



A high-resolution Termination II trace gas record and chronology for a continuous ice core from the Allan Hills Blue Ice Area, Antarctica

Julia Marks-Peterson¹, Edward Brook¹, Jenna A. Epifanio¹, Emily Rice¹, Jeff Severinghaus,² Murat Aydin³, Samantha Gass¹, Jon S. Edwards¹, Austin Carter⁴, Christo Buizert¹, John Higgins⁴, Sarah Shackleton⁵, Jacob Morgan², Michael Kalk¹, Ross Beaudette², Andrei Kurbatov⁶, and Eric Saltzman³

¹College of Earth, Ocean, and Atmospheric Sciences, Oregon State University, Corvallis, OR 97331, USA

²Scripps Institution of Oceanography, University of California – San Diego, La Jolla, CA 92037, USA

³Department of Earth System Science, University of California – Irvine, Irvine, CA 92697, USA

10 ⁴Department of Geosciences, Princeton University, Princeton, NJ 08540, USA

⁵Department of Geology and Geophysics, Woods Hole Oceanographic Institution, Woods Hole, MA 02543, USA

⁶School of Earth and Climate Sciences, University of Maine, Orono, ME 04469, USA

Correspondence to: Julia Marks-Peterson (peterju2@oregonstate.edu)

Abstract. The Allan Hills Blue Ice Area (BIA) in East Antarctica has yielded some of the oldest ice on the planet. The oldest samples, which date to the late Miocene, occur in complexly folded sections and lack stratigraphic continuity. In other parts of the Allan Hills BIA, shallow (~100-200 m) cores have yielded stratigraphically intact sections of younger ice. These cores provide context for the older records and the possibility of easily obtaining large volume samples. The core studied here, ALHIC1903 provides a stratigraphically continuous climate record spanning the Penultimate Glacial Maximum (PGM) to the Last Interglacial Period (LIG), approximately 116 to 149 ka, spanning Termination II (T-II). Here, we present new trace gas records (CO₂, CH₄, and N₂O) across this transition and create a gas-phase chronology for the core by correlating these to well-dated records from the EPICA Dome C (EDC) ice core. We estimate that snow accumulation rates across T-II increased by a factor of 9 (upper bound), corroborating previous findings. Our data show excellent agreement with EDC and other Antarctic cores and provide a new record of N₂O concentrations from the PGM that appears to be free of artifacts that have plagued other N₂O records from this time period. We find a minimum glacial N₂O concentration of 195 ppb at approximately 139 ka. After a short rise, we resolve a drop in N₂O at 136 ka, coincident with the onset of Heinrich Stadial 11 (HS11) and potentially due to a drop in marine emissions caused by a change in ocean circulation. We also report a high-resolution record of parameters that vary due to the characteristics of the depositional site including total air content (TAC), the stable isotopic composition of ice ($\delta^{18}\text{O}_{\text{ice}}$) and of molecular nitrogen ($\delta^{15}\text{N-N}_2$). TAC variability across T-II exhibits the expected relationship with insolation with a possible additional impact of changes in accumulation rate. Using TAC as a first order proxy for barometric pressure suggests a source region elevation of ~2200 m at the LIG, similar to the current elevation of a local dome that may source the BIA. TAC and the other site-specific parameters in ALHIC1903 agree with those in a nearby stratigraphically discontinuous



core (ALHIC1901) containing 4-million-year-old ice, indicating similar depositional settings for both cores and reinforcing ALHIC1903 as a stratigraphically intact analogue for old ice in the region. The excellent preservation of continuous atmospheric records at Allan Hills supports the potential for obtaining high-quality ice core records covering multiple glacial cycles from relatively shallow depths at a location that is relatively easily accessible from McMurdo station.

1 Introduction

Geophysical investigation of the Allan Hills Blue Ice Area (BIA) led to the discovery of a shallow, stratigraphically continuous record of ice spanning Marine Isotope Stage 7 (186-245 ka) to 5 (116-133 ka) (Kehrl et al., 2018; Spaulding et al., 2012, 2013; Yan et al., 2021). The original ice core drilled at this site in 2010-2011 (the S27 core) recorded T-II (~136-129 ka, (Clark et al., 2020)) from high-resolution measurements of $\delta^{18}\text{O}_{\text{ice}}$ and the isotopic composition of molecular oxygen ($\delta^{18}\text{O}_{\text{atm}}$), as well as a limited number of trace gas measurements. Despite very low accumulations rates through the PGM, the S27 record exhibited excellent preservation of both ice and gas signals. A second ice core, ALHIC1903, was drilled at this site in 2019-2020 using the 24 cm diameter Blue Ice Drill (Kuhl et al., 2014), obtaining a 150 m long core. The larger sample volumes provided an opportunity to make high-resolution gas measurements across T-II.

In this study, we measured methane (CH_4), carbon dioxide (CO_2), nitrous oxide (N_2O), $\delta^{18}\text{O}-\text{O}_2$, $\delta^{15}\text{N}-\text{N}_2$ and TAC from the ALHIC1903 ice core across the PGM, T-II, and the onset of the LIG (~129-116 ka) (Fig. 1). We use these data to improve our understanding of the Allan Hills BIA as a paleo archive. Specifically, we first date the ALHIC1903 gas record using CH_4 and $\delta^{18}\text{O}_{\text{atm}}$ matching to high-resolution well-dated EDC records, establishing a gas chronology for future use of this and other cores in the region. We then demonstrate the integrity of the CO_2 and N_2O records, both of which exhibit excellent agreement with previously published records. Using data from two different laboratories, we demonstrate the preservation of an atmospheric N_2O record for the PGM, an interval where other ice core N_2O records suffer from contamination due to dust. We examine millennial-scale features in the N_2O record and show that N_2O responds to Heinrich stadial conditions as has been observed for the last glacial period (Schilt et al., 2013). Third, we investigate site-specific parameters in ALHIC1903 ice (i.e. $\delta^{18}\text{O}_{\text{ice}}$, TAC, and $\delta^{15}\text{N}-\text{N}_2$), discuss their implications for the depositional setting, and compare them to measured parameters from a stratigraphically discontinuous core (ALHIC1901) in the Allan Hills spanning the last 3 million years. Finally, we use TAC to estimate the elevation of the deposition site and conclude that the elevation has stayed stable since the LIG, and possibly through the Pleistocene.

1.1 The Allan Hills Setting

The ALHIC1903 borehole is located at 76.7024° S, 159.3106° E at an elevation of 1960 m in the main ice field of the Allan Hills BIA (Fig. 2). The core site is along a radar track, and analysis of the profile and subsequent ice flow analysis has



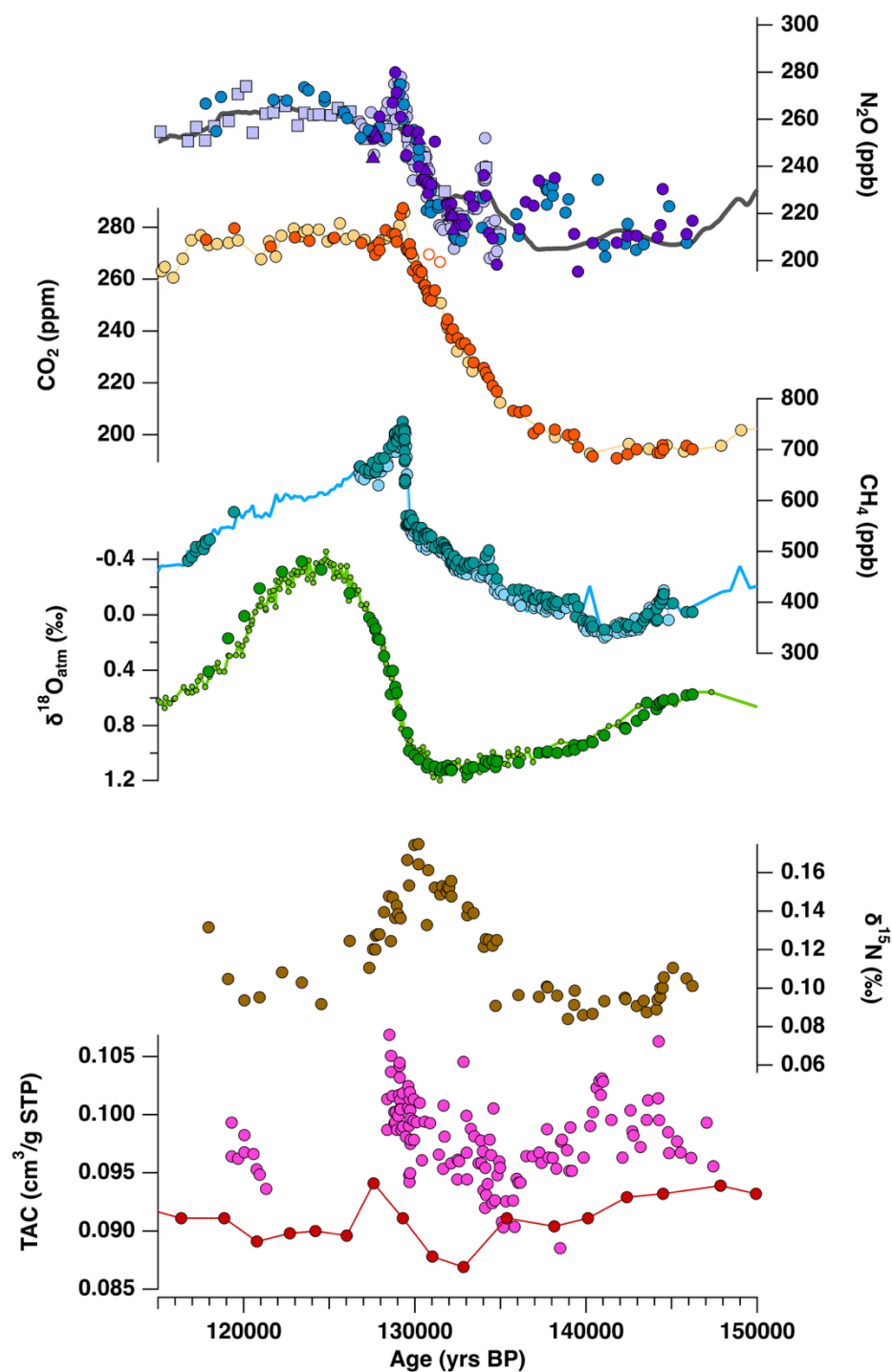
determined the location of the depositional setting to likely be a local ice divide at a surface elevation of 2160 m (Kehrl et al., 2018; Spaulding et al., 2012). In a nearby snow patch, a present-day accumulation rate was estimated at 7.5 mm/yr based on a shallow core spanning the last 660 years (Dadic et al., 2015). However, it should be noted that this value is neither representative of the negative mass balance (i.e. ablation) of the ALHIC1903 coring site, nor the positive accumulation rate of the upstream depositional site. While most ice cores spanning T-II are from the East Antarctic plateau, Allan Hills is nearer to the coast (~80 km away from the Ross Sea coastline) and thus cores from this area can provide insight into regional environmental change across this interval.

Yan et al. (2021) derived accumulation rates from the 225 m S27 core using an inverse firn densification model, inputting the gas-age-ice-age difference calculated from the ice and gas chronologies, and temperatures calculated from stable water isotopes. Derived accumulation rates show a very large increase across T-II from ~2.6 mm/yr at 133.2 ka to ~86 mm/yr at 128.2 ka (Yan et al., 2021). This observation led Yan et al. to hypothesize open water conditions in the Ross Sea leading to enhanced moisture transport to the Allan Hills region, a hypothesis supported by a recent study of dust concentration and provenance from ALHIC1903 (Carter et al., 2026).

Improvements have been made to our understanding of the Allan Hills BIA since the S27 core was collected and analyzed. In addition to the drilling of ALHIC1903, a geophysical survey of the accumulation area suggested that a continuous ice core record extending to potentially 1 million years at >1000 m depth exists in this region (Kehrl et al., 2018). Isochrons can be traced from that location to the site of ALHIC1903/S27. Thus, our high-resolution gas records from the ALHIC1903 ice core can be used to better understand the stratigraphy and potential characteristics of records that could be obtained from coring at the deeper site or in other areas of stratigraphically continuous ice in Southern Victoria Land.



85 1.1.1 Figures



(Figure 1 caption on next page)

Figure 1: Plot of ALHIC1903 gas records (this study) and other ice core records spanning the Penultimate Glacial Maximum to the Last Interglacial Period. 1a: Atmospheric N_2O concentrations from ALHIC1903 (OSU dry extraction, purple circles; OSU wet extraction, purple triangles; UCI dry extraction blue circles) and EDC (light purple circles (Schmidely et al., 2021) and squares (Schilt et al., 2010, 2013). N_2O simulation (Salehnia et al., 2025) in dark gray. 1b: Atmospheric CO_2 concentrations from ALHIC1903 and S27 (dark orange, this study, Yan et al. 2021) and the ice core compilation (light orange, Bereiter et al., 2015). 1c: Atmospheric CH_4 concentrations from ALHIC1903 and S27 (teal, this study, Yan et al., 2021) and EDC (light blue circles, Schmidely et al., 2021; blue line, Loulergue et al., 2008). 1d: $\delta^{18}\text{O}_{\text{atm}}$ records from ALHIC1903 (dark green) and EDC (light green, Bouchet et al., 2023). 1e: $\delta^{15}\text{N}-\text{N}_2$ from ALHIC1903 and S27 (this study, Yan et al., 2021). 1f: TAC from ALHIC1903 (pink, this study) and EDC (red, Raynaud et al., 2007).

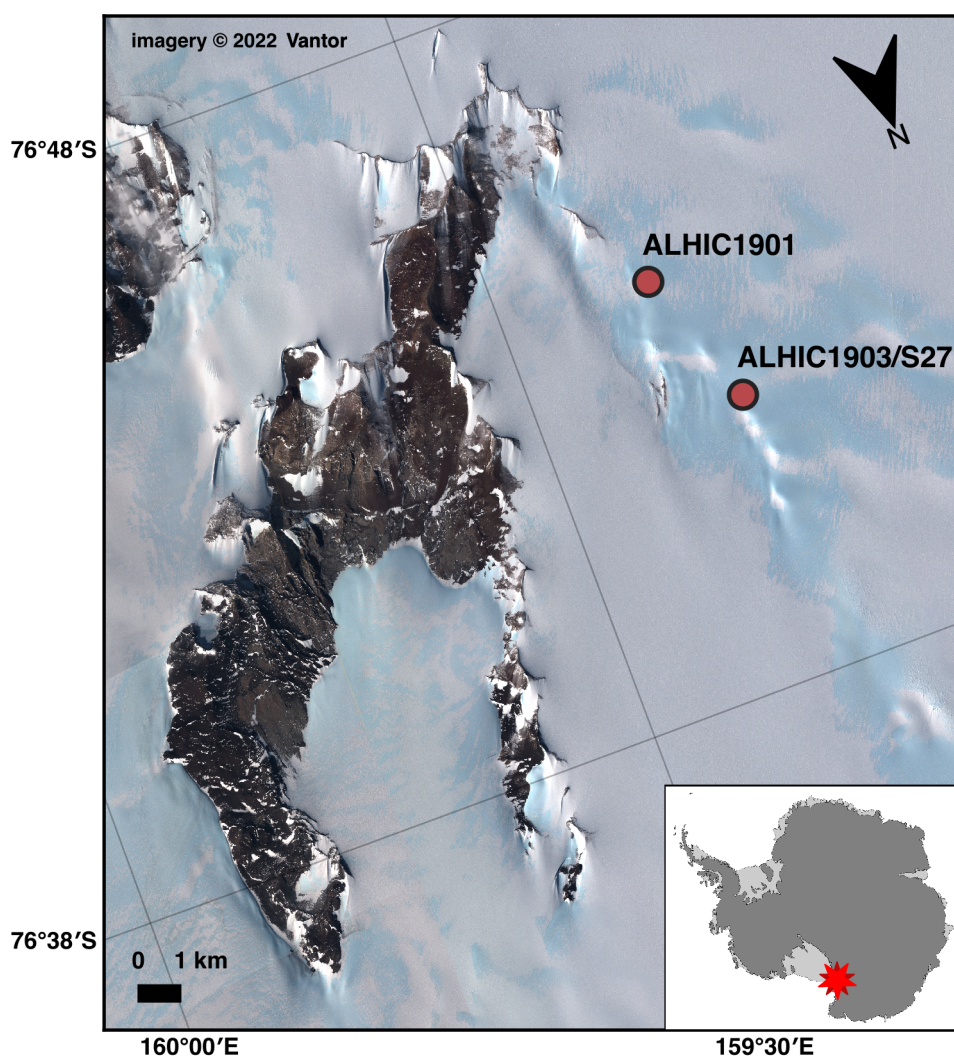


Figure 2: Map of the Allan Hills and locations of shallow drill sites, ALHIC1903 (this study) and ALHIC1901 (stratigraphically discontinuous containing 4-million-year-old ice). Map created using QGIS software and satellite imagery from WorldView-02 and WorldView-03. Imagery © 2022 Vantor. Inset map shows the location of the Allan Hills (red star) within Antarctica using simplified outlines derived from Bedmap2 (Fretwell et al., 2013; Greene et al., 2017).



2 Methods

2.1 CH₄ concentrations

CH₄ concentrations were measured at Oregon State University (OSU) following the method originally described in (Grachev et al., 2009) and updated in (Epifanio et al., 2020; Lee et al., 2020; Mitchell et al., 2011). Briefly, subsamples with dimensions of 10 cm x 6 cm x 3 cm were cut from the core and subsequently cut lengthwise for replicate samples. Individual samples were placed in a glass vacuum flask and attached to an automated vacuum array. The system allows for analysis of 12 flasks sample per day, typically 10 samples and 2 pieces of air-free, laboratory-made “blank” ice. Prior to melting, flasks were kept frozen in a -70 °C ethanol bath and evacuated of laboratory air for one hour using a turbomolecular pump. Samples were subsequently melted by submerging the flasks in a bath of 55 °C water for 30 minutes to release trapped gases. The meltwater was then refrozen in the ethanol bath and the air in the headspace of each flask was expanded into a sample loop and then injected into the gas chromatograph (GC) to be analyzed for CH₄ using a flame ionization detector. Pressure was measured in the sample loop and concentrations were derived from the ratio of peak area to pressure, referenced to a calibrated air standard analyzed eight times at the beginning and end of each day (NOAA Cylinder CB11302, 481.25 ppb). Concentrations are reported on the NOAA4 CH₄ concentration scale (Dlugokencky et al., 2005). Results for each sample were averaged over four repeated expansions to the sample loop.

Reported concentrations were increased by 1.65% to account for the CH₄ remaining in the meltwater. This solubility effect on CH₄ concentrations was empirically derived in (Mitchell et al., 2011) and updated in (Lee et al., 2020). Additionally, any deviations from true concentrations due to small leaks or other contaminations in the analytical setup were quantified by measuring “blank” ice each day along with samples. Before melting the samples, the flasks containing “blank” ice were filled with an amount of air from the NOAA-calibrated air standard that matched the expected amount of air contained in the samples. The difference between measurements from flasks containing “blank” ice and the known standard value was then subtracted from the sample results. Seven different blank corrections were applied to the data associated with different measurement campaigns. Average blank corrections ranged from 7.58 ppb to 14.14 ppb.

Finally, we account for the effect of gravitational settling on gases in the diffusive zone of the firn column (Sowers et al., 1989), which can be estimated using $\delta^{15}\text{N-N}_2$ (Epifanio et al., 2020; Mitchell et al., 2011). Where $\delta^{15}\text{N-N}_2$ data did not exist, we linearly interpolated between measurements. The gravitational fractionation of CH₄ was derived using Eq. (1):

$$\text{CH}_{4,\text{corr}} = \text{CH}_{4,\text{meas}} \left(1 + \Delta M \left(\frac{\delta^{15}\text{N}}{1000} \right) \right), \quad (1)$$

where ΔM is the difference between the molar weight of dry air and CH₄ ($\Delta M = 12.92$ g/mol). $\delta^{15}\text{N}$ varies from 0.084‰ to 0.175‰, and gravitational correction ranges from 0.39 to 1.48 ppb (1 sigma = 0.34 ppb). 136 CH₄ measurements were made



135 from 11 m to 145 m depth (28 singles, 97 in duplicate, 10 in triplicate, and 1 run in quadruplicate). The pooled standard deviation, evaluated from the agreement between replicate samples, is 4.3 ppb.

2.2 Total air content

The total air content (TAC) of ice samples were derived from the methane measurement system using the pressure of the extracted air in the vacuum line, the sample mass, and corrections for temperature of the analytical system (Epifanio et al., 2023). A detailed explanation of the method for deriving TAC can be found in Epifanio et al., 2023, and is briefly described here. We use the ideal gas law to derive volume of air at STP and divide by sample mass to calculate the total amount of air per gram of ice. The method takes advantage of repeated expansions of extracted air from each CH₄ sample during its transfer from glass flask to gas chromatograph. The vacuum line past the GC valve is maintained as a constant temperature and volume and assuming the temperature in the remaining vacuum line and instrument remain stable, the ideal gas law for each expansion can be used to derive TAC. The final equation for total air content is:

$$\text{TAC} = \frac{P_1^2}{R(P_1 - P_2)} \frac{T_{gc} T_{STP}}{m V_{gc} P_{STP}}, \quad (2)$$

where P_1 and P_2 are the measured pressures from the first and second expansions, respectively. T_{gc} and V_{gc} refer to the temperature and volume of the relevant portion of the vacuum line. The ratio of the two was empirically determined to be 12.79 K cm³. R is the universal gas constant (8.314 J mol⁻¹ K⁻¹) and m is the mass of the ice sample. TAC is reported in units of cm³ g⁻¹ at standard temperature and pressure (STP). TAC values ranged from 0.081 to 0.107 cm³ g⁻¹ and the pooled standard deviation of replicates is 0.002 cm³ g⁻¹.

2.3 Other trace gases at OSU (CO₂ and N₂O concentrations)

155 Concentrations of CO₂ and N₂O were analyzed on ice samples for a study reporting measurement of $\delta^{13}\text{C}$ of CO₂, the results of which will be published elsewhere. The analytical method is discussed in detail in (Bauska et al., 2014). Briefly, two subsamples varying from 150 to 300 g were cut adjacent to CH₄ subsamples in the core and placed into individual stainless-steel chamber with a perforated grating surface. The chambers were placed in a chest freezer kept at -60 °C and evacuated of laboratory air using both a scroll pump followed by a turbomolecular pump. The next morning, each chamber was individually pumped again using the turbomolecular pump for 1 hour prior to gas extraction. Individual chambers were then connected to a pneumatic piston system that shakes horizontally on a linear slide and shaves the ice sample to release the air. Each chamber was then attached to a vacuum array fitted with a -100°C water trap and a small aliquot of the extracted air (~3% of the total air) was subsequently expanded into the line and trapped in a stainless-steel tube in an exchange gas cryostat held at 11 K, for later analysis of CO₂ concentration. The rest of the sample air was expanded into the vacuum line through a water trap and

165 CO₂ and N₂O were collected in a -196°C U-shaped stainless-steel trap. The CO₂ and N₂O were then transferred to a small



stainless-steel tube (“cold finger”) that was subsequently attached to the sample inlet of a Thermo Fischer Scientific MAT 253 Plus dual-inlet Isotope Ratio Mass Spectrometer (IRMS) for measurement of $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ of CO_2 .

At the conclusion of sample extraction each day the sample manifold in the cryostat contains three tubes, two filled with samples and one filled with an amount of standard gas that approximately equals the amount of extracted gas from the samples. The manifold was removed from the cryostat and warmed up for 40 minutes before being attached to a vacuum array connected to a gas chromatograph (GC). Each tube was sequentially opened, and air was expanded into the GC to be analyzed for CO_2 concentration. CO_2 measurements were determined by comparing the sample results to a calibrated air standard run on the GC on the same day. CO_2 concentrations were calibrated to the WMO 2007 scale using standard air from a NOAA-calibrated tank with a known concentration of CO_2 , N_2O , and isotopic signature of CO_2 . Raw CO_2 concentrations were corrected for gravitational fractionation and the instrumental blank as measured by the standard from the third tube in the sample manifold. An inverse relationship exists between blank values and pressure in the sample tubes (Fig. S1). We used a linear regression to apply specific blank corrections for various sample pressures. Gravitational enrichment in the CO_2 mixing ratio was corrected for using $\delta^{15}\text{N}\text{-N}_2$ as described in Eq. (1), with a $\Delta M = -15.05$ g/mol. The average blank was 2.31 ppm (1 sigma = 0.57 ppm). CO_2 measurements were made on 58 depths (36 single, and 22 in replicate). The pooled standard deviation of replicates was 1.11 ppm.

N_2O concentrations were derived using results from the dual-inlet IRMS. The method is described in detail in (Assonov and Brenninkmeijer, 2006; Bauska et al., 2014). Briefly, following a sample cycle measurement of $\delta^{13}\text{C}$ of CO_2 , the magnet jumps to m/z 30, to measure the intensity of combined $^{12}\text{C}^{18}\text{O}$ and $^{14}\text{N}^{16}\text{O}$ fragments produced as the sample is ionized in the source. The intensity of m/z 30 is compared to the m/z 44 intensity for both the sample and that of the reference gas of pure CO_2 . The difference between the two signals is used along with empirically-derived parameters for fragmentation yield and the ionization efficiency of N_2O to estimate the $\text{N}_2\text{O}/\text{CO}_2$ ratio of the sample.

A subset ($n=10$) of ALHIC1903 N_2O concentrations were independently measured at OSU via melt-refreeze gas extraction and gas chromatography. Briefly, 5 g samples were placed in pre-chilled glass vacuum flasks and connected to a vacuum extraction line, identical to those used for methane analysis. The evacuation and extraction steps are the same as for CH_4 analysis (see section 2.1). To reduce the effects of solubility, the samples were refrozen slowly with only the bottom 1 cm of the flasks submerged in the ethanol bath. Samples were measured for N_2O by expanding into an Agilent 6890 Gas Chromatograph fitted with an electron capture detector (ECD). N_2O concentrations were determined by comparing the peak area and pressure measurements for the sample to a daily calibration curve made using a NOAA calibrated standard air cylinder (CA08492). The pooled standard deviation of the 10 replicate measurements was 4.48 ppb. Final reported N_2O concentrations from both methods were corrected for the effect of gravitational settling as described in section 2.1, using a $\Delta M = -15.05$ g/mol.



200 2.4 N₂O measurements at UC Irvine

A separate set of samples were analyzed at University of California Irvine (UCI) for measurements of a variety of trace gases; here, we only report N₂O measurements. UCI measurements are based on dry extraction of 400-600 g samples using established procedures, recently outlined by (Aydin et al., 2024). Briefly, samples are loaded into stainless steel vacuum chambers fitted with flat stainless steel cutting surfaces in a walk-in freezer (-23°C), transferred to a -45°C chest freezer, pumped out and flushed with clean N₂ gas for overnight storage under vacuum. The samples are often analyzed the next day, but some samples remain in the -45°C freezer more than one day prior analysis. Storage time in the -45°C freezer does not impact the blanks or the measurements. On the day of the analysis, the vessels are transferred to a -50°C freezer with a cradle connected to a motor to generate linear motion that mechanically releases the gas content of the sample. Background levels of N₂O associated with the extraction vessel, the extraction line, and the analytical system is determined by clean N₂ blanks and are measured before and after extraction. Both the N₂ blanks and the extracted ice core air are condensed into stainless-steel tubes submerged in liquid helium for immediate subsequent analysis. The average N₂O in the pre- and post-blanks is subtracted from the sample analysis before calculation of the mixing ratio. The N₂O blank correction is on the order of 0.1% for a typical ALHIC1903 sample. Reported N₂O concentrations are also corrected for the effect of gravitational settling using $\delta^{15}\text{N}$ as described in section 2.1.

The extracted ice core air is analyzed using a custom preconcentration line coupled to a gas chromatograph and a high-resolution mass spectrometer previously described in the context of other trace gas analyses (Aydin et al., 2024; Nicewonger et al., 2020). Gases are preconcentrated on glass beads inside an 1/8" stainless steel tube cooled down with liquid N₂ and subsequently injected into a GC equipped with an Agilent DB-1 column using He carrier gas at an initial temperature of -50°C, followed by gradual temperature ramping to 150 °C. N₂O is quantified via simultaneous mass spectrometric detection of the NO and ¹⁵NO fragments produced during electron ionization of N₂O and ¹⁵N-N₂O (middle N is labelled) at 30 amu and 31 amu, respectively. The GC-MS system is calibrated with ppb-level working standards prepared at UC Irvine and a known amount of the ppb-level ¹⁵N-N₂O is mixed into every sample during the preconcentration step and injected into the GC-MS along with the sample for simultaneous analysis, providing an internal standard that allows us to track MS sensitivity shifts in real time.

225 2.5 $\delta^{18}\text{O}_2/\text{N}_2$, $\delta^{18}\text{O}-\text{O}_2$, $\delta^{15}\text{N}-\text{N}_2$, $\delta\text{Ar}/\text{N}_2$ measurements

A subset of ALHIC1903 samples were analyzed for $\delta^{15}\text{N}-\text{N}_2$, $\delta^{18}\text{O}-\text{O}_2$, $\delta\text{Ar}/\text{N}_2$ and $\delta\text{O}_2/\text{N}_2$ at Scripps Institution of Oceanography following laboratory methods described in (Severinghaus et al., 2001; Sowers et al., 1989). Air was extracted from a 10-15 g sample using a melt-refreeze method. The sample was then trapped in a stainless-steel submerged in liquid helium. Sample tubes were subsequently warmed to room temperature and analyzed on a Thermo Fischer Scientific Delta V dual-inlet mass spectrometer. Raw $\delta^{18}\text{O}-\text{O}_2$ measurements were corrected for gravitational fractionation and gas loss to derive $\delta^{18}\text{O}_{\text{atm}}$. Here, we use the method and gas loss parameterizations from (Baggenstos et al., 2017), which require inputs of $\delta^{15}\text{N}$



(for gravitational corrections), and $\delta\text{Ar}/\text{N}_2$ and $\delta\text{O}_2/\text{N}_2$ (for gas loss). Gas loss parameters from (Baggenstos et al., 2017) were empirically derived from Taylor Glacier measurements spanning the last 150,000 years. The pooled standard deviation of replicate samples was 0.02‰.

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While a long-term decline of atmospheric O_2 has been detected over the last 800,000 years (Stolper et al., 2016), the effect on $\delta^{18}\text{O}_{\text{atm}}$ over this period is very small, translating to an average shift in $\delta^{18}\text{O}_{\text{atm}}$ of 0.0016‰. Therefore, we assume that atmospheric O_2/N_2 was constant and that measured $\delta\text{O}_2/\text{N}_2$ reflect only fractionations related to firn processes and post-depositional gas loss.

240 3 Chronology development

CH_4 and $\delta^{18}\text{O}_{\text{atm}}$ were used to develop the gas-phase chronology for ALHIC1903 since they are globally well-mixed, and well-resolved records exist from other Antarctic ice cores. Additionally, the gradual change in $\delta^{18}\text{O}_{\text{atm}}$ coupled with the abrupt features of the CH_4 record provide two different and unique constraints for the age model. Tie points (Fig. 3) were manually selected and ages between tie points were linearly interpolated. CH_4 tie points ($n=10$) were chosen by matching either characteristic features in the record (such as abrupt increases and decreases), or in some cases by direct value-matching. Where abrupt CH_4 features did not exist, $\delta^{18}\text{O}_{\text{atm}}$ tie points ($n=7$) were chosen based on value-matching to maintain relatively even spacing between tie points. Ages were synchronized to the Antarctic Ice Core Chronology (AICC) 2023 (Bouchet et al., 2023). The good agreement between ALHIC1903 measurements of CO_2 and N_2O and preexisting ice core records adds confidence to the age model (Fig. 1). Tie point uncertainty includes three contributions that we add in quadrature: 1) uncertainties in the EDC chronology (Bouchet et al., 2023), 2) the impact of analytical uncertainty associated with measurements of CH_4 and $\delta^{18}\text{O}_{\text{atm}}$ from EDC (our reference records) and ALHIC1903 on our ability to value-match these records, and 3) the finite temporal resolution of our record. For item (1), we rely on published AICC2023 uncertainties from Bouchet et al., 2023. For (2), analytical uncertainties were derived by adding the pooled standard deviation of each record in quadrature, and for (3) we derived uncertainty associated with temporal resolution by evaluating the local sample spacing at each tie point.

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For the second source of uncertainty listed above, the $\delta^{18}\text{O}_{\text{atm}}$ pooled standard deviations from EDC (Bouchet et al., 2023, Extier et al., 2018) and ALHIC1903 are 0.03‰ and 0.02‰, respectively. To accommodate the different gas loss corrections made on our $\delta^{18}\text{O}_{\text{atm}}$ record compared to the EDC $\delta^{18}\text{O}_{\text{atm}}$ record, we compared the $\delta^{18}\text{O}_{\text{atm}}$ minimums at ~131.4 ka and added this 0.05‰ difference between the two records as an additional uncertainty. The pooled standard deviation of the CH_4 records from EDC (Schmidely et al, 2021) and ALHIC1903 are 7 ppb and 4.3 ppb, respectively. We translated the combined analytical uncertainty into an age uncertainty by evaluating the derivative of the reference record at each tie point to calculate the tie point uncertainty using Eq. (3):

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$$\sigma_{\text{tie}} = \frac{\sigma_{[\text{gas}]}}{\frac{d_{[\text{gas}]}}{dt}}, \quad (3)$$

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where σ_{tie} is the tie point uncertainty in years, $\sigma_{[\text{gas}]}$ is the tie point uncertainty in ppb (CH_4) or permille ($\delta^{18}\text{O}_{\text{atm}}$) depending on which parameter was used to identify the tie point. Average uncertainty for all tie points ($n=18$) is 824 ± 992 years (Fig. 3). A list of all tie points and associated uncertainties can be found in Supplementary Table 1.

270 The age density is the inverse of the annual layer thickness and is a measure of how compressed an age interval is in a core. Low age densities are commonly preferred for climate reconstruction as they enable higher resolution records. High accumulation rates and small degrees of layer thinning (as found at lower fractional depths in the ice sheet thickness) are both conducive to a larger layer thickness and smaller age densities. Here we compare the age density of the ALHIC1903 ice core to values calculated for other Antarctic cores from Masson-Delmotte et al., 2011. The average age density for ALHIC1903 is 170 yr/m, similar to EPICA Dronning Maude Land (EDML) core and other interior cores like EDC, Vostok, and Dome Fuji (Fig. 4). The other coastal cores (Taylor Dome (TD) and Talos Dome (TALDICE)) exhibit much greater age densities of ~500 and 750 yrs/m, respectively (Fig. 4) (Masson-Delmotte et al., 2011). This is likely due to the greater fractional depth at which this time period is found in the ice core, with generally higher age densities found nearer bedrock. It is remarkable that ALHIC1903 has similar age densities to central Antarctic cores given the relative depth of LIG layers to the total ice thickness. Upstream from ALHIC1903, these LIG layers are found at approximately ~30% of the ice sheet thickness from bedrock (Kehrl et al., 2018), much more similar to EDML and the other interior cores. Thus, we attribute the anomalous position of ALHIC1903 in Figure 4 to surface ice ablation that prevents deep burial at the Allan Hills BIA. It is also possible that the low age density in ALHIC1903 is an apparent one caused by dipping isochrons. Stratigraphic visualization such as electrical conductivity measurements would be needed for further inquiry.

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The ALHIC1903 borehole is 6 m away from S27 (Carter et al., 2026). As a result, we can compare the gas records of the two ice cores to gain further understanding of the stratigraphy of the ice sheet in this area. Plotted versus depth the gas records show good overall agreement, however the abrupt features in the CH_4 record reveal a 3 m offset at the LIG (Fig. 5). The best alignment results from a depth offset that linearly increases toward the surface, from a maximum offset of ~5 m near the surface to no offset around 140 m (Fig. 5 and S2). This observation agrees with layer tracing from previous radar surveys – isochrons nearer the ice surface seem to dip more steeply than those near the bed (Kehrl et al., 2018). This observation places a lower bound of 40° on the dip angle at the surface and a lower bound of 0° at 140 m depth. We applied this depth offset to reconcile the gas records between the two cores. To plot TAC on ice age, we applied the same depth offset to align the two cores and then interpolated the S27 ice chronology (Spaulding et al., 2013) onto ALHIC1903.

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A prior study estimated a glacial-interglacial accumulation rate contrast of at least an order of magnitude at the S27 site (Yan et al., 2021) based on the estimated gas age-ice age difference combined with firn densification modelling. The low accumulation conditions at the Allan Hills BIA fall outside the calibration range of such models, adding uncertainty to the estimates derived using this method. Our gas chronology provides a new constraint on the depth-age scale, which can be used to independently validate the accumulation rate contrast across the glacial termination. We estimate an annual layer thickness for MIS6 (here between 134.8 and 144.2 ka on the gas age scale) of 1.4 mm/yr, and we estimate a peak annual layer thickness for MIS5e (here between 128.9 and 130.8 ka on the gas age scale) of 13.7 mm/yr, or 13.2 mm/yr after correcting for the observed dip angle between S27 and ALHIC1903. This implies a nine-fold increase in the rate of accumulation across Termination II. Note that this is an upper bound given that we do not account for differences in layer thinning experienced by ice in these two depth ranges. For comparison, typical interior Antarctic core sites experience only a doubling of accumulation across terminations. Our estimate independently confirms the large glacial-interglacial contrast identified by Yan et al., 2021.

3.1 Figures

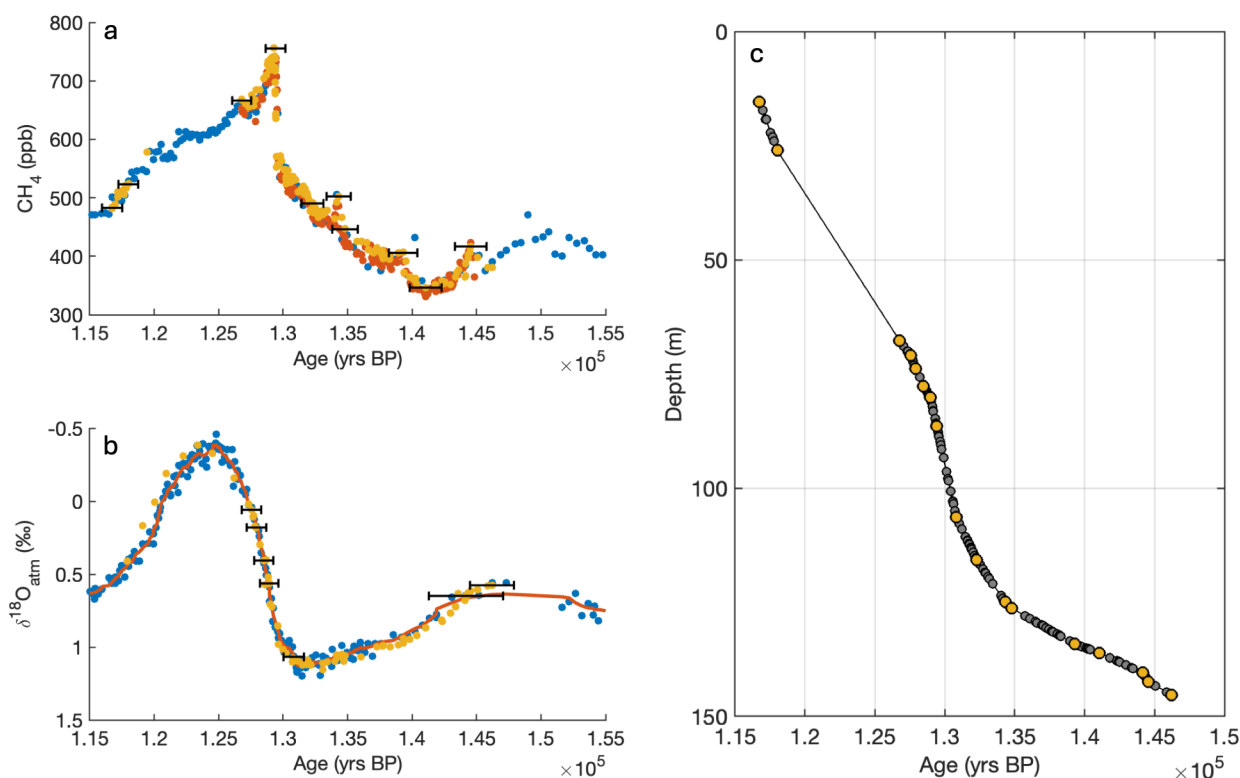
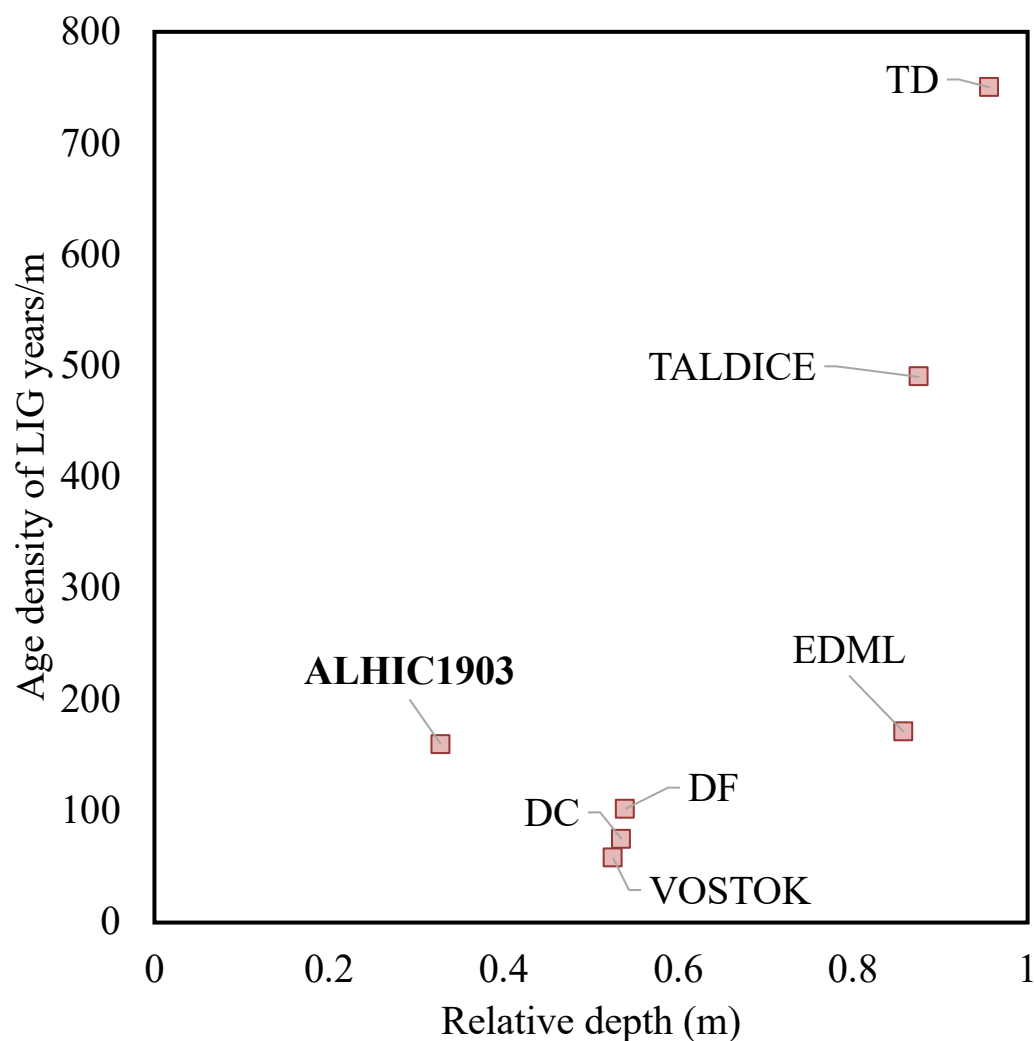


Figure 3: 3a: CH₄ records across T-II from this study (ALHIC1903, yellow) and EPICA Dome C (blue, Louergue et al., 2008; red, Schmidely et al., 2021). Tie points plotted with associated uncertainty in black. 3b: δ¹⁸O_{atm} records across T-II from this study (same as 3a) and EDC (blue, Bouchet et al., 2023; orange, smoothed EDC). 3c: Gas age scale for ALHIC1903 with tie points highlighted as yellow circles. All records are placed on the AICC2023 chronology (Bouchet et al., 2023).



315 **Figure 4: Relative depth of the LIG (referenced to 132.6 ka) where 0 is the surface and 1 is bedrock versus age density (yrs/m) of the LIG in Antarctic ice cores (Vostok, EPICA Dome C (EDC), Dome Fuji (DF), ALHIC1903, EPICA Dronning Maud Land (EDML), Talos Dome (TALDICE), and Taylor Dome (TD). We used an ice thickness of 320 m for ALHIC1903, an estimate provided by Spaulding et al., 2013. Data compiled by Masson-Delmotte et al., 2011, with ALHIC1903 added. All records span 132.6-116 ka).**

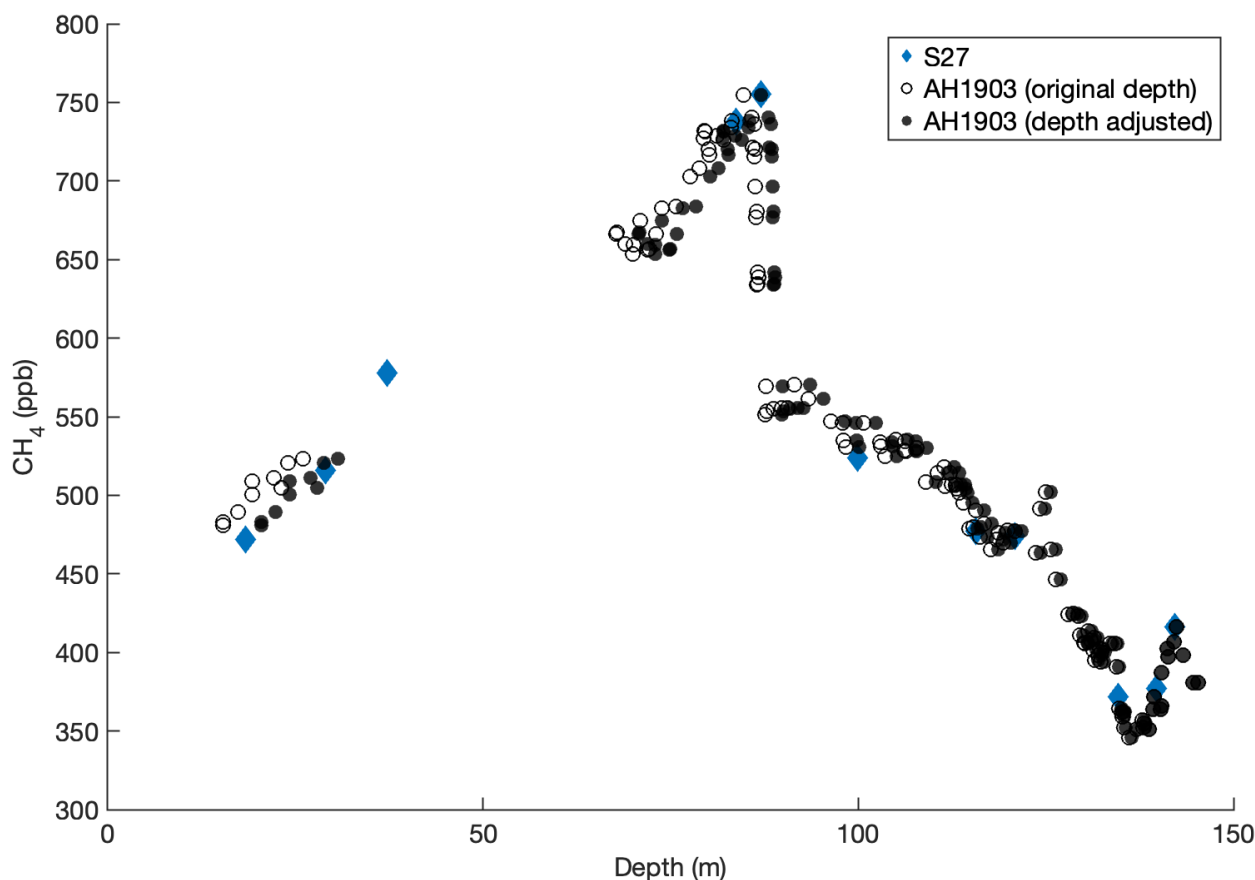


Figure 5: Methane concentrations for the S27 and ALHIC1903 ice cores. S27 CH₄ is plotted on its own depth (blue diamonds) and the ALHIC1903 ice core is plotted on its own depth (open circles) and on an adjusted depth to match S27 (filled circles). The depth adjustment varies from 0 m at ~140 m, to 5.3 m near the surface.

4 Trace gas records from ALHIC1903

4.1 CO₂ and CH₄ records

The ALHIC1903 records of CH₄ and CO₂ show good agreement with other Antarctic ice cores (Fig. 6). In addition to confirming that full glacial cycle variability is preserved for both gases, we also demonstrate that ALHIC1903 can resolve millennial-scale features previously identified in other records across the early penultimate deglaciation. First, our record shows that before the onset of T-II, CO₂ and CH₄ exhibit a small rise at ~139 ka. A recent compilation of calcite $\delta^{18}\text{O}$ from Chinese speleothems recording the intensity of the Asian summer hypothesized this initial CO₂ rise is associated with brief stadial conditions preceding T-II. (Liang et al., 2025).



Second, our record resolves the abrupt increase in CO₂ at beginning of T-II around 137 ka (Fig. 6). This changes is coincident with an increase in the rate of Antarctic warming (Jouzel et al., 2007) and a drop in sea surface temperatures in the North Atlantic (Hodell et al., 2023; Tzedakis et al., 2018), possibly demarcating the onset of Heinrich Stadial 11 (HS11, Fig. 6). As discussed in previous literature, an influx of freshwater into the North Atlantic at this time likely disrupted the AMOC (Men-
 335 et al., 2019), reduced the intensity of the Asian summer monsoon, and increased the strength of southern hemisphere westerlies augmenting the ventilation of carbon-rich waters from the Southern Ocean (Böhm et al., 2015; Deaney et al., 2017; Landais et al., 2013; Liang et al., 2025; Mokeddem and McManus, 2016; Shackleton et al., 2020; Tzedakis et al., 2018).

Third, at 134 ka our record resolves a peak in CH₄ (and N₂O) previously hypothesized as a brief period of AMOC recovery
 340 amidst HS11 (Schmidely et al., 2021). Schmidely et al. note a slight depletion in the calcite δ¹⁸O record from Chinese speleothems at this time which could be due to a strengthening of the Asian summer monsoon linked to a brief reinvigoration of AMOC.

4.2 New constraints on N₂O

Our N₂O measurements across the PGM and early deglacial period appear to be free of excess N₂O that often obfuscates
 345 records from glacial periods with higher dust loads (Fig. S3). Previous N₂O reconstructions exclude N₂O measurements at depths where large scatter is present in adjacent samples and where dust concentrations exceeded 300 ppb (Schilt et al., 2010; Schmidely et al., 2021; Spahni et al., 2005). Though glacial-level measurements from the Talos Dome ice core during MIS 2 were speculated to be free of excess N₂O (Schilt et al., 2010) despite dust concentrations exceeding 300 ppb, the authors concluded that replication from another ice core was necessary to verify absolute N₂O concentrations across this interval. Our
 350 PGM record matches the lowest values of those measured in EDC (Fig. S3) and exhibit no more scatter in adjacent samples than interglacial-level measurements, despite PGM dust concentrations also exceeding the identified threshold (Carter et al., 2026) (Fig. S4). Furthermore, the relationship between dust concentration and N₂O in PGM samples is not statistically significant ($R^2 = 0.08$, $p \gg 0.05$, Fig. S5).

355 Additionally, an interlaboratory comparison of ALHIC1903 N₂O from Oregon State University and University of California, Irvine shows very good agreement (Fig. S3). Thus, we believe ALHIC1903 provides a trustworthy record of atmospheric N₂O through the PGM and early termination. At present, it is not clear why Allan Hills glacial ice is free of *in situ* N₂O production. A previously proposed mechanism for N₂O production in ice is the abiotic reduction of NO₃⁻ to NO₂⁻ via metal ions like Fe²⁺ that are present in glacial dust (Soussaintjean et al., 2025). Possibly the Allan Hills BIA has a different dust composition from
 360 other cores plagued with N₂O artifacts, though we cannot evaluate this hypothesis without measurements of Fe²⁺ from the ALHIC1903 ice core. The relatively close proximity of Allan Hills to Talos Dome may result in similar dust composition at both sites and therefore low excess *in situ* N₂O. Additionally, the low accumulation rate at the Allan Hills results in surface snow having long exposure in the photic zone where UV photochemistry results in removal of nitrate (Mulvaney et al., 1998;



Mulvaney and Wolff, 1994). Reductions in the concentration of the precursor chemicals may further reduce *in situ* N₂O
365 production

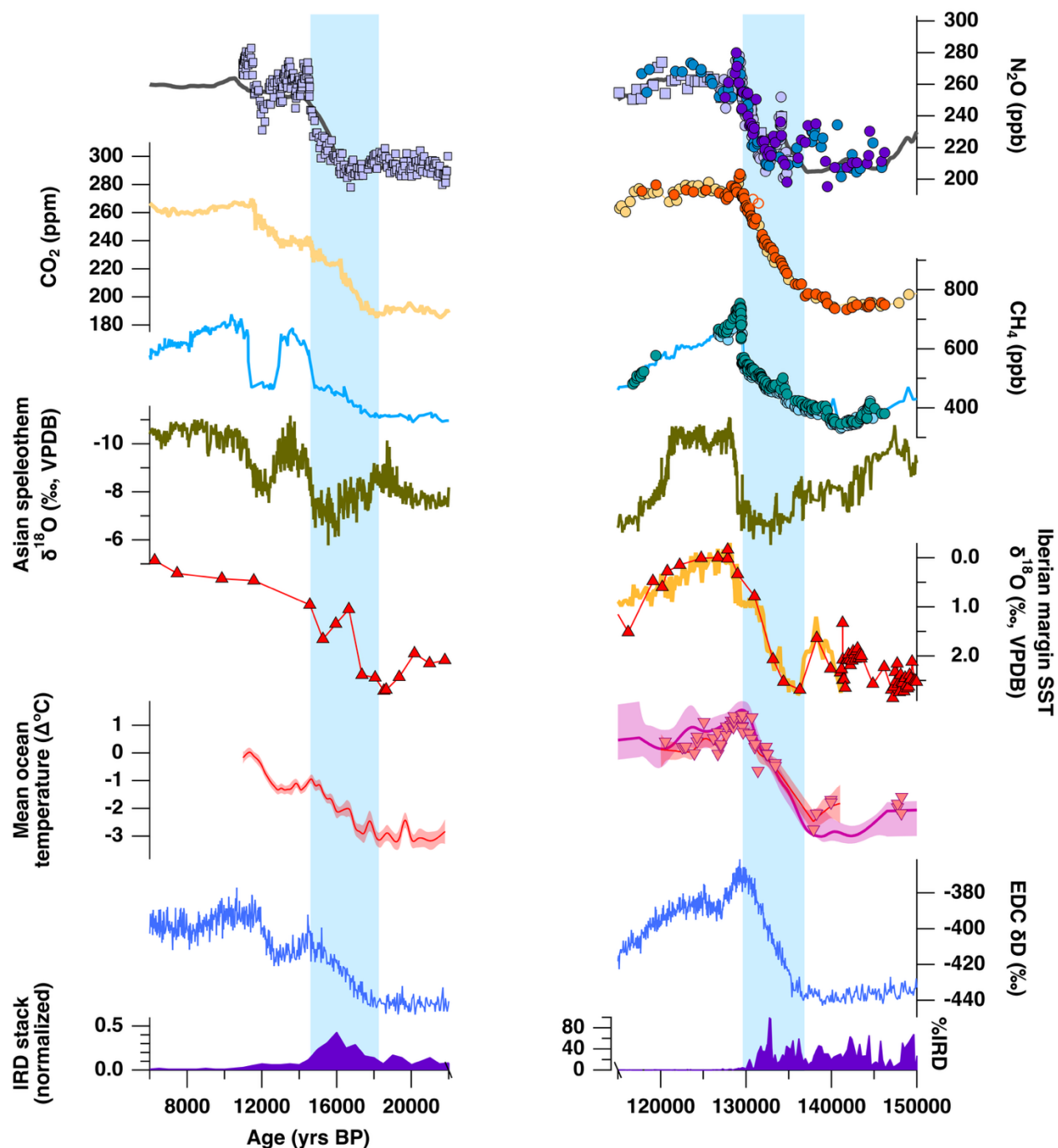
During the PGM, the lowest N₂O concentrations (195-210 ppb) are found from 142-139 ka, very similar to the N₂O
concentrations observed during the Last Glacial Maximum (Schilt et al., 2010) (Fig. 6). These glacial-level concentrations are
also in agreement with an N₂O reconstruction based on machine learning techniques from (Salehnia et al., 2025), though we
370 seem to capture millennial-scale variability that is not found in their record (Fig. 6). Like CO₂ and CH₄, N₂O begins to increase
at 139 ka. However rather than a continuous rise, N₂O exhibits two significant reversals during the early deglaciation, one at
~137 ka and the other at 134 ka. Though the feature at 134 ka has been described by Schmidely et al., 2021, to our knowledge
the N₂O decline at 137 ka has not been examined. Here, we provide a discussion of this prominent feature in the early deglacial
N₂O record of T-II.

375 At 137 ka, N₂O declines by approximately 40 ppb. This feature aligns with the onset of HS11 as discussed in section 4.1 (Fig.
6). A decline in N₂O at the onset of stadial conditions agrees with previous observations from Schilt et al., 2013, who identified
especially low N₂O concentrations during Heinrich stadials over the last glacial period. Simulations from a coupled model
including dynamic biogeochemical cycling showed a link between AMOC variability and atmospheric N₂O (Schmittner and
380 Galbraith, 2008). When AMOC is weakened in the model N₂O concentrations drop, predominantly due to a decrease in
productivity from the equatorial Pacific and Indian oceans. The authors hypothesize the drop in N₂O emissions is linked to
rapid changes in the oxygen cycling of the upper Indo-Pacific oceans. Specifically, they show a weakened AMOC causes an
increase in ventilation and thus dissolved oxygen in the subsurface, reducing denitrification rates. Stott et al., 2002 also
examined the relationship between stadials and a reduction in eastern equatorial Pacific upwelling, hypothesizing that
385 atmospheric N₂O drops during stadials from a disruption of this equatorial marine flux (Stott et al., 2002).

Schmittner and Galbraith, 2008 also show that the amplitude of N₂O change is associated with the duration of AMOC
shutdown. This is similar to a previous observation that the duration of a Dansgaard-Oeschger event correlates to N₂O
amplitude over the last glacial period (Flückiger et al., 2004). Comparing the last two deglaciations reveals similar N₂O
390 variability associated with Heinrich Stadial 1 (HS1) and HS11 (Fig. 6), where a minimum N₂O concentration coincides with
stadial conditions before concentrations begin rising to interglacial-level values (Fig. 6). The larger amplitude of the N₂O
decline at T-II stands in contrast to the smaller decline during T-I and suggests the duration of AMOC disruption associated
with HS11 was larger or longer in duration. According to Schmittner and Galbraith 2008, an AMOC shutdown of >1000 years
is required to produce an N₂O drop as large as HS11.



395 4.3 Figures



(Figure caption on next page)



Figure 6. Climate records of Termination I and II. 6a-c: N_2O , CO_2 , and CH_4 (same as Figure 1). 6d: stacked $\delta^{18}\text{O}$ record of calcite from Asian speleothems (dark yellow, Cheng et al., 2009). 6e: $\delta^{18}\text{O}$ from planktic foraminifera (*G. bulloides*) from the Iberian Margin core U1385 (red triangles, Hodell et al., 2023) and core MD01-2444 (yellow line, Tzedakis et al., 2018). 6f: Two records of mean ocean temperature from Taylor Glacier ice cores (pink spline and upside-down triangles (Bereiter et al., 2018; Shackleton et al., 2020)) and from the EDC ice core (purple spline, (Grimmer et al., 2025)). 6g: EDC record of δD of ice (Jouzel et al., 2007), and 6h: ice-rafted debris records from the North Atlantic (Lisiecki and Stern, 2016; Mokeddem and McManus, 2016). The blue shading corresponds to HS1 (left, (Hodell et al., 2017)) and HS11 (right, (Deaney et al., 2017)).

5 Ice core properties and the depositional setting of Allan Hills ice cores

5.1 Implications for old ice

The good agreement between our new trace gas records and preexisting Antarctic records corroborates previous observations from the S27 ice core that the full range of well-mixed atmospheric gas concentrations and isotopic variations across a glacial cycle are preserved (Spaulding et al., 2013; Yan et al., 2021), ruling out (long) hiatuses in snow accumulation. These observations have implications for interpreting records from nearby (within 10 km) stratigraphically discontinuous ice cores (Fig. 2) containing samples as old as the late Miocene (Higgins et al., 2015; Shackleton et al., 2025; Yan et al., 2019), in which full glacial-interglacial variability has yet to be identified (i.e. the ALHIC1901 core). The ALHIC1903 ice core provides evidence that the full range of glacial-interglacial variability is deposited and recorded upstream in Allan Hills ice cores, but there are numerous reasons why the ALHIC1903 ice core could be an imperfect analogue to the old ice nearby. For example, the ice could have originated from a different source region or the site conditions at the accumulation area (i.e. elevation, vapor transport pathways, wind speed, etc.) could have varied through time. Such differences would be evident in parameters that vary with depositional settings, like total air content (TAC), stable isotopes of ice ($\delta^{18}\text{O}_{\text{ice}}$), and $\delta^{15}\text{N}-\text{N}_2$. We thus examined these measured parameters in ALHIC1903 and ALHIC1901 and compared them to measurements from other Antarctic cores across the T-II interval (or T-I for cores that do not extend through T-II).

For all three parameters, ALHIC1901 and ALHIC1903 measurements plot close to one another relative to other Antarctic ice core sites (Fig. 7). δD values for both cores are slightly heavier than ice from the Antarctic interior (EDC, South Pole, and Dome Fuji) and more depleted than sites from the West Antarctic side of the Ross Ice Shelf (Siple Dome and Roosevelt Island). Both Allan Hills ice core sites exhibit much lower $\delta^{15}\text{N}-\text{N}_2$ values relative to other Antarctic cores, and total air content values are larger than sites from higher elevation (EDC and Dome Fuji) and smaller than the lower elevation RICE core. Results from two tailed t-tests reveal the differences in mean δD and $\delta^{15}\text{N}-\text{N}_2$ between ALHIC1901 and ALHIC1903 are statistically significant at the 1% significance level ($p \ll 0.01$). These differences may be due to differences in conditions between the late Pliocene/early Pleistocene and T-II. Heavier δD values measured in ALHIC1901 relative to ALHIC1903 follow the expected global cooling trend across the Pleistocene (Shackleton et al., 2025), and the lighter $\delta^{15}\text{N}-\text{N}_2$ values in ALHIC1901 may indicate different site conditions at the Allan Hills in these earlier warm periods. For example, smaller $\delta^{15}\text{N}-\text{N}_2$ in ALHIC1901 could arise from a change in wind speed that causes an apparent change in the length of the firn column due to a deepening of



the convective zone (Severinghaus et al., 2010), or lower snow accumulation rates at that site. For TAC, the difference in mean values for the two cores are not statistically significant. Further discussion on TAC can be found in the next section.

5.2 TAC insight into source elevation

435 Though an imperfect proxy, our record of the total air content from ALHIC1903 provides an opportunity to consider the elevation of the depositional site. In addition to barometric pressure, TAC is believed to be controlled by pore volume at bubble close-off, which is influenced by site conditions such as local summer insolation and accumulation rate (Eicher et al., 2016; Epifanio et al., 2023; Raynaud et al., 2007; Vudayagiri et al., 2025). We first examined the orbital and millennial-scale variability in our record. Using the method established by Raynaud et al., 2007, we assessed the relationship between the orbital-scale variation in TAC and local (76.7° South) integrated summer insolation (ISI). We applied a lowpass filter to the TAC record with a cutoff of 5000 years and analyzed the correlation to ISI with an insolation threshold of 380 W/m² (Fig. 8). Though the minimum TAC value aligns with the ISI peak at ~136 ka, as expected, we observe a weak anticorrelation between the two parameters ($R^2 = 0.2$, $p \ll 0.01$), likely because our record is too short to resolve multiple orbital-scale features and because of millennial-scale variability imposed on the orbital-scale trends. Using this correlation, we calculated the TAC predicted from ISI. We then examined the residual variability in TAC by calculating the misfit from this correlation and subtracting it from the mean (Fig. 8). The weak correlation with ISI causes a residual change of +0.003 cm³/g across T-II (Fig. 8).

Previous studies have also identified a positive correlation between TAC and accumulation rate due to its effect on the depth or porosity at which bubble close-off occurs in the firn (Courville et al., 2007; Eicher et al., 2016; Epifanio et al., 2023; Gregory et al., 2014). An accumulation rate time series was calculated for the nearby S27 ice core (Yan et al., 2021), showing a large increase across the deglaciation, from 135-128 ka that we confirm in this study. This is synchronous with the increase in ISI and could explain the large residual increase in TAC after detrended the record for ISI effects (Fig. 8). An increase of +0.003 cm³/g is in the range of the millennial-scale anomalies found at glacial terminations in the TAC record of the EDC ice core (Guilluy et al., 2026) and South Pole ice core (Epifanio et al., 2023).

To examine the source elevation implied by our TAC record, we used the modern spatial calibration between elevation and TAC from Antarctic ice core sites compiled by Martinerie et al., 1992 (Fig. 9). The average elevation predicted from the full TAC record (ranging 146-117 ka) is 2500 ± 200 m (Fig. 9). Since the record produced by Martinerie et al., 1992 was developed from modern data during an interglacial climate state, the best elevation estimate from the ALHIC1903 TAC record likely comes from LIG-aged ice. Additionally, the TAC value at the LIG does not change after detrending using the correlation with ISI (Fig. 8). Limiting our elevation approximation to only include samples from the LIG, we estimate a source elevation of 2400 ± 150 m (Fig. 9). The residual 0.003 cm³/g possibly caused by accumulation rate changes adds another 170 m of elevation uncertainty.



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Another source of uncertainty comes from the fraction of TAC lost from the cut-bubble effect, or the amount of air lost from bubbles cut during subsampling (Martinerie et al., 1990). This correction is usually <3% for polar ice cores (Epifanio et al., 2023; Mitchell et al., 2015)). A recent study of bubble characteristics from the ALHIC1901 ice core shows very low bubble number densities (Stoll et al., 2026) ranging from 35-120 bubbles/cm³ which likely increases the cut-bubble correction. We show the correction ranges from 3.8-5.6% for the ALHIC1901 ice core (see Supplement, Fig. S6). Without information about bubble number density and size from the ALHIC1903 ice core, we use the mean value from ALHIC1901 (4.7%). It should be noted that this is likely an upper limit for the ALHIC1903 ice core since it is less complexly folded than ALHIC1901. A 4.7% addition to our TAC measurements shifts the estimated source elevation at the LIG down to $\sim 2150 \pm 150$ m. This predicted elevation is very similar to the current elevation of the local dome (2160 m).

The range of TAC from the ALHIC1901 ice core is larger and thus produces a larger spread of predicted elevations (Fig. 9). Removing TAC outliers (points that fall beyond 2σ from the mean) narrows this range slightly, producing a predicted elevation of 2550 ± 400 m. Applying the same cut-bubble correction of 4.7% reduces the mean elevation by ~ 250 m. Albeit an imperfect proxy, our TAC measurements imply the elevation of the source region may have remained quite stable across the Pleistocene (Fig. 9).



5.2.1 Figures

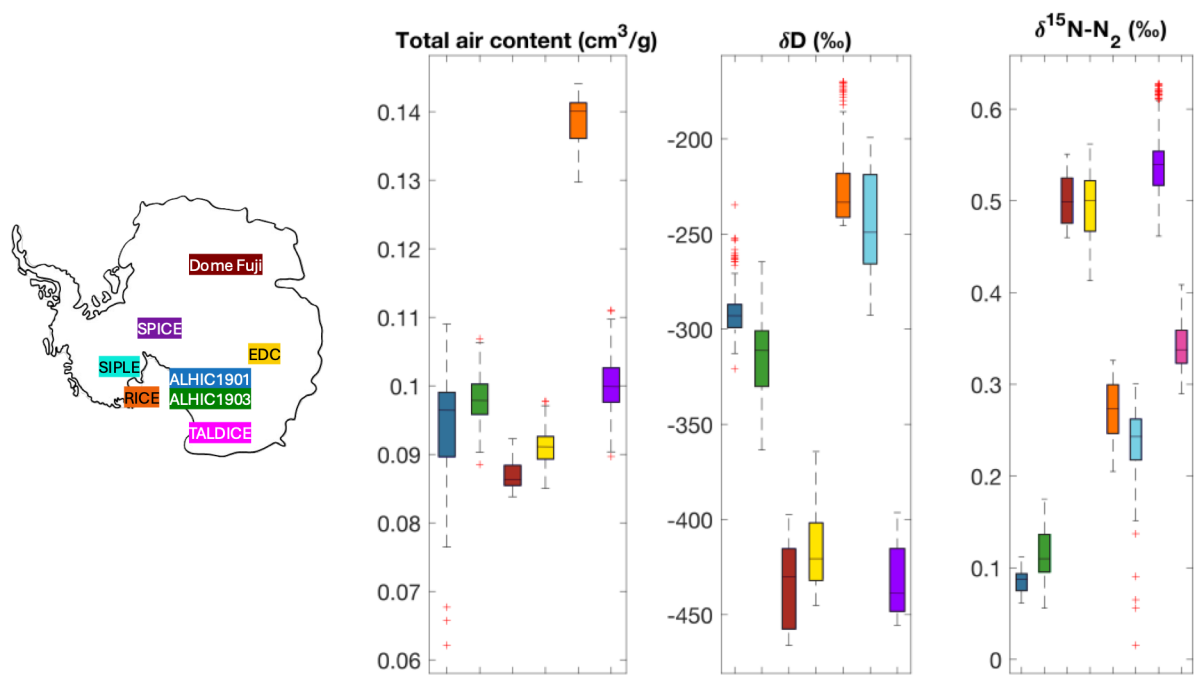


Figure 7. Box-and-whisker plots of TAC, δD of ice, and $\delta^{15}N-N_2$ from various Antarctic ice core sites from the Antarctic interior (EDC (Jouzel et al., 2007; Landais et al., 2013; Raynaud et al., 2007), Dome Fuji (Oyabu et al., 2020; Uemura et al., 2018), South Pole (SPICE) (Epifanio et al., 2020, 2023; Steig et al., 2021)) and the margins (Siple Dome (Brook et al., 2005; Severinghaus et al., 2009), Talos Dome (TALDICE, Masson-Delmotte et al., 2011; Crotti et al., 2021), and Roosevelt Island (RICE, Lee et al., 2020)). Data shown for 147-116 ka for ice cores that span TII (EDC, Dome Fuji, TALDICE) to match the ALHIC1903 age range. For cores that do not extend to TII (RICE, SPICE, Siple), we plot data that spans the last glacial maximum and deglaciation from 23-11.7 ka. Each parameter for ALHIC1903 (this study) plots closely to that of the nearby core containing 4-million-year-old ice (ALHIC1901) relative to other sites.

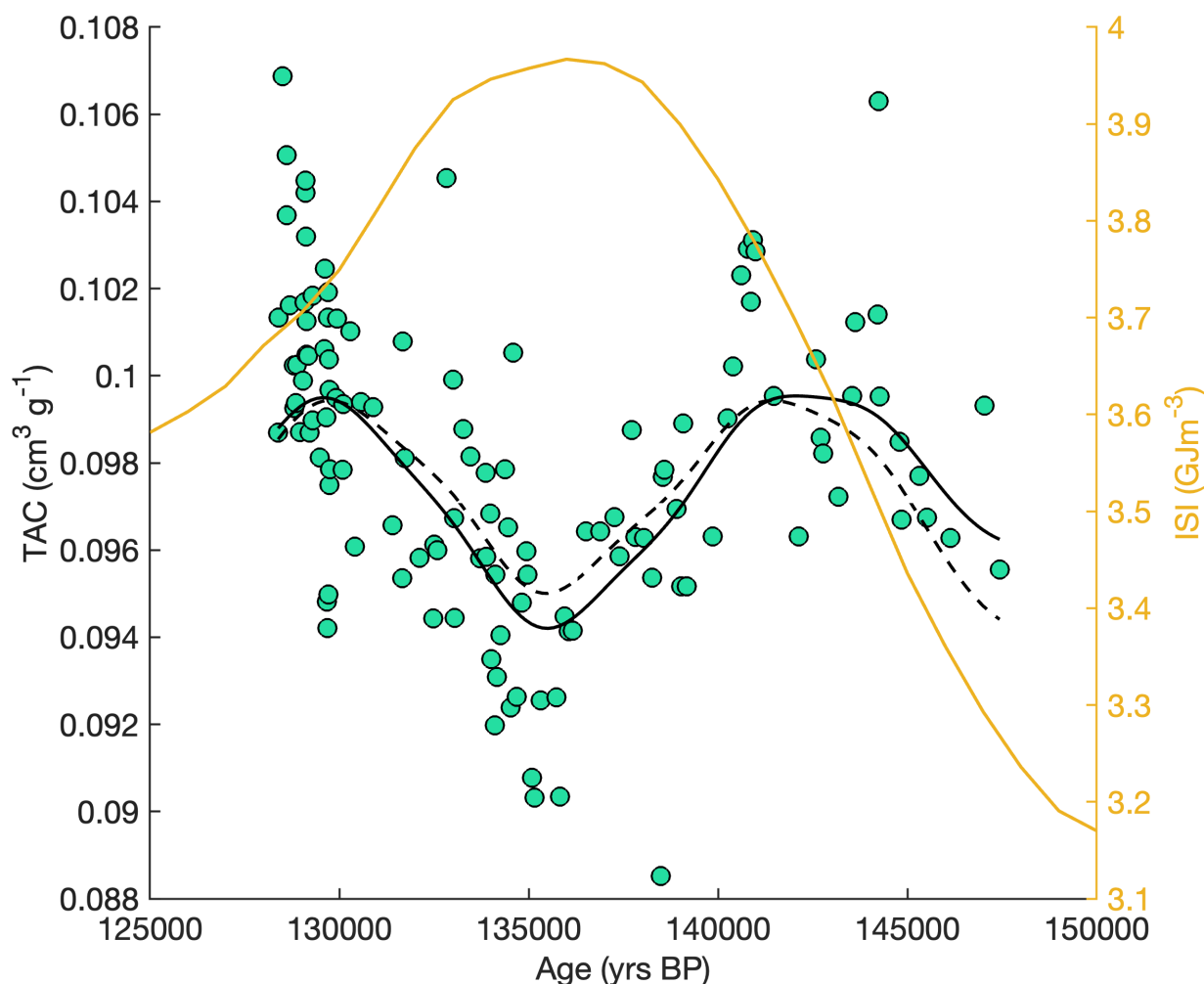


Figure 8. The ALHIC1903 TAC record (green circles) and filtered record using a Butterworth lowpass filter with a cutoff of 5000 years (black line). Dashed black line represents the filtered TAC after removing TAC predicted from the correlation with ISI. Integrated summer insolation (ISI) (yellow line, (Berger and Loutre, 1991)) in units of GJ/m³ for ALHIC1903 drill site latitude and a daily insolation threshold of 380 W/m². TAC is plotted on ice age.

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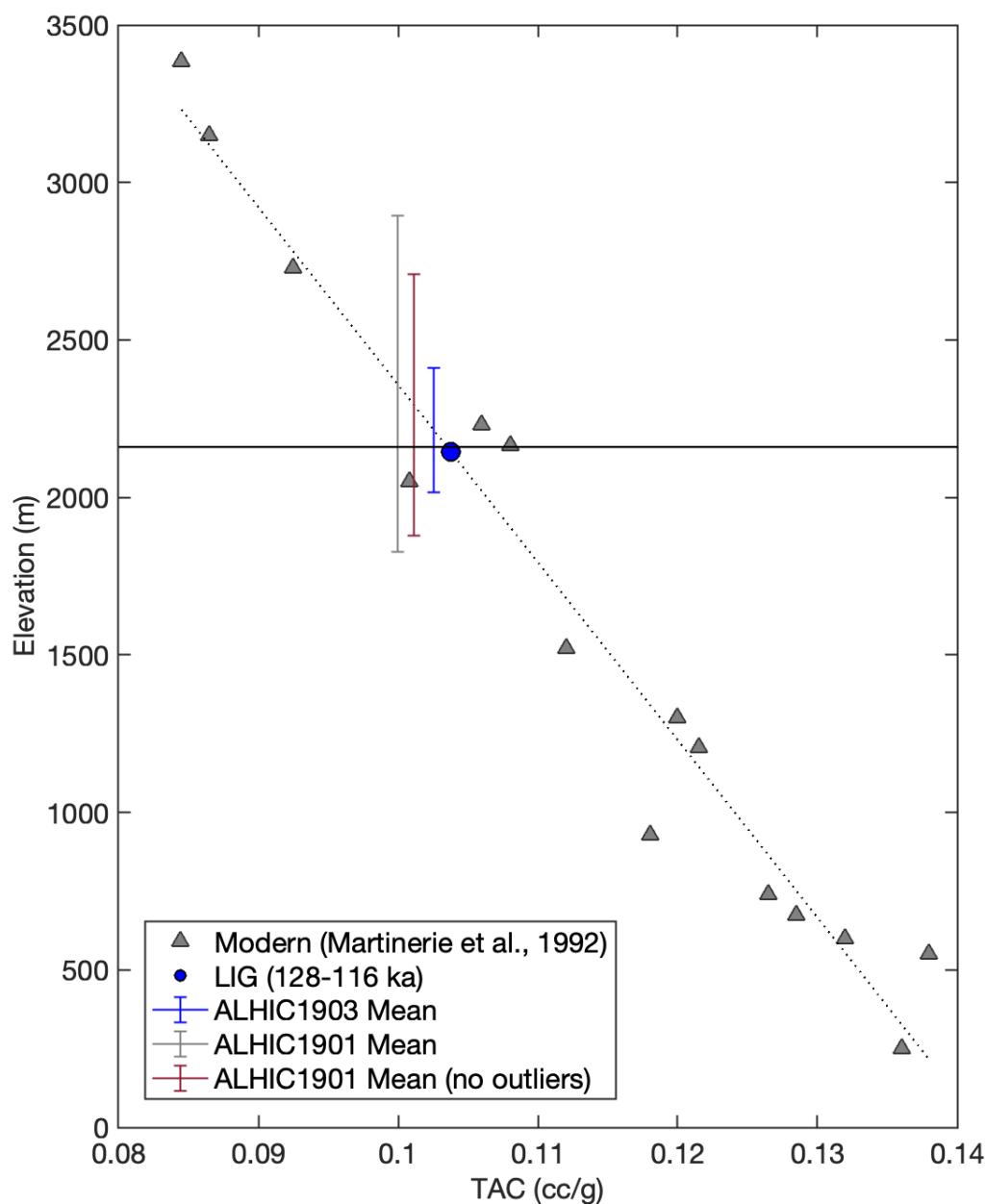


Figure 9. Total air content for modern Antarctic sites plotted against elevation (data taken from Martinerie et al. 1992). Using the relationship between elevation and TAC ($R^2 = 0.97$, $p < 0.01$), and applying a cut-bubble correction of 4.7%, the average LIG value of ALHIC1903 TAC (blue) suggests an elevation of the source region during the LIG was approximately 2150 m. The average TAC from ALHIC1901 (full record in grey, excluding outliers in maroon) suggests an elevation of ~2300 m. Error bars represent one standard deviation from the mean.



6 Conclusion

We show here that a shallow ice core from the Allan Hills Blue Ice Area in Antarctica preserves a pristine record of atmospheric composition for the Penultimate Glacial Maximum to the Last Interglacial Period (146-117 ka). We establish a gas-phase timescale for this core, based on matching atmospheric CH_4 and $\delta^{18}\text{O}_{\text{atm}}$ to well-dated ice core records from well-dated interior cores. We estimate that snow accumulation rates increased by a factor of 9 (upper bound) across T-II. We demonstrate that millennial-scale variability in trace gases can be resolved and that the core preserves pristine glacial-level N_2O concentrations, providing new insight into N_2O variability in the PGM and onset of T-II. Specifically, our record exhibits a substantial drop in N_2O at the onset of HS11, similar to the variability during Termination I and other Heinrich stadials. This new observation reinforces the notion that N_2O is sensitive to AMOC variability. We also investigate the site-specific properties of the ALHIC1903 ice core to further our understanding of its source region. We show that the Pliocene-aged ice in the nearby ALHIC1901 ice core exhibits similar site-specific characteristics to ALHIC1903 relative to other Antarctic cores. Small differences in $\delta^{18}\text{O}_{\text{ice}}$ and $\delta^{15}\text{N-N}_2$ may result from climatic changes across time periods such as local temperatures, vapor transport, and/or wind speeds. Total air content values suggest that the elevation of the depositional setting has been relatively stable through the LIG and possibly the late Pliocene/early Pleistocene. Taken together, our work validates the utility of ALHIC1903 as a stratigraphically intact analog for old ice studies at the Allan Hills. Future use of the ALHIC1903 ice core and the work presented here can help our understanding of signal preservation at the nearby continuous drill site (Kehrl et al., 2018) potentially containing 1-million-year old ice.

Code and data availability

All data will be made publicly available on the US Antarctic Program Data Center upon publication.

Supplement link

The link to the supplement will be included by Copernicus, if applicable.

Author contributions

E.B., J.H., and M.B. designed the research; J.M-P, J.A.E., A.C., J.H., S.S., J.M., J.S., M.A., E.S., S.G., E.R., M.K., R.B., and A.K. collected the ice cores and performed the measurements; J.M-P, E.B., C.B., A.C., M.A., E.R., S.S., J.S.E., and J.A.E. analyzed the data, J.M-P, E.B., C.B., A.C., M.A. prepared the manuscript; and all authors contributed to improving the final paper.



Competing interests

At least one of the (co-)authors is a member of the editorial board of Climate of the Past.

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activities in ice coring and analysis, Antarctic exploration and ice sheet modeling, education, knowledge transfer, and broadening participation. All Center components and participants are essential to its success and contribute to the shared infrastructure that underpins all COLDEX products, including this publication.

Review statement

- 560 The review statement will be added by Copernicus Publications listing the handling editor as well as all contributing referees according to their status anonymous or identified.

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