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# Interannual variability in net accumulation on the Greenland Ice Sheet: Observations and implications for mass balance measurements

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**Abstract**. Nine 24-year accumulation records from the Summit region in central Greenland are analyzed to separate the effects of spatial noise and interannual fluctuations on the variability in each core. The study shows that both processes are equally important, with standard deviations of 25 mm water equivalent per year and 24 mm water equivalent per year, respectively. A comparison with estimates of surface roughness based on high-resolution laser altimetry of the surface indicates that in the studied region the spatial noise can be reliably estimated from surface roughness. The response of the ice-sheet surface to the interannual fluctuations can be estimated using a simple zero-dimensional ice-sheet response model. For the Summit region of central Greenland, a change in surface elevation of ~20 mm water equivalent per year measured over a 5-year period, can be attributed with 95% confidence to a trend in climate. This probability decreases rapidly as the observation period is shortened. For intervals greater than ~5 year, the probability depends only weakly on the measurement interval. This suggests an optimum spacing of ~5 years between repeat elevation measurements.

## 1. Introduction

One of the potential impacts associated with global warming is a rise in global sea level caused by melting of the polar ice masses and thermal expansion of the oceans. Over the past century, global sea level has risen 10-25 cm, but accounting for this rise is difficult because of the large uncertainties in estimates of the mass balances of the ice sheets [*Warrick et al.*, 1996]. Thus resolving the current state of balance of the different components of the cryosphere is needed before predictions about future trends can be issued.

Recently, repeat laser and radar altimetry has been used to measure directly the change in surface elevation of the polar ice sheets. With repeat altimetry, elevation changes over parts of the ice sheet can be determined over short time intervals. For example, *Zwally et al.* [1989] compare data from Seasat (1978) and Geosat (1985) radar altimeters and infer a spatially-averaged thickening of the ice sheet in southern Greenland of 0.233 ± 0.041 m/year for the period 1978-1985. A recent reassessment of the satellite altimetry data by *Wingham* [1995] arrives at essentially the same thickening rate reported by *Zwally et al.* [1989]. However, *Davis et al.* [1998] find thickening rates about an order of magnitude smaller. Using repeat airborne laser altimetry, *Thomas et al.* [1995] find that the western part of the Greenland Ice Sheet south of Jakobshavn Isbræ appears to have thickened by up to 2 m over the period 1980-1993.

Relatively large changes in surface elevation or inferred ice thickness measured over a short period need not be indicative of any trend in ice volume. Such changes could also reflect the response of the ice sheet to random fluctuations in surface accumulation. How such random variations can lead to short-term elevation changes is discussed by *Van der Veen* [1993] using a simple "zero-dimensional" ice-sheet model developed by *Oerlemans* [1981]. The most important parameters for evaluating the natural fluctuations in ice-sheet surface elevation are the standard deviation of the temporal fluctuations in accumulation and the horizontal scale of these fluctuations.

Records of annual net accumulation contain noise associated with temporal variability (short-term fluctuations in snowfall due to climate variability) and spatial noise (small-scale spatial differences in the amount of snow deposited). Because it may be expected that the spatial noise is unimportant when considering the average accumulation over entire drainage basins (because the random component associated with spatial noise cancels when averaging over a sufficiently large number of measurements), it becomes necessary to separate these effects from interannual climate fluctuations, which may be considered spatially invariant over large areas. Only the temporal fluctuations in accumulation, associated with synoptic-scale meteorological phenomena, are relevant for the interpretation of short-term elevation changes. The problem is to extract the climatic temporal fluctuations from the noise-contaminated accumulation record.

To estimate the spatial variability in net accumulation due to surface microrelief requires multiyear observations of net accumulation over a larger area. Such observations can be acquired from repeat annual accumulation measurements along stake networks or from time series derived from neighboring ice cores covering a climatically representative area of the ice sheet. In this study, nine 24-year accumulation records are evaluated to estimate the spatial variability contained in these records.

In 1987, nine firn cores of ~17 m length were recovered near Summit in central Greenland as part of the Greenland Ice Sheet Project 2 (GISP2) site selection [*Bolzan and Strobel*, 1994]. The core sites cover an area of about 150 × 150 km, which also includes the United States (GISP2) and European (Greenland Ice Core Project) deep-drilling sites (Figure 1). Drill depth was carefully measured before and after drilling runs in order to accu-



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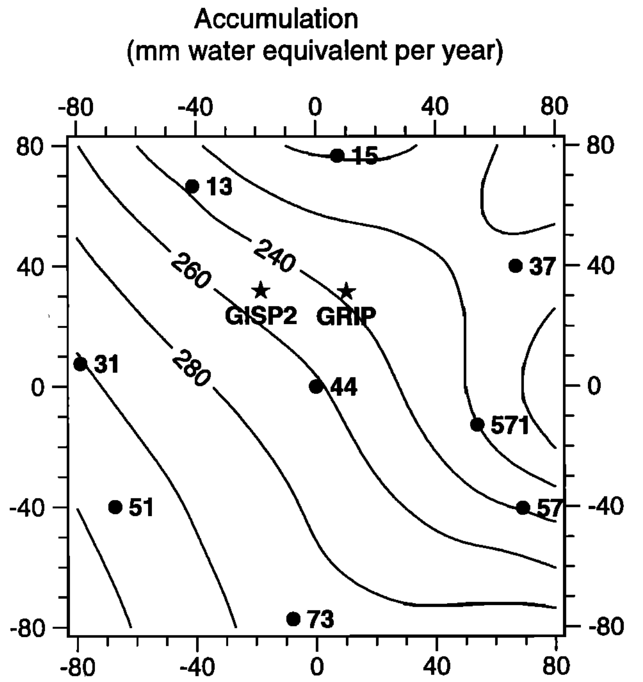


**Figure 1.** Contours of constant accumulation rate. The black dots represent the nine sites where the shallow cores were retrieved; the two stars indicate the location of the European (GRIP) and United States (GISP2) deep boreholes.

rately determine core depth; core loss was assigned using the algorithm developed by *Whillans and Bolzan* [1988]. Upon return to the Byrd Polar Research Center, densities were measured on each recovered core segment, and the cores were sampled for oxygen isotope and gross β activity measurements. Further details are given by *Bolzan and Strobel* [1994].

For each core, the stable oxygen-isotope ratio and gross β activity were measured as a function of depth, to identify and date annual layers in the firn. Peaks in the isotope ratio may be identified with summer, and troughs may be identified with winter. By measuring the layer thickness from isotopic peak to peak (or trough to trough), net accumulation from one summer to the next (or between two successive winters) can be obtained. In the present study, the summer-to-summer accumulation rates for the period 1963-1987, common to all nine records, are used (Figure 2). There is no persistent trend in accumulation that is common to all cores [*Bolzan and Strobel*, 1994], so the average distribution may be considered constant. The spatial distribution of the longer-term average is represented by the contours in Figure 1; this pattern is discussed by *Bolzan and Strobel* [1994].

## 2. Variations in Net Accumulation

The net accumulation is the amount of snow deposited at the sampling site over the course of 1 year and preserved in the stratigraphic record. This quantity shows large variability, both in space and in time, due to topographic effects, snow redistribution by wind action, and synoptic-scale fluctuations in precipitation. Consequently, time series of accumulation rate derived from adjacent ice cores may show considerable differences.

Spatial variations in net accumulation arises from the interaction between surface winds and the local topography of the snow surface. Two different length scales should be distinguished,

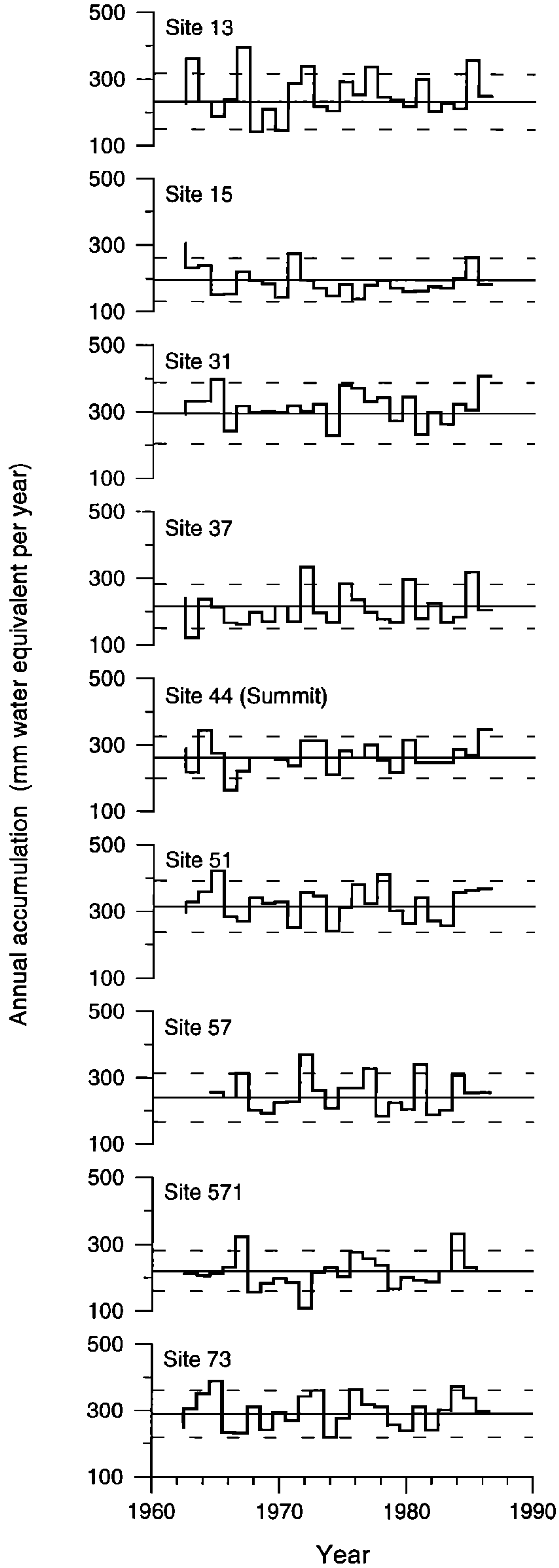


**Figure 2.** Annual values of accumulation rate for the nine Summit sites determined from the layer thickness between successive summer peaks in oxygen-isotope values. Horizontal dashed lines indicate one standard deviation from the core average.

namely microrelief, being topographic features that are smaller than a few kilometers (wind drifts, sastrugi), and macrorelief, which consists of surface undulations with wavelengths ranging from 5 to 30 km. The height of features associated with microrelief is generally too small to affect winds in the surface boundary layer immediately above the snow surface. Further, these features are transient with a typical lifetime of 1 year or less. This microrelief contributes to the random noise observed in accumulation records. Macrorelief, however, tends to persist over many years, and the air flow in the surface boundary layer conforms to these surface undulations, resulting in a feedback between snow drift and surface topography [*Whillans*, 1975], leading to a spatial accumulation distribution that may persist over many years. This spatial pattern can lead to trends in accumulation histories from ice cores, which could be erroneously interpreted as indicator of a changing climate.

Spatial noise in net accumulation due to microrelief manifests itself as temporal fluctuations in a record obtained at a sampling site. This is because small-scale fluctuations in space are not uniform in time but randomly distributed over the glacier surface. For example, during 1 year, a sastrugi may develop and be preserved in the stratigraphic record, but there is no a priori reason to expect comparable sastrugi to form at this site in the next year. Thus the stratigraphic record will be characterized by noise due to the random occurrence of surface microrelief.

Interannual variability in precipitation is caused by varying numbers of storms and fluctuations in their tracks. The modeling study of *Bromwich et al.* [1993] indicates a high degree of year-to-year variability in precipitation over Greenland, with changes of 10% or more from one year to the next occurring fairly frequently. Periods of relatively low precipitation are associated with a systematic shift away from the ice sheet as well as a decrease in the activity of the dominant storm tracks. According to the model results of *Bromwich et al.* [1993], the spatial scale of these variations is several hundreds of kilometers, indicating that large parts of the ice sheet are affected similarly by a change in storm activity. This means that at least part of the temporal variability in accumulation may be considered uniform over regions comparable in size to individual drainage basins.

Interannual fluctuations can be estimated independently if climate observations are available. For the Greenland Ice Sheet, this was done by *Bromwich et al.* [1993] for the period 1963-1988, using as input data the analyzed geopotential height fields produced by the National Meteorological Center. Their precipitation model includes a parameterization of the synoptic activity at the 500 mbar level, plus orographic forcing by the ice-sheet surface profile. Available climate data span only the last 30 years or so; thus, for records going back much farther in time, temporal variations must be considered the unknown quantity to be extracted from the core data.

## 3. Method

There are three processes contributing to the interannual fluctuations in net accumulation shown in Figure 2, namely spatial noise, climate variability, and random measurement errors. These processes may be described by three independent normally distributed stochastic terms with zero mean. That is, the annual net accumulation, $A(\vec{x},t)$, at horizontal postion $\vec{x}$ and time $t$ can be written as

$$A(\vec{x},t) = A_o(\vec{x},t) + A_s(\vec{x},t) + A_c(\vec{x},t) + \varepsilon \tag{1}$$

where $A_o$ is the long-term average accumulation, including slowly varying climate trends and effects from large-scale surface undulations; $A_s$ is the spatial noise associated with small-scale (~1 km or less) microrelief of the snow surface; $A_c$ is the contribution from fluctuations associated with the synoptic-scale climate variability, which are assumed to be uniform over large areas; and $\varepsilon$ is the measurement error. The standard deviation of the observed record is then

$$\sigma^2 = \sigma_s^2 + \sigma_c^2 + \sigma_\varepsilon^2 \tag{2}$$

The standard deviation due to measurement error ($\sigma_\varepsilon$) can be estimated independently from the accuracy with which individual layers can be identified and measured, while the standard deviation of the actual records ($\sigma$) can be determined from the core records. By averaging the nine core records, it is also possible to evaluate the standard deviations associated with spatial and temporal variability, respectively, thus allowing for a consistency check of the estimated standard deviations.

### 3.1. Measurement Uncertainty

One source of uncertainty in determining annual accumulation values is ambiguity in dating the isotopic seasonal features. This uncertainty is difficult to quantify and can arise if there are years with multiple isotopic peaks or if there are missing years. Here measurement of gross beta activity provided reference horizons which were used to date the isotopic peaks. This dating was unambiguous, except for one summer peak, 1976, in the core 15 profile which seemed to be missing as a result of core loss. The location of this peak was assigned to center of a core loss zone, but this dating ambiguity affects only the summer-to-summer accumulation rates for 1975-1976 and 1976-1977.

Once the isotopic profiles are dated, the largest source of uncertainty in measuring annual accumulation values is specifying the timing of the summer horizon in each annual layer [see *Bolzan and Strobel*, 1994]. While the nominal length of each accumulation year is 1 year, there is some uncertainty in the timing of the occurrence of summer peaks in the oxygen-isotope record. This peak corresponds to the summer peak in air temperature. Available Automatic Weather Station data [Stearns and Weidner, 1991] show the warmest air temperatures generally occuring in June and July, although similar temperatures can also occur in August. Assuming that the most positive isotope values are associated with the warmest air temperatures, then isotopic peaks refer to snow deposited between June and August. Making the additional assumption that the occurrence of temperature maxima can be described by a Gaussian probability distribution, the corresponding standard deviation of the timing of each stratigraphic summer horizon is ~1 month. Because annual accumulation rates are determined from two consecutive summer peaks, the standard deviation of the length of accumulation years is 1.4 months. Using an average value of 250 mm water equivalent per year for the accumulation rate, the standard deviation from measurement error is 30 mm water equivalent per year.

### 3.2. Interannual Climate Variability

To evaluate the standard deviation of the interannual variability, the assumption is made that interannual climate fluctuations are uniform over the region. The annual deviation from the long-term mean at each site is averaged over the entire region. Denoting spatial averages by <...>, equation (1) becomes

$$< A(\vec{x},t) - A_o(\vec{x},t) > = < A_s(\vec{x},t) > + A_c(t) + < \varepsilon > \tag{3}$$

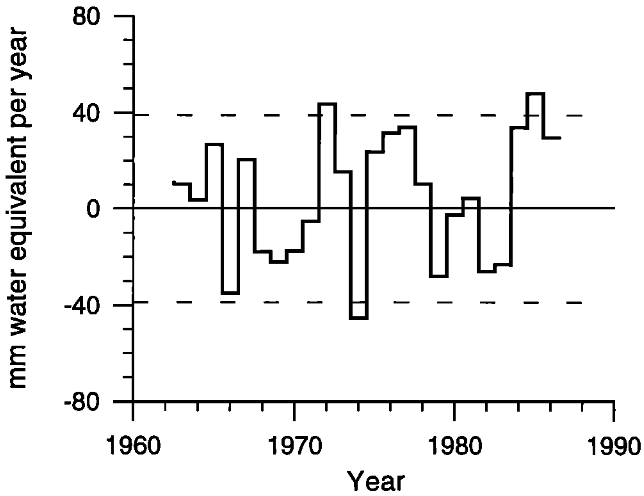


**Figure 3.** Areally averaged deviation of the accumulation rate from the long-term trend. Horizontal dashed lines indicate the standard deviation.

The left side of equation (3) is evaluated by averaging the nine records, with the long-term mean subtracted. The result of this calculation is shown is Figure 3.

To compute $\sigma_c$ the standard deviations associated with each of the terms in equation (3) are considered. The first and third terms on the right-hand side are the sums of nine independent normally distributed stochastic terms. The standard deviations of these sums are $\sigma_s / \sqrt{9}$ and $\sigma_\varepsilon / \sqrt{9}$ , respectively. Hence

$$\sigma^2_{<A>} = \sigma_c^2 + \frac{1}{9}\left(\sigma_s^2 + \sigma_\varepsilon^2\right) \tag{4}$$

The standard deviation of the time series shown in Figure 3 is 28 mm water equivalent per year. Because the standard deviation associated with spatial noise, $\sigma_s$ has yet to be determined, equation (4) cannot be used by itself to infer the interannual variability.

### 3.3. Spatial Noise

Spatial variability can be estimated for each year by subtracting the spatially averaged record shown in Figure 3 from the actual records at each site, to eliminate the interannual variability. That is, a new time series, $\tilde{A}(\bar{x},t)$ , is defined as

$$\begin{aligned}\tilde{A}(\bar{x},t) &= A(\bar{x},t) - < A(\bar{x},t) - A_o(\bar{x},t) > \\ &= A_o(\bar{x},t) - A_s(\bar{x},t) - < A_s(\bar{x},t) > + \varepsilon - < \varepsilon >\end{aligned} \tag{5}$$

Noting that $A_o(x,t)$ is constant [*Bolzan and Strobel*, 1994], the corresponding relation between the standard deviations is

$$\sigma^2_{\tilde{A}} = \frac{10}{9}\left(\sigma_s^2 + \sigma_\varepsilon^2\right) \tag{6}$$

For each of the nine cores, $\sigma_{\tilde{A}}$ can be evaluated. Averaging these values gives $\sigma_{\tilde{A}} = 41$ mm water equivalent per year. With $\sigma_\varepsilon = 30$ mm water equivalent per year, the standard deviation of the spatial noise is $\sigma_s = 24$ mm water equivalent per year.

## 4. Results

With $\sigma_s$ known, the standard deviation of the climate fluctuations can now be estimated from equation (4). The result is $\sigma_c = 25$ mm water equivalent per year. Thus interannual climate fluctuations and surface microrelief contribute equally to noise in accumulation records from central Greenland.

As a consistency check, the three standard deviations determined above can be used to calculate the standard deviation of the actual records and the result compared with the standard deviation derived directly from the measurements. The measured standard deviation of the nine records varies between 28 mm water equivalent per year and 81 mm water equivalent per year. The reason for this variation is the relative shortness of the records. A more accurate estimate requires a much longer record. Equivalently, the average of the standard deviations of the nine cores may also be considered a reliable estimate of the standard deviation in net accumulation in this region and is 50 mm water equivalent per year. This value agrees with the estimate based on the sum of the squares of the three standard deviations estimated above (46 mm water equivalent per year).

There are few direct measurements of surface roughness available. *Palais* [1980] used elevation measurements along a bamboo stake network near Dome C, East Antarctica, to infer a standard deviation of the microrelief of ~ 20 mm we. Similar values for the roughness associated with sastrugi are reported by *Whillans* [1978] along the Byrd Station Strain Network and by *Hulbe and Whillans* [1994] for a strain grid near the UpB camp on ice stream B, West Antarctica. These estimates of surface roughness are based on measurements of the exposed length of stakes placed typically ~500 m apart.

With the advent of remote sensing, it has become feasible to estimate the roughness of the snow surface more accurately over larger areas. For the Summit region, results from laser altimetry were provided by W. Krabill. A fuller description of the interpretation of these data is given elsewhere [*Van der Veen et al.*, 1998], but the main result of importance to the present study is the derived standard deviation of the small-scale surface features. The altimetry data allow the surface to be mapped with a horizontal resolution of a few centimeters. For the perturbations from the large-scale trend, the standard deviation is 16 mm water equivalent. Since each stratigraphic layer is bounded by two such random surfaces, the standard deviation of the corresponding stratigraphic noise in accumulation rate is 23 mm water equivalent per year [*Van der Veen et al.*, 1998]. This value agrees well with the estimate obtained in the present study from the nine accumulation records and is also compatible with the previous estimates referred to above.

## 5. Implications for Repeat Altimetry

The Summit cores indicate a standard deviation in accumulation rate due to temporal fluctuations of 25 mm water equivalent per year. This corresponds to ~10% of the long-term average in the region. In response to these fluctuations, the ice-sheet surface elevation will exhibit short-term variability. As shown by *Van der Veen* [1993], the probability that the average rate of change in surface elevation measured over $n$ years that can be attributed to fluctuations in accumulation exceeds the value $\Delta_1$ is given by

$$P(\Delta_1) = \frac{1}{2}\left[1 - \mathrm{erf}\left(\frac{\Delta_1}{\sigma_c}\sqrt{\frac{n}{2}}\right)\right] \tag{7}$$

where $\sigma_c$ represents the standard deviation of interannual fluctuations in accumulation rate. Because the mean of the accumulation fluctuations is zero, this probability function is symmetric, $P(\Delta_1) = P(-\Delta_1)$, and $P(0) = 0.5$.

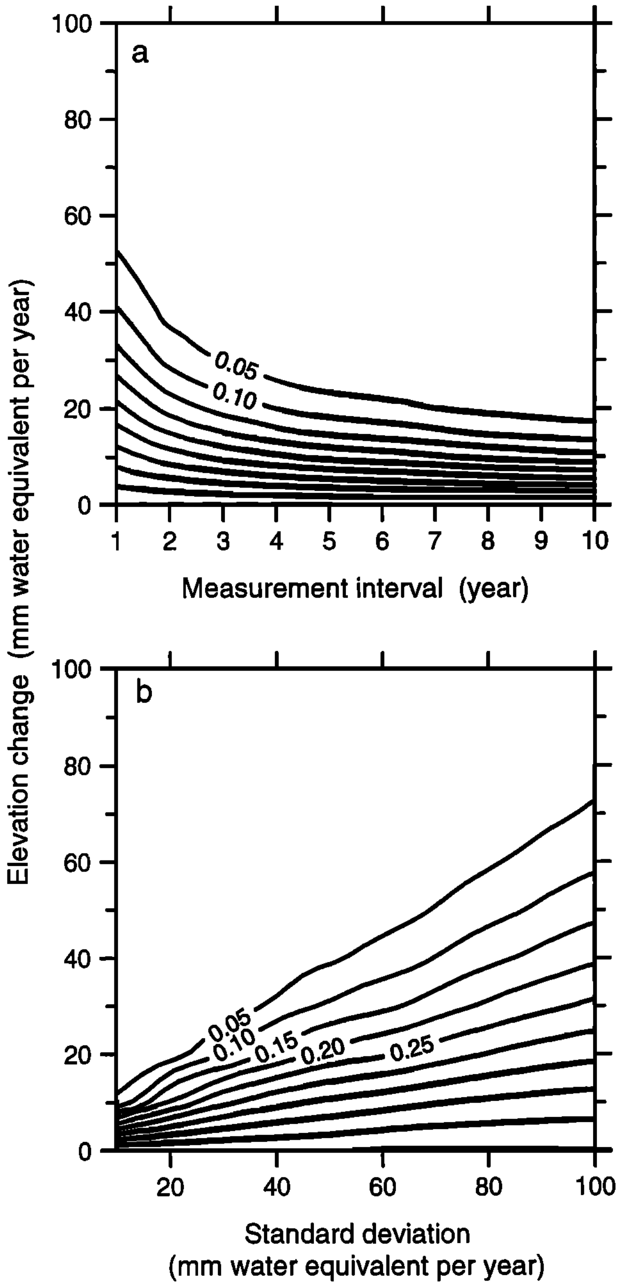


**Figure 4.** Contours of equal probability (interval: 0.05 or 5%) (a) as a function of elevation change and measurement interval for $\sigma_c$ = 25 mm water equivalent per and (b) as a function of the elevation change and interannual variability for an observation period of 5 years.

Using equation (7), the likelihood of a change in surface elevation being associated with fluctuations in accumulation rate can be assessed. For $\sigma_c$ = 25 mm water equivalent per year, the result is shown in Figure 4. The first feature to note is that increasing the measurement period has the greatest effect for small values of $n$, but, for $n \geq 5$, the probability becomes almost independent of the length of the observation period. Second, for the Summit region, the 5% probability level is ~20 mm water equivalent per year ($n \geq 5$ year). That is, a change in surface elevation larger than 20 mm water equivalent per year can, with 95% confidence, be considered climatically significant.

## 6. Conclusions

The analysis of variability in annual accumulation rate presented in this study was aimed at arriving at an estimate for the interannual variability. This requires separating the effects of spatial variability and measurement errors from the actual variability of an accumulation record. Where multiple cores are available, such as near Summit in central Greenland, spatial noise can be estimated by considering the deviation of each core record from the spatial average. However, in many instances, only one core is available. In that case, the spatial noise can be estimated from measurements of small-scale surface roughness. For the Summit region, results from laser altimetry agree well with the estimate obtained in the present study from the nine accumulation records and is also compatible with the previous estimates referred to above.

For the region considered, interannual variability associated with climate fluctuations and small-scale spatial noise (sastrugi) contribute about equally to the variability in net accumulation (standard deviation of 25 mm water equivalent per year and 24 mm water equivalent per year, respectively). While a number of studies has been undertaken to estimate interannual variability from stratigraphic records (summarized by *Van der Veen* [1993]), these studies do not adequately separate temporal from spatial fluctuations. Consequently, estimates for $\sigma_c$ quoted in the literature span an order of magnitude and tend to be larger than the value derived here.

The elevation of the ice-sheet surface will show short-term variability in response to the interannual fluctuations in accumulation. For central Greenland, the 5% probability level is ~20 mm water equivalent per year measured over a period of 5 years. That is, if a change of 20 mm water equivalent per year in surface elevation is observed over a period of five years, one can, with 95% confidence, attribute this change to a trend in climate, instead of being the response to interannual fluctuations. For intervals greater than ~5 years, the confidence level becomes almost independent of the measurement period. This suggests an optimum interval of ~5 years for conducting repeat altimetry measurements of the surface elevation of the polar ice sheets.

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