robertson*, 1967; *Beaumont*, 1965], but the only possible effect of this error on the reconstruction would be a time lag in the reconstructed event. Large pillows, such as the one at RCEW, respond faster. Snow pillows also exhibit diurnal variations of pressure, caused by daily fluctuations in temperature, atmospheric pressure, and radiation that cause water to leave the pack [*Penton and Robertson*, 1967]. Errors associated with the influence of the snow pillow on soil heat fluxes are discussed by *Johnson and Schaefer* [2002] and are most common when there is a steep temperature gradient between the soil and the atmosphere: that is, early in the winter or late in the spring melt season. Another problem associated with snow pillows is bridging. This occurs when snow on the pillow is supported by snow outside of the pillow and is associated with snow that has undergone thawing and refreezing [*California Department of Water Resources*, 1976]. Large diameter (3 m) snow pillows, such as the one at RCEW, are not highly susceptible to errors in bridging conditions.

[20] The Reynolds Mountain station snow pillow is taken as truth for this study because it closely matches the SWE of the snow course during most years (Figure 3). Exceptions are that during three snow seasons (1986–1987, 1990–1991, 1994–1995) the snow pillow slightly underestimates the peak SWE of the snowpack, compared to the snow course, and during the early part of 1989, the SWE of the snow pillow appears to lag that of the snow course. A snowfall time series is determined by the positive differences of consecutive daily SWE means divided by the number of 20-min model time steps per day (i.e., daily precipitation is spread evenly over 24 hours). Daily mean differences are used to avoid spurious observations due to the diurnal pressure fluctuations of the pillow. Defining snowfall in this way means that snow depth (SWE) and snowfall are internally consistent. The depth of snow on the snow pillow is estimated using density from the snow course. Throughout this study true snowfall and snow depth are defined in this way. Additionally, snow densities for events previous to the first snow course SWE measurement in December are assumed to have density equal to that of the first observation in that season, so these events can also be reconstructed. The snow pillow at RCEW has a diameter of 3 m and therefore is not highly susceptible to many of the errors discussed above. However, during the years where the snow pillow underestimates or shows a lag in SWE (Figure 3), there were midwinter warming events which may be the cause of measurement error and/or small-scale heterogeneity in SWE. Errors in the observations and the model runs (described below) are summarized in Table 1.

### 3.3. NASA Catchment-Based Land Surface Model

[21] The model used here is the NASA Seasonal-to-Interannual Prediction Project Catchment-based Land Surface Model (NSIPP CLSM) developed by *Koster et al.* [2000] and *Ducharne et al.* [2000]. The model is conceptually based on a TOPMODEL framework, developed by *Beven and Kirkby* [1979], wherein points of hydrologic similarity (these are points within a catchment that saturate,

**Table 1.** Summary of Errors Discussed in Section 3

| Error | Subcategory | Description |
|---|---|---|
| $E_{phys}$ | Type A | all errors associated with the model physics and forcing/restoring observations |
| $E_{phys}$ | Type B | heterogeneity in true quantity that is not captured in measurement |
| $E_{phys}$ | | instrumental errors |
| $E_{num}$ | | errors introduced by restoring to observed snow depth |
| $E_{tot}$ | | sum of all errors |

discharge, and in other ways respond similarly to meteorological forcing) are identified by a topographically based index. This topographical index provides the fundamental unit of hydrological response and is used to represent heterogeneity in soil moisture and other prognostic variables. The snow model used by CLSM is that of *Lynch-Stieglitz* [1994] and the details of the snow physics are described in that paper. 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.

[22] The major modification to the model of *Lynch-Stieglitz* [1994] is made by modeling albedo explicitly. 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 kg $m^{-3}$), 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]. This model has been used successfully in small and large-scale applications [*Lynch-Stieglitz*, 1994; *Rind et al.*, 1997; *Stieglitz et al.*, 1999, 2000, 2001, 2003; *Déry et al.*, 2004, 2005a, 2005b].

[23] A number of forcing variables are required to drive the model. At RCEW, surface air temperature, relative humidity, incoming shortwave radiation, wind speed, precipitation, atmospheric pressure, snow depth, and SWE are all measured. In the Arctic, typically only surface air temperature, precipitation, and snow depth are measured. In the next section, a sensitivity test will be described which quantifies the effect of forcing the model with a reduced number of observed atmospheric variables for the model reconstructions. For the experiments at RCEW, liquid precipitation is estimated from the gauge using the *Hamon* [1973] correction. For liquid precipitation this correction is minimal. The phase delineation between liquid and solid is taken to be 0°C. Testing shows that the model runs for RCEW are only moderately sensitive to the delineation for freezing point. The sensitivity at any particular site will depend on how many days the temperature hovers near freezing during precipitation events at that station. Below 0°C, precipitation is taken from the snow pillow as described in the previous section. All reconstructions at this site are conducted between the first snowfall following 15 September and 1 April. True snowfall after 1 April is taken from the dual gauge system (with the *Hanson et al.* [1999] correction) because a significant amount of mixed phase precipitation occurs during this month.

[24] Wind is not measured at gauge height, so it must be corrected using a wind profile method [*Golubev et al.*, 1992]. For this method, it is necessary to estimate a surface roughness coefficient, which is approximated as $Z_0$ = 0.01 m for snow in winter and $Z_0$ = 0.03 m for short grass in summer. This is a typical method which is also used when applying gauge corrections if observations were taken at a nonstandard height. The mean annual wind speed at RCEW is 4.14 m $s^{-1}$ with a daily standard deviation of 2.53. Surface air temperature, relative humidity, wind speed, and incoming shortwave radiation are also observed at this site and used to drive the model. Mean annual temperature at Reynolds Mountain station is 5.1°C with a daily standard deviation of 8.9. On days that snowfall occurs (an average of 115 days per snow season), mean snowfall from the snow pillow is 1.75 mm $d^{-1}$, with a daily standard deviation of 5.19. Downwelling longwave radiation is estimated as: $\sigma(T_{surf} - 20)^4$ wherein the term subtracted from surface air temperature accounts for absorptivity of the atmosphere [*Jones*, 1992] and sigma is the Stefan-Boltzmann constant ($\sigma = 5.67 \times 10^{-8}$ W $(m^2 K^4)^{-1}$).

### 3.4. Control Run and Experiments

[25] For this method development at RCEW, model runs are conducted as a column. All runs are spun up with all available forcing data (surface air temperature, liquid and solid precipitation, relative humidity, wind speed, incoming shortwave, downwelling longwave, and atmospheric pressure) from perpetual 1984 to ensure the correct soil heat fluxes to and from the snowpack. The purpose of the control run is to demonstrate the model's ability to reproduce the observed snow depth and SWE, given the observed meteorological forcings. In this run, the model is run in its normal forward mode (no restoring) using the hourly observations from 1984–1995 at Reynolds Mountain station. Precipitation falls at the site; when it falls consistently as snow, it accumulates and compacts. Sublimation, condensation, and evaporation may occur. The snowpack ripens and finally it ablates through melting.

[26] In the first experiment (the first reconstruction run), the model is forced with the same rain, snowfall, and meteorological observations as in the control. In this case the snowfall is considered to be a first guess for true snowfall. Any initial guess can be used; however, using the true precipitation as the initial guess will allow us to separate the errors associated with model physics ($\mathbf{E_{phys}}$, see below) from those associated with the numerical scheme of the reconstruction ($\mathbf{E_{num}}$). In this first experiment, model snow depth is compared to observed snow depth at the end of each time step. The difference between these two depths is defined as excess (**S**, ±). The excess (**S**) precipitation added to the first guess is considered to be the actual precipitation that must have fallen to produce the observed snow depth, given the model physics. This reconstruction of solid precipitation can be represented mathematically in the following way:

$$\mathbf{P_{rec}} = \mathbf{P_{ini}} + \mathbf{S} - (\mathbf{E_{phys}} + \mathbf{E_{num}}) \quad (1)$$

$$\mathbf{P_{rec}} = \mathbf{P_{ini}} + \mathbf{S} - (\mathbf{E_{tot}}) \quad (2)$$

**Table 2.** Summary of Experiments Described in Section 3

| Model Run | Purpose | Results | Associated Errors |
|---|---|---|---|
| Control | assess if model can reproduce observed snow depth and SWE | Figure 4 | $E_{phys}$[a] |
| Experiment 1 | basic snowfall reconstruction | Figures 5 and 10 | $E_{phys}$[a] and $E_{num}$ |
| Experiment 2 | reconstruction for error analysis | Figures 6 and 7 | $E_{num}$ |
| Experiment 3 | "Arctic" sensitivity test with reduced forcing | Figures 8 and 9 | $E_{phys}$[a] and $E_{num}$ |

[a]$E_{phys}$ can be further divided into heterogeneity (type A) and instrumental errors (type B) associated with snow depth used in restoration plus imperfect model physics.

where reconstructed precipitation, based on restoration to observations ($\mathbf{P_{rec}}$), is equal to the first guess ($\mathbf{P_{ini}}$), plus the excess snowfall ($\mathbf{S}$, $\pm$), minus model errors. $\mathbf{P_{rec}}$ is always taken to be $\geq 0$. All units are in mm SWE. Error is a raw difference, not root mean squared (RMS) and total errors ($\mathbf{E_{tot}}$) are the sum of errors in the model physics ($\mathbf{E_{phys}}$) and those numerical errors associated with the restoring ($\mathbf{E_{num}}$).

[27] Because snowfall from the snow pillow is considered to be truth here, excess precipitation is zero ($\mathbf{S} = 0$) and reconstructed precipitation is equal to the first guess minus the model errors:

$$\mathbf{P_{rec}} = \mathbf{P_{ini}} - \left(\mathbf{E_{phys}} + \mathbf{E_{num}}\right) \tag{3}$$

Finally, after precipitation has been reconstructed in this way, the model snow depth is restored to the observed snow depth. The SWE and snowpack heat content are recalculated based on the new depth, and then the model advances to the next time step.

[28] The excess precipitation added to the first guess is assumed to have a density of freshly fallen snow (150 kg m$^{-3}$). SWE and heat content are recalculated after the restoration. At times the excess is negative, which occurs when the model overestimates the true snow depth. In this case SWE is subtracted from the initial guess ($\mathbf{P_{ini}}$) but $\mathbf{P_{rec}}$ is positive or zero. At times the model might underestimate snow depth and this would wrongly be compensated by a snowfall event, according to this reconstruction method. This is one possible source of error that would fall under $\mathbf{E_{phys}}$. It is important to remember that in this proof of concept study, the model is using true snowfall as the first guess. Without a snowfall defined as truth, it would be impossible to validate this method. At a nontest site (i.e., in the Arctic), the initial snowfall guess can be based on gauge data or positive daily snow depth changes. Sensitivity testing shows that an initial guess performs similar to the perfect guess, so long as snow depth or SWE changes are measured at least as frequently as daily and physical changes in the snowpack are occurring on a longer timescale. Conceptually, all of the experiments are inverse methods: snow depth is known from observation, model physics are given, and therefore it is possible to reconstruct true precipitation.

[29] A second experiment (the second reconstruction run) is conducted to separate the two sources of error ($\mathbf{E_{phys}}$ and $\mathbf{E_{num}}$) in experiment one. The model is run again in a reconstruction mode, but instead of restoring to observed snow depth, the model is restored to the depth produced by the control run, calculating a new reconstructed precipitation ($\mathbf{P_{ctl}}$). In this way, remaining errors should be attributed to the process of restoring to the snow depth data because physical errors have been eliminated by design. The difference between the total errors and the numerical errors should thus be the original errors associated with imperfect model physics. This experiment is illustrated in the following way:

$$\mathbf{P_{ctl}} = \mathbf{P_{ini}} - \mathbf{E_{num}}. \tag{4}$$

Rearranging equation (1) and adding the definition of $\mathbf{E_{tot}}$ from the text:

$$\mathbf{E_{tot}} = \mathbf{E_{phys}} + \mathbf{E_{num}} = \mathbf{P_{ini}} - \mathbf{P_{rec}}. \tag{5}$$

Combining equations (3), (4), (5), it follows that

$$\mathbf{P_{rec}} - \mathbf{P_{ctl}} = -\mathbf{E_{phys}} \tag{6}$$

$$\mathbf{E_{num}} = \mathbf{E_{tot}} + \mathbf{P_{rec}} - \mathbf{P_{ctl}}. \tag{7}$$

In this way, two sources of error have been isolated.

[30] A third experiment (the third reconstruction run) is conducted to show that this method can be used with historical records from the Arctic. Few snow depth and meteorological stations in the Arctic measure energy fluxes. This experiment is conducted with the minimum amount of meteorological and snow data that are available for the Arctic from the NSIDC Historical Soviet Daily Snow Depth [*Armstrong*, 2001], the Environment Canada daily snow depth and temperature databases (see http://www.ec.gc.ca/envhome.html), and the Alaskan and European domains in Global Daily Climatology Network, version 1 (http://www.ncdc.noaa.gov/oa/climate/research/gdcn/gdcn.html): surface air temperature and snow depth. The purpose of this experiment is to see if the reconstruction is sensitive to the removal of the other forcing variables. This is a run similar to the first experiment, except that no incoming radiation measurements are used to drive the snow physics. Surface air temperatures, rather than the surface energy balance, are used to drive the snow physics. Temperature of the first snow layer is made equal to the temperature of the surface air temperature. This is an extreme case; for some historic Arctic records surface energy observations are available. For clarity, these experiments, location of results, and associated errors have been summarized in Table 2.

## 4. Results

[31] The results from these four simulations are a control run, a reconstruction run (experiment one), a reconstruction run that estimates model error (experiment two), and a

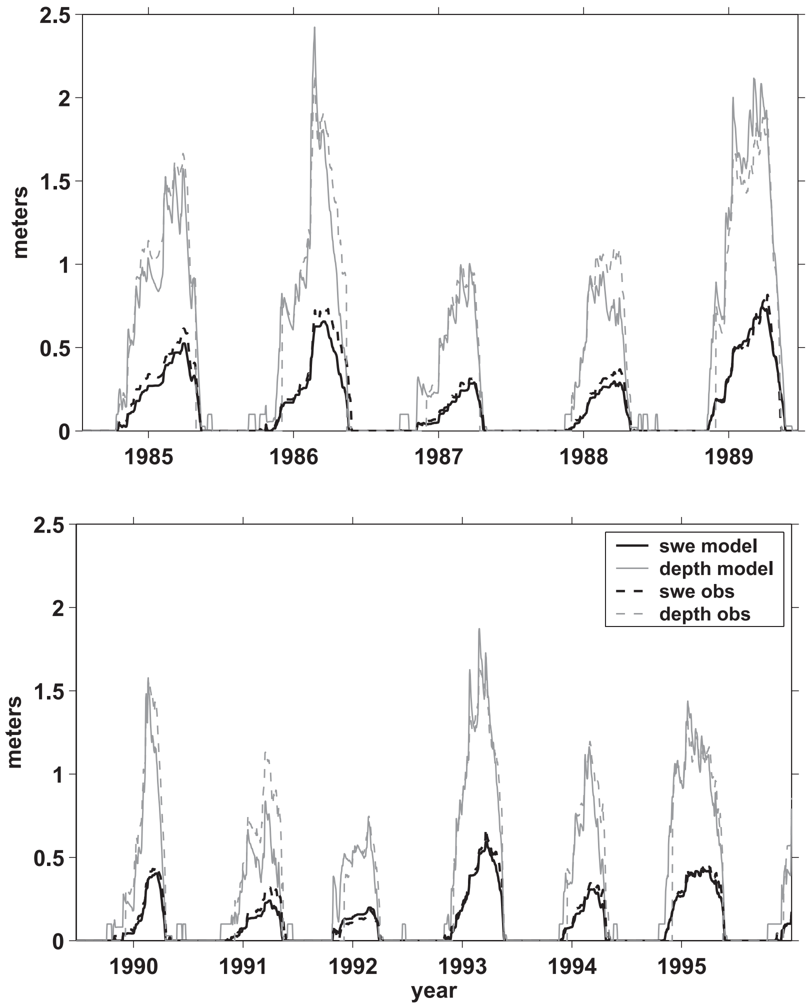


**Figure 4.** The control run of the model is plotted alongside observations for RCEW. Solid thick line is modeled meters of SWE, and dashed thick line is observed. Solid thin line is modeled snow depth, and dashed thin line is observed snow depth from a combination of the snow course (density) and snow pillow (SWE).

sensitivity test of the reconstruction run (experiment three). Results from the control run are shown in Figure 4. The snow depth and SWE produced by the CLSM match observations closely. There are several years when the model overestimates or underestimates the observed snow depth or SWE during a period of compaction and ablation. These discrepancies, in addition to discrepancies between the snow course and the snow pillow (due to instrument error and heterogeneity), will impact the reconstruction. As mentioned in section 3, there are also shortcomings in the model physics. All of these errors are embodied in $E_{phys}$. Comparison of SWE on days when it is observed on both the snow course and snow pillow can yield a general estimate of type A plus type B errors (RMS = 33.7 mm, $r^2$ = 0.98) between the two measurement systems, but these cannot be separated or isolated from incorrect model physics on an hourly or daily basis. However, with the above exceptions, all other years capture the approximate timing of the growth and ablation of the snowpack, with good estimates of the maximum snow depth and SWE. This control run shows that the model physics are capable of reproducing observed snow depth and SWE and therefore the CLSM is an appropriate tool for the method described here.

[32] Results from the first experiment are shown in Figure 5. Cumulative reconstructed precipitation is plotted against cumulative snowfall from the snow pillow and one of the corrected gauges. As expected, the reconstructed precipitation is close, in most instances, to the "true" snowfall recorded by the snow pillow, with the exception of snow years 1984–1985, 1987–1988, and 1989–1990 when the reconstructed precipitation is greater. In the cumulative reconstruction plot (Figure 5), there are times when there is no accumulation on the snow pillow and the cumulative snowfall flattens out, but the model keeps reconstructing snow. Most notable is 1984–1985, the most problematic reconstruction year. During the first few snow days of 1985, the test site was subjected to temperatures above freezing and rain on snow. What happened during these times is that the pack densified/ablated and the way the "real" snowpack densified/ablated was slightly different than the model snowpack. In this case, the model pack densified/ablated slightly faster than the "real" snowpack and therefore extra snowfall was reconstructed during the restoration to add height to the model snowpack.

[33] What we are calling the "true" snow depth is an hourly SWE from the snow pillow constrained by a snow

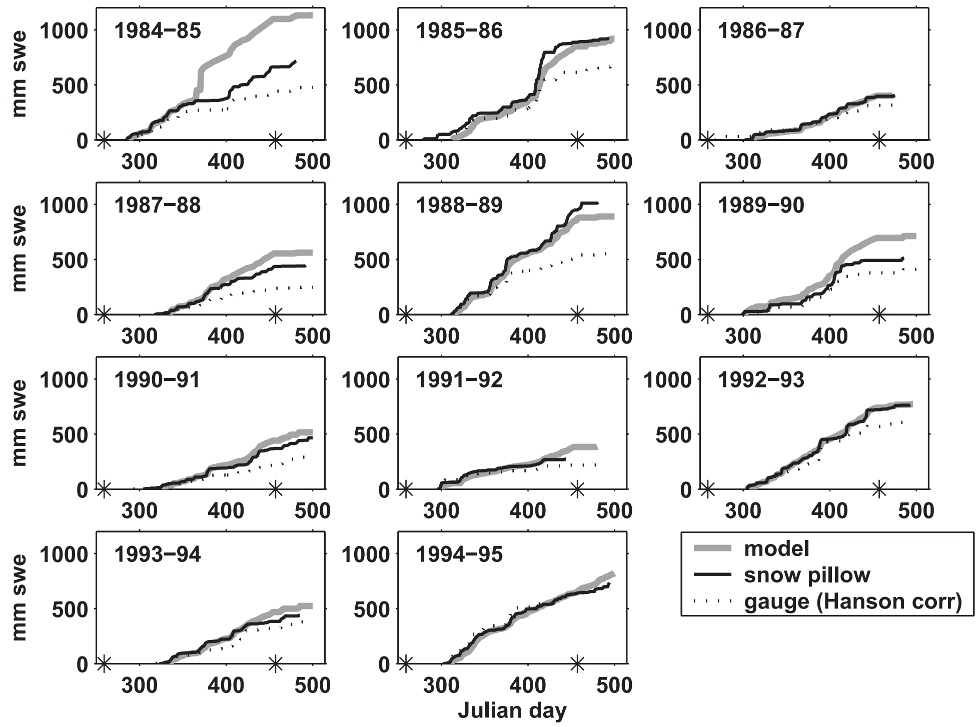


**Figure 5.** Cumulative reconstructed snowfall at RCEW is plotted with thick line. Observations from the snow pillow and the dual gauge with the Hanson correction are plotted with a thin line and a dashed line, respectively. Stars mark 15 September and 1 April, the period of reconstruction.

course density that has been observed every two weeks and then interpolated. Referring back to Figure 3, the hatches that mark snow course measurements are not capturing detailed ablation dynamics. In 1984–1985 there are two measurements in the whole end-season ablation period. The period 1984–1985 also has the dramatic midseason ablation event. Even though there was a snow course measurement during this time, it is not capturing the high-frequency changes. This is precisely when the major error occurs during the 1984–1985 reconstruction (Figure 5) and similar errors occur during early densification/ablation events in 1987–1988 and 1989–1990. The frequency of the snow course measurements constrains the accuracy of snow depth in this experiment. This will not be a problem in the Arctic sites where we have daily snow depth measurements. It should also be added that the model's densification scheme works very well at the test site, on average, and sensitivity testing of the densification parameters have shown that the published parameters in CLSM produce the most accurate control run at RCEW.

[34] Results from the second experiment are shown in Figure 6. Here the physical errors have been eliminated by restoring to the control run. Errors have been separated and plotted in Figure 7. Errors are greatest (an average of 1.6 mm $d^{-1}$) in January, which is the month of highest deposition of snow at RCEW, and again in March (an average of 1.4 mm $d^{-1}$). Numerical errors associated with the data assimilation are small, with a peak average of 0.4 mm $d^{-1}$ in March. Further analysis shows no significant correlation between physical errors and either wind speed or temperature, but a small inverse correlation between errors and amount of snow.

[35] Results from the third experiment are shown in Figure 8. At this site, estimating the surface energy budget with surface air temperature produces a cumulative reconstructed snowfall that is equal to or only slightly higher than a run using the full surface energy budget. Errors associated with this experiment are shown in Figure 9. Like the physical errors in the control run, errors in the third experiment peak in January and March at 2.3 mm $d^{-1}$ and 2.1 mm $d^{-1}$ respectively, somewhat higher (in a positive direction) than errors from the first experiment, in which more observed meteorological parameters were used to constrain evolution of the snowpack.

## 5. Discussion and Conclusion

[36] Several communities of researchers depend on a reliable Arctic snowfall record to validate their work. Developers of unbiased remote sensing algorithms need a reliable ground truth. Modelers who study past, present, and future climate simulations need time series of reliable snowfall data for validation of their methods. Finally, any researcher who considers gridded reanalysis data sets as an observation field should be aware of the biases in what they are defining as the observation. This particularly applies to those researchers who are trying to identify mechanisms of Arctic climate change and their consequences. This method looks promising for reconstructing historic snowfall in the Arctic for these and other applications.

[37] This method avoids both the systematic and some unsystematic sources of error associated with gauges because it is based on snow depth observations. Figure 10 shows mean cumulative error over the snow season. Snowfall reconstructed in the method described

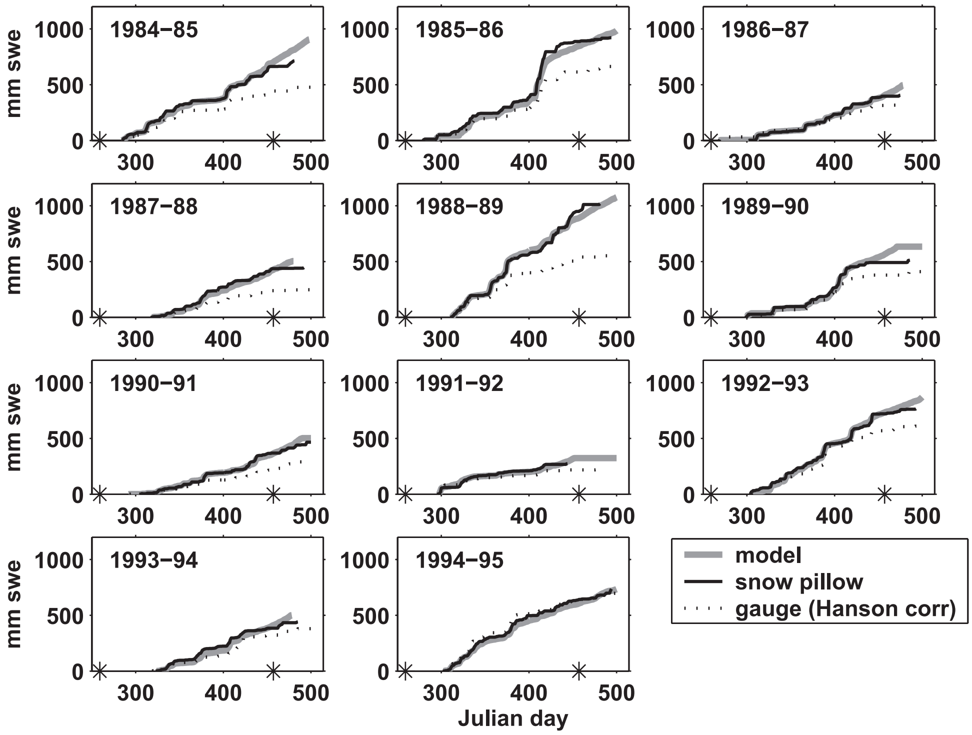


**Figure 6.** Cumulative snowfall is plotted with a thick line for a reconstruction (experiment 2) in which snow depth is restored to the control run, for the purpose of estimating errors. Observations from the snow pillow and the dual gauge with the Hanson correction are plotted with a thin line and a dashed line, respectively. Stars mark 15 September and 1 April, the period of reconstruction.

here shows 30% lower RMS error than corrected gauges. If the outlier of 1984–1985 were removed, that RMS error would be further reduced by 50%. In addition to being smaller, the mean annual cumulative reconstruction error is positive (29 mm SWE), while the mean of corrected gauges error is strongly negative (−77 mm SWE). Reconstructed snowfall is significantly greater than snowfall from corrected gauge records at RCEW and thus accounts for more snowfall than gauge corrections can recover.

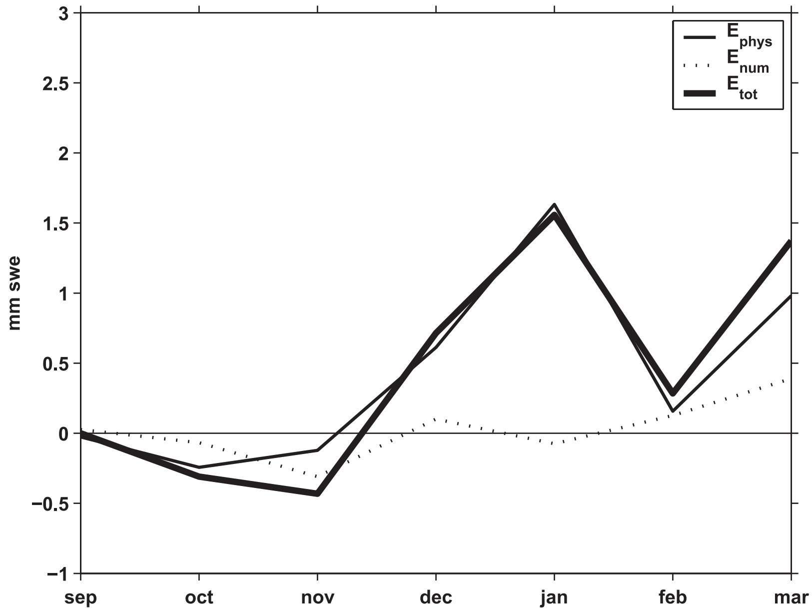


**Figure 7.** Mean daily errors by month are shown here for the snow season at RCEW. Errors in the physics are plotted with a thin line, errors in the numerics are plotted with a dotted line, and total errors are plotted with a thick line.

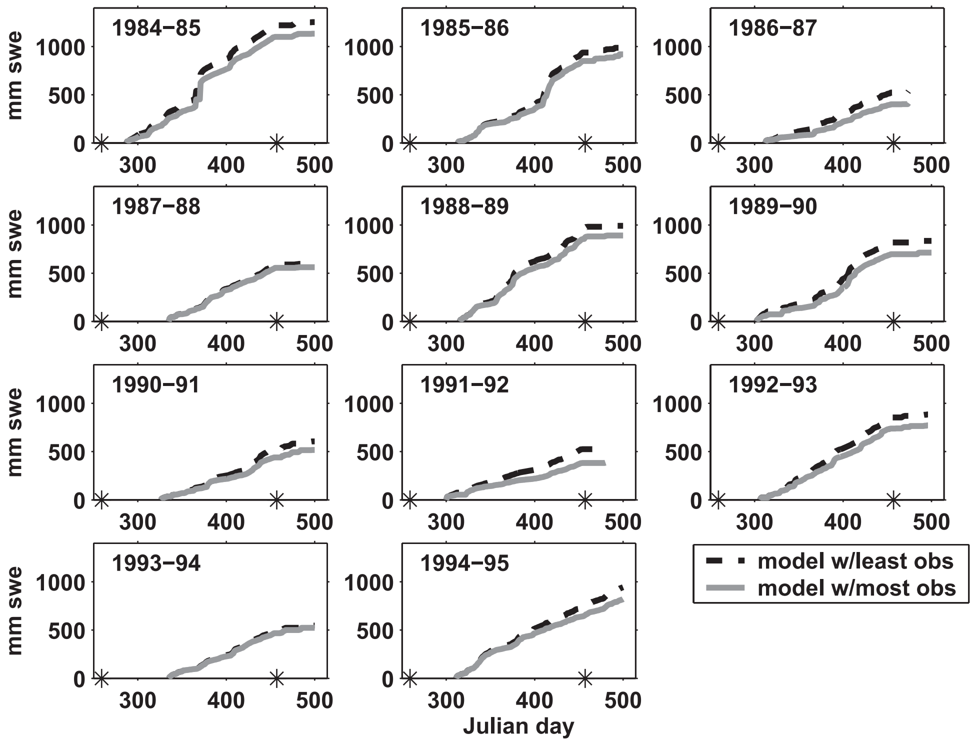


**Figure 8.** Cumulative snowfall reconstructions for RCEW are shown for the runs using the greatest number of observed surface energy budget (SEB) parameters and the fewest number (only surface air temperature). This model run tests the sensitivity of the reconstruction to the number of observed SEB parameters used.

[38] Analysis shows that the physical model errors seem to peak during months of high deposition, but an overestimate of 1.6 mm $d^{-1}$ when there is over 1.5 m of snowpack is only 3 % error by the end of the month. The significance of these errors in model physics for the monthly or annual snowfall budget in the Arctic would differ significantly between low-, medium-, and high-deposition sites. In general, the Arctic receives less snowfall than the RCEW Reynolds Mountain site. Errors related to amount of snowfall and the likelihood of the snowpack having been

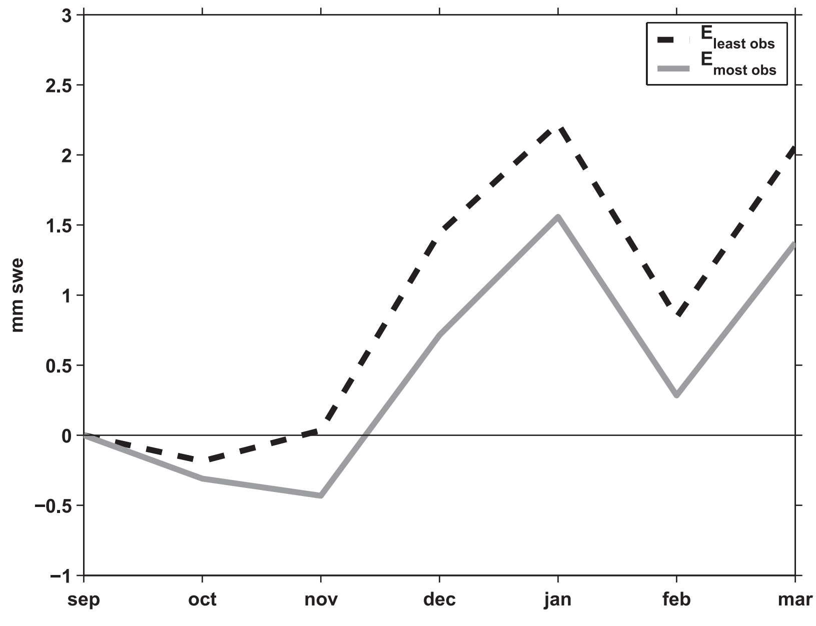


**Figure 9.** Mean daily errors by month shown for the snow season. Total errors for the case of surface air temperature as a proxy for observed surface energy budget (SEB) parameters are plotted with a dashed line, and total errors for the case of best available observed SEB parameters are plotted with a solid line.

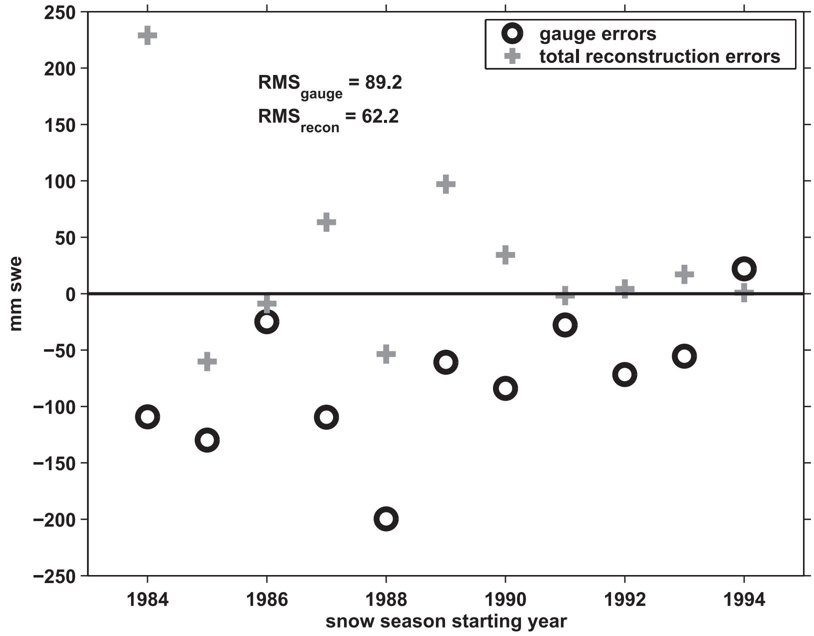


**Figure 10.** Cumulative errors from gauges (circles) and from the model (crosses) are averaged over the snow season of each year. The root-mean-square is calculated from this annual mean of cumulative error and is in units of millimeters SWE.

impacted by blowing snow will be discussed further in the Arctic paper.

[39] Like most snow models, CLSM does poorly when there is only a trace amount of snow on the ground (less than 13 mm). While at RCEW this only occurs at the very start of the season, days with trace amounts of snow on the ground may be more common in the Arctic. However, many gauges do not even record trace events. So as long as trace amounts of snow depth are measured by rulers, reconstruction will estimate snowfall more accurately than gauges. Reconstructions for periods of trace snow on the ground can be verified using assumptions of constant density and daily changes in trace snow depth. It should be mentioned that trace events are also a problem for gauges, particularly those that measure volumetrically [*Goodison et al.*, 1998], due to wetting losses.

[40] The other time of year when shortcomings in model physics create errors is in the spring season (March and later at RCEW). This is the time of year when the snowpack is ablating and there may be some rain-on-snow and mixed precipitation-on-snow events, which are difficult for any snow model to capture. As with errors during peak accumulation, it is difficult to predict how these errors will affect Arctic reconstructions. Significance of errors associated with rain-on-snow events, mixed precipitation events, and midseason melt will depend on the number of these events at any given site and the number over a given catchment. Finally, the model performs better when more meteorological observations are used, which is not possible at all stations in the Arctic. Sensitivity analysis showed, however, that reasonable results are possible even when only surface air temperature observations are available.

[41] The advantages of this method far outweigh the limitations. When applied to the Arctic, this method will utilize historical records of temperature and snow depth from Canada, Europe, Alaska, and the Former Soviet Union. In doing so, a record of snowfall is reconstructed that is independent from corrected snow gauges and based on the physics of the snowpack. When a pan-Arctic database is created, many other studies can be done based on this new solid precipitation record.

[42] A new method for reconstructing snowfall is presented here which, when applied to the Arctic, could ameliorate many of the problems associated with current estimates of the freshwater budget there. In this work, the importance of Arctic snowfall for climate and the water is discussed and the problem of gauge undercatch is described. Next, a site description and information about the model are provided and true precipitation is defined for validation of the proof of concept. A control run and three experiments are conducted. The control run shows that the model can skillfully reproduce observed snow depth and SWE. The first experiment shows that true precipitation can be reconstructed with a small amount (3% per month) of model error. It also shows that both true snowfall and reconstructed snowfall are significantly larger than snowfall from corrected gauges. The second experiment shows that most of the error from the first experiment can be attributed to imperfect model physics during the months of the highest amount of snow deposition. A third experiment shows that this method can be expected to produce accurate results in the Arctic, even where limited radiation measurements are available. Finally, this method is shown to be relatively simple, accurate, and portable to any number of future investigations.

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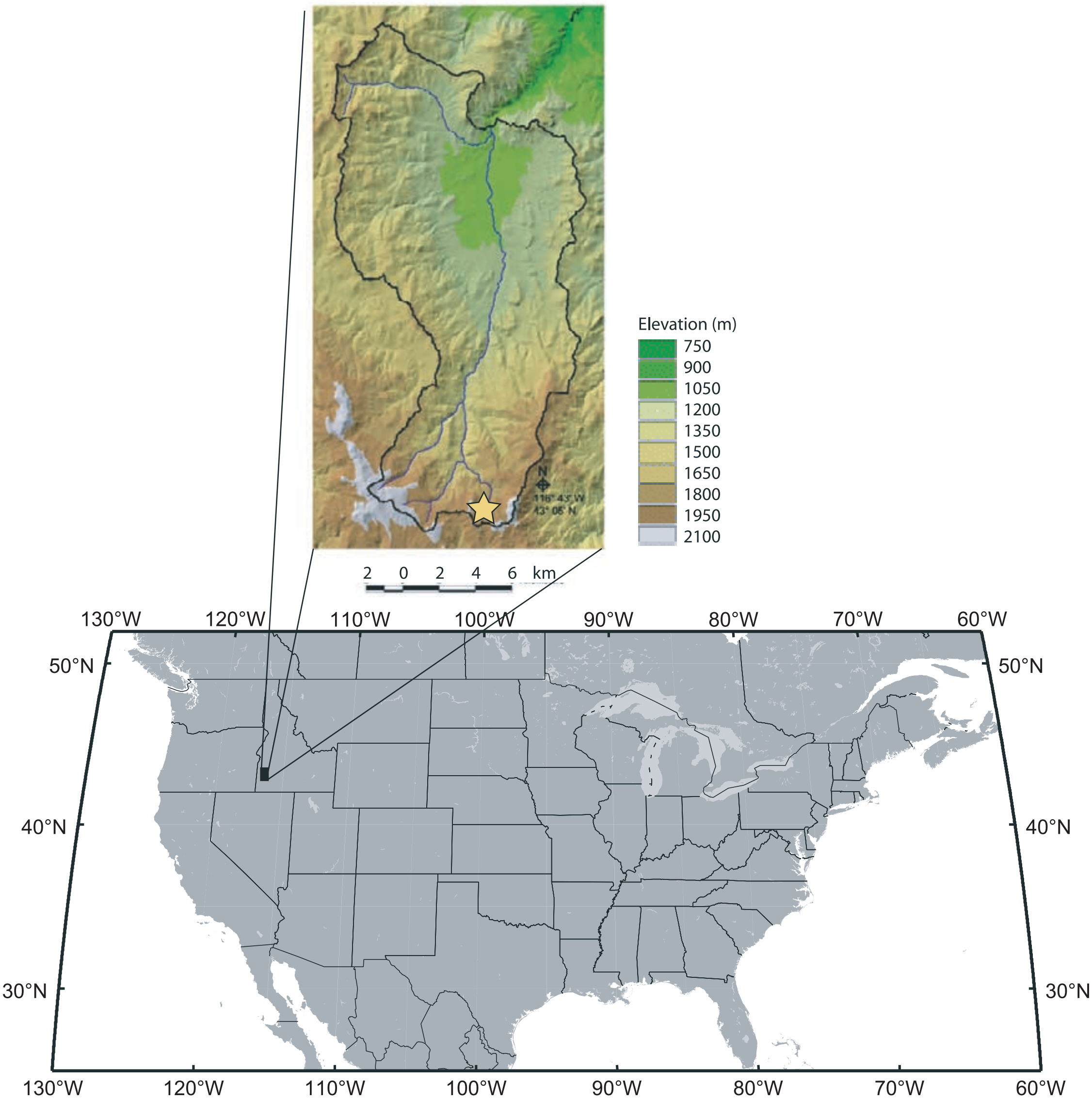


**Figure 2.** Reynolds Creek Experimental Watershed in southwestern Idaho. The Reynolds Mountain station is marked with a star. The catchment map is modified from *Seyfried et al.* [2000].