The WAIS Divide deep ice core WD2014 chronology – Part 1: Methane synchronization (68–31 ka BP) and the gas age–ice age difference


1 College of Earth, Ocean, and Atmospheric Sciences, Oregon State University, Corvallis, OR 97331, USA
2 Department of Geography, University of California, Berkeley, CA 94720, USA
3 Scripps Institution of Oceanography, University of California, San Diego, La Jolla, CA 92093, USA
4 Quaternary Research Center and Department of Earth and Space Sciences, University of Washington, Seattle, WA 98195, USA
5 Department of Geosciences and Earth and Environmental Systems Institute, Pennsylvania State University, University Park, PA 16802, USA
6 US Geological Survey, Boulder, CO 80309, USA
7 Institute of Global Environmental Change, Xi’an Jiaotong University, Xi’an 710049, China
8 Department of Geology and Geophysics, University of Minnesota, Minneapolis, MN 55455, USA
9 Desert Research Institute, Nevada System of Higher Education, Reno, NV 89512, USA

Correspondence to: C. Buizert (buizertc@science.oregonstate.edu)

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Abstract. The West Antarctic Ice Sheet Divide (WAIS Divide, WD) ice core is a newly drilled, high-accumulation deep ice core that provides Antarctic climate records of the past ∼ 68 ka at unprecedented temporal resolution. The upper 2850 m (back to 31.2 ka BP) have been dated using annual-layer counting. Here we present a chronology for the deep part of the core (67.8–31.2 ka BP), which is based on stratigraphic matching to annual-layer-counted Greenland ice cores using globally well-mixed atmospheric methane. We calculate the WD gas age–ice age difference (Δage) using a combination of firn densification modeling, ice-flow modeling, and a data set of δ15N-N₂, a proxy for past firn column thickness. The largest Δage at WD occurs during the Last Glacial Maximum, and is 525 ± 120 years. Internally consistent solutions can be found only when assuming little to no influence of impurity content on densification rates, contrary to a recently proposed hypothesis. We synchronize the WD chronology to a linearly scaled version of the layer-counted Greenland Ice Core Chronology (GICC05), which brings the age of Dansgaard–Oeschger (DO) events into agreement with the U/Th absolutely dated Hulu Cave speleothem record. The small Δage at WD provides valuable opportunities to investigate the timing of atmospheric greenhouse gas variations relative to Antarctic climate, as well as the interhemispheric phasing of the “bipolar seesaw”.

1 Introduction

Deep ice cores from the polar regions provide high-resolution climate records of past atmospheric composition, aerosol loading and polar temperatures (e.g., NGRIP community members, 2004; EPICA Community Members, 2006; Wolff et al., 2006; Ahn and Brook, 2008). Furthermore, the coring itself gives access to the ice sheet interior and bed, allowing investigation of glaciologically important processes such as ice deformation (Gundestrup et al., 1993), folding (NEEM community members, 2013), crystal fabric evolution...
(Gow et al., 1997), and geothermal heat flow (Dahl-Jensen et al., 1998). Having a reliable ice core chronology (i.e., an age–depth relationship) is paramount for the interpretation of the climate records and comparison to marine and terrestrial paleoclimate archives.

The West Antarctic Ice Sheet Divide (WAIS Divide, WD) ice core (79.48°S, 112.11°W; 1766 m above sea level; −30 °C present-day mean annual temperature) was drilled and recovered to 3404 m depth (WAIS Divide Project Members, 2013). Drilling was stopped 50 m above the estimated bedrock depth to prevent contamination of the basal hydrology. Due to high accumulation rates of 22 cm ice a⁻¹ at present and ~10 cm ice a⁻¹ during the Last Glacial Maximum (LGM), the WD core delivers climate records of unprecedented temporal resolution (Steig et al., 2013; Sigl et al., 2013) as well as gas records that are only minimally affected by diffusive smoothing in the firn column (Mischler et al., 2009; Mitchell et al., 2011, 2013; Marcott et al., 2014). The combination of high accumulation rates and basal melting at the WD site results in ice near the bed that is relatively young (~68 ka) compared to cores drilled in central East Antarctica.

In WD, annual layers can be identified reliably for the upper 2850 m of the core, reaching back to 31.2 kaBP (thousands of years before present, with present defined as 1950 CE). Below 2850 m depth an alternative dating strategy is needed. Several methods have been employed previously at other deep ice core sites. First, orbital tuning via ΔO₂/N₂ has been applied successfully to several Antarctic cores (Bender, 2002; Kawamura et al., 2007). However, an age span of only ~3 precessional cycles in WD, in combination with the low signal-to-noise ratio of ΔO₂/N₂ data, makes this technique unsuitable for WD. The uncertainty in the orbital tuning is about one-fourth of a precessional cycle (~5 ka), making it a relatively low-resolution dating tool. Second, in Greenland, ice-flow modeling has been used to extend layer-counted chronologies (e.g., Johnsen et al., 2001; Wolff et al., 2010). This method requires assumptions about past accumulation rates, ice flow, and ice sheet elevation. Particularly for the oldest WD ice, the resulting uncertainty would be substantial. Third, several radiometric techniques have been proposed to date ancient ice. Radiocarbon (¹⁴C) dating of atmospheric CO₂ trapped in the ice is unsuitable as it suffers from in situ cosmogetic production in the firn (Lal et al., 1990), and the oldest WD ice dates beyond the reach of ¹⁴C dating. Other (absolute radiometric) dating techniques, such as recoil ²³⁴U dating (Aciego et al., 2011), ⁸¹Kr dating (Buizert et al., 2014a), or atmospheric ⁴⁰Ar buildup (Bender et al., 2008), currently suffer from uncertainties that are too large (~20 ka) to make them applicable at WD.

Instead, at WD we use stratigraphic matching to well-dated Greenland ice cores using globally well-mixed atmospheric methane (CH₄) mixing ratios (Blunier et al., 1998; Blunier and Brook, 2001; Blunier et al., 2007; Petrenko et al., 2006; EPICA Community Members, 2006; Capron et al., 2010). This method is particularly suited to WD because of the small gas age–ice age difference (ΔAge, Sect. 3) and the high-resolution, high-precision CH₄ record available (Sect. 2.1). The method has three main sources of uncertainty: (i) the age uncertainty in the records one synchronizes to, (ii) ΔAge of the ice core being dated, and (iii) the interpolation scheme used in between the CH₄ tie points. We present several improvements over previous work that reduce and quantify these uncertainties: (i) we combine the layer-counted Greenland Ice Core Chronology (GICC05) and a recently refined version of the U / Th-dated Hulu speleothem record (Edwards et al., 2015; Reimer et al., 2013; Southon et al., 2012) to obtain a more accurate estimate of the (absolute) ages of abrupt Dansgaard–Oeschger (DO) events (Sect. 4.4); (ii) we combine firn densification modeling, ice-flow modeling, a new WD δ¹⁵N-N₂ data set that spans the entire core, and a Monte Carlo sensitivity study to obtain a reliable Δage estimate (Sect. 3); and (iii) we compare four different interpolation schemes to obtain an objective estimate of the interpolation uncertainty (Sect. 4.5).

This work is the first part in a series of two papers describing the WD2014 chronology for the WD core in detail. The second part describes the development of the annual layer count from both multi-parameter chemistry and electrical conductivity measurements. The WD2014 chronology is currently the recommended gas and ice timescale for the WD deep core, and as such it supersedes the previously published WDC06A-7 chronology (WAIS Divide Project Members, 2013).

2 Methods

2.1 Data description

Measurements of water stable isotopes. Water isotopic composition (δ¹⁸O and δD = δ²H) was measured at IsoLab, University of Washington. Procedures for the deep section of the core are identical to those used for the upper part of the core reported in WAIS Divide Project Members (2013) and Steig et al. (2013). Measurements were made at 0.25 to 0.5 m depth resolution using laser spectroscopy (Picarro L2120-i water isotope analyzer), and normalized to VSMOW-SLAP (Vienna Standard Mean Ocean Water – Standard Light Antarctic Precipitation). The precision of the measurements is better than 0.1 and 0.8 ‰ for δ¹⁸O and δD, respectively.

Measurements of CH₄. Two CH₄ data sets were used for WD. The first is from discrete ice samples, and was measured jointly at Pennsylvania State University (0–68 ka, 0.5–2 m resolution) and Oregon State University (11.4–24.8 ka, 1–2 m resolution). Air was extracted from ~50 g ice samples using a melt–refreeze technique, and analyzed on a standard gas chromatograph equipped with a flame-ionization detector. Corrections for solubility, blank size and gravitational enrichment are applied (Mitchell et al., 2011; WAIS Divide
Project Members, 2013). The second data set is a continuous CH$_4$ record measured by coupling a laser spectrometer to a continuous flow analysis setup (Stowasser et al., 2012; Rhodes et al., 2013; Chappellaz et al., 2013), and was measured jointly by Oregon State University and the Desert Research Institute (Rhodes et al., 2015). The continuous data set is used to identify the abrupt DO transitions, as it provides better temporal resolution and analytical precision. Both records are reported on the NOAA04 scale (Dlugokencky et al., 2005). Analytical precision in the CH$_4$ data (2σ pooled standard deviation) is around 3.2 and 14 ppb for the discrete data from Oregon State University and Pennsylvania State University, respectively, and 3 to 8 ppb for the continuous CH$_4$ data, depending on the analyzer used (Rhodes et al., 2015): the 14 ppb stated for the PSU discrete data may be an overestimation, as depth-adjacent (rather than true replicate) samples were used in the analysis.

Measurements of $\delta^{15}$N. Atmospheric N$_2$ isotopic composition ($\delta^{15}$N) was measured at Scripps Institution of Oceanography, University of California. Air was extracted from ~12 gram ice samples using a melt–refreeze technique, and collected in stainless steel tubes at liquid-He temperature. $\delta^{15}$N was analyzed using conventional dual-inlet isotope ratio mass spectrometry (IRMS) on a Thermo Finnigan Delta V mass spectrometer. Results are normalized to La Jolla (California, USA) air, and routine analytical corrections are applied (Sowers et al., 1989; Petrenko et al., 2006; Severinghaus et al., 2009). Duplicates were not run for most $\delta^{15}$N data in this study, but the pooled standard deviations of Holocene WD $\delta^{15}$N data sets with duplicate analyses are 0.003‰ (Orsi, 2013). We conservatively adopt an analytical uncertainty of 0.005‰ for this data set to allow for other sources of error.

Measurements of [Ca]. Ca concentrations in the ice were measured at the Ultra Trace Chemistry Laboratory at the Desert Research Institute via continuous flow analysis. Longitudinal samples of ice (approximately 100 cm × 3.3 cm × 3.3 cm) were melted continuously on a melter head that divides the meltwater into three parallel streams. Elemental measurements were made on meltwater from the innermost part of the core with ultra-pure nitric acid added to the melt stream immediately after the melter head; potentially contaminated water from the outer part of the ice is discarded. Elemental analysis of the innermost meltwater stream is performed in parallel on two inductively coupled plasma mass spectrometers (ICPMS), each measuring a different set of elements; some elements were analyzed on both. The dual ICPMS setup allows for measurement of a broad range of 30 elements and data quality control (McConnell et al., 2002, 2007). Precision of the Ca measurements in WD glacial ice is estimated to be ±3‰, with a lower detection limit of 0.15 ng g$^{-1}$. Continuous Ca and CH$_4$ measurements are done on the same ice, and are exactly co-registered in depth.

2.2 Firn densification model description

Air exchange with the overlying atmosphere keeps the interstitial air in the porous firn layer younger than the surrounding ice matrix, resulting in an age difference between polar ice and the gas bubbles it contains, commonly referred to as Δage (Schwander and Stauffer, 1984). Here we use a coupled firn–densification–heat–diffusion model to calculate Δage back in time (Barnola et al., 1991; Goujon et al., 2003; Schwander et al., 1997; Rasmussen et al., 2013), constrained by measurements of $\delta^{15}$N of N$_2$, a proxy for past firn column thickness (Sowers et al., 1992). The model is based on a dynamical description of the Herron–Langway model formulated in terms of overburden load (Herron and Langway, 1980), which is solved in a Lagrangian reference frame. This model has been applied previously to the Greenland NEEM, NGRIP, and GISP2 cores (Rasmussen et al., 2013; Seierstad et al., 2015; Buizert et al., 2014b), where it gives a good agreement to the Goujon densification model (Rasmussen et al., 2013; Goujon et al., 2003). The model allows for the inclusion of softening of firn in response to impurity loading (Horhold et al., 2012), following the mathematical description of Freitag et al. (2013a). The equations governing the model densification rates are given in Appendix A.

The model uses a 2-year time step and 0.5 m depth resolution down to 1000 m, the lower model boundary. A thick model domain is needed because of the long thermal memory of the ice sheet. At WD, downward advection of cold surface ice is strong due to the relatively high accumulation rates, and the geothermal gradient does not penetrate the firn column (Cuffey and Paterson, 2010). We further use a lock-in density that equals the mean close-off density (Martinierie et al., 1994) minus 17.5 kg m$^{-3}$ (as in Blunier and Schwander, 2000) and an empirical parameterization of lock-in gas age based on firm air measurements from 10 sites (Buizert et al., 2012, 2013).

We furthermore use the steady-state version of the Herron–Langway model (Herron and Langway, 1980) in performing sensitivity studies (Sect. 3.2) and the dynamical Arnaud model (Arnaud et al., 2000; Goujon et al., 2003) to validate our Δage solution.

2.3 Temperature reconstruction and ice-flow model

Our temperature reconstruction (Fig. 1a) is based on water ΔD, a proxy for local vapor condensation temperature, calibrated using a measured borehole temperature profile (following Cuffey et al., 1995; Cuffey and Clow, 1997) and, for the last 31.2 ka, adjusted iteratively to satisfy constraints on firm thickness provided by $\delta^{15}$N and by the observed layer thickness λ(z). Using $\delta^{18}$O rather than ΔD in the temperature reconstruction leads to differences that are negligibly small. This borehole temperature calibration approach is possible at WD because the large ice thickness and relatively high accumulation rates help to preserve a memory of past
temperatures in the ice sheet. A coupled 1-D ice-flow–heat-diffusion model converts surface $T(t)$ into a depth profile for comparison to measured borehole temperatures. The 1-D ice-flow model calculates the vertical ice motion, taking into account the surface snow accumulation, the variation in density with depth, and a prescribed history of ice thickness. Vertical motion is calculated by integrating a depth profile of strain rate and adding a rate of basal melt. As in the model of Dansgaard and Johnsen (1969), the strain rate maintains a uniform value between the surface and a depth equal to 80% of the ice thickness, and then varies linearly to some value at the base of the ice. This basal value is defined by the “basal stretching parameter” $f_b$, the ratio of strain rate at the base to strain rate in the upper 80% of the ice column. The basal ice is melting, so part of the ice motion likely occurs as sliding. The along-flow gradient in such sliding is unknown and thus so too is the parameter $f_b$. We overcome this problem by making both the current ice thickness and the basal melt rate free parameters when optimizing models with respect to measured borehole temperatures. Because the basal melt rate and $f_b$ affect the vertical velocities in similar fashion, the optimization constrains a combination of melt rate and $f_b$ that is tightly constrained by the measured temperatures. Thus we find that varying $f_b$ through a large range, from 0.1 to 1.5, changes the reconstructed LGM temperature by less than 0.2°C. Effects of the prescribed ice-thickness history are likewise minor; assuming a 150 m thickness increase from the LGM to 15 ka changes the reconstructed LGM temperature by less than 0.2°C compared to a constant thickness. Note that the 1-D flow model used here is simpler than the one used by Cuffey and Clow (1997) in that it does not attempt to calculate changes in the shape of the strain rate profile; the unknown basal sliding motion at the WD site negates the usefulness of such an exercise.

One output of the 1-D flow model is the strain history of ice layers as a function of depth and time. The cumulative

Figure 1. Modeling $\Delta$age for WAIS Divide. (a) Past temperatures reconstructed from water $\delta$D, calibrated to the borehole temperature profile. (b) Past accumulation rates as reconstructed by the firn densification inverse model (red), and from the annual-layer count (black). (c) $\delta^{15}$N data (black dots) with densification model output (green). (d) $\Delta$age calculated using the densification model (orange) and using the Parrenin $\Delta$depth method (black) with constant 4 m thick convective zone and no correction for thermal $\delta^{15}$N fractionation. (e) Modeled thinning function from ice-flow model (solid), and a simple Nye strain model for comparison (dashed); the Nye thinning function, which has a uniform strain rate as a function of depth, is given as $f_\lambda(z) = (H - z)/H$ with $H$ the ice sheet thickness (Cuffey and Paterson, 2010, p. 616).
strain is represented by the thinning function $f_b(z)$ (Cuffey and Paterson, 2010), the ratio of annual-layer thickness at depth in the ice sheet to its original ice-equivalent thickness at the surface when deposited. The modeled thinning function is shown in Fig. 1e (solid line). In the deep part of the ice sheet, $f_b(z)$ becomes increasingly uncertain as the unknown basal melt rate and $f_b$ become the dominant controls. Here we optimize the model by comparing accumulation rates derived from $f_b(z)$ with those implied by a firm densification model and the measured $\delta^{15}$N of N$_2$ (Sect. 3.1). While this has little effect on the temperature history reconstruction, it provides an important constraint on calculated basal melt rate, an interesting quantity for ice dynamics studies. Our analysis of basal melt rates and further details of the temperature optimization process and 1-D flow modeling will be presented elsewhere.

3 The gas age–ice age difference ($\Delta$age)

3.1 The WD2014 $\Delta$age reconstruction

The firm densification forward model uses past surface temperature $T(t)$ and accumulation $A(t)$ as model forcings, and provides $\Delta$age($t$) and $\delta^{15}$N($t$) as model output.

For the past 31.2 ka, WD has an annual-layer-counted chronology; for this period the annual-layer thickness $\lambda(z)$ provides a constraint on past accumulation rates via $\lambda(z) = A(z) \times f_b(z)$. WD accumulation reconstructed from $\lambda(z)$ is plotted in black in Fig. 1b.

Prior to 31.2 ka we have no such constraint on $A(t)$, and an alternative approach is needed. We use the densification model as an inverse model, where we ask the model to find the $A(t)$ history that minimizes the root-mean-square (rms) deviation between measured and modeled $\delta^{15}$N, given the $T(t)$ forcing. The $\delta^{15}$N data and model fit are shown in Fig. 1c, the $A(t)$ history that optimizes the $\delta^{15}$N fit is shown in Fig. 1b (red), and the modeled $\Delta$age is shown in Fig. 1c (orange). The optimal $A(t)$ history is estimated in two steps. First, we make an initial estimate $A_{\text{init}}(t)$ for the past accumulation history. Second, we adjust the $A(t)$ forcing by applying a smooth perturbation $\xi(t)$ such that $A(t) = [1 + \xi(t)] \times A_{\text{init}}(t)$; an automated algorithm is used to find the curve $\xi(t)$ that optimizes the model fit to the $\delta^{15}$N data. For the last 31.2 ka we obtain a good agreement between $A$ obtained from $\lambda(z)$ and the modeled $f_b(z)$ (Fig. 1b, black) and $A$ obtained from the inverse method (red). The solution we present here is therefore fully internally consistent, i.e., the $A$ and $T$ histories used in the firm densification modeling are the same as those used in the ice-flow modeling, and they provide a good fit to both the $\delta^{15}$N data and borehole temperature data. WD does not suffer from the $\delta^{15}$N model–data mismatch that is commonly observed for East Antarctic cores during the glacial period (Landais et al., 2006; Capron et al., 2013).

We base our $A_{\text{init}}$ values on $\lambda(z)$ for the past 31.2 ka; prior to that we use the common assumption that $A$ follows $\delta^{18}$O (i.e., Clausius–Clapeyron scaling); the fit to the $\delta^{15}$N data is optimized for $A = 24.2 \times \exp[0.1263 \times \delta^{18}$O]. To test the validity of the Clausius–Clapeyron assumption, we additionally run the scenario $A_{\text{init}}(t) = 0.22$ m a$^{-1}$ (i.e., constant accumulation at present-day level). The $A(t)$ and $\Delta$age reconstructed under both $A_{\text{init}}$ scenarios are similar at multi-millennial timescales (Fig. 2). In the layer-counted interval (< 31.2 ka BP), $A$ obtained from $\lambda(z)$ and $\delta^{18}$O is significantly coherent at all timescales longer than 3000 years, but not at higher frequencies. This is equivalent to the variability resolved in the $A_{\text{init}}(t) = 0.22$ m a$^{-1}$ scenario above. We conclude that the WD $\delta^{15}$N data support the idea that $A$ follows $\delta^{18}$O on multi-millennial timescales. However, there may not be a strong relationship at timescales less than a few thousand years, as is clear from the abrupt $A$ increase around 12 ka seen in $\lambda(z)$ that is not reflected in $\delta^{18}$O (Fig. 1a and b).
consistency between the upper and deeper part of the core we use the \( \Delta \text{age} \) values obtained with the inverse densification model for the entire core.

Recently, another \( \delta^{15} \text{N} \)-based approach has been suggested that uses \( \Delta \text{depth} \), rather than \( \Delta \text{age} \), in reconstructing gas chronologies (Parrenin et al., 2012). This method removes the dependence on \( T(t) \) and replaces this with a dependence on the thinning function \( f_s(z) \). Note that this method is very successful in the upper part of an ice core, where \( f_s(z) \) is well constrained, but not very reliable near the base, where \( f_s(z) \) is highly uncertain. Therefore, the firm densification modeling approach should be considered to be more reliable at WD during marine isotope stages (MIS) 2 through 4. Results from the \( \Delta \text{depth} \) method are plotted in black in Fig. 1c, and generally show good agreement with the firm modeling approach. A notable exception is the 60–65 ka interval, where the \( \Delta \text{depth} \) method overestimates the \( \Delta \text{age} \) due to the fact that we have to compress \( \lambda(z) \) strongly in order to fit age constraints derived from DO 18 (Sect. 4.5).

Last, we want to point out that the \( \delta^{15} \text{N} \) data support an early warming at WD, as reported recently (WAIS Divide Project Members, 2013). WD \( \delta^{15} \text{N} \) starts to decrease around 20.5 ka BP, suggesting a thinning of the firn column. The \( \lambda(z) \) (as derived from the layer count) shows that accumulation did not change until 18 ka BP, at which point it started to increase (which would act to increase the firm thickness). The most plausible explanation for the \( \delta^{15} \text{N} \) decrease around 20.5 ka BP is therefore an early onset of West Antarctic deglacial warming, in agreement with increasing impurity content. In this section we evaluate the sensitivity of the model output to all of these parameters. We performed 1000 model runs in which the model parameters were randomly perturbed. The spread in \( \Delta \text{age} \) model results is used to calculate the WD2014 age uncertainty.

### 3.2 \( \Delta \text{age} \) sensitivity study

Besides \( A \) and \( T \) there are several model parameters that have the potential to influence the model outcome; these are the convective zone (CZ) thickness (Sowers et al., 1992; Kawamura et al., 2006), surface density \( (\rho_0) \), and sensitivity to ice impurity content. In this section we evaluate the sensitivity of the model output to all of these parameters. We performed 1000 model runs in which the model parameters were randomly perturbed. The spread in \( \Delta \text{age} \) model results is used to calculate the WD2014 age uncertainty.

**Convective zone thickness.** In the WD2014 model run (Sect. 3.1) we use a constant 3.5 m CZ, corresponding to the present-day situation (Battle et al., 2011). In the sensitivity study we vary the CZ by one of two methods: (1) we let the CZ be constant in time; its thickness is set by drawing from a Gaussian distribution with 3.5 m mean and 3.5 m 2\( \sigma \) width (i.e., 95 % probability of drawing a value in the 0–7 m range). (2) We let the CZ be a function of accumulation rate (Dreyfus et al., 2010), \( \text{CZ} = 3.5 + k \times (A - 0.22) \); we draw \( k \) from a Gaussian distribution with mean of \(-10\) and a 2\( \sigma \) width of 40 (at an LGM \( A \) of 10 cm a\(^{-1}\) this gives a CZ of 0–10 m thickness). In both methods, whenever CZ values are selected that are smaller than 0 m, the CZ thickness is set to 0 m instead. For each of the 1000 model runs in the sensitivity study we randomly selected either of the two methods.

**Surface density.** In the WD2014 model run we use past surface densities \( (\rho_0) \) as given by the parameterization of Kaspers et al. (2004). In the sensitivity study we add a constant offset to the Kaspers values, the magnitude of which is drawn from a Gaussian distribution of zero mean and a 2\( \sigma \) width of 60 kg m\(^{-3}\). This range corresponds to the full range of observed \( \rho_0 \) variability in Kaspers et al. (2004).

**Past temperatures.** Model temperature forcing is constrained by \( \delta^D \) and measured borehole temperatures. There is, however, a range to the solutions allowed by the borehole temperature and ice-flow model; here we use the upper and lower extremes of this range, determined by Monte Carlo analysis using uncertainties of input variables. The scenarios were chosen to provide the maximum \( T \) range for the glacial period rather than for the Holocene, because we are interested in the uncertainty in the methane synchronization (68–31.2 ka BP). In the sensitivity study we use \( T(t) = T_{\text{optimal}}(t) + k \times \Delta T(t) \), where \( T_{\text{optimal}} \) is the forcing used in the WD2014 model run (Fig. 1a), \( \Delta T(t) \) is half the difference between the maximum-\( T \) and minimum-\( T \) scenarios, and \( k \) is drawn from a Gaussian distribution of zero mean and unit 2\( \sigma \) width (giving 95 % probability that \( T(t) \) is within the extreme range identified from the borehole, Fig. 3a).

\( \delta^{15} \text{N} \) **uncertainty.** We conservatively adopt an analytical uncertainty of 0.005 ‰ for this data set; in addition, the interpretation of \( \delta^{15} \text{N} \) in terms of firm thickness is subject to further uncertainty due to irregular firm layering and the stochastic nature of bubble trapping, as was observed for other atmospheric gases such as CH\(_4\) (Etheridge et al., 1992; Rhodes et al., 2013). For each run of the sensitivity study, we therefore perturb each of the individual \( \delta^{15} \text{N} \) data points by adding an offset that is drawn from a Gaussian distribution of zero mean and a 2\( \sigma \) width of 0.015 ‰.

**Impurity-enhanced densification.** Following recent work we include the possibility that increased glacial impurity loading could have enhanced densification rates (Horhold et al., 2012; Freitag et al., 2013a). We use the mathematical formulation of Freitag et al. (2013a), in which the activation energy of the sintering process is a function of the Ca concentration in the firn. The value of \( \beta \), the sensitivity to Ca, is drawn from a Gaussian distribution with 0.0015 mean and a 2\( \sigma \) width of 0.0015. The topic of impurity-enhanced densification is discussed in detail in Sect. 3.3.

The **A** and \( \Delta \text{age} \) scenarios found in the sensitivity study are shown in Fig. 3b and c, respectively. The shaded areas in Fig. 3b and c give the total range of solutions, as well as the \( \pm 2 \sigma \) and \( \pm 1 \sigma \) confidence intervals. Note that the total range of solutions will depend on the number of model runs (here 1000) but that the position of the \( \pm 2 \sigma \) and \( \pm 1 \sigma \) envelopes will not. To investigate the distribution of values, we include
Figure 3. Δage sensitivity study. Shades of blue give the confidence intervals as marked; the black curves represent the values used in the WD2014 chronology; the red curve gives an alternative Δage solution using the Arnaud densification model. (a) Temperature forcing of the densification model. (b) Reconstructed accumulation rates. (c) Reconstructed Δage; note the reversed scale. Histograms of Δage distribution are shown for (d) 60 kaBP, (e) 40 kaBP, and (f) 20 kaBP. Distribution mean and 2σ uncertainty bound is stated in each panel.

Histograms of Δage at 20kyr intervals (Fig. 3d–f). Based on the sensitivity study, we estimate the WD Δage to be 521 ± 120 years (2σ uncertainty) at the LGM (∼20 kaBP). The Δage value of 351 ± 73 years at 40 kaBP gives a representative Δage for MIS 3; Δage at 60 kaBP is 262 ± 50 years.

Additionally, we have repeated our Δage reconstruction using the firn densification physics described by Arnaud et al. (2000) rather than the Herron–Langway description used so far; the Arnaud model provides the physical basis for the commonly used firn densification model of Goujon et al. (2003). More details on the implementation of the Arnaud model are given in Appendix A. Δage found using the Arnaud model is plotted in red in Fig. 3c. Averaged over the entire core, Δage found with the Arnaud model is 19 years (about 7%) smaller than Δage from the Herron–Langway model. The root-mean-square (rms) difference between both solutions is 35 years, corresponding to 0.63 times the 2σ uncertainty found in the sensitivity study. Both solutions are thus found to be in good agreement. The Herron–Langway approach is preferred because the internally consistent solution of temperature, accumulation, and ice flow associated with it is in better agreement with borehole temperature data than the solutions associated with the Arnaud model. Furthermore, the Herron–Langway model is more successful in simulating the magnitude of the δ15N response to the accumulation anomaly at 12 ka (not shown), suggesting it has a more realistic sensitivity to accumulation variability.
3.3 Impurity softening of firn?

Recent work suggests a link between densification rates and impurity content (for which [Ca$^{2+}$] is used as a proxy) in polar firn (Horhold et al., 2012; Freitag et al., 2013a). Here we measured total [Ca] by ICP-MS, but at WD nearly all Ca is in the form of Ca$^{2+}$. The influence of the impurity sensitivity ($\beta$) (see Eq. A6 in the Appendix) on $\Delta$age at WD is shown in Fig. 4. The sensitivity recommended by Freitag et al. (2013a) from investigating present-day firn packs is $\beta = 1 \times 10^{-2}$. We reconstructed $A$ and $\Delta$age with the firn densification inverse model using five values of $\beta$ ranging from $\beta = 0$ (red) to $\beta = 1 \times 10^{-2}$ (blue) in steps of $2.5 \times 10^{-3}$ (shades of deep purple). Black curves give $A$ and $\Delta$age from ice-flow modeling and $\lambda(z)$.

Figure 4. Impurity enhancement of densification rates at WD. Densification modeling results for (a) accumulation rates and (b) $\Delta$age. We use Ca sensitivities $\beta = 0$ (red) through $\beta = 1 \times 10^{-2}$ (blue), in steps of $2.5 \times 10^{-3}$ (shades of deep purple). Black curves give $A$ and $\Delta$age from ice-flow modeling and $\lambda(z)$.

4 Constructing the WAIS Divide WD2014 chronology

4.1 Annual layer count (0–31.2 ka)

A first layer-counted chronology for the upper 2800 m of the WD core based on electrical conductivity measurements (ECM), named WDC06A-7, was presented by WAIS Divide Project Members (2013). The WAIS chronology presented...
in this work, WD2014, uses an updated layer count for the upper 2850 m, based on new data and analyses that have become available since publication of WDC06A-7. These updates are as follows:

1. a reassessment of the dating in the upper 577 m (2.4 ka) using high-resolution multi-parameter chemistry data in combination with automated layer detection algorithms (Winstrup et al., 2012);
2. a reassessment of the dating between 577 and 2300 m (2.4–15.3 ka) using high-resolution multi-parameter chemistry data in combination with ECM;
3. a reassessment of the dating between 2300 and 2800 m (15.3–29.5 ka) using ECM and dust particle measurements, with the ECM having increasing importance with depth;
4. an extension of the annual-layer dating between 2800 and 2850 m (29.5–31.2 ka) using ECM.

Details on the updated WD layer count and the layer counting methodology are presented in part 2 of the WD2014 papers.

4.2 Methane synchronization (31.2–68 ka)

For the deep part of the core where an annual-layer count is not available, we date WD by synchronization to well-dated Northern Hemisphere (NH) climate records of abrupt DO variability using the WD record of globally well-mixed CH$_4$ (Fig. 5). This process consists of several steps:

1. Determine the midpoint of the abrupt DO transitions in WD CH$_4$, NGRIP $\delta^{18}$O, and Hulu speleothem $\delta^{18}$O.
2. Assign a gas age to the WD CH$_4$ tie points (i.e., the DO transitions).
3. Apply the WD $\Delta$age (Sect. 3) to find the corresponding ice age at the depth of the CH$_4$ tie points.
4. Interpolate between the ice age constraints to find the WD depth–age relationship.
5. Redo the $\Delta$age calculations on the new ice age scale.
6. Repeat steps 3–5 iteratively until the depth–age relationship is stable within 1 year. At WD this happened after three iterations. These steps are described in more detail in the following sections.

4.3 Establishing the midpoint in abrupt DO transitions

The procedure for determining the midpoint of the abrupt DO warming transitions is depicted in Fig. 6. For each of

Figure 5. Records of abrupt DO climate variability, (a) revised Hulu Cave speleothem $\delta^{18}$O record on Hulu chronology with U/Th ages above the time series (red dots), (b) NGRIP ice core $\delta^{18}$O on 1.0063 × GICC05 chronology, and (c) WD CH$_4$ on WD2014 (discrete data). DO numbering is given in the bottom of the figure following Rasmussen et al. (2014). White dots denote the midpoints of the stadial–interstadial transitions; the orange vertical lines show the timing of the NGRIP tie points (on 1.0063 × GICC05). For DO 3, 4, and 5.1 the WD2014 chronology is based on annual-layer counting, and minor timing differences exist between WD and NGRIP.
The midpoints of abrupt interstadial terminations were determined in the same fashion (WD CH4 and NGRIP only). Tables 1 and 2 give the results for NH warming and NH cooling, respectively.

4.4 Synchronizing WD to a NGRIP–Hulu hybrid chronology

Abrupt DO variability is expressed clearly in a great number of NH climate records (Voelker, 2002). For the purpose of methane synchronization, our interest is in high-resolution records that express the abrupt DO events very clearly, and are furthermore exceptionally well dated. We here use a combination of two such NH records (Fig. 5), namely the Greenland NGRIP δ18O record (NGRIP community members, 2004), and a refined version of the Hulu Cave speleothem δ18O record (Edwards et al., 2015; Reimer et al., 2013; Southon et al., 2012) with improved resolution and additional dating constraints (see Wang et al., 2001, for the original, lower resolution Hulu δ18O record). The DO events are resolved most clearly in the NGRIP δ18O record, which is available at 20-year resolution. We use the GICC05modelext chronology for this core, which is based on annual-layer counting back to 60 ka BP and ice-flow modeling for ice older than 60 ka (Rasmussen et al., 2006; Svensson et al., 2006; Wolff et al., 2010). While annual-layer counting provides accurate relative ages (e.g., the duration of DO interstadials), it provides relatively inaccurate absolute ages due to the cumulative nature of counting uncertainty (Table 1). The refined Hulu δ18O record also shows the abrupt DO events in high temporal resolution (Fig. 6). The speleothem chronology is based on U/Th radiometric dating, providing much smaller uncertainty in the absolute ages than GICC05 (Table 1). The reason for selecting this record over other speleothem records is the large number of U/Th dates, the low detrital Th at the site, and the high sampling resolution of the δ18O record (Wang et al., 2001). In the Hulu data, as in other records of DO variability, the interstadial onsets are more pronounced and abrupt than their terminations. We therefore only use the timing of the former as age constraints, as they can be established more reliably. The onset of NH interstadial periods as expressed in Hulu δ18O is given in Table 1.

In both the NGRIP and Hulu Cave δ18O records we have determined the ages of the midpoints of the DO transitions (Fig. 6; Table 1); a plot of their difference (Hulu age minus NGRIP age) is shown in Fig. 7, where the error bars denote the root sum square of the NGRIP and Hulu midpoint determination uncertainty (Sect. 4.3). The Hulu ages are systematically older than the NGRIP ages, and the age difference increases going further back in time. Note that the Hulu–NGRIP age difference is smaller than the stated GICC05 counting uncertainty (832 to 2573 years) but larger than the Hulu age uncertainty (92 to 366 years). A linear fit through these data, forced to intersect the origin, is given by
Table 1. Overview of CH$_4$ tie points for NH warming events. WD ages printed in boldface are assigned as part of the CH$_4$ synchronization; all other ages are on their independent chronologies.

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0.0063 \times \text{GICC05 age}, suggesting that the GICC05 annual-layer count on average misses 6.3 out of every 1000 layers. Because of this observation we use a linearly scaled version of the GICC05 chronology (GICC05 \times 1.0063) as the target chronology for methane synchronization. This approach has several advantages. First, it respects both the superior relative ages (i.e., interval durations) of GICC05, as well as the superior absolute ages of the Hulu chronology. Second, it is very simple to convert between the WD2014 and GICC05 chronologies (CH$_4$-synchronized section of the chronology only); one simply needs to divide WD2014 ages by 1.0063 (and add 50 years to convert to the b2k reference date). Third, it still allows for direct synchronization of WD CH$_4$ to the NGRIP $\delta^{18}$O record, providing more tie points than direct synchronization to the Hulu record would. Note that the GICC05 \times 1.0063 target chronology respects the Hulu age constraints in an average sense only; the age of individual events differs between Hulu and our target chronology by up to 180 years. Our approach therefore represents only a first-order correction of a growing offset between GICC05 and Hulu; nonlinear temporal changes in the counted dating error may exist from one tie point to the next (Fleitmann et al., 2009).

The exercise of finding the transition midpoints and determining the GICC05–Hulu scaling factor was performed by two of the authors (J. P. Severinghaus and C. Buizert), independently of each other. The scaling factors obtained were 1.0063 and 1.0064, respectively, showing that, to first order, this result is insensitive to (subjective) judgment in identifying the transitions. The difference between the Hulu ages and 1.0063 \times GICC05 ages are all within the stated Hulu 2\sigma dating error (Table 1). Consequently, our chronology is not in violation of any Hulu constraint as it respects the Hulu 2\sigma error at all of the tie points. In deriving the scaling we have assumed that the abrupt DO transitions observed in NGRIP and Hulu are simultaneous, which is not necessarily true. The
4.5 Interpolation between age constraints

We can assign a gas age to each of the depths where an abrupt WD CH$_4$ transitions occurs; we do this for DO 4.1 through DO 18, i.e., the events prior to 31.2 ka BP (the onset of the WD layer count). The gas age we assign is equal to 1.0063 times the GICC05 age for the same event, with 25 years subtracted to account for the slight CH$_4$ lag behind Greenland $\delta^{18}O$. By adding $\Delta$age (Sect. 3) to this gas age we assign an ice age. These assigned ages are printed in boldface in Tables 1 and 2.

To obtain a continuous depth–age relationship between these ice age constraints, we have to apply an interpolation strategy. This task amounts to estimating the annual-layer thickness $\lambda(z)$ along the depth part of the core. The simplest approach is to assume a constant accumulation rate in between the age constraints; this is shown in Fig. 8b for the case where we use the age constraints from NH warming events only (black) or the age constraints from both NH warming and cooling events (red). The disadvantage of this approach is that it results in discontinuities in $\lambda(z)$ (the first derivative of the depth–age relationship), which we consider highly unrealistic. A more realistic approach is therefore to assume that $\lambda(z)$ is continuous and smooth (Fudge et al., 2014); Fig. 8e shows two scenarios in which we use a spline function to estimate $\lambda(z)$, where again we have applied age constraints from NH warming events only (orange) or age constraints from both NH warming and cooling events (blue).

For comparison, past $A$ obtained from the firm densification model (Sect. 3) is plotted in green (Fig. 8b). While the $\delta^{15}N$-based $A$ follows the synchronization-based $A$ estimates broadly, the millennial-scale details do not agree. We want to point out that this is not unexpected, since both methods have their imperfections. In particular, any errors in the (stretched) GICC05 age model or in our modeled thinning function $f(z)$ will strongly impact the synchronization-based $A$ estimates in Fig. 8b. The discrepancy is pronounced between 60 and 65 ka, where we have to strongly reduce $\lambda(z)$ in order to fit the age constraint(s) from DO 18, while $\delta^{15}N$ provides no evidence for low $A$ during this interval.

For the WD2014 chronology we have applied the smooth $\lambda(z)$ interpolation scheme using all age constraints (i.e., both NH warming and cooling events). The midpoint detection uncertainty is comparable for all events and systematically smaller at the start of interstadial periods than at the terminations (Tables 1 and 2). For short interstadials (e.g., DO 9) this leads to a large relative uncertainty in the event duration, and thereby a large uncertainty in the implied accumulation rates (Fig. 8b). We force the interpolation to fit all NH warming constraints perfectly, yet relax this requirement for NH cooling constraints to prevent large swings in $\lambda(z)$ for the short-duration events. The WD2014 chronology fits the NH warming and NH cooling age constraints with a 0- and 16-year rms offset, respectively. Because the duration of (inter)stadial periods is well constrained in the layer-counted GICC05 chronology, using both the NH warming and NH cooling tie points results in a more robust chronology. The duration of (inter)stadial periods is 0.63 % longer in WD2014 than in GICC05, which is well within the stated GICC05 counting error of 5.4 % (31.2–60 ka interval).
4.6 Age uncertainty

The age uncertainty we assign to the deep part (> 2850 m) of the WD2014 chronology has four components.

The first source of uncertainty is the $\Delta$age calculation; we use the 2σ uncertainty obtained in the $\Delta$age sensitivity study (Sect. 3.2). The second source of uncertainty is the choice of interpolation scheme used to obtain a continuous chronology; here we use the standard deviation between the four different interpolation schemes of Fig. 8b as an uncertainty estimate. The third source of uncertainty is the difficulty in determining the timing of the abrupt events in the time series; we use the uncertainty in the midpoint evaluation (root sum square of WD CH$_4$ and NGRIP $\delta^{18}$O estimates). The last source of uncertainty is the age uncertainty in the hybrid NGRIP–Hulu chronology that we synchronize to. We use the stated Hulu age uncertainty plus 50 years to account for possible leads or lags in the NGRIP–Hulu $\delta^{18}$O phasing, plus the absolute value of the offset between the Hulu ages and the 1.0063 $\times$ GICC05 ages. For DO events where we do not have reliable Hulu age estimates (Table 1), we set the uncertainty to the Hulu age uncertainty of the nearest event, plus the uncertainty in the interval duration specified by the GICC05 layer count. For example, for DO 14 we do not have a reliable Hulu age estimate, and we use the Hulu age uncertainty of DO 16.2 (226 years) plus the uncertainty in the DO 14 to DO 16.2 interval duration on GICC05 (209 years), giving a total of $226 + 209 = 435$ years.
Table 2. Overview of CH$_4$ tie points for NH cooling events. WD ages printed in boldface are assigned as part of the CH$_4$ synchronization; all other ages are on their independent chronologies

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The uncertainties (2σ values) are plotted in Fig. 8c (log scale). We assume these four uncertainties to be independent, and use their root sum square as the total uncertainty estimate on the WD2014 ice age scale (Fig. 8c, black curve). Note that the fourth source of uncertainty is only relevant when considering absolute ages; when evaluating relative ages (e.g., between WD ice and WD gas phase, or between WD and NGRIP), this last contribution does not need to be considered. For the deepest WD ice (3404 m depth) we thus find an age of 67.7 ± 0.9 ka BP.

5 Discussion

While the WAIS Divide ice core does not extend as far back in time as deep cores from the East Antarctic Plateau, its relatively high temporal resolution (due to the high snow accumulation rate) makes it an ice core of great scientific value. WD accumulation rate during the LGM (∼ 10 cm a$^{-1}$ ice equivalent) is still higher than the present-day accumulation rate at the EPICA (European Project for Ice Coring in Antarctica) Dronning Maud Land core (7 cm a$^{-1}$), which is generally considered a high-accumulation core (EPICA Community Members, 2006). With 68 ka in 3404 m of core, the core average λ is 5 cm a$^{-1}$, at the onset of the last deglaciation (18 ka BP) λ is around 4 cm a$^{-1}$, and near the bed λ is around 0.4 cm a$^{-1}$. This high temporal resolution provides the opportunity for obtaining very detailed climatic records.

High accumulation rates also result in a small Δage. Figure 9 compares Δage between several Antarctic cores (note the logarithmic scale). Δage at WD is approximately one-third of the Δage at EPICA DML (EDML) and Talos Dome (TALDICE), and one-tenth of the Δage at EPICA Dome C (EDC), Vostok, and Dome Fuji. Because the uncertainty in the Δage (or Δdepth) calculation is typically on the order of 20 %, a smaller Δage allows for a more precise interhemispheric synchronization with Greenland ice core records using CH$_4$. The small WD Δage uncertainty during MIS 3 allows for investigation of the phasing of the bipolar seesaw (Stocker and Johnsen, 2003) at sub-centennial precision (WAIS Divide Project Members, 2015).

In comparing the shape of the Δage profiles, there are some interesting differences (Fig. 9). It is important to realize that not all the Δage histories shown were derived in the same way; WD and Dome Fuji Δage were derived using densification models, and the other four were derived using the Δdepth approach (Parrenin et al., 2012) and a Bayesian inverse method that includes a wide range of age markers (Veres et al., 2013). We will therefore focus on comparing the WD and Dome Fuji results. Δage at WD shows more pronounced variability than at Dome Fuji, particularly during MIS 3. The reason is that the glacial firn pack at Dome
Fuji is about 4000 years old, and consequently the firn column integrates over 4000 years of climate variability, thereby dampening the Δage response to millennial-scale climatic variability. At WD the glacial firn layer is only about 350 years old, and therefore the firn is in near equilibrium with the millennial-scale climate variations. This difference in response time is also obvious during the deglaciation, where WD Δage transitions from glacial to interglacial values between 18 and 14.5 kaBP, while Dome Fuji takes more time (18–10 kaBP). Surprisingly, EDML Δage does not show a strong deglacial Δage response, unlike all the other cores.

The relatively small Δage at WAIS Divide also allows for precise investigation of the relative timing of atmospheric greenhouse gas variations and Antarctic climate (Barnola et al., 1991; Pedro et al., 2012; Caillon et al., 2003; Parrenin et al., 2013; Ahn et al., 2012). Recent works suggest that during the last deglaciation the rise in atmospheric CO₂ lagged the onset of pan-Antarctic warming by approximately 0 to 400 years (Pedro et al., 2012; Parrenin et al., 2013). This Antarctic warming around 18 kaBP is presumably driven by the bipolar seesaw, as it coincides with a reduction in Atlantic overturning circulation strength as seen in North Atlantic sediment records (McManus et al., 2004). The WD Δage at 18 ka (gas age) is 515 ± 91 years (2σ), much smaller than at central East Antarctic sites such as EPICA Dome C, where Δage is approximately 3850 ± 900 years (Veres et al., 2013, with the Δage uncertainty taken to be the difference between the gas age and ice age uncertainties). The precision with which one can determine the relative phasing of climatic (i.e., δ¹⁸O of ice) and atmospheric signals is set by the uncertainty in Δage (or equivalently, the uncertainty in Δdepth). High-resolution WD records of CO₂ and CH₄ (Marcott et al., 2014) place the onset of the deglacial rise in the atmospheric mixing ratio of these greenhouse gases on the WD2014 chronology at 18 010 and 17 820 years BP, respectively. However, evaluating the relative phasing of CO₂ and Antarctic climate is complicated by the observation of asynchronous deglacial warming across the Antarctic continent (WAIS Divide Project Members, 2013). Attempts to capture the climate–CO₂ relationship in a single lead-lag value may be an oversimplification of deglacial climate dynamics.

An important next step will be to synchronize the WD chronology with other Antarctic cores via volcanic matching and other age markers (e.g., Severi et al., 2007; Sigl et al., 2014). Because of the annual-layer count and possibility of tight synchronization to Greenland ice cores, WD could contribute to an improved absolute dating of Antarctic cores, as well as improved cross-dating between cores. Such cross-dating could help inform the WD chronology as well, particularly in the deepest part of the core, where the ice is potentially highly strained, as suggested by the interpolation difficulties in the 60–65 ka interval (Fig. 8b). With a synchronized chronology, WD could improve the representation of West Antarctic climate in Antarctic ice core stacks (Pedro et al., 2011; Parrenin et al., 2013), and provide a more refined pan-Antarctic picture of the climate–CO₂ relationship.

6 Summary and conclusions

We have presented a first chronology for the deep (> 2850 m) section of the WAIS Divide ice core, which is based on stratigraphic matching to Greenland ice cores using globally well-mixed methane. We use a dynamical firn densification model constrained by δ¹⁵N data to calculate past Δage, and find that Δage was smaller than 525 ± 120 years for all of the core. Using high-resolution WD records of atmospheric CH₄, we synchronize WD directly to Greenland NGRIP δ¹⁸O for the abrupt onset and termination of each of the DO interstadials. To each event we assign an age corresponding to 1.0063 times its GICC05 age, which brings the ages in agreement with the high-resolution U/Th-dated Hulu speleothem record. The uncertainty in the final chronology is based on the uncertainties in (i) the Δage calculations, as evaluated with a sensitivity study; (ii) the interpolation strategy, as evaluated by comparing four different interpolation methods; (iii) determining the timing of events in the
different time series; and (iv) the ages of the hybrid NGRIP–Hulu chronology we are synchronizing to.

Due to the combination of a small age and a high-resolution methane record, the WAIS Divide ice core can be synchronized more precisely to Greenland records than any other Antarctic core to date. This is important when investigating interhemispheric climate relationships such as the bipolar seesaw. The small WD age furthermore provides valuable opportunities for precise investigation of the relative phasing of atmospheric greenhouse gas variations and Antarctic climate.

Appendix A: Densification physics

The densification rates used in this work are based on the empirical steady-state model by Herron and Langway (1980) (the H-L model). We use the H-L model with minor modifications that allow it to be run dynamically (i.e., with time-variable T and A) and to include the softening effect of impurities following Freitag et al. (2013a). The H-L model divides the firn column in two stages, separated at the critical density \( \rho_c = 550 \, \text{kg m}^{-3} \), occurring at the critical depth \( z_c \).

For the upper firn (\( \rho \leq \rho_c \), stage 1), the densification rates are given by

\[
\frac{d\rho}{dt} = k_1 A (\rho_{ice} - \rho),
\]

with

\[
k_1 = 11 \exp \left( -\frac{E_1}{RT} \right),
\]

where \( E_1 = 10.16 \, \text{kJ mol}^{-1} \) is the activation energy for stage 1 and \( R \) is the universal gas constant. Because both the sinking velocity of deposited layers (\( w = dz/dt \)) and the densification rate scale linearly with \( A \), the resulting density–depth profile \( \rho(z) \) in stage 1 becomes independent of \( A \), and sensitive to \( T \) variations only.

For the deeper firn (\( \rho > \rho_c \), stage 2), we use Eq. (4c) from Herron and Langway (1980), which was first derived by Sigfús J. Johnsen. This equation gives the densification rate in terms of overburden load, which allows the model to be run dynamically. The stage 2 densification rates are given by

\[
\frac{d\rho}{dt} = k_2 \frac{(\sigma_z - \sigma_z') (\rho_{ice} - \rho)}{\ln ((\rho_{ice} - \rho_c)/(\rho_{ice} - \rho))},
\]

with

\[
k_2 = 575 \exp \left( -\frac{E_2}{RT} \right),
\]

where \( E_2 = 21.4 \, \text{kJ mol}^{-1} \) is the activation energy for stage 2 and \( \sigma_z \) denotes the firn overburden load at a given depth in \( \text{Mg m}^{-2} \):

\[
\sigma_z = \int_0^z \rho(z') dz' / 1000.
\]

Note that we divide by 1000 to convert from \( \text{kg m}^{-3} \) to \( \text{Mg m}^{-2} \), the units used by Herron and Langway (1980).

We use the mathematical description by Freitag et al. (2013a) to include the hypothesized firm softening effect of impurities. In this approach an increasing Ca concentration, as a proxy for mineral dust content, lowers the activation energy of firm, thereby enhancing densification rates. This is tantamount to stating that dusty firm behaves as if it were “warmer” than its climatological temperature. The H-L activation energies of Eqs. (A2) and (A4) are modified by [Ca] in the following way:

\[
E^{\text{Ca}} = E^{\text{HL}} \times \alpha \left[ 1 - \beta \ln \left( \frac{[\text{Ca}]}{[\text{Ca}]_{\text{crit}}} \right) \right],
\]

where \( E^{\text{Ca}} \) and \( E^{\text{HL}} \) are the Ca-modified and original H-L activation energies, respectively, \( [\text{Ca}]_{\text{crit}} = 0.5 \, \text{ng g}^{-1} \) is the minimum concentration at which impurities affect densification, and \( \alpha \) and \( \beta \) are calibration parameters. Whenever \([\text{Ca}](z) < [\text{Ca}]_{\text{crit}}\), we set \([\text{Ca}](z) = [\text{Ca}]_{\text{crit}}\).

The parameter \( \beta \) sets the sensitivity to dust loading, and \( \alpha \) is a normalization parameter that is included to account for the fact that the original H-L model was calibrated without the impurity effect. Consequently, if \( \beta > 0 \), one needs to compensate by setting \( \alpha > 1 \) to preserve the original H-L calibration. The work by Freitag et al. (2013a) recommends \( \beta = 0.01 \) and \( \alpha = 1.025 \) (which yields \( E^{\text{Ca}} = E^{\text{HL}} \) at \([\text{Ca}] = 5.73 \, \text{ng g}^{-1} \)).

Using the recommended value of \( \alpha = 1.025 \) at WD provides a poor fit to observations of present-day firm density and close-off depth. The optimal fit to present-day WD observations is obtained using an activation energy equal to \( 1.007 \times E^{\text{HL}} \); this is in between the values suggested by Herron and Langway (1980) and Freitag et al. (2013a). In the experiment presented in Fig. 4 we changed the dust sensitivity \( \beta \); it is clear that we need to simultaneously change \( \alpha \) to keep the model well-calibrated to present-day conditions. Due to the fact that the mean late Holocene WD [Ca] is around 0.8 ng g\(^{-1}\), we let \( \alpha = 1.007/(1 - \beta \ln(0.8/0.5)) \) in the experiment of Fig. 4. This approach ensures that the present-day \( E^{\text{Ca}} \) is invariant with \( \beta \), and equals \( E^{\text{Ca}} = 1.007 \times E^{\text{HL}} \). This means that whatever value we choose for \( \beta \), we will obtain a good fit to the present-day age, \( \delta^{15}\text{N} \), and \( A \) values that are well known from direct observations (Battle et al., 2011).

To validate the H-L model age simulations, we repeated the firm modeling using the densification physics of Arnaud et al. (2000), which is also the basis of the model by Goujon et al. (2003). Our implementation of the Arnaud model is based on the description in the latter paper, with one modification at the critical density that we outline here.
In the Arnaud model, densification in the stage 1 follows the work of Alley (1987), and is given by

\[
\frac{dD}{dt} = \gamma \left( \frac{P}{D^2} \right) \left( 1 - \frac{5}{3} D \right), \quad (A7)
\]

with \( D \) the relative density \( D = \rho / \rho_{ice} \), \( P \) the overburden pressure, and \( \gamma \) a scaling factor used to make the densification rates continuous across the critical density \( D_c \). Stage 2 densification is given by

\[
\frac{dD}{dt} = k_A \left( D^2 D_c \right)^{\frac{1}{2}} \left( \frac{a}{\pi} \right)^{\frac{1}{2}} \left( \frac{4 \pi P}{3 a Z D} \right)^3, \quad (A8)
\]

with

\[
k_A = 4.182 \times 10^4 \exp \left( - \frac{E_A}{RT} \right), \quad (A9)
\]

where \( a \) is the average contact area between the grains, \( Z \) is the coordination number, and \( E_A \) is the activation energy (60 kJ mol\(^{-1}\)). Arnaud densification rates for stage 3 (\( D \geq 0.9 \)) are describe elsewhere (Goujon et al., 2003; Arnaud et al., 2000).

The difficulty in implementing this model is the following. The densification rates of Eqs. (A7–A8) exhibit a discontinuity at the critical density \( D = D_c = 0.6 \) that cannot be remedied with the scaling factor \( \gamma \). On approaching \( D_c \), densification rates given by Eq. (A7) go to zero (due to the inclusion of the term \((1 - \frac{5}{3} D)\), while densification rates given by Eq. (A8) go to infinity because the contact area \( a \) equals zero at \( D = D_c \). Clearly neither equation gives a realistic result at \( D = D_c \). Therefore, in our implementation of the Arnaud model we use the H-L densification rates of Eq. (A1) instead of Eq. (A7) in stage 1. We take the onset of stage 2 to be the density at which Eqs. (A1) and (A8) intercept, thus avoiding the singularity in Eq. (A8). This approach has the additional advantages of removing dependence on ad hoc scaling factor \( \gamma \) and introducing realistic temperature dependence for stage 1 densification. Because stage 1 spans just the top 10–20 % of the firn column, the modification has only a minor influence on the overall behavior of the Arnaud model. The Goujon model code avoids the singularity in Eq. (A8) by extending stage 1 to \( D_c + \varepsilon \) (Anaïs Orsi, personal communication, 2014), a procedure not described in Goujon et al. (2003).

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