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Snow accumulation variability and random walk: how to interpret changes of surface elevation in Antarctica

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Abstract

Because snow accumulation above an ice sheet is a cumulative process, a random fluctuation of snow accumulation can lead to significant variability in ice sheet mass or elevation, without being linked to a long-term climatic change. Moreover, ice sheet mass balance cannot be measured directly. We can determine the ice sheet surface elevation through altimetric data but, due to the densification process, the snow accumulation variability affects the surface elevation variations more than the mass variations. We investigated the effect of the recent snow accumulation variability on both ice sheet mass and volume at the time scale of the satellite era. There is more than 10% chance of measuring an artificial trend greater than 15% of the mean accumulation rate, from a 10-year elevation series. In our simplified model, the impact of snow accumulation variability only depends on the ratio between the snow densification time and the variability time period: It decreases with the densification rate and increases with the variability frequency, so that the induced error is maximum for high-frequency noise in the central part of Antarctica. Finally, our results showed that knowledge of snow accumulation variability for the past 5–10 years is needed to eliminate the delayed induced error.

Keywords: ice sheet; mass balance; snow accumulation variability; stochastic process

1. Introduction

Climate variability is one of the limiting factors when interpreting a short time series for climate change. Depending on the effect of a given random parameter on a given system, the impact in terms of long-term change and the risk of misinterpreting any temporal series can be more or less crucial. For instance, a Gaussian noise added on to a trend only decreases the correlation. The induced standard error decreases with the length of the series. If, however, some feedback processes act on the system, the problem is more serious, as is the case for sea ice or continental glaciers affected by the Earth's albedo. It is also more serious if the noise, even a Gaussian one, contributes to a cumulative effect, e.g., if each random

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fluctuation is added to the previous ones. Indeed, in this case, the system undergoes fluctuations that can be compared with a random walk, and it can move far away from its initial equilibrium without any cause acting on long term. Any terrestrial reservoirs are affected by the cumulative effect of random fluctuations on their input, in particular, ice sheets, such as those in Antarctica or Greenland.

The ice sheet mass is controlled by the balance between the flow that evacuates the ice and the snow falling each year that feeds the ice sheet. This snow accumulation gets a large variability [1]. As in the case of a random walk, the ice sheet surface may move far away from its initial value, except that the long dynamic response will slowly restore the volume to its initial value. The time required for an ice sheet to return to equilibrium after a perturbation is the relaxation time; this varies from 100 years near the coast to up 100,000 years in the central areas [2]. Previous studies have already investigated the long-term effect (namely, at the time scale of the relaxation time) of short-term snow accumulation variability of ice sheet volume [1–3]. All these studies concluded that this phenomenon leads to a change in elevation of the ice sheet, which is not obviously related to a long-term volume change, without any climatic significance but which may nevertheless contribute significantly to sea level change at the human scale.

Moreover, the estimation of ice sheet mass variations is not a straightforward process. Although we can now determine ice sheet elevation fluctuations accurately, using satellite altimetry data, volume and mass variations differ because of changes in snow density in the firn. (Changes in bedrock elevation, of the order of 1 mm/year at most, can be neglected.) Changes of firn densification are due to changes in accumulation rates, surface temperature or surface density.

The effect of the intra- and interannual variations of temperature on the firn densification process has already been studied [4,5]. The authors found that for Greenland Summit, most of the densification and consequent surface lowering occur within three months in late spring/early summer. Moreover, in central Antarctica and central Greenland, elevation changes caused by firn densification changes can be similar in magnitude to those caused by natural fluctuations in the mass [6].

In our work, we compared the impact of accumulation variation on (1) ice sheet mass variations and (2) ice sheet elevation variations, the latter being evaluated by taking the firn densification process into account. We took accumulation variations either random or periodic. In this later case, we expressed the impact of the accumulation perturbation with respect to its time period and to the firn densification time period. We tried to quantify the length of satellite series of elevation data, which would be needed to be able to infer real mass balance changes.

We focussed on the Antarctica ice sheet, which represents 90% of the terrestrial ice and whose annual balance of 2280 Gt (or km^3; [7]), corresponds to about 6.5 mm of sea level.

2. Snow accumulation variability and ice sheet mass fluctuations

We shall not consider here the problem of firn densification: The snow that falls on the ice sheet has a density equal to the ice density. Over a short period, the fluctuation H_m' of the ice sheet elevation H_m with respect to the equilibrium shape is directly linked to the fluctuations b' of the accumulation rate b, expressed in ice equivalent:

$$\mathrm{d}H_m' / \mathrm{d}t = b' \tag{1}$$

(The m subscript is used here because elevation and mass changes are proportionally linked.)

In Antarctica, sublimation or evaporation contributes to less than 5% of the surface balance [8]. We will thus consider here that snow accumulation is equal to snow precipitation.

Few stations allow for the analysis of long time series of year-to-year precipitation variations. At the Orcadas station in the South Orkney Islands, north-east of the Antarctica Peninsula, where precipitation-gauge observations have been made since 1907, the standard deviation of the accumulation rate is found to be up to 34% [9]. The nine stations studied by [9] show variations from 15% to up to 45% (at Vostok) of the annual average precipitation. For instance, Morgan et al. [10] analysed a 200-year-old ice core near Law Dome and also found an interannual variability of 25%. We shall consider here an average year-to-year variability of 25% with respect to a mean value of 16.6 cm/year water equivalent (w.e.; [7]).

The stochastic process affecting the temporal series of ice elevation measurements is given by the integration of Eq. (1):

$$H'(t) = \int_{-\infty}^{t} G(t')\mathrm{d}t' \quad (2)$$

$G(t')$ represents the variation of the accumulation rate. We suppose here that it is a random series following a Gaussian distribution of root-mean-square (RMS) 25% of the mean accumulation rate. Some results of the games theory of chance or of the Markov chains are rather commonplace and easy to demonstrate. On the contrary, other results cannot be determined intuitively. For instance here, as the time increases, the elevation will remove further away from its initial position. The probability of it returning to its original value decreases with the time.

In Fig. 1, we show the fluctuations of the ice sheet mass for an average accumulation rate of 16.6 cm/year. This was done by applying a stochastic process of root-mean-square (RMS) of 4 cm/year. As evidence, the cumulative effect of the noise can lead to a coherent signal over a few decades.

To estimate the chance of having an artificial trend over 10 years, we launched 100,000 runs of 10 years and statistically analysed their trend. The average trend is of course close to 0, but the RMS is 1.4 cm/year or approximately 8% of the mean accumulation rate. This means that there is about 30% chance of having an artificial trend greater than 1.4 cm/year (in absolute value) or about 20% of chance of having an artificial trend greater than 10% of the mean accumulation rate (in absolute value).

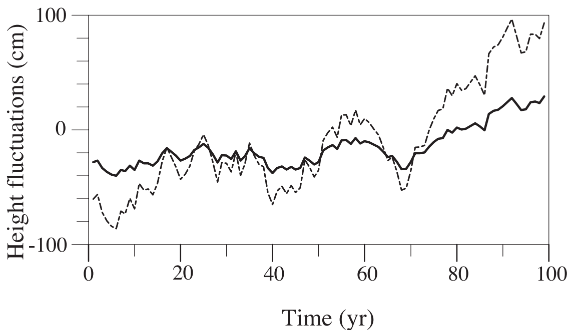

Fig. 1. Elevation fluctuations due to random snow accumulation variations of 25% of the nominal accumulation rate (namely, 16.6 cm/year), with an instantaneous densification of the firn (solid line) or a density in the firn given by our simplified model (dotted line, for k=0.01 year^{-1}). The process is initialised for 1000 years before the time t=0 of the plot.

3. Snow accumulation variability and ice sheet volume fluctuations

3.1. Firn densification model used

Snow falls with a mean density ρ_s of around 0.35 g/cm^3. With time, the snow sinks into the snowpack, densification occurs, and snow is slowly transformed into ice. But the densification process and rate are still poorly known.

Depending on the age and depth of the snow, three stages of densification may be distinguished [11]. During the first stage (ρ between ~0.3 and ~0.6 g/cm^3), densification is mainly a structural rearrangement of grains by grain-boundary sliding. During the second stage (ρ between ~0.6 and ~0.9 g/cm^3), densification is due to the grains deformation and leads to closed air bubbles. During the third stage, densification is driven by a pressure lag between ice and air bubbles. The vertical profile of density derived from a shallow ice core clearly reveals these different stages [12].

For our evaluation of ice elevation changes, we used the simplified and empirical model of Herron and Langway [13], which gives the densification rate for a snow particle sinking into the firn:

$$\frac{\mathrm{d}\rho}{\mathrm{d}t} = K(T)b^{a}(\rho_{i} - \rho) \quad (3)$$

Where ρ is the density of the snow particle and ρ_i is the ice density. $K(T)$ is an Arrhenius relation depending on the temperature T. The mean accumulation rate b intervenes at a power a, that can be taken as 1 within the whole profile [4]. We can show that the integration of Eq. (3) gives:

$$\rho(t) = \rho_{i} - (\rho_{i} - \rho_{s})\exp(-kt) \quad (4)$$

where

$$k = K(T)b \quad (5)$$

and ρ_s is the surface snow density.

Compaction velocity then depends on initial snow density, internal temperature and average snow accumulation.

The values for the parameters of the Arrhenius function $K(T)$ (preexponential and activation energy) are not very well known. They are usually taken to be constants independent of temperature. Herron and Langway [13] derived empirical values for the activation energy varying from 10 to 20 kJ/mol, while Alley [14] obtained a constant value of 41 kJ/mol, and Arnaud [11] used a constant value of 60 kJ/mol. Zwally and Li [5] used a relation in which both activation energy and the preexponential factor are temperature dependant. The activation energy is found to decrease from 100 kJ/mol at -10 °C to 20 kJ/mol at -50 °C. These values agree fairly well with the time needed to transform firn into ice, which varies from 100 years in Greenland to a few hundred years or more than one thousand years in the central part of the Antarctica, corresponding to k varying from 0.02 to 0.002 year^{-1}. For instance, this last value fits the first hundred years of the density profile measured at Vostok [11] very well.

Li and Zwally [4] have shown that the history of the densification rate is strongly dependant on the season of the snow deposition. They used data from a drilling site at Berkner Island, which can be considered to be a typical climatic site in Antarctica, with a mean temperature of -27 °C and a mean accumulation rate of 0.18 cm/year. They estimated a value of k for the first 4 years (see Eq. (5)) of 0.1 year^{-1} for a snow layer falling in the early spring, k=0.07 year^{-1} for a late-winter layer and k=0.045 year^{-1} for a late autumn layer.

The range of k for the upper layers (e.g., for the first thousand years) seems to vary between 0.1 and 0.002 year^{-1}. In the following, we will use a reference value of 0.01 year^{-1} and perform a sensitivity test within the whole range.

Of course, the Herran and Langway model is not the more realistic firn densification model nowadays. In particular, the Arnaud model [11] is based on a physical description of the three densification stages described above. But we used this simplified model for the following reasons: (1) The aim of our work is to qualitatively investigate the snow accumulation variability effect on ice sheet variations, but not to quantify it exactly. (2) The upper part of the firn plays the most important role in the difference between ice sheet mass and elevation changes. The Herron–Langway model seems to give reasonable results in this part. (3) The Herron–Langway model is an empirical one, based on firn profiles comparison. It is thus only designed to study a firn in a steady state. Here, we consider perturbations around a steady state (high frequency variability of the accumulation), and it is why this model is still valid.

3.2. Volume variations

We assume that the ice sheet is in a steady state, meaning that the initial snow density, internal temperature and snow accumulation are constant. In this case, one can show that the change in ice elevation due to snow accumulation variability $G(t')$ can be expressed as a convolution between the accumulation rate anomaly and the compaction function. Then, Eq. (2) expressed in terms of elevation change and no longer in terms of ice mass can be written as:

$$H'(t) = \int_{-\infty}^{t} G(t')/\rho(t-t')\mathrm{d}t' \tag{6}$$

Fig. 2a illustrates the time during which a perturbation in accumulation rate can persist in the difference between measured and ice-equivalent elevation, for four different values representing the acceptable range of variations. If one assumes a perturbation b' occurring at time t_0, the induced artefact in terms of surface elevation decreases from $b'(t_0)/\rho_s$ at t=t_0 to $b'(t_0)/\rho_i$ after a long time. Half of the artefact hardly exists after 6 years for the higher densification rate value (k=0.1) and after 10 years for k=0.05. On the contrary, for a lower densification rate, corresponding to a cold temperature and a weak accumulation rate, a large part of the artefact still exists a few centuries after the initial perturbations.

In terms of height fluctuations, the derivative of Eq. (6) is given by:

$$\mathrm{d}H(t)/\mathrm{d}t = G(t)/\rho_s + \int_{-\infty}^{t} G(t')\frac{\mathrm{d}\rho/\mathrm{d}t(t-t')}{\rho^2}\mathrm{d}t' \tag{7}$$

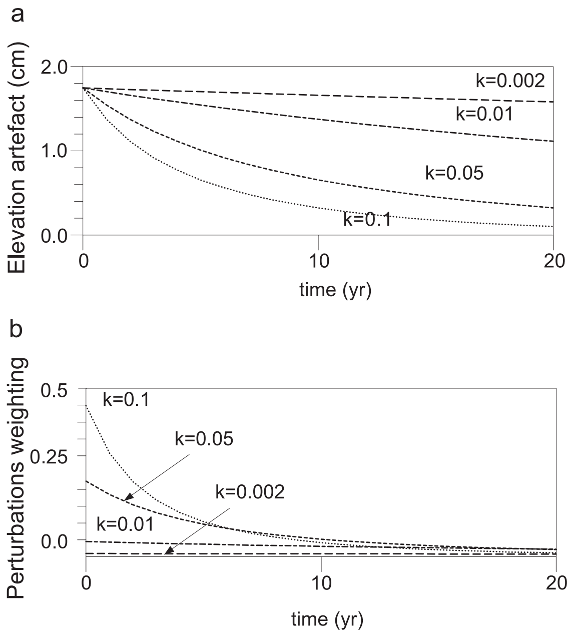

Fig. 2. Evolution of the artefact in ice sheet mass measurement after a simple excess of accumulation of 1 cm/year during 1 year (namely, the $1/\rho(t)-1/\rho_i$ function), for different densification rates (see k in Eqs. (4) and (5)) (a) and of the weighting function of the perturbation (see $f(t)$ in Eqs. (7) and (8)) (b).

The time derivative of the surface elevation is then due to the instantaneous perturbation plus a term due to the densification of the past perturbation weighted by a function $f(t-t')$ such that:

$$f(t) = \frac{\mathrm{d}\rho/\mathrm{d}t(t)}{\rho^2} = \frac{k(\rho_i - \rho)}{\rho^2} \tag{8}$$

This function strongly decreases with the densification rate (see Fig. 2b). However, for a high rate of snow densification, the recent history plays an important role. The weight of the last year's perturbation is 0.45 for k=0.1, while it is 0.05 for k=0.01. After 10 years, the weight decreases to 0.07 for k=0.1, while it has almost not changed for lower densification rates.

3.3. Periodic variations of accumulation

One can try to analyse the impact of snow accumulation variability with respect to its frequency. Let us consider the Gaussian noise $G(t')$ as the sum of sinusoidal functions $A\cos(\omega t)$, in which ω is the angular velocity (2π times the frequency) and A an amplitude. For a given wavelength, Eq. (6) now reads:

$$H(t) = \int_{-\infty}^{t} \frac{A\cos(\omega t')}{\rho_i(1 - \alpha\exp(-k(t-t')))}\mathrm{d}t' \tag{9}$$

where α is a function of snow and ice densities:

$$\alpha = \frac{(\rho_i - \rho_s)}{\rho_i} \tag{10}$$

Knowing that the expression $1/(1-\alpha\exp(-k(t-t'))$ can be developed into a power series of $\exp(-k(t-t'))$, Eq. (9) can be written as:

$$H'(t) = A\sum_{n=0}^{\infty}\alpha^n \int_{-\infty}^{t}\cos(\omega t')\exp(-kn(t-t'))\mathrm{d}t' \tag{11}$$

Using the formula:

$$\int_{-\infty}^{t}\cos(\omega t')\exp(-\kappa(t-t'))\mathrm{d}t' = \frac{1}{\kappa^2+\omega^2}(\kappa\cos(\omega t) + \omega\sin(\omega t))$$

We obtain

$$H(t) = \frac{A}{\omega\rho_i}D\cos(\omega t) + \frac{A}{\omega\rho_i}C\sin(\omega t) \tag{12}$$

$$\frac{\mathrm{d}H}{\mathrm{d}t}(t) = \frac{A}{\rho_i}C\cos(\omega t) - \frac{A}{\rho_i}D\sin(\omega t) \tag{13}$$

with

$$C = \sum_{n=0}^{\infty}\frac{\alpha^n}{(k/\omega)^2 n^2 + 1} \quad D = \sum_{n=0}^{\infty}\frac{\alpha^n (k/\omega)n}{(k/\omega)^2 n^2 + 1}$$

Taking into account the densification process in case of a sinusoidal perturbation of amplitude A then leads to elevation changes of the same wavelength, but with an amplitude B and a phase φ such that:

$$B = \frac{A}{\rho_i}\sqrt{C^2 + D^2} \quad \cos\varphi = \frac{C}{\sqrt{C^2 + D^2}}$$

$$\sin\varphi = \frac{D}{\sqrt{C^2 + D^2}} \tag{14}$$

B increases with α and then decreases with the initial snow density ρ_s. Between a snow density of 0.3 and 0.35 g/cm^3, α increases by a factor of 7%.

We can immediately see that B only depends on the ratio k/ω that a high-frequency noise in a fast densification area gives the same impact as a low frequency noise in a slow densification area.

For rapid accumulation variations with respect to the densification (k/ω<<1), the amplitude B is approximately equal to A/ρ_s, and the phase ω, to 0. On the contrary, for slow accumulation variations, with respect to the densification (k/ω>>1), the amplitude B is approximately equal to A/ρ_i, and the phase φ, to 0. One should however note that our simple model is not appropriate for slow accumulation variations because in such cases, the densification rate k would vary with the accumulation rate, with faster densification when accumulation is stronger.

The phase φ is equal to 0 for k/ω>>1 or k/ω<<1 and is maximum for k=ω, where it is equal to Arctan(α/(1$-\alpha$)). For example, for ρ_s=0.35 and ρ_i=0.92, we obtain $\varphi \approx \pi/3$, an upper bound for the phasing.

3.4. Random fluctuations of accumulation

In Fig. 1, we superimposed the ice elevation fluctuations due to random snow accumulation variability from Eq. (6), with the ice elevation expressed in w.e. (or ice mass fluctuation) derived from Eq. (2), for k=0.01 and for the same random fluctuations with a prior initialisation of 1000 years. As evidence, the fluctuations are bigger, the artificial trends are greater, and the short-scale noises are enhanced with respect to those in ice mass fluctuations.

We performed sensitivity tests to investigate the effect of the densification rate value on ice elevation fluctuations (Fig. 3). Most of the sensibility lies in the large-scale pattern. The 10-year scale features are practically the same for all the values considered.

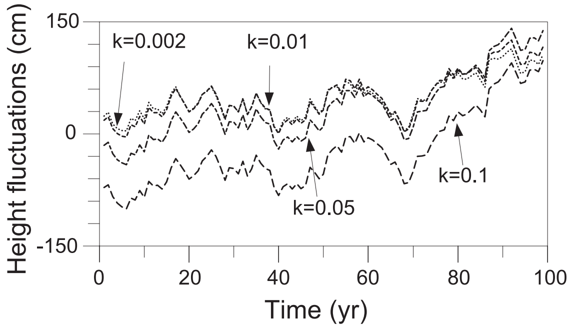

Fig. 3. Elevation fluctuations due to random snow accumulation for different densification rates (see k in Eqs. (4) and (5)). The random fluctuations are the same as for Fig. 1. The process is initialised for 1000 years before the time t=0 of the plot.

Depending on the frequency of the perturbation, the impact of densification is, more or less, important. For low-frequency perturbations, when ω is weak, the impact of the random perturbation is weak on the trend restitution and very dependant on the exact densification rate, with which it decreases. On the contrary, for high-frequency noise, the relative enhancement of the random perturbation is great but is not very dependant on the densification rate. Moreover, for low densification rates (k/ω<<1), the impact of the variability mostly depends of the initial density.

We perform the same statistical tests as described in the previous section, with the same $G(t')$ values. The average trend estimated with a 10 years series for 100,000 runs and for the nominal densification rate (0.01 year^{-1}) is also close to 0, but the RMS is 3.49 cm/year. This means that we have about a 30% chance of having an artificial absolute trend greater than this value that represents 20% of the considered mass balance. As suggested by the sensitivity tests, these values only decrease slightly when the densification rate increases: For the higher densification rate considered here (k=0.1), the RMS of the trend is 2.76.

4. Discussion

It is clear from the previous section that knowledge of ice elevation trends alone may lead to an erroneous interpretation of ice sheet mass balance. In Fig. 4a, we have plotted the histogram of the measured or ice-

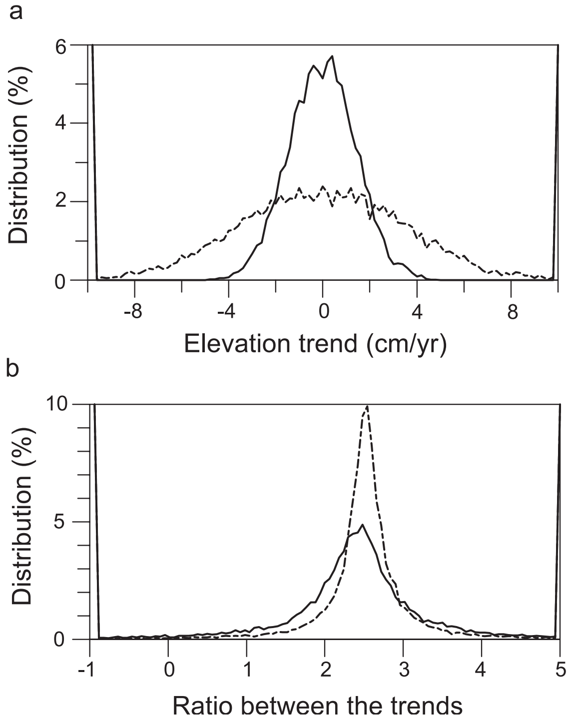

Fig. 4. (a) Histogram of the ice equivalent elevation trend (solid line) and of the measured elevation trend (dotted line, k=0.01 year^{-1}) as estimated after applying a stochastic process. (b) Ratio between the measured and ice equivalent elevation trends for k=0.002 year^{-1} (dotted line) and k=0.01 year^{-1} (solid line).

equivalent elevation trends. As has already been seen in Fig. 1, the distribution of the measured elevation trend is larger than for the ice equivalent elevation trend. In Fig. 4b, we plotted the ratio between these two trends. The mean value is 2.64, and the most likely value is 2.54, i.e., the ratio between the ice and snow densities. For the reference value, the RMS of the ratio between elevation and mass is 0.3; even with a constant densification rate, one can find a ratio between ice equivalent and measured trends of between 2 and 3. In addition, as shown in Fig. 4b, 3–4% of the ratios have negative values, meaning that the sign of the ice equivalent and measured trends can even be opposite. Finally, when the densification time decreases, the histogram becomes narrower around 2.54.

A serious consequence of that elevation–mass difference is that one can attribute the difference between the measured mass and elevation trends to other physical mechanisms, such as postglacial rebound, that acts both on ice elevation and ice mass.

Note that we do not know exactly the dominant wavelengths of the variability. Oerlemans [1] shows that the snow accumulation variability is maximum for data averaged over 30 years. Remy et al. [2] compiled a few accumulation rates derived from ice cores or stations and confirmed this value. If the wavelength of the dominant variability is 30 years (ω=0.2 year^{-1}), then the relative frequency variability is high, everywhere in Antarctica.

One can therefore try to estimate the time during which the past variability still influences the estimation of the elevation trend, or the time period for which knowledge of the past variability is needed to correct for the effect of the snow accumulation variability. Such a study can be performed in a place for which shallow ice core data are available, but such information can only be deduced for the whole continent with the help of satellite observation, which means for a shorter period of time.

Assuming knowledge of the past variability and the densification process or assuming that the past variability is null amounts to the same thing. We thus applied a stochastic process with the same rule as previously (initialisation for a few thousands years, mean accumulation rate of 16.6 cm/year with a variability of 25%) and then stop it ($G(t)$=0) a few years before we estimated the trend. The RMS of the elevation trend is estimated for a few thousand different runs. The evolution of the RMS is shown in Fig. 5 with respect to the time during which the stochastic process is stopped. The RMS falls from the initial value to around 1 cm/year over 5 to 10 years, depending on the densification rate, before reaching a slowly decreasing slope. Note that if the stochastic process is stopped during only a year, the RMS of the

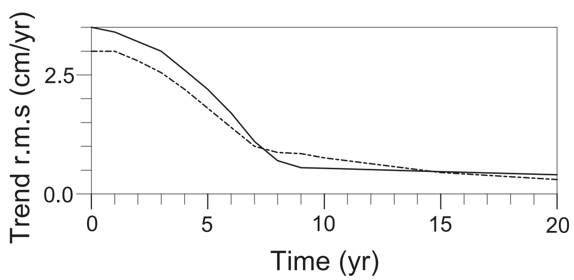

Fig. 5. Evolution of the RMS of the elevation trend with respect to time during which the stochastic process is stopped, for k=0.002 (dotted line) and 0.01 year^{-1} (solid line)

measured trend is not really diminished. If one wants to significantly decrease the uncertainties in the elevation trend estimation, one has to know the past variability for the 5 to 10 last years.

5. Conclusion

The stochastic variations of the accumulation rate above an ice sheet may lead to a significant trend in the ice sheet mass during several decades without any long-term climatic relationship. If one considers a noise of 25% of the mean accumulation rate, one has a 20% chance of finding an artificial absolute elevation trend greater than 10% of the accumulation rate (expressed in ice equivalent).

Moreover, snow accumulation variability affects the estimation of ice sheet mass variation through ice sheet elevation variation. When the densification process is also taken into account, i.e., by observing variations in elevation, the measured trend is enhanced. One now has a 30% chance of having an artificial absolute trend greater than 20% of the accumulation rate. The scatter of the induced trends is such that the sign of the ice equivalent trend (or mass trend) may be opposite to the sign of the measured elevation trend. The impact of this climatic noise increases when the frequency of the variability decreases. For low-frequency perturbation, the impact of the elevation trend is only poorly dependant on the densification rate value.

One needs to know the 'recent past' variability to eliminate the residual error due to snow densification, above all in area of high densification. At least, 10 years are needed to reduce the induced error by a factor of 6. A longer series does not significantly reduce the error for the trend estimation.

Our study, based on a simplified densification model, is valid for short-term variations of accumulation (i.e., shorter than the densification time). For long-term or intermediate variations of accumulation, a future analysis should be done using a more realistic firn densification model.

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