Impact of Local Insolation on Snow Metamorphism and Ice Core Records

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Abstract: Local insolation is a major component of the energy balance at the surface of an ice sheet and causes temperature gradient metamorphism (TGM) of snow and firn. TGM is one of the dominant processes changing the structure of dry snow. We present a physically based model that calculates insolation-induced relative changes in TGM in the past. The results indicate that TGM at Dome Fuji varied by up to a factor of 2 over the past 350ka, and is driven predominately by the precession-band variability in local summer solstice insolation. At Dome Fuji, the impact of glacial-interglacial temperature changes on TGM is almost fully compensated by synchronous, opposite changes in accumulation rate, which determines the exposure time of a snow layer to TGM. Even small remaining temperature signals in TGM can cause phase shifts between TGM and local summer solstice insolation of several ka. This directly affects the accuracy of orbitally tuned ice core time scales using O_2/N_2 or total air content records, as this dating method is based on the assumption of synchronicity between TGM and insolation. It must be assumed that the strong variability in TGM will also be reflected in physical and chemical ice core records by e.g. modulating the volatilization of reversibly deposited species including the stable isotopes of water. Sublimation and thus accumulation rates are also closely linked to TGM, affecting the concentrations also of irreversibly deposited non-volatile impurities. Thus, the effect of a local, post-depositional contribution of TGM on ice core records must be quantified prior to their interpretation in terms of larger scale climate variability in the orbital frequency bands.

Key words: Local insolation, snow, firn, temperature gradient metamorphism, ice core records, Dome Fuji.

1. Introduction

Local insolation, i.e. the amount of solar energy received at the top of the atmosphere per area and time (in units of W m^{-2}) is a key component of the energy balance at the Earth's surface. Even though snow covered areas reflect ~90% of the incoming radiation, solar radiation will directly or indirectly modulate snow temperatures and temperature gradients in the snow and firn (e.g. [1]) from diurnal to annual and orbital time scales. Snow and firn temperatures and corresponding gradients can potentially influence the preservation of most if not all parameters measured in an ice core: Temperature gradients lead to vertical water vapour fluxes and a very efficient metamorphism of the snow called temperature gradient metamorphism (TGM). TGM directly affects a wide range of physical properties of snow, firn and ultimately ice such as e.g. grain size, density, optical, mechanical and thermal properties (e.g. [2-5]). The post-depositional exchange of reactive chemical species (i.e. reversibly deposited species, such as e.g. HCHO, H_2O_2, HNO_3, MSA, HCl) between the air and the snow is modulated by snow temperatures and snow metamorphism [6-8].

Sublimation rates also depend on snow temperature and affect snow accumulation rates (i.e. precipitation – sublimation) leading to an enrichment of irreversibly deposited impurities in ice (e.g. aeolian dust, Ca^{2+}, Na^{+}) and a depletion of volatile impurities [9]. Sublimation also influences the preservation of the stable water isotopes (^{18}O, deuterium) and thus affects the most common ice core paleo-thermometer. In the case of ice core records of photochemically active species, these indirect, post-depositional effects of insolation are superimposed on the direct effect through photochemistry: Gas- and snow phase photochemical reactions both in the atmosphere and at the snow-air interface are by definition directly driven by solar insolation [6] and changes in the latter are thus potentially reflected in corresponding ice core records.

Whether the above-mentioned effects of variations in local insolation are actually relevant for the interpretation of a specific ice core record depends on the scientific question asked. In most cases also the environmental conditions such as temperature and accumulation rate at the drill site need to be taken into account. Thus an assessment on a case-by-case basis is required for a quantitative interpretation of ice core records.

The potential impact of variations of local insolation on TGM has not yet been quantified in a physically stringent way. The presented work is a first step towards closing this gap. Using a physically based model we estimate the impact of climatic changes (temperature and accumulation rate) and local insolation on the "total TGM" (tTGM) of the snowpack over the past 350 ka at Dome Fuji (77ºS, 40ºE, 3810 m.a.s.l.). We define tTGM as the integrated water vapour flux through a snow layer

from its deposition to the time it is buried below the depth where significant temperature gradients exist. The resulting time series of modelled relative changes in tTGM then provides a basis to investigate the impact of changes in local insolation on physical and chemical snow and ice properties.

Dome Fuji conditions have been used in this study due to the availability of the necessary data. The results, however, are, qualitatively, representative of similar low accumulation sites on the East Antarctic plateau such as Dome C (75ºS, 123ºE, 3233 m.a.s.l.) and Vostok (78ºS, 106ºE, 3488 m.a.s.l.).

2. Physical processes

2.1 Local insolation

Local solar insolation not only significantly varies over diurnal and annual cycles, but also on orbital time scales. The orbitally-driven variations in local insolation can be accurately calculated over the past few millions of years [10, 11]. Local summer insolation variance is predominately driven by the 20 ka precession cycle (more than 50% for the seasonal average, ~80% for summer solstice, i.e. December 21st insolation) with additional contributions from the 40 ka obliquity and 100 ka eccentricity cycles [12]. Daily average summer solstice (December 21st) insolation at Dome Fuji varies by as much as 31% (463-607 W m^{-2}) relative to the minimum value over the past 1Ma (Figure 1).

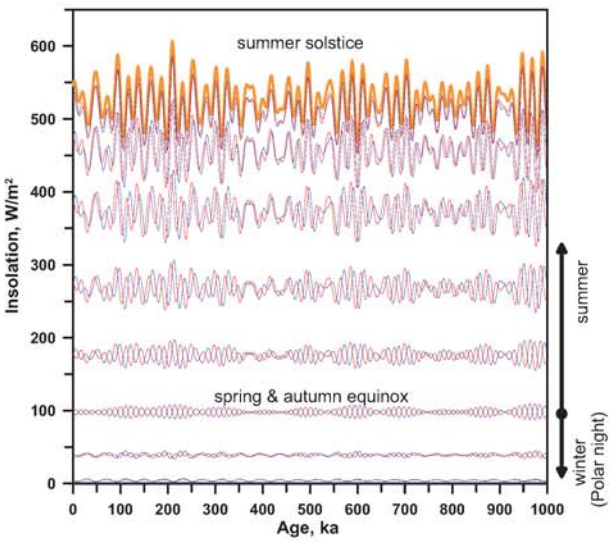


Figure 1: Calculated variations of the local daily average insolation at Dome Fuji over the past 1 Ma [11]. Each curve represents the variation of the local daily average insolation on a specific day of the year. The top, thick orange line is the insolation at 270 degrees from the March equinox and corresponds roughly to the summer solstice. The red curves are the local daily average insolation plotted every 15 degrees true longitude (i.e. every ~15.2 days) thereafter, i.e. moving from summer towards winter. Analogously, the blue curves correspond to the local daily average insolation moving back up in steps of 15 degrees true longitude from Polar winter to summer (note that all the Polar winter curves are invisible because the insolation is zero, see also Figure 2).

Such variations obviously significantly affect the energy balance at the snow surface and thus snow metamorphism. It is important to note that both, precession and obliquity predominantly affect the seasonal distribution of insolation on Earth, but, at least on a global scale, not significantly the annual average insolation [10]. Thus, the higher the summer (spring) insolation, the lower the winter (autumn) insolation at a specific site (e.g. Figure 1, red and blue curves at ~100 W m^{-2} correspond to autumn and spring insolation). At high latitudes, however, the annual mean insolation is co-varying with summer insolation, because the winter insolation cannot drop below zero. It is also interesting to note that not only the seasonal contrast but also the lengths of the seasons vary, as does the date of the maximum summer insolation. For instance, the summer insolation at Dome Fuji peaks 12 days earlier at 215.5 ka compared to 204 ka and December 21st insolation does strictly speaking not represent maximum summer insolation as often assumed (Figure 2).

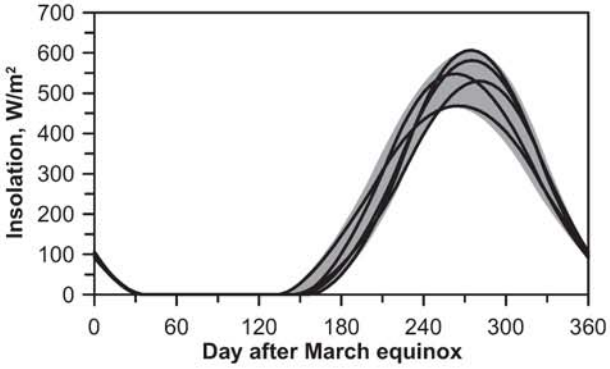


Figure 2: The full range of calculated local daily average insolation in the course of a year at Dome Fuji over the past 1Ma (grey area). Randomly chosen examples of five annual cycles are plotted in black to illustrate the variability in the width and height of the annual insolation cycles (calendar years have been transformed into standard years of 360 model days to be compatible with the model output. One model day corresponds to 1 degree change in the true longitude of the sun and thus 365.24/360 calendar days).

2.2 Air and snow temperatures

Most of the TGM takes place in the top few meters of the firn, i.e. where the temperature gradients are strongest (e.g. [1, 2, 5]). The latter are driven by the diurnal to annual temperature variations and are almost completely attenuated in the top ~0.5 m to ~10 m of the firn respectively [5, 13]. Temperature gradients are also induced by climatic changes and the geothermal heat flux. However, these gradients are much smaller and therefore not considered in this study.

Surface snow temperatures are closely linked to near-surface air temperatures through the exchange of

sensible heat at the interface, although other processes such as radiative heating, and cooling or latent heat exchange also affect surface snow temperatures. However, it is important to note that at Dome Fuji measured surface snow temperatures closely follow air temperatures from the Automatic Weather Station and no obvious (within a few degrees) "solid-state greenhouse" effect was observed (Fig. S1, in electronic supplements to [14]), in accordance with previous observations [15]. We therefore assume that the measured air temperatures are representative of surface snow temperatures and can be used as boundary values to drive the heat transfer in the firn.

We thus determined an empirical relationship between local insolation and air and surface snow temperatures on diurnal and annual time scales. For this we compared calculated insolation values with measured air temperatures at Dome Fuji [16].

2.2.1 Annual temperature cycle

Figure 3 shows the very close linear relationship (r^2=0.97) between calculated daily mean insolation at Dome Fuji under current conditions and the 1994-2001 averages of daily air temperatures. Measured air temperatures lag the insolation by ~6 days, significantly less than the ~30 days observed over the sea or snow free continental areas [12]. This is due to the relatively low heat uptake as a result of the high albedo, low heat conductivity and low heat capacity of snow.

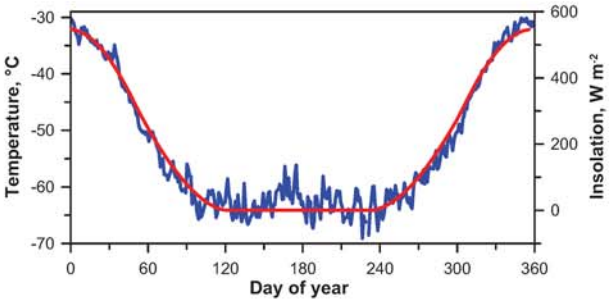


Figure 3: Relationship between calculated local daily insolation (red curve, right axis) and measured averaged daily air temperature [16] (blue, left axis) at Dome Fuji. R^2=0.97 for a temperature lag of 6 days using daily average temperatures from 1994-2001 (calendar years have been transformed into standard years of 360 days to be compatible with the model output).

The observed relationship of air (and thus surface snow) temperature to changes in insolation could now be used to calculate past changes of the annual temperature cycle relative to current values by simply scaling daily mean temperatures with corresponding relative changes in daily insolation throughout the year. However, the caveat of this approach is that it would lead to a significant underestimation of the changes in the actual amplitudes of the annual temperature cycle: Because insolation can not take values below zero, the winter temperatures would be kept at a constant value in this approach. This, however, is not realistic. As mentioned earlier, higher average summer insolation coincides with lower average winter insolation and correspondingly lower average winter temperatures. Average lower southern hemispheric winter temperatures, however, will also translate to the high southern latitudes through meridional heat flux irrespective of the presence of direct sunlight. We thus adjust the temperatures during the dark months to counteract the impact of insolation changes during the sunlit period in order to get realistic amplitudes of the annual temperature cycle. This is done by scaling the annual temperature cycle such that both the annual mean temperature and the calculated summer maximum temperature estimated based on insolation changes remain unchanged, while winter temperatures are adjusted. This annual temperature cycle is then superimposed on the measured annual mean temperature derived from the stable isotope record. With this approach the orbital variability observed in the annual mean temperature [14] is implicitly taken into account in the model. It should be noted, that at Dome Fuji isotope temperatures drop with increasing 65ºS summer solstice insolation, which is attributed to a climatic signal originating in the northern hemisphere [14]. Further, the obliquity signal in the isotope temperature record lags the corresponding annual mean insolation variability by several ka, suggesting that it is not predominately due to local insolation but to a larger scale climatic signal [14].

2.2.2 Diurnal temperature cycle

Superimposed on the annual is the diurnal temperature cycle, which leads to strong temperature gradients in the top ~0.5 m of the snow [5, 13]. The diurnal cycles of air and surface snow temperatures are expected to be closely linked to temporal changes in the irradiation at the surface and thus solar elevation angle. At Dome Fuji the diurnal amplitude of the latter is essentially constant throughout the period of 24-hour sunshine and decreases once the sun starts to set below the horizon (Figure 4a). However, in terms of solar energy available at the surface, the air mass factor has to be considered too: the lower the elevation angle, the longer the path through the atmosphere and accordingly the lower the irradiation (Figure 4b). This leads to the somewhat counterintuitive result that the diurnal amplitudes of direct irradiation are maximal on the day the sun first sets or rises rather than at the summer solstice (Figure 4 and 5).

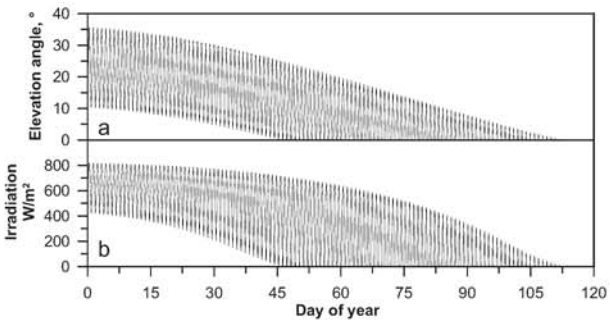


Figure 4: Diurnal amplitudes of (a) calculated solar elevation angles (e.g. [17]) and (b) calculated global solar radiation at the surface considering the zenith-angle dependent air mass factor from summer to polar winter [18].

Diurnal temperature amplitudes extracted from the Dome Fuji AWS data [16] indeed reflect this feature (Figure 5) revealing a close relationship between the calculated diurnal amplitudes of direct irradiation and air temperature (r^2=0.79 for 5 day averages from 1994-2001). Note that the diurnal temperature amplitudes in winter are not zero. This is a result of the observed high-frequency temperature variability during this period (Figure 3), which is in this way accounted for in the model.

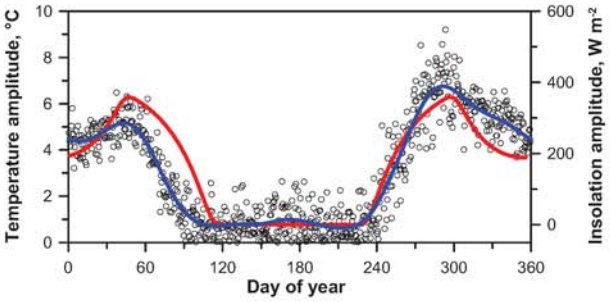


Figure 5: Diurnal amplitudes of measured air temperature [16] (black dots: daily averages over the period 1994 through 2001 based on 3 hourly air temperature data) and calculated local irradiation (red) assuming clear sky conditions at Dome Fuji. The blue line is the spline through the measurements.

Measured temperature amplitudes are higher in austral spring compared to autumn, an asymmetry that is not reproduced in the estimate based on calculated zenith angles. Further, the measured amplitudes drop off earlier in autumn than the calculated ones. The origin of these differences between the calculated irradiation and measured temperature amplitudes may be due to an asymmetry in the humidity of the atmosphere in autumn compared to spring, affecting the radiation and energy balance at the surface. However, whether this is indeed the case remains to be investigated.

2.3 Total Temperature Gradient Metamorphism (tTGM)

The water vapour pressure over ice strongly increases with increasing temperature (Clausius-Clapeyron relationship, e.g. [17]). Thus, temperature gradients in the firn cause gradients in water vapour concentrations in the firn air and thus corresponding fluxes from warmer to colder layers. This redistribution of water from warmer to colder layers is the origin of Temperature Gradient Metamorphism (TGM) [4, 19]. Here we define total Temperature Gradient Metamorphism (tTGM) as the time integrated absolute value of the water vapour flux through an individual snow layer from its deposition on the snow pack surface to its burial below 30 m depth (i.e. to well below the depth where significant temperature gradients are present [5]). By taking the absolute value, the sign of the flux is assumed to be unimportant and we are not considering the net loss or gain of ice in an individual snow layer. Thus, tTGM is a measure of the total water mass that has been recycled (i.e. condensed and re-evaporated) while moving through a specific snow layer driven by temperature gradients.

3. Model description

Conceptually the model is very similar to the one described in [7, 20], but adapted to calculate water vapour fluxes. It is a time-dependent, one-dimensional physically based, finite differences model driven by the boundary values temperature, accumulation rate, and local insolation (Figure 6).

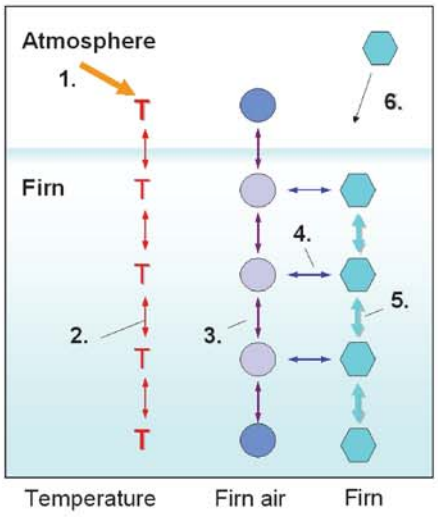


Figure 6: Schematic of total Temperature Gradient Metamorphism (tTGM) model: 1. Local insolation-air temperature; 2. Heat transfer; 3. Water vapour fluxes; 4. Water vapour pressure; 5. Densification model; 6. Accumulation rates

In the current setup, the model calculates the water vapour flux profiles every 500 years, for 2 years, at an effective time resolution of 1.5 hours and 1cm depth

resolution from the surface to 30 m depth using the Crank-Nicolson method [21]. The first year serves as a spin-up to avoid an impact of the initial values on the result and only the second model year is used for calculating tTGM. The latter is done by integrating the total water vapour fluxes of the individual layers from the surface to 30 m depth, while weighting them with their residence time at their depth. The residence time depends on the snow accumulation rate and firn densification. This approach is equivalent to integrating TGM of an individual snow layer from the surface to the time it reaches the depth of 30 m, but is computationally much more efficient. A depth of 30 m was chosen to assure no temperature gradient from the annual cycle was present at the model boundary.

The initial temperature profiles at the beginning of the 2-year model runs are calculated assuming a sinusoidal annual and diurnal temperature cycle. Then, heat transfer is modelled driven by prescribed air temperatures (see below) based on a calculated density profile [22] using annual mean temperature and accumulation rate (both deduced from $\delta^{18}O$ [14, 23, 24]) according to [7, 22] while accounting for corresponding temperature dependent thermal conductivities [5] (see appendix A in [22] for a detailed description of the calculations using the Crank-Nicholson method). Water vapour pressures over ice are calculated according to the Clausius-Clapeyron relation (e.g. [17]). The temperature, pressure and porosity dependent effective diffusion coefficients for water vapour in firn air are calculated according to [25]. The water vapour fluxes resulting from the temperature gradients were then calculated according to Fick's law.

The empirical relationships between calculated local insolation and measured annual and diurnal air temperature amplitudes shown in Figures 3 and 5 were used in the model to determine the relative changes of annual and diurnal air temperature cycles in the past based on calculated changes in local daily insolation.

The resulting estimated annual and diurnal air temperature cycles were then superimposed on the annual mean temperature deduced from the ice core isotope record and used as boundary values driving the temporal evolution of the modelled firn temperature profiles.

4. Results and discussion

The range of estimated annual cycles in air temperatures at Dome Fuji over the past 1 Ma are plotted in Figure 7a and the range of diurnal temperature amplitudes in Figure 7b. Insolation variations change the amplitude of the annual temperature cycle by up to ~4.5˚C (~30%) relative to the minimum value. Similarly, the amplitudes of the diurnal temperature cycle vary by about 1.5˚C (up to 50%). Those values should be considered lower limits because a potential contribution from a 'solid-state greenhouse' effect in the near surface snow would increase the amplitudes of the annual and diurnal surface snow temperature cycle. Although this is expected to be small ([15], see also 2.2.), the radiative heating of the top few millimetres of the snow might lead to a more vigorous metamorphism in this layer and promote a stratification of the snow. The impact of the latter on tTGM has not been considered here and remains to be investigated. However, given that effectively only 500 year averages of tTGM are considered, potential effects of higher-frequency density and morphological variations of the snowpack should largely average out.

Also diffusion of water molecules on the ice surfaces is neglected, as it is expected to be small compared to vapour diffusion in most conditions [26]. Further, the diurnal temperature amplitudes, and thus tTGM, are likely somewhat underestimated because they are based on 3 hour air temperature data. Assuming a sinusoidal diurnal cycle this averaging will affect the amplitudes by less than 10%, and has no significant effect on the relative changes in tTGM. In our approach we assume that the current empirical relationships between insolation and diurnal and annual temperature cycles at Dome Fuji remained unchanged over time, and remain valid specifically also during the past colder glacial periods. This assumption is supported by the fact that annual average temperatures in past glacials were only ~8˚C colder than at present [14, 24]. This is only a fraction of the total temperature range spanning >45˚C from winter minimum to summer maximum noon temperature over which the empirical relationships have been developed. They therefore also cover most of the temperature range encountered during the glacials except for the coldest temperatures in winter. The latter, however, are not significantly contributing to tTGM as discussed in more detail below.

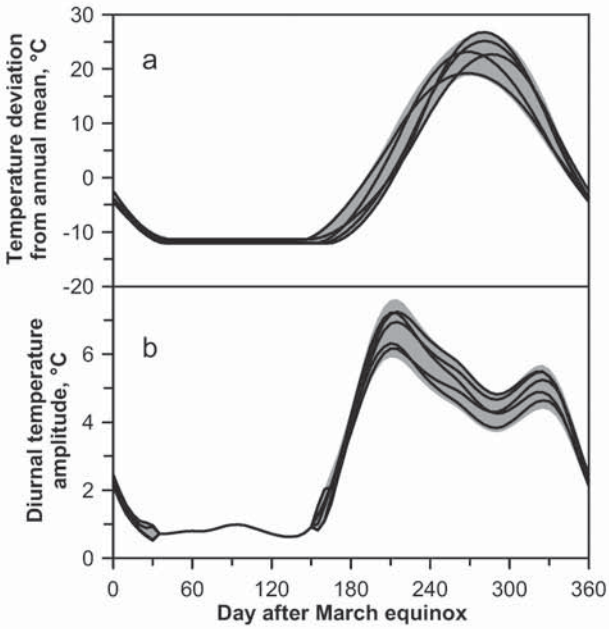


Figure 7: Range of calculated annual temperature cycles over the past 1 Ma at the snow surface (a). Corresponding range of the evolution of the

amplitudes of the diurnal temperature cycle in the course of the year (b). The thick lines are examples of individual curves.

As mentioned earlier, the high frequency diurnal temperature cycle only propagates about 0.5 m into the snowpack while most of the temperature gradients from the annual temperature cycle are attenuated in the top meters of the firn [5]. Thus, TGM is most vigorous near the surface (Figure 8).

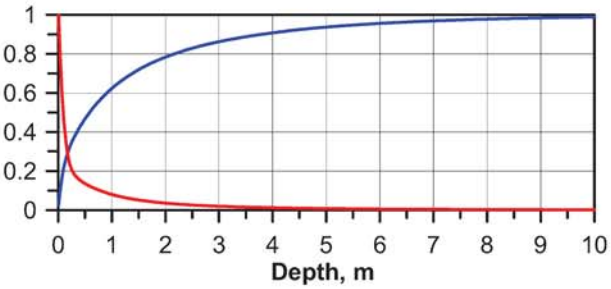


Figure 8: Depth profile of normalized modelled total water vapour fluxes in the course of a year for present-day conditions in red and in blue their corresponding cumulative contribution to tTGM (i.e. the sum of the water vapour fluxes over the full profile, see text).

Nevertheless, it is interesting to note that due to the very low accumulation rate at Dome Fuji on the order of ~30 kg m^{-2} a^{-1} an individual snow layer is exposed to insolation-driven TGM for several decades.

The red curve in Figure 9B depicts the resulting modelled tTGM at Dome Fuji over the past 350 ka.

Figure 9 A: Local daily average December 21st (~summer solstice) insolation at Dome Fuji (orange, left axis) and seasonal average summer insolation, i.e. the average insolation between spring and autumn equinoxes(green, right axis). B: Relative changes in modelled tTGM with accumulation rate from (D) in red and in blue with an accumulation rate reduced by 5 kg m^{-2} a^{-1}. The thin black line is modelled tTGM with the accumulation rate set to a constant, present-day, value. All three time series have been divided by their mean values. C: Isotope temperature [14, 24] and D: accumulation rate [23].

There is a strong correlation (r^2=0.69) between tTGM and December 21st insolation. The correlation between tTGM and seasonal summer average insolation is r^2=0.59, only marginally larger than the correlation between December 21st and seasonal summer average insolation (r^2=0.55). The higher correlation of tTGM with December 21st insolation is the result of the nonlinear dependence of water vapour pressure over ice on temperature combined with the high diurnal temperature amplitudes during that time of year. This results in a much more vigorous TGM around the summer solstice compared to autumn spring or winter, when TGM is minimal, given the colder temperatures in combination with smaller diurnal temperature amplitudes during these periods (Figure 10).

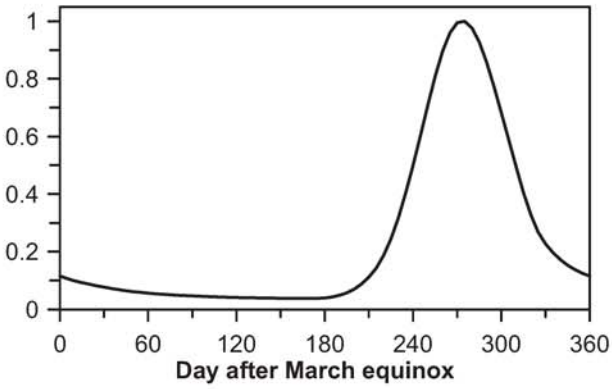
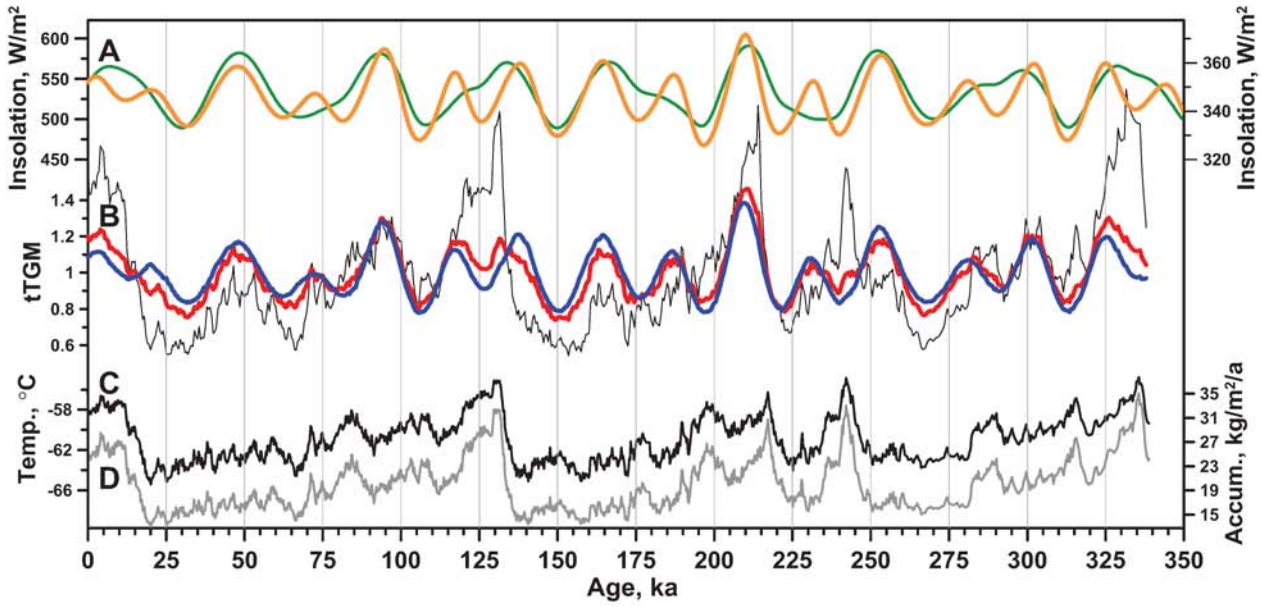


Figure 10: Normalized modelled total water vapour flux in the firn column in the course of the year for present day conditions. The sum of the total water vapour flux over the year corresponds by definition to tTGM (see text). Note that the higher values in autumn compared to spring are due to the summer



heat wave diffusing into deeper depths in the firn inducing higher water vapour fluxes in these layers. Thus the evolution of the total water vapour flux in the firn column in the course of a year does not directly reflect insolation or temperature at the surface (Figures 2 and 7).

Therefore, the tTGM variability is dominated by the mostly precession-driven variability of the local insolation around the summer solstice rather than the more obliquity-driven linear seasonal average summer insolation.

Quite striking is the fact that the impact of the climatic temperature changes between glacial and interglacial periods are almost negligible in tTGM, despite the very strong dependence on temperature. The correlation between tTGM and isotope temperature is only r^2=0.19. The reason for this is the fact that higher temperatures are coincident with higher accumulation rates (Fig. 9). Higher accumulation rates reduce the time a snow layer is exposed to TGM. Thus, the impact of the ~8°C local temperature change between interglacial and glacial periods at Dome Fuji (Figure 8C, [14, 24]) is largely compensated by the corresponding doubling of the accumulation rate causing a correspondingly reduced exposure time. This is visualized by the model run where the accumulation rate was kept constant over time at the present-day value, resulting in a strong temperature signal in tTGM (Figure 9B).

Because the model calculates relative changes of tTGM with respect to a well-calibrated initial state (present-day conditions), it is very insensitive to inherent uncertainties in the model parameters. Thus, varying the absolute values of model parameters within their range of uncertainty does not significantly affect the relative changes in tTGM. For instance there is no significant difference in the result whether the measured or calculated cycle of diurnal temperature amplitudes are used (Figure 5).

However, the model is quite sensitive to changes in the relationship between local temperature and accumulation rate. Both are derived from the stable isotopes of water in the ice core. Their relationship is defined by glaciological models [23, 27] and is, like tTGM, a direct result of the temperature dependence of water vapour pressure over ice i.e. the Clausius-Clapeyron relation (e.g. [28]). Thus, the fact that the effects of changes in temperature and accumulation rate tend to compensate each other originates from this intrinsic physical link. However, the degree to which they cancel each other will vary between sites as a function of the actual local temperature-accumulation relationship. Even on the East Antarctic plateau, there might be significant differences. For instance, at Dome C the current local temperature is ~1.65°C higher while at the same time the average accumulation rate is ~10% lower (sometimes more than 35% during glacial periods) compared to Dome Fuji [23].

A second model run (blue curve in Figure 9B) illustrates the impact of a changed temperature-accumulation rate relationship: the published accumulation rate at Dome Fuji was lowered by a constant amount of 5 kg m^{-2} a^{-1}. This corresponds to an average reduction of the accumulation rate by ~23%, which is within the uncertainty of reconstructed accumulation rates [23, 27] and moves it closer to the values determined for Dome C [23]. In line with the discussion above, in this second run the imprint of climatological temperature changes on tTGM is further reduced, resulting in an almost pure December 21st insolation signal (r^2=0.99 between tTGM and December 21st insolation). This curve is also much smoother because temperature-induced noise is almost absent (r^2=0.002 between tTGM and isotope temperature) and also phase shifts with respect to December 21st insolation are reduced.

5. Conclusions

We estimated the relative changes in temperature gradient snow metamorphism at Dome Fuji over the past 350 ka for two scenarios, one with the published temperature-accumulation rate relation and one with a lowered accumulation rate. In both cases tTGM closely follows local summer insolation but is not identical to e.g. December 21st insolation or a seasonal summer average (Fig. 9). tTGM represents a complex, non-linearly weighted average of the local insolation throughout the year, modulated by compensating effects of temperature and accumulation changes.

For instance, there are significant phase shifts between tTGM and December 21st insolation of up to several ka (clearly visible e.g. during the penultimate termination around 130 ka in Fig. 9). Such phase shifts are directly relevant for the orbital tuning of ice core time scales using O_2/N_2 or total air content records: Both records reveal a strong similarity with local December 21st or seasonal average summer insolation and are thought to be related to TGM [14, 29]. Orbitally tuned ice core time scales have been created under the assumption that O_2/N_2 [14] and total air content [29] are synchronous with local December 21st or a seasonal summer average insolation, respectively. However, the tTGM model results suggest that this assumption is not necessarily correct and the timescales could be improved further by tuning them to a tTGM model output.

The tTGM is estimated to vary by up to 95% relative to the minimum value over the past 350 ka. Relative changes of the local insolation are thus amplified by about a factor of 3 in tTGM. It must be assumed that these changes will be reflected to some extent in physical and chemical ice core records by e.g. modulating the volatilisation of reversibly deposited species including the stable isotopes of water. Sublimation and therefore accumulation rates are also linked to tTGM, and will affect the concentrations of essentially all impurities measured in ice cores including the non-volatile, irreversibly deposited ones [9].

For instance, at Dome Fuji sublimation varies throughout the year while removing on the order of 50% of the precipitation in summer and depositing an additional ~25% of mass during the remainder of the year [30]. Removing 50% of the summer precipitation at Dome Fuji would result in a decrease of the annual mean $\delta^{18}O$ value of the snow by ~1.8‰ corresponding to a ~2.3°C colder isotope temperature [31]. Thus, assuming sublimation scales with tTGM, it would imprint an insolation signature in the Dome Fuji $\delta^{18}O$ record that is anticorrelated to local insolation (and therefore following northern hemisphere insolation) with an amplitude corresponding to ~1.2°C. The impact of the additional 25% of the winter precipitation condensing on the surface is less clear: If its isotopic value reflects near surface temperature, it would lower the isotope temperature by an additional ~0.5°C, resulting in a total impact of insolation variability in the precession and obliquity frequencies on the Dome Fuji $\delta^{18}O$ record of ~2.8°C peak-to-trough, similar in magnitude to the observed variations [14, 32]. These very rough estimates suggest that the sublimation effect might not be negligible. If in situ studies confirm that such a seasonal sublimation bias on isotopic records does exist, then it would be valuable to correct for its effect when interpreting precession-band variations of the Dome Fuji, Dome C, and Vostok isotope records. Such variations have been interpreted as a support for the hypothesis that the northern hemisphere is pacing Antarctic climate [14], although it has also been suggested that they are due to the changing duration of the southern hemisphere summer [32].

Whether the above-mentioned effects of TGM variability on ice core records are indeed significant remains to be assessed in detail case-by-case. Thus, caution is warranted when interpreting apparent imprints of insolation in ice core records in terms of large-scale climatic processes. It first has to be assured that such imprints are not a result of local, post-depositional effects driven by TGM.

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Editor's comments

To better understand the arguments on the relation between the local insolation and the metamorphism of dry snow proposed in this paper, readers are recommended to consider also the following comments from one of reviewers, S. Fujita:

The arguments given in this paper is based on the model developed by the present authors to calculate insolation-induced relative changes in the total vapor flux in the past. Applying this model to the Dome Fuji, Antarctica, the authors claim that the orbitally tuned time scale used for dating the Dome Fuji ice core [Kawamura et al., 2007] may need corrections by up to 4 kyr. This model assumes that the predominant process for the snow metamorphism is the total vapor flux driven by a temperature gradient in snow and firn. However, common assumption in the earlier studies was indeed importance of intense insolation and temperature in addition to temperature gradient, considering an initial imprint from solar radiation at the surface. Earlier studies have shown how summer insolation initially causes rapid densification and hardening at the summit region of polar ice sheets (e.g., Koerner [1971]). It is also known that radiative heating is concentrated in the top several millimeter of the ice sheet in Antarctic snow [Brandt and Warren, 1993]. Thus, initial solar imprinting seems to affect formation of stratification strongly. In addition, some of earlier studies have estimated that molecular diffusion through surface of ice also accelerate the sintering phenomena (e.g., Figure 4 in Maeno and Ebinuma, [1983]) and it is known diffusion process depends on temperature (e.g., Petrenko and Whitworth, [1994]). Therefore, we should evaluate interplay of the initial imprint (i.e., formation of strata) and further metamorphism given to the strata within the firn to better understand the orbital tuning of ice core time scale and underlying physical processes.

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