



Characterization of complex basal layering in an ancient Allan Hills ice core

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Abstract. Ice cores from the Allan Hills blue ice area in East Antarctica contain the oldest known direct samples of the ancient atmosphere, with ages extending through the Pleistocene (0.01–2.58 Ma), Pliocene (2.58–5.33 Ma), and into the late Miocene (>5.33 Ma). These cores are challenging to interpret because their timescales are compressed and discontinuous; understanding sub-meter-scale layering and stratigraphic continuity is therefore critical for robust climate interpretation. To provide insight into the composition and origins of sub-meter-scale layering in Allan Hills ice, we apply a coordinated sampling approach to two sections of the ALHIC1901 ice core, located 3 m and 5 m above the bed and dated to 1.3–1.8 Ma. This study is the first high-resolution characterization of ice of this age in the Allan Hills, and so provides critical insight into the composition and variability within this ice. First, we use three-dimensional electrical conductivity measurements (3D ECM) to identify planar and steeply dipping layers across the large-diameter (241 mm) cores. We then construct a common dip-adjusted depth scale and apply it to measurements of trace elements, ions, particle size and concentration, greenhouse gases, major gas (N₂, O₂) isotope and elemental ratios, water-isotope ratios, and ⁴⁰Ar_{atm} ages. Replicate measurements from different positions within the core cross section agree closely once aligned on the dip-adjusted depth scale, indicating coherent planar layering. Furthermore,



$^{40}\text{Ar}_{\text{atm}}$ dating demonstrates that the layers are consistent with isochronal layering at dating precision. Age differences can exceed 100 kyr across decimeter-scale layering. While some climate signal is preserved across layers in water isotope ratios, we conclude that the primary source of variability across most measurements in this near-bed Allan Hills ice is incorporated basal material. Multiple layers enriched in basal material exhibit distinct properties, suggesting that basal material is incorporated into multiple discrete ice packets rather than a single homogeneous unit. This study demonstrates a novel approach—using high-resolution, multi-proxy measurements to characterize steeply dipping layering—with broad applicability to future work at the Allan Hills and other ice core sites.

1 Introduction

Ice cores provide valuable records of Earth's past atmospheric composition alongside other paleoclimate proxies (Chappellaz et al., 1993; Petit et al., 1999; EPICA community members, 2004). These archives have been central to understanding the relationship between atmospheric greenhouse gases and climate through the late Quaternary (Petit et al., 1999; Lüthi et al., 2008; Kawamura et al., 2017). Recent work has begun to extend direct atmospheric measurements through the Pleistocene (0.01–2.58 million years before present, or Ma) and into the Pliocene (2.58–5.33 Ma), opening new opportunities to study Earth-system behavior during climate shifts including the mid-Pleistocene transition and in the warmer Pliocene climate (Westhoff et al., 2026; Higgins et al., 2015; Yan et al., 2019; Shackleton et al., 2025).

Blue ice areas (BIAs) are a promising frontier in the search for old ice. BIAs are regions of exposed glacier ice resulting from negative surface mass balance, where sublimation rates exceed snow accumulation. The Allan Hills BIA (76.7° S, 159.3° E; 1900–2000 m above sea level) is among the most studied BIAs, with extensive shallow ice core drilling (up to 327 m depth) conducted over the past decade. While some Allan Hills cores preserve stratigraphically continuous records in younger (90–250 ka) ice (Spaulding et al., 2013; Yan et al., 2021b; Carter et al., 2026), an increasing number of cores located outside of the main ice flow exhibit complex, discontinuous depth-age relationships, containing packets of ice up to 6 Ma, substantially older than existing 800-kyr continuous records (Higgins et al., 2015; Yan et al., 2019; Shackleton et al., 2025). These cores have enabled new constraints on past climate, including greenhouse gas concentrations (Yan et al., 2019; Marks-Peterson et al., 2026) and mean ocean temperature to 3 Ma (Shackleton et al., 2026), and local temperature extending to 6 Ma (Shackleton et al., 2025).

1.1 Interpreting climate signals in disturbed stratigraphy

Allan Hills cores provide valuable insight into climate evolution since 6 Ma and refinements of processes and methods tailored for discontinuous stratigraphy will further improve interpretations. Traditional dating approaches, including annual layer counting, ice thinning models, and orbital tuning, rely on continuous chronologies and are therefore not applicable to most Allan Hills cores. Instead, these cores are dated using the $^{40}\text{Ar}_{\text{atm}}$ geochronometer, which exploits the monotonic increase in atmospheric ^{40}Ar due to the decay of geologic ^{40}K (Bender et al., 2008). This method requires large (>500 g) ice volumes and yields discrete gas age constraints on individual samples rather than a continuous depth–age relationship.



45 A primary challenge to interpreting Allan Hills ice is aligning samples on a common age scale. In conventional ice cores, parallel high-resolution measurements represent the same age at a given depth (accounting for Δage , the age difference between ice and enclosed gases). However, at the Allan Hills, a combination of complex bed topography, ice flow, and deformation (Nesbitt and Brook, 2023; Stoll et al., 2026) means horizontal layering cannot be assumed. In shallow ice cores, layers have been observed to dip up to 69° (Kirkpatrick et al., 2025), which can cause substantial stratigraphic offset in wide-diameter
50 cores. Therefore, in Allan Hills cores obtained using the 241-mm diameter “blue ice drill” (BID) (Kuhl et al., 2014), a single layer can be offset by more than 60 cm in depth across the width of the core (Kirkpatrick et al., 2025), and samples taken at the same depth might represent dramatically different ages, and so climate states. The standard approach has been to co-locate $^{40}\text{Ar}_{\text{atm}}$ samples alongside other measurements, and assign the $^{40}\text{Ar}_{\text{atm}}$ age to the proxy measurements on adjacent samples. While minimizing the distance between samples minimizes potential offset from dipping layers, steeply dipping layers still
55 introduce uncertainty when assigning sample ages.

A second challenge is uncertainty in the degree of temporal averaging within individual Allan Hills samples. A single $^{40}\text{Ar}_{\text{atm}}$ measurement may represent either an average across one continuous time interval or between non-consecutive periods. Prior Allan Hills studies have observed reduced variability in the older, deeper ice, and therefore describe older samples as “long snapshots” which exhibit reduced variability relative to “short snapshots” in younger ice (Marks-Peterson et al., 2026;
60 Shackleton et al., 2025). The absence of clear glacial–interglacial variability in the older samples suggests smoothing over timescales exceeding 40 kyr in the “long snapshot” regime. We identify two potential mechanisms for smoothing in Allan Hills ice. The first, likely the primary driver of reduced variability in “long snapshot” ice, is smoothing within packets of similar ages. This is likely driven by extreme thinning within the packet, with either diffusion within the packet or large-volume samples averaging across preserved cm-scale variability also contributing. The second is smoothing between packets of very different
65 ages. While not a major concern in stratigraphically continuous ice cores, Allan Hills cores exhibit discontinuous depth–age relationships with abrupt transitions exceeding 100 kyr within a single meter. Again, this averaging might be driven either by a diffusional or physical mixing process, or by large-volume samples capturing both sides of a transition. Distinguishing between these mechanisms requires high-resolution measurements that can resolve variability both within individual ice packets and across abrupt stratigraphic transitions.

70 A third challenge for the interpretation of Allan Hills ice is post-depositional alteration of the greenhouse gas record in near-basal ice. Elevated CO_2 concentrations in many of the deepest and oldest Allan Hills samples have been attributed to post-depositional metabolic production (Marks-Peterson et al., 2026). Measurements of $\delta^{13}\text{C}-\text{CO}_2$ (Marks-Peterson et al., 2026) and $\Delta^{17}\text{O}-\text{O}_2$ (Banerjee et al., 2026) enable reconstruction of atmospheric CO_2 even in ice with elevated CO_2 concentrations, but have higher uncertainties than direct CO_2 measurements. The association of elevated CO_2 and depleted $\delta^{13}\text{C}-\text{CO}_2$ with
75 proximity to bedrock suggests a basal source involving oxidation of organic carbon. Improved understanding of where basal material is incorporated into the ice column, and how this, in turn, influences the gas composition of the ice, is therefore highly relevant to key paleoclimate records.



1.2 Understanding sub-meter-scale variability

Addressing the challenges posed by the unusual characteristics of Allan Hills ice requires coordinated, high-resolution sam-
80 pling across multiple proxies on a shared depth scale. Existing ice core sampling strategies are insufficient to accomplish this
goal in cores with steeply dipping layers. In this study, we develop a method to integrate constraints on layer geometry into
sample cutting and depth tracking, enabling the accurate co-location of parallel datasets even in cores with steeply dipping
layers. We then apply this method to characterize layering in two high-quality sections of the ALHIC1901 core. We use three-
dimensional electrical conductivity measurements (3D ECM) to identify steeply dipping layers, and integrate water-isotope,
85 chemistry, particle, greenhouse-gas, major-gas, and $^{40}\text{Ar}_{\text{atm}}$ age measurements into a novel dip-adjusted depth framework.

2 Methods

Below we discuss a high-resolution sampling approach facilitated by 3D ECM. We first describe the ice core samples them-
selves (Sect. 2.1), followed by the 3D ECM method for determining layer structure (Sect. 2.2), and a dip-adjusted coordinate
90 nate system which we use to accurately align samples across the core relative to layering (Sect. 2.3). Within the context of this coor-
dinate system, we provide a summary of the sampling strategy for each measurement (Sect. 2.3.1–2.3.5). Finally, we describe a
wide range of analytical methods used to measure water stable isotope ratios, particles, major ions, trace elements, greenhouse
gases, major gas isotopes and elemental ratios, and $^{40}\text{Ar}_{\text{atm}}$ (Sect. 2.4.1–2.4.5).

2.1 ALHIC1901 Sections 228_4 and 230_4

This study focuses on two sections of the ALHIC1901 core (76.732376 °S, 159.356125 °E). ALHIC1901, drilled with the
95 241-mm diameter BID system in the 2019–2020 field season, is the most heavily sampled Allan Hills core to date (Shackleton
et al., 2026, 2025; Marks-Peterson et al., 2026; Choi et al., 2026; Stoll et al., 2026). The oldest dated sample from the core
is approximately 4 Ma, and the core demonstrates a discontinuous age-depth relationship, with multiple age inversions. In
this study, we focus on two sections 5 m and 3 m above the bed, referred to as sections 228_4 and 230_4, respectively. The
first number of the section ID describes the drilling run and the second gives the position of this sample within the sections
100 recovered on that run. Each section is approximately 300 mm in length.

The selected sections are the two longest minimally fractured intervals within the depth range where ice older than 1.5 Ma
has been identified in ALHIC1901, and so provide a unique opportunity to combine multiple sets of high-resolution samples
to constrain sub-meter-scale variability. Poor core quality has been the primary limiting factor preventing the development
of continuous, high-resolution measurements at the Allan Hills. While core quality in shallow ice ($< \sim 80$ m) is often good
105 (Kirkpatrick et al., 2025; Hudak et al., 2026), deeper ice recovered by dry drilling systems like the BID often consists of short
(~ 300 mm) intact sections interspersed with fractured ice and rubble. Notably, recent efforts using drilling fluid in the 2025–
2026 field season improved core quality considerably, and may enable more extensive continuous high-resolution sampling on
future cores.

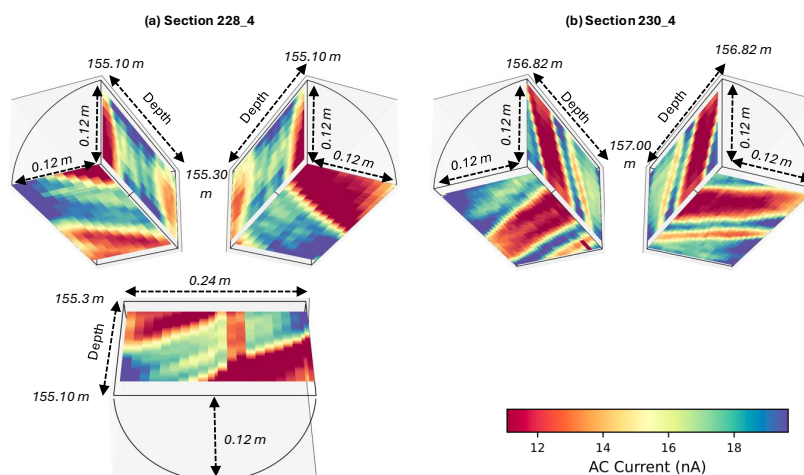


Figure 1. To-scale depiction of 3D AC ECM measurements on ALHIC1901 sections 228_4 and 230_4. Blue colors indicate high conductivity, and red colors indicate lower conductivity. On section 228_4, 3D ECM measurements were completed on both faces of two adjacent quarter-core cuts and the face of the opposite half-core cut. On section 230_4, 3D ECM measurements were limited to just two quarter-core cuts. Note the strongly dipping layering, which is consistent across opposite sides of the same cut.

2.2 3D electrical conductivity measurements

Electrical conductivity measurements in three dimensions (3D ECM) can be used to constrain layer geometries in large-diameter ice cores (Kirkpatrick et al., 2025). This method requires completing multitrack ECM measurements on two perpendicular planes to provide a 3D picture of layer geometry. In this study, ECM measurements are completed along the full width of a half core. This half core is then quartered, and the two resulting half-width faces are both analyzed (Fig. 1). ECM tracks on each face are separated by 5 or 10 mm. This results in a total of 9 tracks on the half-width faces, 20 tracks on the full-width face of section 228_4 (see Fig. 1), and 31 tracks on the full-width face of section 230_4. Both alternating current (AC, 100 Hz) and direct current (DC) ECM scans are completed, but as the ice is relatively homogeneous to DC ECM, only the AC ECM data are used in layer characterization. The computational determination of layer orientation is described in Kirkpatrick et al. (2025). Briefly, the apparent dip angle on a single face is determined by finding the maximum correlation coefficient between pairs of ECM tracks across a wide range of potential dip angles. The angle is defined relative to the core axis, such that a dip angle of 0° corresponds to a layer that is parallel with the geoid. Here, the method is modified slightly to accommodate the short length of these core sections. Rather than finding a maximum correlation coefficient across all possible pairs of ECM tracks, only pairs of adjacent tracks are aligned. This ensures sufficient overlap between the pair of tracks, even at steep dip angles. While only the apparent dip in both measured planes is necessary for adjusting sample depths, we also compute the true dip of the layering to inform understanding of the ice stratigraphy.



125 2.3 Ice core sampling, 3D dimension tracking, and dip adjustment

Traditionally, the position of a sample within an ice core is referenced only in the depth dimension, which is sufficient when layering is near-horizontal. We refer to this depth as the “core depth”. We map the position of a sample in 3D, and determine the depth at which the layer containing a sample intersects the central axis of the core (Fig. 2); we refer to this as the “dip-adjusted depth”. This framework accounts for non-horizontal layering, enabling the comparison of different sets of samples at different positions within the core on a shared depth scale.

We adopt a Cartesian coordinate system, where the location of each corner of the sample in the cross-core plane is recorded as a distance from the center of the core in the x- and y-directions (Fig. 3). For each sample, an “effective center” is calculated, which is simply the volume-weighted average x- and y-coordinates for the cross-section of each sample. With the apparent dip in two dimensions calculated from ECM, it is possible to calculate the dip-adjusted depth of an individual sample. This calculation is described in Eq. (1), where x_{eff} and y_{eff} represent the effective center coordinates of a given sample; α_x and α_y represent the dip angle in the x and y planes; d_{adj} is the dip-adjusted depth; and d is the true core depth of the sample. We compute d_{adj} for the top, center, and bottom depths of all samples.

$$d_{\text{adj}} = d + x_{\text{eff}} \cdot \tan(\alpha_x) + y_{\text{eff}} \cdot \tan(\alpha_y) \quad (1)$$

2.3.1 Water-isotope samples

140 Measurements of water stable isotope ratios (hereafter “water isotopes” for brevity) require only small sample volumes (≤ 1 g) and are not sensitive to gas loss or fracturing at the outer edge of the core. Therefore, it is possible to make replicate sets of 1-cm resolution measurements to gain relevant paleoclimate information and validate the 3D ECM layer orientations. Multiple sets of high-resolution water-isotope data (triplicate in section 228_4 and duplicate in section 230_4) were designed to be offset in both the x- and y-dimensions in both cut plans (Fig. 3 and 4).

145 2.3.2 Particle and chemistry samples

Sampling strategies for chemistry and particle content vary between the two sections. The cut plan for section 228_4 included duplicate major-ion chemistry and particle-size sticks and a single trace-element stick, each decontaminated and sub-sampled discretely. As with water isotopes above, this plan was designed to evaluate the ability of 3D ECM to align parallel datasets even across steeply dipping layering. In section 230_4, two parallel sticks were cut for major ions, and one of these was also measured for trace elements. Both were melted on the Dartmouth continuous-flow analysis (CFA) system (Osterberg et al., 2006) and sub-sampled using a fraction collector. A third stick was sub-sampled discretely for particle-size measurements.

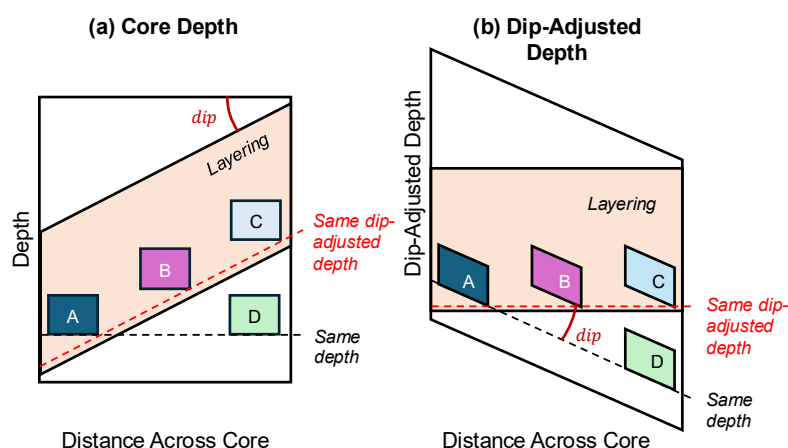


Figure 2. In our dip-adjusted depth framework, the top and bottom core depths of a sample are adjusted to reflect the impact of dipping layers on the position of a sample relative to the center axis of the core. This illustration represents the flat face of a half-core section, where depth is on the y-axis and distance across the width of the core is on the x-axis. In a), the true positions of four samples (A–D) are shown, along with a dipping layer in orange. Here, note that samples A and D share the same core depth, even though their relative position to layering is very different. In b), the samples and layering are shown in a dip-adjusted framework. Now samples A, B, and C, which contain the same stratigraphy, are at the same dip-adjusted depth.

2.3.3 Greenhouse gas samples

Both sections share a similar greenhouse gas sampling plan, with one stick for CO₂ and one for CH₄. In section 228_4, the cut plan resulted in a CH₄ stick which was 50 mm wide in the steeply dipping x-plane; the cuts between individual samples were made at an angle matching the dip, minimizing averaging across layering. Meltwater from section 230_4 CH₄ samples was retained, and used to provide a third set of water isotope measurements for this section.

2.3.4 Major gas isotope and elemental ratio samples

We cut a single stick in section 228_4 for major gas isotope and elemental ratio sampling. Insufficient remaining ice volume in section 230_4 prevented similar analysis in this section.

160 2.3.5 Argon-40 dating samples

Achieving large sample volumes (>500 g) for ⁴⁰Ar_{atm} age measurements while minimizing sampling across multiple dipping layers required a complex cut plan, rather than the straight sticks used for other sampling. Due to limited remaining ice in section 230_4, ⁴⁰Ar_{atm} work was possible only on section 228_4. We cut a total of six samples: two each from above, inside, and below the pronounced layer of high AC ECM conductivity (Fig. 5). Each cut was made parallel to the observed dip angle

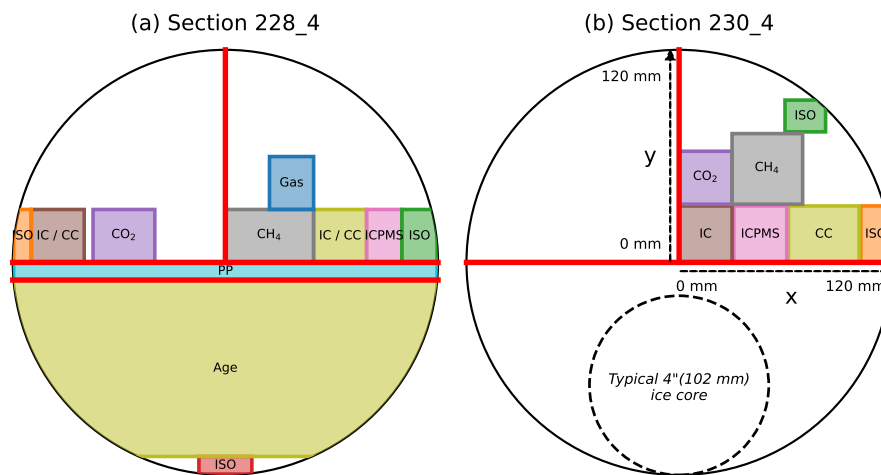


Figure 3. The position of all samples in this study in the cross-core dimension, where up (shallower depth) is defined as coming out of the page toward the reader. Here “ISO” refers to water-isotope samples, “IC” to ion chromatography, “CC” to Coulter Counter particle measurements, “ICPMS” to trace element measurements with inductively coupled plasma-mass spectrometry, “PP” to physical properties (not included in this study, but discussed in Stoll et al. (2026) and future work), “CO₂” to CO₂, “CH₄” to CH₄ and total air content, “Gas” to major gas isotopes and elemental ratios, and “Age” to argon age measurements. Red lines indicate the location of 3D ECM measurements. The x and y dimensions of the Cartesian coordinate system are shown in section 230_4. Each cut is assigned its own color, and this color is maintained in subsequent figures (Fig. 4, Fig. 5, Fig. 6, and Fig. 8).

165 in the steeply dipping x-dimension (Fig. 1). However, the cuts did not match layer dip in the y-dimension, as it is difficult to cut at non-perpendicular angles in two dimensions simultaneously. This is acceptable because the y-dimension dip angle in section 228_4 is moderate, and results in only minimal sampling across layers.

Accurately tracking the large and irregular volume of these ⁴⁰Ar_{atm} samples requires a more sophisticated approach than the effective-center method described above for use on small-volume samples. We begin by computing the 3D extent of each sample. Each bandsaw cut is a plane, and so can be described by a 3D planar equation. Likewise, the outer diameter of the core can be described by a cylindrical equation. We determined coefficients for each equation. Each ice sample can then be bounded by the equations corresponding to each cut. In a Monte-Carlo approach, we randomly place many ($N = 10^8$) points in a regular cuboid bounded by the outer limits of the core section. This cloud of points is filtered using the equations describing the boundary of the sample, such that only points which fall within a given sample remain. These discrete points can then be shifted by Eq. (1), to account for dip-adjustment. The set of shifted points provides a distribution of the relative weighting of the sample over dip-adjusted depth, from which we compute a weighted-average depth. We validated the calculated positions



by comparing the measured water-isotope composition of each $^{40}\text{Ar}_{\text{atm}}$ sample with the high-resolution water-isotope samples (Sect. 3.1 and Fig. 5). Notably, this method with cuts following layer dip requires significantly more time than conventional sample preparation. The 6 $^{40}\text{Ar}_{\text{atm}}$ samples in this study required a full day to cut, as opposed to only approximately 2 hours for traditional cutting.

2.4 Analytical techniques

The analytical methods applied in this study largely follow prior work. A summary of each method is included below.

2.4.1 Water-isotope analytical methods

Stable water isotope ratios were analyzed at the University of Washington using either a Picarro L2130-i or a L2140-i cavity-
 185 ring-down laser spectrometer following Steig et al. (2014) and Schauer et al. (2016). Analyses include four reference waters measured after every 10th sample. Six total injections were made for each sample and reference water vial, with the average of the last three injections used as the measurement outcome. The reference waters (Table A1) cover a range of Antarctic values and are created from excess Holocene and glacial ice from West Antarctic and South Pole cores (Steig et al., 2013, 2021). These reference waters were previously calibrated to the VSMOW-SLAP (Vienna Standard Mean Ocean Water - Standard
 190 Light Antarctic Precipitation) water-isotope scale (Steig et al., 2014). Sample calibrations were based on a three-point linear regression, with the fourth reference water treated as an unknown and used for quality control. Data are reported as deviations (δ) in ‰ from VSMOW. Precision (1 standard deviation) of individual measurements is $<0.5\text{‰}$ for δD and $<0.05\text{‰}$ for $\delta^{18}\text{O}_{\text{ice}}$. Accuracy (1 standard deviation) relative to the VSMOW-SLAP scale is $<0.6\text{‰}$ for δD and $<0.1\text{‰}$ for $\delta^{18}\text{O}_{\text{ice}}$, based on the quality-control reference water.

2.4.2 Chemistry and particle analytical methods

All samples for impurity geochemistry were prepared and decontaminated at Dartmouth College. Three sticks from Section 228_4 were sub-sampled discretely (in 3–4 cm increments) and cleaned inside a HEPA-filtered laminar flow hood in a freezer. For the two ice sticks designated for ion chromatography (IC) and Coulter Counter measurements, outer layers (~ 4 mm) were scraped from each side of every sample using a ceramic blade, and all samples were subsequently rinsed with Optima-grade
 200 methanol. Samples were kept frozen and transported to Colby College, where each was rinsed with MQ water immediately prior to melting. Immediately following melting, a 10-mL aliquot for Coulter Counter measurements was taken and diluted with 10 mL of electrolyte solution, and the remaining (undiluted) meltwater was collected and refrozen for subsequent IC analysis. Coulter Counter measurements were made immediately following dilution using a Multisizer 3 with a $50\text{ }\mu\text{m}$ aperture. The stick designated for inductively coupled plasma mass spectrometry (ICP-MS) measurements was sub-sampled similarly, except
 205 that the cutting board and ceramic knives were pre-cleaned with HNO_3 . ICP-MS samples were melted immediately following decontamination.



One stick from section 230_4 was prepared identically to the 228_4 IC/Coulter Counter stick, although only Coulter Counter particle distributions were measured for this section. Two additional sticks from section 230_4 were melted using the Dartmouth CFA system, which separates pristine interior meltwater from potentially contaminated exterior surfaces (Osterberg et al., 2006). For the first CFA stick, two sets of meltwater fractions were collected at 1.5 cm-resolution for ICP-MS and IC analysis, respectively. For the second CFA stick, meltwater was collected at ~0.9-cm resolution for IC analysis only. Due to challenges associated with melting and measuring very coarse particles, continuous measurements of particle counts and conductivity were occasionally interrupted and contain gaps and unrealistic variability; these data are not discussed further here.

We additionally measured major ions and trace elements on ALHIC1903 (76.702435 °S, 159.310603 °E), a core taken near the main ice flow area that contains a continuous chronology spanning the MIS 6–5e transition (Spaulding et al., 2013; Carter et al., 2026). These samples provide glacial and interglacial baselines for comparison with the ALHIC1901 data. The major ion samples were prepared and cleaned discretely at Dartmouth College following a similar method to the IC/Coulter Counter preparation of 228_4 samples. The trace element samples were melted and decontaminated using the Dartmouth CFA similarly to the 230_4 samples.

For all ICP-MS samples, 1% (v/v) HNO₃ was added and allowed to leach for six weeks prior to analysis following established practice (Koffman et al., 2014). Measurements were made at the University of Maine using a Thermo Scientific Element XR High Resolution Inductively Coupled Plasma Mass Spectrometer (HR-ICP-MS) equipped with the jet interface, using a nickel jet sample cone and a nickel X skimmer cone to maximize sensitivity. An Elemental Scientific Apex 2 system with a PFA-ST-Apex nebulizer (self-aspirated at 100 μL min⁻¹) was used for sample introduction.

Major ions on section 228_4 were analyzed at the University of Maine using a Thermo Scientific Dionex Ion Chromatograph ICS-6000 system with suppressed conductivity detection. Anions (Cl⁻, NO₃⁻, SO₄²⁻) were measured using a Dionex IonPac AS-11 column with a 100 μL sample loop and 12 mM KOH eluent. Cations (Ca²⁺, Na⁺, Mg²⁺, K⁺) were measured using a Dionex IonPac CS12A column with a 100 μL sample loop and 20 mM methanesulfonic acid eluent. Two ion chromatographs were paired with a Dionex AS-HV autosampler for simultaneous anion and cation analysis. Calibration curves bracketed the expected concentration range and were validated using Environment Canada's ION-92 standard diluted to fall within the calibration range.

Major ions on section 230_4, as well as ALHIC1903, were measured at South Dakota State University using ion chromatography. Anion analysis (Cl⁻, NO₃⁻, SO₄²⁻) was performed on a Dionex ICS-1100 system equipped with an IonPac AS11 column using 4 mM NaOH eluent and an ADRS 600 suppressor. Similarly, cation analysis (Na⁺, K⁺, Mg²⁺, Ca²⁺) was performed on a Dionex ICS-1500 system equipped with an IonPac CS12a column using 11 mM H₂SO₄ eluent and a CDRS 600 suppressor. Both systems utilized a 250 μL sample injection loop at a flow rate of 0.40 mL/min and a run time of 9.0 minutes. Sample injection was managed using a Dionex AS-AP autosampler.

Given the presence of significant amounts of entrained basal material across all measured ions including Na⁺ in the ALHIC1901 data (see Discussion), we are not able to reliably derive relative contributions of sea-salt and non-sea-salt sources to Na⁺, SO₄²⁻, Ca²⁺, and Cl⁻ because this derivation requires the assumption that Na⁺ is sourced exclusively from sea salt. We



note that throughout the Results and Discussion, ions (measured by IC) are written with their ionic charges (e.g., Ca^{2+}), trace elements (measured by ICP-MS) are indicated with a TE subscript (e.g., Ca_{TE}), and when referring to an element without distinguishing between measurement methods we omit both the charge and the subscript.

245 We calculate the net ionic balance from major ion measurements as a proxy for acidity/alkalinity (Eq. 2), where concentrations are given in $\mu\text{eq L}^{-1}$. These major ions are typically dominant in Antarctic ice (Legrand and Mayewski, 1997), although we note that basal material may result in additional, unmeasured ions impacting the ionic balance.

$$\text{Net Ionic Balance} = \sum \text{Anions} - \sum \text{Cations} = ([\text{Cl}^-] + [\text{NO}_3^-] + [\text{SO}_4^{2-}]) - ([\text{Mg}^{2+}] + [\text{Ca}^{2+}] + [\text{Na}^+] + [\text{K}^+]) \quad (2)$$

We perform a principal component analysis on trace element measurements of each stick, using Cs, Ba, Ce, Pb, Mn, Co, Ni, 250 Mg, Al, Fe, K, S, and Na. These elements are all well above detection limits even in ALHIC1903 interglacial (MIS 5e) ice, and they span a range of sources and mineralogies. The first principal component captures 66% of the variance, and is applied in the analysis below as a single metric to capture the overall trace element content of each sample.

2.4.3 Greenhouse gas analytical methods

CO_2 concentrations were measured using a dry extraction method that employs a needle crusher and gas chromatography. 255 Methods are described in Ahn et al. (2009) and Wendt et al. (2024). CH_4 concentrations were measured following the wet extraction technique developed in Grachev et al. (2009) and further described in Lee et al. (2020a) and Epifanio et al. (2020). Total air content was derived from the CH_4 measurement technique following the method described in Epifanio et al. (2023).

2.4.4 Major gas isotope and ratio methods

Major gas isotope and elemental ratios analyzed here are $\delta^{15}\text{N}-\text{N}_2$, $\delta^{18}\text{O}-\text{O}_2$, $\delta\text{O}_2/\text{N}_2$, and $\delta\text{Ar}/\text{N}_2$. Measurements were 260 done at OSU using established techniques and methods (Sowers et al., 1989; Petrenko et al., 2006; Severinghaus et al., 2009). In short, 10-15 g gas samples were cut on a band saw, with ~ 4 mm trimmed from all edges to minimize the effect of gas loss. Samples were placed in glass vacuum flasks and evacuated for 60 minutes while chilled in a -50°C ethanol bath. Next, gas was extracted using a melt-refreeze method, and trapped cryogenically in stainless steel sample tubes placed in a 12K closed-cycle cryogenic cold trap system (Lake Shore Cryotronics). Sample tubes were warmed to room temperature for a minimum of 4 265 hours before being expanded one by one into the sample bellow of a dual inlet Isotope Ratio Mass Spectrometer (Thermo Fisher Scientific Delta Q) for analysis of isotope and elemental ratios. Measurements were made against an in-house working standard. Corrections were applied for pressure imbalance between the sample and standard bellows, and for the chemical slope of O_2 and N_2 on $\delta^{15}\text{N}-\text{N}_2$, $\delta^{18}\text{O}-\text{O}_2$, respectively, following ?. All ice samples were then calibrated to contemporary Corvallis air collected in glass flasks on the roof of Burt Hall.

270 Measurements of $\delta^{15}\text{N}-\text{N}_2$ are used to correct $\delta^{18}\text{O}-\text{O}_2$, $\delta\text{O}_2/\text{N}_2$, and $\delta\text{Ar}/\text{N}_2$ for gravitational fractionation in the firn (Sowers et al., 1989). A gas loss correction was also applied to gravitationally corrected $\delta^{18}\text{O}-\text{O}_2$ ($\delta^{18}\text{O}_{\text{grav}}$) to account for associated isotopic fractionation, as defined by Severinghaus et al. (2009):



$$\delta^{18}O_{atm} = \delta^{18}O_{grav} + a \cdot \delta O_2/N_{2,grav} + b \cdot \delta Ar/N_{2,grav} \quad (3)$$

where coefficients a and b are -0.0119 and 0.0155 , respectively, as defined in Baggenstos et al. (2017).

275 The analytical line used for these measurements was recently constructed at OSU based on similar systems elsewhere, and a more detailed description of the setup will be provided in a future publication. On each measurement day, six ice samples and one Corvallis air atmospheric standard were analyzed. Reproducibility of replicate measurements of Corvallis air (1σ) is 0.003‰ , 0.006‰ , 0.01‰ , and 0.04‰ for $\delta^{15}N-N_2$, $\delta^{18}O-O_2$, $\delta O_2/N_2$, and $\delta Ar/N_2$, respectively.

2.4.5 Argon dating analytical methods

280 Following prior work at the Allan Hills (Higgins et al., 2015; Yan et al., 2019; Shackleton et al., 2025), argon dating of the ice core samples is based on the deficit of ^{40}Ar in ancient air trapped in ice compared to the modern atmosphere. This method takes advantage of the contemporary degassing rate of radiogenic ^{40}Ar produced by the decay of ^{40}K in Earth's crust and mantle and its subsequent outgassing into the atmosphere over geologic time (Bender et al., 2008). To extract noble gases from an ice sample, about 500–600 g of ice is slowly melted in a stainless-steel vessel under vacuum and the entrapped air is frozen
 285 in silica gel in a glass tube, using liquid nitrogen (Shackleton et al., 2025). This air is then exposed to a getter alloy (Al-Zr) heated at $900^\circ C$ for 45 min to remove reactive gases (N_2 , CO_2 , O_2 , CO , etc.) leaving behind only the noble gases. The noble gases (Ne, Ar, Kr, Xe) are cryogenically transferred into a stainless steel tube cooled to 10–12 K. The sample is then heated to room temperature for a minimum of 4 hours, and attached to an external port of the dual inlet system on a Thermo 253 Plus isotope ratio mass spectrometer. The $\delta^{40/38}Ar$ ratio of a sample is corrected for gravitational fractionation taking place in the
 290 firn layers using the $\delta^{38/36}Ar$ ratio (Bender et al., 2008).

The resulting $^{40}Ar_{paleoatm}$ is normalized to the modern atmosphere, and the age is calculated following the method developed by Bender et al. (2008) considering the rate of atmospheric $^{40}Ar_{atm}$ increase ($-0.066 \pm 0.007 \text{‰ Myr}^{-1}$). The analytical uncertainty of this method is 66.08 kyr (1σ) and the uncertainty on the absolute age in samples >600 kyr is $\pm 11\%$ (1σ) due to uncertainty in the calibration (Shackleton et al., 2025).

295 We collect the meltwater from each $^{40}Ar_{atm}$ sample for water isotope measurements (following the methods described in Sect. 2.4.1) and validation of the Monte-Carlo dip-adjustment approach (see Sect. 3.1).

3 Results

In this section, we present the results from each of the sets of measurements. We begin by assessing the layer structure and age using ECM, $^{40}Ar_{atm}$, and water isotope data. We then present results for chemistry, particle, greenhouse gas, and major gas
 300 isotopes and elemental ratio measurements.



3.1 Layer structure and age results

3D ECM measurements indicate strong layering in both sections of ice (Fig. 1). Layer dip estimates in both core sections are similar: 38.6° in 228_4 and 38.8° in 230_4. These layer dip estimates are similar to a lower-precision 48° stratigraphy matching estimate between the ALHIC1901 and ALHIC1503 cores (Marks-Peterson et al., 2026) and a 28° bedrock slope estimate from ground-penetrating radar (Nesbitt and Brook, 2023; Kirkpatrick et al., 2025). Due to poor ice quality, a consistent azimuth could not be maintained across the two core sections, resulting in different apparent dip angles on individual ECM faces. The figures in this paper (except for the 3D visual in Fig. 1) show a cross section in the x-plane, where section 228_4 dips at 36.8° and section 230_4 dips at 27.8° . While figures plotted on dip-adjusted depth (Figs. 4, 5, 6, 8) show ECM data only in the x-plane, dip-adjusted depth accounts for dip in both x and y dimensions.

The dip angle calculations presented above are validated by water-isotope sampling results. Fig. 4 demonstrates that triplicate sets of high-resolution water-isotope data are not well aligned when plotted over core depth but do align when plotted on dip-adjusted depth. This is most clearly evident in section 228_4, a result of the steep apparent dip in the x-plane and the wide separation of water isotope samples in the cut plan on this section. In section 228_4, we note a small offset in overlapping $\delta^{18}\text{O}_{\text{ice}}$ measurements. While we cannot rule out variability in $\delta^{18}\text{O}_{\text{ice}}$ across a layer driving these changes, they could also be the result of a combination of analytical uncertainty or cm-scale errors in depth registration.

As with the water isotope results, $^{40}\text{Ar}_{\text{atm}}$ dates align for samples within the same layer (Fig. 5). The two samples below the layer of high AC conductivity are dated to 1.65 and 1.58 Ma, within the 0.066 Ma relative dating error. Likewise, the two samples in the layer of high AC conductivity, which are dated to 1.68 and 1.78 Ma, fall within the uncertainty of each other. The two samples above the ECM layer are dated to 1.37 and 1.53 Ma, and are separated by over twice the 1σ relative measurement error. However, these two samples have distinct dip-adjusted depth distributions, which may reconcile the age difference between the two samples.

Measuring the water isotope composition of meltwater from each of the $^{40}\text{Ar}_{\text{atm}}$ age samples provides an opportunity to test the accuracy of large-volume sample dip-adjustment. The relative distribution of each $^{40}\text{Ar}_{\text{atm}}$ age sample over dip-adjusted depth is constrained by the Monte-Carlo approach (Fig. 5). Computing a weighted average of the high-resolution water isotope record according to the dip-adjusted depth distribution of each $^{40}\text{Ar}_{\text{atm}}$ age sample yields a synthetic $\delta^{18}\text{O}$ value for each $^{40}\text{Ar}_{\text{atm}}$ age sample. This synthetic estimate of the $\delta^{18}\text{O}$ value can then be compared to a real measurement of the $\delta^{18}\text{O}$ meltwater from the corresponding $^{40}\text{Ar}_{\text{atm}}$ -dated sample. The two numbers are in good agreement: for the five $^{40}\text{Ar}_{\text{atm}}$ samples with meltwater $\delta^{18}\text{O}$ measurements, the difference between the synthetic and measured values is within the analytical error of the $\delta^{18}\text{O}$ data. This result provides evidence that we have accurately determined the dip-adjusted position of the large $^{40}\text{Ar}_{\text{atm}}$ age samples relative to other, higher-resolution measurements.

3.2 Chemistry results

Duplicate sets of major ion measurements alongside acid-leached trace element measurements in section 228_4 provide an additional opportunity to test the dip-adjusted depth method (Fig. 6). As described in the methods (Sect. 2.4.2), we note that

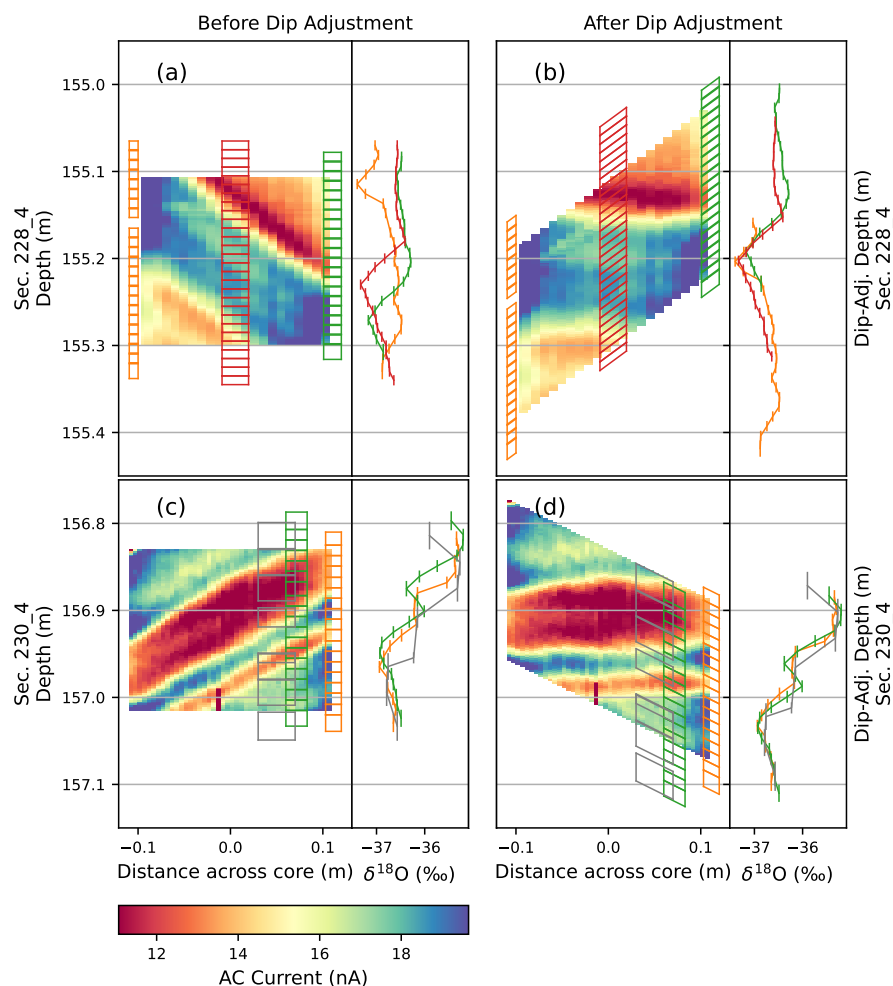


Figure 4. $\delta^{18}\text{O}_{\text{ice}}$ results from section 228_4 (top) and section 230_4 (bottom) demonstrate that ECM layering can be used to align other proxies, including those which do not directly affect electrical conductivity. The left column, (a) and (c), show data plotted on core depth, and the right column, (b) and (d), show data after dip-adjustment, as in Fig. 2. In all four sub-figures, the left subplot shows ECM data across the full width (x plane) of a half-core cut, with core depth (or dip-adjusted depth) on the y-axis and the distance across the core on the x-axis. The location of three sets of water isotope measurements is shown with the blue, green, and yellow boxes. The right subplot shows the $\delta^{18}\text{O}_{\text{ice}}$ on the same depth scale. The triplicate $\delta^{18}\text{O}$ records only align after the dip-adjustment is completed in both section 228_4 and 230_4.

ions are written with their ionic charges (e.g., Ca^{2+}), trace elements are indicated with a TE subscript (e.g., Ca_{TE}), and when referring to an element without distinguishing between measurement methods we omit both the charge and the subscript. We find good alignment in replicate samples of most ion measurements (Cl^- , NO_3^- , and Ca^{2+}). Where comparisons to trace element measurements are available (e.g., Ca^{2+} vs. Ca_{TE} , SO_4^{2-} vs. S_{TE}), there is likewise strong alignment across measurement methods in most cases. An exception is Mg, which is consistently higher in ICP-MS measurements than in IC

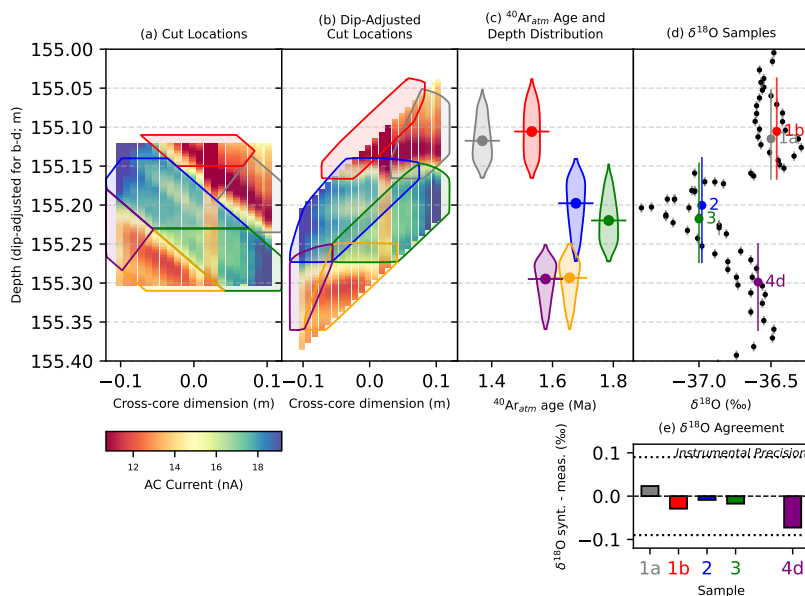


Figure 5. (a) shows the extent of the six $^{40}\text{Ar}_{\text{atm}}$ age samples cut from section 228_4 in this study overlaid on ECM (as with Fig. 4). (b) shows the same cut dimensions in dip-adjusted space. The overlap between samples is a result of dip-adjustment in both the x-plane (shown) and y-plane (into the page). (c) shows the age of each sample. The width of the distribution represents the portion of the $^{40}\text{Ar}_{\text{atm}}$ sample at a given dip-adjusted depth, the dot shows the weighted-average depth, and the horizontal bar shows the relative dating uncertainty. (d) Small black circles represent the high-resolution $\delta^{18}\text{O}_{\text{ice}}$ samples presented in Fig. 4, larger colored circles represent the $\delta^{18}\text{O}_{\text{ice}}$ of the $^{40}\text{Ar}_{\text{atm}}$ samples, and the colored bar represents the full depth span of the $^{40}\text{Ar}_{\text{atm}}$ samples. No $\delta^{18}\text{O}_{\text{ice}}$ data are available for sample 4c (yellow). (e) shows the difference between the measured $\delta^{18}\text{O}_{\text{ice}}$ of each $^{40}\text{Ar}_{\text{atm}}$ sample compared to a synthetic calculation of each sample based on its depth distribution and high-resolution $\delta^{18}\text{O}_{\text{ice}}$ samples. All estimates fall within the $\delta^{18}\text{O}_{\text{ice}}$ sample uncertainty, increasing confidence in our tracking of $^{40}\text{Ar}_{\text{atm}}$ sample dimensions.

measurements. The first principal component of the selected ICP-MS measurements (Cs, Ba, Ce, Pb, Mn, Co, Ni, Mg, Al, Fe, K, S, and Na; see Section 2.4.2) is at a maximum in the layer of high AC conductivity in section 228_4, matching the pattern of most major ion measurements.

Some measurements do show inhomogeneities across the width of the core. For example, Na and $\text{SO}_4^{2-}/S_{\text{TE}}$ data generally align, although several high Na and $\text{SO}_4^{2-}/S_{\text{TE}}$ measurements within the high-AC layer of 228_4 are not consistent with other measurements at the same dip-adjusted depth.

ECM results provide an additional constraint on chemistry. While AC ECM results show significant variability and coherent layer structure, the DC ECM results are homogeneous in section 228_4 with very low DC conductivity. With the same electrode design and device configurations as a previous study on shallow Allan Hills ice (Kirkpatrick et al., 2025), the DC ECM conductivity on section 228_4 is two orders of magnitude lower. Given that DC ECM is driven by H^+ concentrations (Wolff,

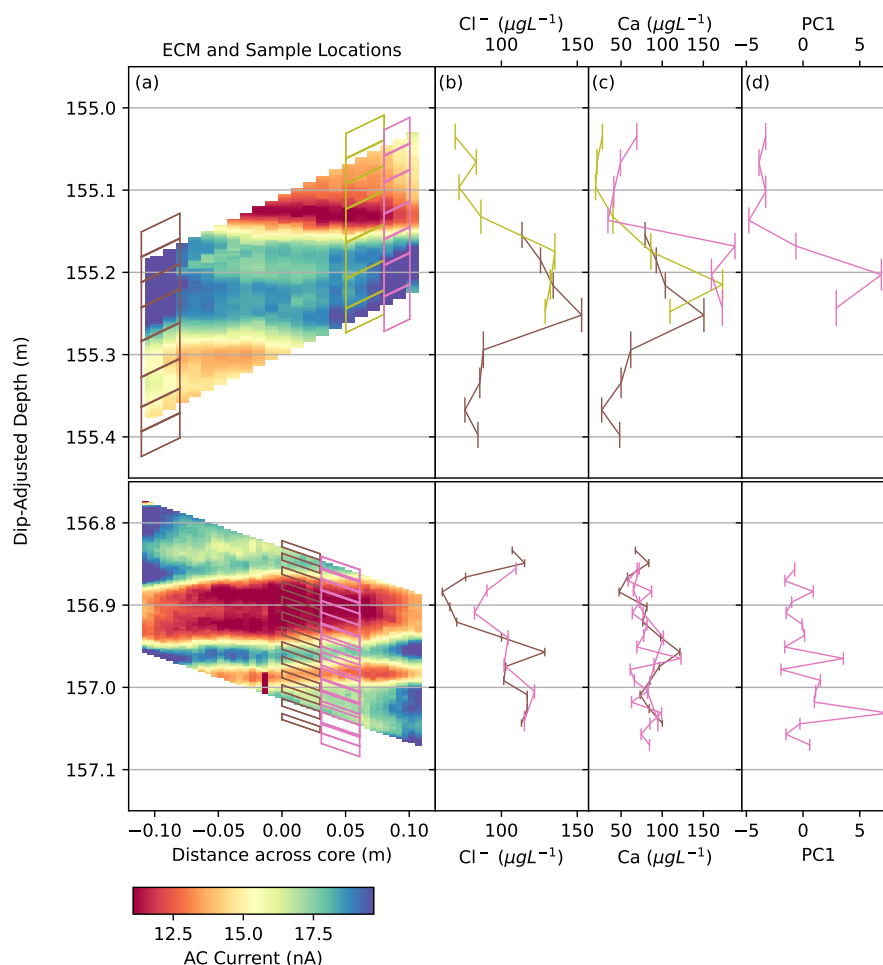


Figure 6. As in Fig. 4 but with impurity measurements from sections 228_4 and 230_4. (a) shows the location of each sample overlaid on ECM data on dip-adjusted depth, with the color of the sample matching other panels. (b) shows Cl^- concentrations, (c) shows Ca^{2+} and Ca_{TE} concentrations, and (d) shows the first principal component of the ICP-MS data (PC1; see Methods).

2000), this implies alkaline conditions in section 228_4. This is consistent with the measured ionic balance ($\sum \text{anions} - \sum \text{cations}$), which is largely neutral or negative across 228_4, indicating heightened alkalinity. DC ECM data on section 230_4 were collected with different electrodes and cannot be directly compared to other data, although they also show a relatively homogeneous signal.

In section 228_4, trace element and ion concentrations are elevated within the layer of high AC conductivity, older ages, and depleted $\delta^{18}\text{O}$ values. The relationship between ECM layering and chemistry is weaker in section 230_4. Some impurity species (including Ca) appear to increase in the layers of high AC conductivity, but there is no common structure across many measurements as in section 228_4.

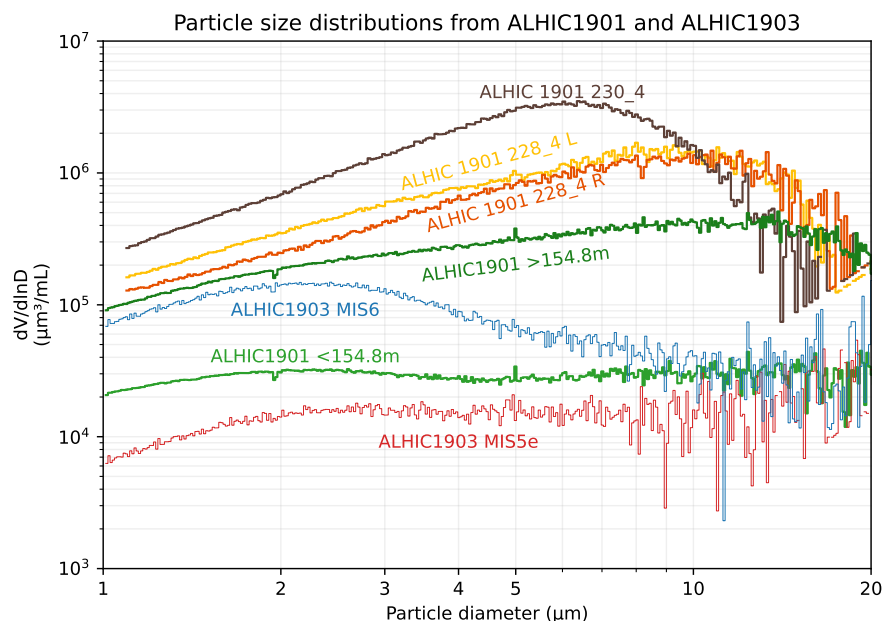


Figure 7. Coulter Counter particle size distribution for each set of data from sections 228_4 and 230_4 is shown alongside other Allan Hills Coulter Counter data to contextualize these results. The particle size distributions for MIS5e (orange) and MIS6 (brown) in nearby ALHIC1903 provide an example of typical glacial-interglacial variability (Carter et al., 2026). Distributions from a separate set of measurements from this core show a clear difference between measurements in the upper portion of the core (<154.8 m; dark green) and the deepest 5 m (>154.8 m; bright green) (Choi et al., 2026). Note that particle diameters are shown on a logarithmic scale.

3.3 Particle results

Coulter Counter measurements identify high particle concentrations across all deep ALHIC1901 samples, with a large fraction of coarse particles (Fig. 7). The average concentrations of the two 228_4 sticks are $5509 \mu\text{g L}^{-1}$ and $4797 \mu\text{g L}^{-1}$, respectively, while the 230_4 stick has a higher average concentration of $10458 \mu\text{g L}^{-1}$. There is substantial cm-scale variability in particle concentrations and sizes. The 228_4 samples ($N = 15$ across the two sticks) range from 1465 – $16024 \mu\text{g L}^{-1}$, while the 230_4 samples ($N = 7$) range from 1167 – $22623 \mu\text{g L}^{-1}$. For context, typical glacial and interglacial particle concentrations from the nearby ALHIC1903 core are on the order of $500 \mu\text{g L}^{-1}$ and $20 \mu\text{g L}^{-1}$ respectively (Carter et al., 2026). The peaks in the volume normalized particle size distributions of sections 228_4 and 230_4 are above $9 \mu\text{m}$ and $5 \mu\text{m}$ respectively, similar to the $\sim 10 \mu\text{m}$ peak in ALHIC1901 samples below 154.8 m (Choi et al., 2026). However, these peaks are significantly higher than the peak of the ALHIC1901 data above 154.8 m (Choi et al., 2026), as well as peaks in both interglacial (MIS5e) and glacial (MIS6) concentrations in the nearby ALHIC1903 core (Carter et al., 2026), which all fall below $3 \mu\text{m}$. Unlike the acid-leach elemental and major ion data, the Coulter Counter particle concentrations do not clearly align on dip-adjusted depth where duplicate data are available in section 228_4.

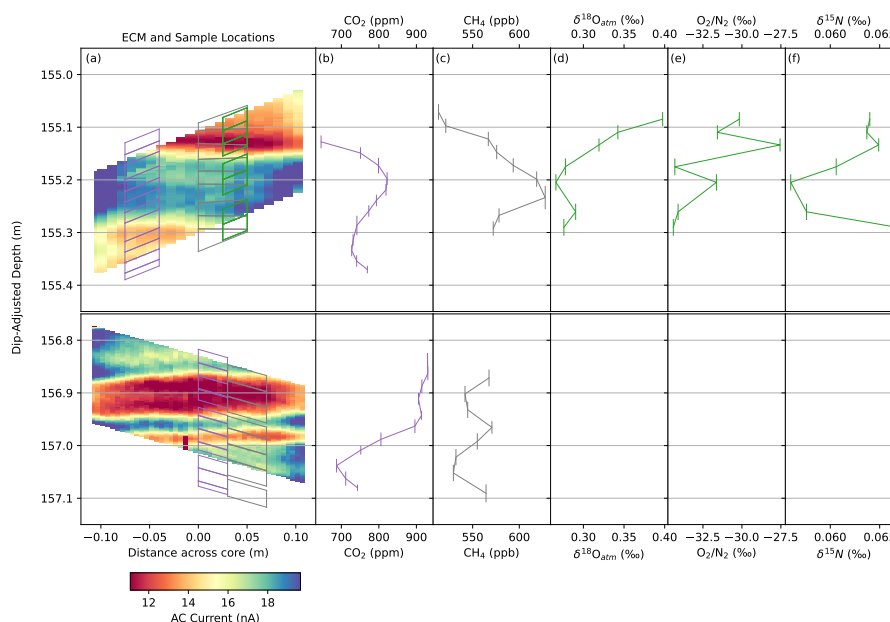


Figure 8. Gas sampling results. As with prior figures, the location of samples is shown over dip-adjusted ECM data for section 228_4 (top) and section 230_4 (bottom) in (a). Subplots (b), (c), (d), (e), and (f) show CO_2 , CH_4 , $\delta^{18}\text{O}_{\text{atm}}$, $\delta\text{O}_2/\text{N}_2$, and $\delta^{15}\text{N}$ data, respectively. The vertical bar represents the dip-adjusted depth range covered by each individual sample.

370 3.4 Greenhouse gas results

CO_2 concentrations range from 646–822 ppm in section 228_4 and 687–914 ppm in section 230_4 (Fig. 8). CH_4 concentrations are 513–618 ppb in section 228_4 and 529–570 ppb in section 230_4. The samples with the highest CO_2 and CH_4 concentrations fall within the layer of high AC conductivity and depleted $\delta^{18}\text{O}_{\text{ice}}$ in section 228_4. In section 230_4, the opposite pattern is observed, with the lowest CO_2 and CH_4 values falling within layers of high AC conductivity and depleted $\delta^{18}\text{O}_{\text{ice}}$ values. Total air content measurements range from 0.085–0.101 $\text{cm}^3 \text{g}^{-1}$ in section 228_4 and 0.093–0.104 $\text{cm}^3 \text{g}^{-1}$ in section 230_4, within the variability previously observed in the bottom 30 m of ALHIC1901 (Marks-Peterson et al., 2026).

3.5 Major gas isotope and elemental ratios results

We find that the major gas isotope and elemental ratios follow the layering seen in other analyses (note that all data are from section 228_4). The $\delta^{15}\text{N}$ isotopic ratios range from 0.056‰ to 0.067‰, with the lowest values coincident with the layer of high AC ECM conductivity. $\delta^{15}\text{N}$ is commonly interpreted as a proxy for the past thickness of the (diffusive zone in the) firn column (Sowers et al., 1992), and the observations suggest a diffusive zone thickness ranging from 11 to 14 m. While this is anomalously thin compared to most observations across Antarctica, it is very typical for observations in Blue Ice Areas such as Allan Hills (Yan et al., 2019) and Taylor Glacier (Baggenstos et al., 2017).



$\delta^{18}\text{O}_{\text{atm}}$ values range from 0.27‰ to 0.40‰, with the lowest values coincident with the layer of high AC ECM conductivity.

385 In the age range spanned by sample 228_4, atmospheric $\delta^{18}\text{O}_{\text{atm}}$ values are between 0 and 0.1 ‰ with the more depleted values in older ice possibly reflecting trends in surface ocean $\delta^{18}\text{O}$ related to changes in global ice volume (Marks-Peterson et al., 2026; Shackleton et al., 2026). The 228_4 $\delta^{18}\text{O}_{\text{atm}}$ thus appears artifactually enriched, which we attribute to oxidation or other O_2 loss processes (Sect. 4.1.3). The relationship between $\delta^{18}\text{O}_{\text{atm}}$ and ^{40}Ar age in 228_4 does appear to have the correct sign, though we refrain from interpreting this due to the aforementioned enrichment.

390 Last, we look at the elemental ratios $\delta\text{O}_2/\text{N}_2$ and $\delta\text{Ar}/\text{N}_2$. Both are typically negative in ice cores, reflecting the small molecular diameter of O_2 and Ar that results in their loss from the ice during bubble close-off at the firn-ice transition and core storage (Huber et al., 2006; Severinghaus and Battle, 2006). The $\delta\text{O}_2/\text{N}_2$ and $\delta\text{Ar}/\text{N}_2$ caused by such fugitive gas loss typically scale; in pristine samples from ALHIC1901 we observe a relative slope of around 0.27 (see Appendix B). In section 228_4 we find $\delta\text{O}_2/\text{N}_2$ that are much more depleted than the typical scaling with $\delta\text{Ar}/\text{N}_2$ would imply. This shows

395 an additional process acting to remove O_2 from the ice, which we attribute to oxidation of organic material (respiration) and perhaps non-organic impurities. Respiration of organic material would be consistent with the enriched CO_2 seen in these and other deep ALHIC1901 samples (see Appendix C), where the added CO_2 was shown to have a $\delta^{13}\text{C}-\text{CO}_2$ of -29‰ (Marks-Peterson et al., 2026). We estimate an O_2 deficit of around 2900–3500 ppm, which is several times more than expected from stoichiometry given the 400–500 ppm of in situ CO_2 production.

400 4 Discussion

In this discussion, we first assess the completed measurements and conclude that the layering is likely caused by the integration of basal material rather than the aeolian deposition of particles. We then discuss how the strong and dipping layering observed in ALHIC1901 sections 228_4 and 230_4 might impact paleoclimate interpretations of this ice. We end the discussion with suggestions for future measurements, both at the Allan Hills and other locations where layering may be disturbed.

405 4.1 Basal material as a driver of layer composition

Given that sections 228_4 and 230_4 are within 5 meters of the bed, the impact of integrated basal material is a strong hypothesis for the origin of the observed variability across ECM, particles, chemistry, and gas measurements. In this section, we evaluate this hypothesis within the context of our sampling results. We first discuss how particle and chemistry results provide evidence of layering driven by either basal (subglacial) or proximal aeolian (supraglacial) material and compare the

410 results with the penultimate glacial–interglacial transition measured in the nearby ALHIC1903 core (Carter et al., 2026). We then discuss the elevated greenhouse gas concentrations and major gas isotope and elemental ratio results. We conclude that incorporation of basal material is more likely than aeolian deposition.



4.1.1 Particle constraints on basal material

As shown in Fig. 7, particles in sections 228_4 and 230_4 are substantially coarser and more concentrated than aeolian glacial-
 415 interglacial dust variability observed in the nearby ALHIC1903 core (Carter et al., 2026). The basal ALHIC1901 samples
 exhibit elevated concentrations of particle volume at coarse diameters with maxima broadly centered near 9 and 5 μm diameters
 for sections 228_4 and 230_4, respectively. In contrast, typical glacial ice is dominated by fine particles around 2 μm , while
 interglacial ice from ice cores drilled on the peripheral portion of the East Antarctic Ice Sheet is characterized by consistently
 elevated particle volume in the 2–6 μm diameter range (Carter et al., 2026). In interpreting Victoria Land ice-core particle
 420 records, Delmonte et al. (2010) suggest that particles $>5 \mu\text{m}$ are unlikely to result from typical distal dust sources, and are only
 consistent with very local sources. Furthermore, particle counts at the coarse maxima in the basal ice are nearly an order of
 magnitude higher than the glacial ice maxima and nearly two orders of magnitude higher than interglacial ice (Fig. 7). Together,
 these observations are at odds with typical aeolian transport of dust to the Allan Hills, although they do not definitively rule
 out aeolian transport from a very local source region which might produce both large particle size distributions and total
 425 concentrations.

A recent study of ALHIC1901 dust particle concentrations (Choi et al., 2026), which are included in Fig. 7, supports the
 conclusion that particles in sections 228_4 and 230_4 are basal in origin. Choi et al. (2026) identify a shift to much higher
 particle concentrations with a coarser size distribution in the bottom 5 m of the core, and the data from sections 228_4 and
 230_4 are consistent with this basal signal.

430 4.1.2 Chemistry constraints on basal material

Elemental and major ion concentrations throughout 228_4 and 230_4, and especially in the layers of elevated AC conductiv-
 ity, exceed expectations from aeolian deposition based on observed glacial–interglacial variability. To provide a constraint on
 aeolian deposition, we present new major ion measurements from the nearby stratigraphically continuous ALHIC1903 core
 spanning MIS6 (glacial) and MIS5e (interglacial) ice (Fig. 9). The ALHIC1903 ion measurements are consistent with other
 435 Antarctic ice cores, demonstrating elevated impurities during the glacial periods and reduced impurities during interglacial pe-
 riods (Bigler et al., 2006; Wolff et al., 2010; Lambert et al., 2008). The ALHIC1903 measurements are therefore representative
 of expected aeolian contributions to the ice in the Allan Hills. Against this baseline, the magnitude and structure of impurity
 concentrations in ALHIC1901 sections 228_4 and 230_4 point to a different dominant source of impurities.

Ca in Antarctic ice, for example, comprises contributions from both marine (sea-salt) and terrestrial (mineral dust) compo-
 440 nents (Röthlisberger et al., 2002), which are each efficiently dissolved during atmospheric transport and/or when ice is melted
 prior to analysis (de Angelis et al., 1992). As such, following a simple sea-salt correction, Ca is commonly used as a proxy
 for aeolian dust deposition in Antarctica (e.g., Lambert et al., 2012). However, Ca concentrations in sections 228_4 and 230_4
 are up to a factor of approximately 4 greater than the measured ALHIC1903 MIS6 glacial values and are highest within the
 high AC ECM layer of section 228_4 (Fig. 6 and Fig. 9(b)). Therefore, these concentrations point to either a basal or highly
 445 proximal aeolian source of Ca.

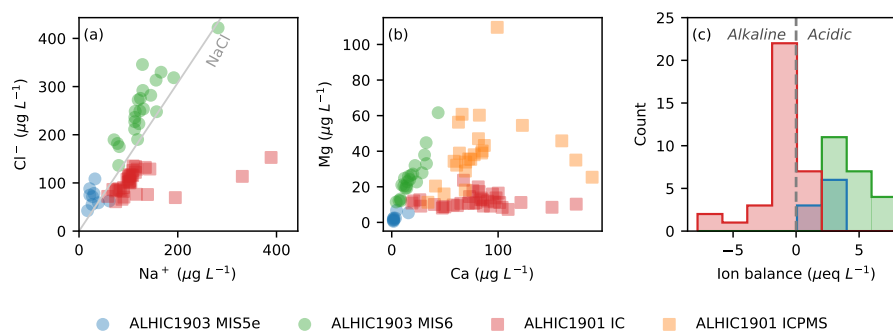


Figure 9. Here we compare the chemistry of ALHIC1901 sections 228_4 and 230_4 with measurements from ALHIC1903 in MIS5e and MIS6 ice. (a) shows the Cl^- - Na^+ relationship, with a 1:1 NaCl mixing line in grey. (b) shows the Mg - Cl^- relationship. We differentiate measurements with IC and ICP-MS as the Mg concentrations are distinct. (c) shows the distribution of net ion balance ($\sum \text{anions} - \sum \text{cations}$) between the two cores. Note that these ALHIC1901 samples do not fall within the range of glacial-interglacial variability represented by ALHIC1903 across all three subplots.

The $\text{Na}^+:\text{Cl}^-$ relationships in ALHIC1901 sections 228_4 and 230_4 are also distinct from typical variability over glacial–interglacial timescales in nearby cores (Fig. 9). In ALHIC1903, the $\text{Na}^+:\text{Cl}^-$ relationship falls roughly along a 1:1 molar ratio consistent with sea salt (NaCl) input in both glacial (MIS6) and interglacial (MIS5e) periods. Sections 228_4 and 230_4 are both enriched in Na^+ relative to the sea-salt ratio, especially in 228_4. This divergence is indicative of either an additional
 450 non-aeolian Na^+ source or a post-depositional loss of Cl^- only. We suggest that a subglacial input of Na^+ is most likely, as the mechanism for Cl^- loss at depth—ionic migration along liquid-like grain boundaries (Traversi et al., 2009)—would be expected to produce broader perturbations to the ionic record than are observed here, although we cannot completely exclude some degree of differential ion mobility.

Notably, acid-leached Mg_{TE} concentrations are substantially higher than the ionic Mg^{2+} concentrations (Fig. 9(b)). This
 455 pattern is striking because other measurements where both IC and ICP-MS data are available, such as Ca_{TE} and Ca^{2+} , share similar ranges in section 228_4. The simplest interpretation is that most Mg is present in particulate, largely insoluble Mg-bearing silicates that do not readily dissolve in meltwater and therefore do not appear in IC measurements. This interpretation is consistent with mineral entrainment from local bedrock geology (Barrett et al., 1986).

We also note that the ion balance of both sections measured in ALHIC1901 is significantly lower than ALHIC1903, as
 460 shown in Fig. 9(c). The negative measured ionic balance in most ALHIC1901 samples implies the addition of alkalinity (e.g., CO_3^{2-} or HCO_3^-), whereas the aeolian measurements from ALHIC1903 lack this alkalinity. Changes to ice alkalinity can be relevant for in-ice CO_2 alteration (Anklin et al., 1995; Smith et al., 1997), so these results may be meaningful for the elevated CO_2 measured in ALHIC1901 (Marks-Peterson et al., 2026).



4.1.3 Greenhouse gas constraints on basal material

465 A final line of evidence supporting the basal material hypothesis comes from the greenhouse gas and major gas elemental records. All measured CO₂ concentrations (uniformly above 650 ppm) are well above mean atmospheric CO₂ in shallower pristine Allan Hills samples from 1.3–1.8 Ma (approximately 225 ppm; Marks-Peterson et al., 2026). They are also above the generally below 300 ppm atmospheric CO₂ reconstructed from marine and terrestrial proxies from that time (?). Post-depositional CO₂ production is documented in some ice cores, through both organic and inorganic processes. Inorganic
470 processes involve CaCO₃ reactions, typically in acidic ice, and are most frequently observed in Greenland where CaCO₃ concentrations in glacial ice are much higher than most Antarctic ice (but not necessarily Allan Hills basal ice) (Anklin et al., 1997; Smith et al., 1997). Metabolic processes are also possible, particularly in ice with higher dust and impurity concentrations (Montross et al., 2014; Souchez et al., 1993). Altered CO₂ has been widely observed at the Allan Hills (Yan et al., 2019; Marks-Peterson et al., 2026; Banerjee et al., 2026). A combination of $\delta^{13}\text{C}$ and O₂/N₂ measurements suggests that the primary
475 production mechanism for CO₂ in the bottom 30 m of the ALHIC1901 core is metabolic oxidation of entrained organic material (Marks-Peterson et al., 2026). While $\delta^{13}\text{C}$ –CO₂ measurements are not available in sections 228_4 and 230_4, the depleted O₂ observed in section 228_4 supports that the process producing CO₂ in this section is the same as the process described across a broader range of the core by Marks-Peterson et al. (2026).

This metabolic process must be reliant on a source of organic carbon unique to certain depth ranges of this core, either from
480 an aeolian source or from basal material. The $\delta^{13}\text{C}$ –CO₂ compositions (Marks-Peterson et al., 2026) presented in Fig. 10 help to differentiate these two sources. $\delta^{13}\text{C}$ –CO₂ abruptly decreases roughly 5 m above the bed, just above sample 228_4 and at approximately the same depth as the increase in particle concentrations identified by Choi et al. (2026). The drop in $\delta^{13}\text{C}$ –CO₂, combined with the elevated CO₂ levels near the bottom of the core (Fig. 10), implies increased post-depositional CO₂ production in the ice closest to the bed, whereas portions of the ice spanning similar ages further from the bed are less
485 impacted. The impact of CO₂ alteration is greatest near the bed, suggesting that a basal source is more likely than an aeolian source. An aeolian source would be expected to have a more uniform impact throughout the ice column.

Additional evidence that the source of elevated CO₂ is basal comes from the relationship between CO₂ and $\delta^{18}\text{O}_{\text{ice}}$ in sections 228_4 and 230_4. Low $\delta^{18}\text{O}_{\text{ice}}$ is associated with more elevated CO₂ in section 228_4, and less elevated CO₂ in section 230_4 (see Appendix Fig. D1). If CO₂ production was the result of increased aeolian dust transport due to changes
490 in climate, like cooler glacial periods, then the relationship between elevated CO₂ and $\delta^{18}\text{O}_{\text{ice}}$ would be expected to remain consistent. The opposite result here suggests that post-depositional CO₂ production is not related to climate.

Measured 228_4 CH₄ values are slightly elevated above the approximately 500 ppb concentrations in pristine samples of similar ages (Marks-Peterson et al., 2026), and are highest at the same dip-adjusted depths as the most elevated CO₂ concentrations. Greenland cores exhibit elevated CH₄ concentrations coincident with layers of higher aeolian-deposited dust
495 and impurities (Lee et al., 2020b; Rhodes et al., 2013). Studies have also identified elevated CH₄ in basal and near-basal ice, presumably as a result of the integration of basal material (Tung et al., 2005). While isotopic constraints on post-depositional alteration of CH₄ at the Allan Hills are not yet available, CH₄ values accompanying altered CO₂ measurements are likely also

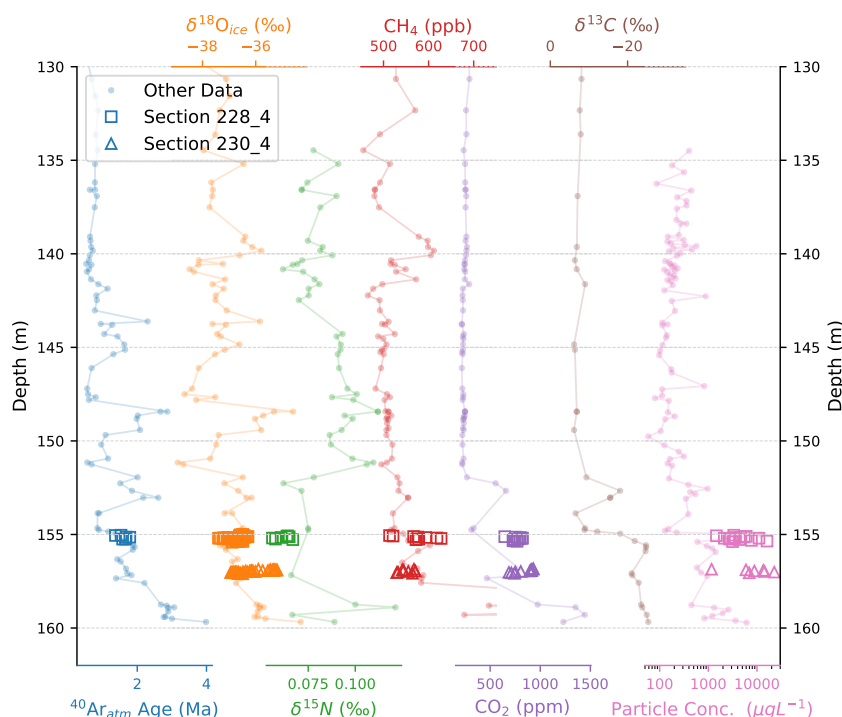


Figure 10. Here we show published $^{40}\text{Ar}_{\text{atm}}$ age, $\delta^{18}\text{O}_{\text{ice}}$ (Shackleton et al., 2025), $\delta^{15}\text{N}$, CH_4 , CO_2 , $\delta^{13}\text{C}-\text{CO}_2$ (Marks-Peterson et al., 2026), and particle concentration measurements (Choi et al., 2026) from the bottom 30 m of ALHIC1901, with circular markers. We show measurements from section 228_4 (open squares) and section 230_4 (open triangles) on the same depth scale. Note that for many measurements ($\delta^{18}\text{O}$, CO_2 , and CH_4), the variability within the two short sections in this study represents much of the variability seen within the published lower-resolution record.

altered (Marks-Peterson et al., 2026; Yan et al., 2019). Therefore, as with the CO_2 results, the elevated CH_4 concentrations in our samples are consistent with the integration of basal material.

500 4.1.4 Differentiating basal and local aeolian sources

Our analysis indicates that the source of the layers of high particle, impurity, and greenhouse gas concentrations is likely the integration of basal material. The layering seen in the ECM reflects different degrees of basal entrainment, rather than a single band of material in otherwise clean ice. Chemistry results suggest that the particle composition is well outside the glacial–interglacial variability in the nearby ALHIC1903 core, and both the CO_2 and particle concentrations are consistent with measurements in the deepest 5 m of this core, but not ice of similar ages further above the bed. However, given the uncertain flow histories of Allan Hills ice (Stoll et al., 2026), we cannot definitively rule out that the ice in the deepest portion of ALHIC1901 accumulated in a different area than ice preserved in the ALHIC1903 record or the upper portions of the



ALHIC1901 record. In this case, an accumulation area for this deepest ice which is very proximal to a nunatak or other ice-free area might enable local aeolian transport of large, coarse dust particles, increasing impurity and greenhouse gas concentrations.

510 Differentiating the two competing hypotheses of very short transport of material from nearby nunataks or the integration of material from basal processes is inherently challenging, given that the fingerprint of both processes across particle, chemistry, and gaseous measurements might be similar. The strongest evidence that locally sourced aeolian deposition is not responsible for higher particle concentrations is the strong correspondence between the proximity of the bed and elevated CO₂ (Marks-Peterson et al., 2026), increased particle concentrations, and larger particle sizes (Choi et al., 2026). Therefore, while the sum
515 of the evidence points to a basal source for the layering observed in this study, research at the Allan Hills should continue to interrogate the origin of particles, impurities, and elevated greenhouse gas measurements in these cores. In this study, we characterize the chemistry of the suspected basal material. Future work may be able to match these measurements with samples from nearby nunataks or subglacial samples (from subglacial coring). Cosmogenic nuclide dating of large rock fragments (>20 mm) in some Allan Hills cores might provide an alternative approach—basal material should have exposure ages significantly
520 older than the ice, while aeolian-deposited material should match the ice age.

4.2 Implications for Allan Hills paleoclimate records

The complex glaciological setting of the Allan Hills preserves unique records of ancient ice, but also adds complexity to its paleoclimatic interpretation. In this section, we discuss how our results improve understanding of four sources of uncertainty for Allan Hills ice: 1) age offsets for ice from the same depth; 2) temporal averaging in samples; 3) drivers of variability in
525 samples of similar ages; and 4) the heterogeneity of layers of elevated basal material. We conclude the section by discussing applications to other ice cores that might be impacted by similar processes.

4.2.1 Coordinating samples by layering rather than depth

Our results from section 228_4 show that the combination of steep layer dip (38°) and large (>100 kyr) age differences between layers means that ice at the same core depth may be offset by as much as 19 cm in dip-adjusted depth and represent very
530 different ages. We also observe differences in the composition between layers at the same depth, including CO₂ variations of more than 100 ppm, order-of-magnitude differences in the concentration of impurities like Ca, and $\delta^{18}\text{O}_{\text{ice}}$ changes of more than 1 ‰. The results of this study demonstrate that dip-adjusted depths significantly improve confidence in co-registration of multiple samples in large-diameter ice cores. We also note that the confirmation of dipping layers of significantly different ages may reconcile previous observations of age offsets in excess of 100 kyr between ⁴⁰Ar_{atm} measurements made at the same core
535 depth in the ALHIC1901 core (Shackleton et al., 2025).

4.2.2 Temporal averaging of samples

A critical additional uncertainty in Allan Hills cores is the degree to which a single sample might average ice of multiple ages, either across a continuous climate chronology or across temporal discontinuities. Prior Allan Hills studies (Marks-Peterson



et al., 2026; Shackleton et al., 2026, 2025) refer to “long snapshots” and “short snapshots”. “Short snapshots”, found in ice
540 younger than 900 ka, appear to preserve significant glacial-interglacial variability. Conversely, “long snapshots”, including
some younger samples and all samples older than 900 ka, demonstrate reduced variability, and appear to capture an average
climate state. This study, by combining multiple measurements at high resolution, provides new insight into the mechanisms
which may drive reduced variability in older Allan Hills ice.

The strong sub-meter-scale variability identified in both ALHIC1901 sections 228_4 and 230_4 suggests that sampling
545 across ice of different ages may contribute to smoothing of climate signals in the “long snapshot” regime. In both sections, lay-
ers are often less than 10 cm thick, and dipping. Given the large volume of $^{40}\text{Ar}_{\text{atm}}$ samples, it is almost certain that a sampling
plan devised for these sections without ECM or other constraints on layer composition and geometry would include ice from
multiple layers in $^{40}\text{Ar}_{\text{atm}}$ samples. Sampling across multiple layers, separated by over 100 kyr in age, might contribute to the
smoothing in the “long snapshot” regime. However, large-volume samples crossing abrupt transitions are unlikely to explain
550 the reduced variance in smaller-volume samples like $\delta^{18}\text{O}_{\text{ice}}$ and CO_2 . Prior work by Marks-Peterson et al. (2026) explored
a similar question by assessing the variability in high-resolution CO_2 and CH_4 measurements alongside six $^{40}\text{Ar}_{\text{atm}}$ samples
with ages less than 0.8 Ma. This analysis showed minimal variability in both CO_2 and CH_4 along each $^{40}\text{Ar}_{\text{atm}}$ sample,
supporting the conclusion that sample dimensions were unlikely to average across ice of different ages in younger samples.

Instead of being driven by sample dimensions, this study suggests that the observed reduced variance in the “long snap-
555 shot” regime is driven by reduced variability in the ice itself. This is most clearly supported by the cm-resolution $\delta^{18}\text{O}_{\text{ice}}$
measurements, the highest resolution measurements of their type in Allan Hills ice of this age. Even at this high resolution, the
variability in these samples (on the order of 1 ‰) is only a fraction of the full range of glacial-interglacial variability observed
in the ‘short snapshot’ regime (roughly 8 ‰). Rather than sampling dimensions being responsible for the reduced variability, it
appears that within packets of ice of similar ages, either extreme thinning or diffusion results in the homogenization of glacial-
560 interglacial variability, resulting in the reduced variability within “long snapshots”. This is consistent with prior observations of
reduced variability in “long snapshot” ice (Marks-Peterson et al., 2026; Shackleton et al., 2026, 2025), but the higher-resolution
measurements in this study confirm that variability is not masked by the larger sample sizes in prior studies.

While it is clear that the “long snapshot” samples integrate over multiple glacial-interglacial cycles in the time domain, it is
remarkable that the layering of distinct ice units of dissimilar age or origin remains so well preserved in the spatial domain.
565 We observe coherent spatial variations on length scales of ~ 10 cm in multiple ice core proxies, including >100 ppm changes
in CO_2 over 10 cm, >110 ppb changes in CH_4 over 17 cm, and >1 ‰ in $\delta^{18}\text{O}_{\text{ice}}$ over 10 cm. Additionally, computing the
difference between the average ages above, in, and below the layer of high AC ECM conductivity in section 228_4 shows
transitions of 280 kyr over 9.9 cm and 114 kyr over 8.8 cm. While there is evidence of some diffusion or mixing between
these different packets, with intermediate values of CO_2 and $\delta^{18}\text{O}_{\text{ice}}$ within the 10 cm transition between packets, these results
570 place an upper bound on any estimates of smoothing occurring between packets of different ages. It is unlikely that such sharp
gradients can be preserved over million(s) of years in the face of diffusion and other mixing processes, which suggests that
these layers are actively thinning or deforming in a way that places these dissimilar ice units into such close proximity.

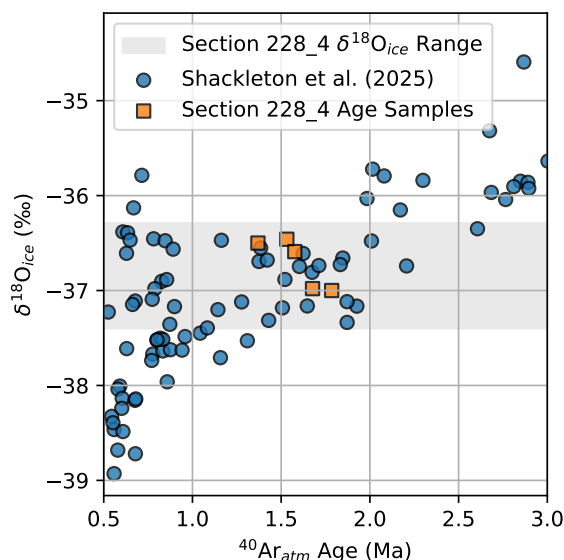


Figure 11. While $\delta^{18}\text{O}_{\text{ice}}$ measurements presented by Shackleton et al. (2025) show a general cooling trend over time, there is significant variability even in samples of similar ages. $\delta^{18}\text{O}_{\text{ice}}$ and $^{40}\text{Ar}_{\text{atm}}$ measurements from section 228_4 show similar variability, and high-resolution $\delta^{18}\text{O}_{\text{ice}}$ measurements capture a wider range of $\delta^{18}\text{O}_{\text{ice}}$ values.

4.2.3 Drivers of variability in samples of similar ages

In section 228_4, we find an enriched $\delta^{18}\text{O}_{\text{ice}}$ signal in the two younger layers, and depleted $\delta^{18}\text{O}_{\text{ice}}$ in the older layer. This
 575 $\delta^{18}\text{O}_{\text{ice}}$ result, taken at face value, implies a gradual warming from approximately 1.8 Ma to 1.5 Ma, with a linear regression
 on the five data points showing $\delta^{18}\text{O}_{\text{ice}}$ increasing by 1.5 ‰ per Myr forward in time. This is opposite to the cooling trend in
 Allan Hills $\delta^{18}\text{O}_{\text{ice}}$ over the same time period (Shackleton et al., 2025), and also contradicts cooling trends in mid-Pleistocene
 Allan Hills mean ocean temperature (Shackleton et al., 2026) and global climate records (e.g., Lisiecki and Raymo, 2005).
 The unexpected age-temperature relationship is also not readily explained with either dip-adjusted or time-averaging effects
 580 discussed above, as they are both measured on the same ice sample, which indicates some other source of signal variability
 is likely present. Instead, the apparent short-term warming trend appears to be noise on the long-term cooling trend. This
 conclusion is informed by existing Allan Hills data. As shown in Fig. 11, the spread in the $\delta^{18}\text{O}_{\text{ice}}$ -age relationship in section
 228_4 is similar to the range of variability in samples of similar ages from other samples in ALHIC1901 and other Allan Hills
 cores (Shackleton et al., 2025).

585 Multiple hypotheses might explain why samples at similar ages—even in long snapshot ice which appears to represent long-
 term averages—demonstrate temperature variability. One explanation might be changes to the glacial-interglacial weighting
 of individual ice packets. Increased weighting towards glacial or interglacial ice, perhaps resulting from changes to ice flow or
 surface mass balance, could result in ice from similar time periods and climate states exhibiting different long-term averages.



Another possible explanation is changes to the source elevation. Given the uncertain flow history of Allan Hills ice, it is possible that ice of similar ages in ALHIC1901 and other Allan Hills cores comes from multiple source elevations, which could also result in a range of average temperatures even in ice of similar ages. Integrating other measurements from this study aids in differentiating these hypotheses. Notably, the older layer in section 228_4 with more negative $\delta^{18}\text{O}_{\text{ice}}$ also has lower $\delta^{15}\text{N}$, suggesting a thinner firn column during colder periods as is observed throughout East Antarctica due to the reduced snow accumulation rates at such times (Caillon et al., 2003; Buizert et al., 2021). This supports the hypothesis that weighting towards a cooler climate state in the older layer of section 228_4 might drive the observed age- $\delta^{18}\text{O}_{\text{ice}}$ relationship, but more evidence across a range of proxies is needed to confirm this hypothesis. However, in this study, the dominant basal signal prevents such an analysis on the greenhouse gas, chemistry, and particle measurements.

4.2.4 Variation in basal material over small spatial scales

The concentration of basal material in distinct layers provides useful context for the interpretation of Allan Hills records. Fig. 10 shows data from this study within the context of other published data from the bottom 30 m of the ALHIC1901 core. While most high particle concentrations (Choi et al., 2026), depleted $\delta^{13}\text{C}-\text{CO}_2$, depleted $\delta\text{O}_2/\text{N}_2$ and elevated CO_2 (Marks-Peterson et al., 2026) measurements occur in the bottom 5 m of the core, there are sections of ice higher above the bed which exhibit the same indications of presumed basal interaction. One example is at 7.5 m above the bed, where both particle concentrations and $\delta^{13}\text{C}-\text{CO}_2$ appear similar to the 5 m closest to the bed. Marks-Peterson et al. (2026) also identified altered CO_2 concentrations between 120–140 m depth. We propose that the observations of localized alteration, even tens of meters above the bed, may be due to distinct layers with higher concentrations of basal material similar to those observed in this study. Based on recent work (Stoll et al., 2026), simple shear is dominant in (at least) the deepest 15 m of ALHIC1901 with small-scale folding observed. Jacobson and Waddington (2004) demonstrate that a layer perturbation in a simple shear regime will become a recumbent fold and elongate in the direction of flow, eventually forming layers parallel to the shear direction. Together with the steep bedrock and different concentrations of impurities, folding on different scales may be transporting basal material to ice layers meters, or tens of meters, above the bed. Notably, Hudak et al. (2026) identify a repeated sequence at 42–45 m depth in the nearby ALHIC2201 core that is consistent with the same folding mechanism, indicating these folding processes may also be relevant further above the bed.

This study also indicates that layers with more basal material do not all share the same climate origin. While low $\delta^{18}\text{O}_{\text{ice}}$ is associated with elevated CO_2 in section 228_4, the opposite pattern is observed in section 230_4 (Appendix Fig. D1). Likewise, Marks-Peterson et al. (2026) identify altered CO_2 across a range of $\delta^{18}\text{O}_{\text{ice}}$ and ages. These results suggest that basal material is not entrained upwards from a homogeneous basal ice packet but rather is integrated into multiple distinct packets of ice.

4.3 Future work at the Allan Hills and beyond

By identifying bands of concentrated basal material, this study highlights the need for a deeper understanding of the mechanisms responsible for entraining basal material into ALHIC1901 and other Allan Hills ice cores. Discrete layers of integrated



basal material have been observed in both Greenland (e.g., Bender et al., 2010; Souchez et al., 1994) and Antarctic ice cores (e.g., Gow et al., 1979). Related research has also explored mechanisms for the mechanical incorporation of basal material (Souchez et al., 1995) through flow-induced mixing. The subject of basal material in ice cores is also a focus of ongoing analysis at the Beyond EPICA ice core site (Westhoff et al., 2026). More work integrating geophysical studies, ice core analysis including physical properties, and modeling is needed to constrain the nature of basal material integration at the Allan Hills. Key questions include the timescale of mixing and folding, whether these processes occur locally or upstream, and how far above the bed these impacts extend.

The methods presented in this paper also have broad application to ongoing work in cores where dipping layers may result in significant depth offsets across the width of the core. Future work will apply these techniques to layering in other Allan Hills cores, at younger and older ages, and at varying distances from the bed. This will enable the high-resolution, multi-measurement constraints on layering presented in this study to be extended to the full range of Allan Hills ice-core samples.

The impacts of complicated and disturbed stratigraphy may be most pronounced at the Allan Hills because of the glaciological setting, the strong thinning, and the large-diameter ice cores, but traditional deep ice cores are not immune from similar complications. Folding, steeply dipping layers and/or hiatuses have been identified in the near-bed ice of many deep ice cores. Examples includes GRIP, GISP2 (Taylor et al., 1993), and NEEM (NEEM community members, 2013) in Greenland and EDML (Faria et al., 2010), SkyTrain (Mulvaney et al., 2023), Dome Fuji (Kawamura et al., 2017), Byrd (Gow et al., 1979), and Roosevelt Island (Lee et al., 2020a) in Antarctica. Even with small-diameter cores, measurements on either side of a core may be offset in age where there is steeply dipping layering or folding. Even moderate layer dip can produce substantial age offsets across a typical ice-core diameter. For example, in ice with a 20 kyr m^{-1} age density, similar to projections for the Beyond EPICA core (Chung et al., 2025), a 50° dip across a 10 cm diameter core would result in an effective age difference of over 1 kyr; comparable dips have been observed at Dome Fuji (Kawamura et al., 2017). This will be particularly relevant in recent and future oldest ice cores where layering in the deepest ice is both valuable and highly thinned, and so any effect from dipping layers might have an outsized impact on aligning parallel sets of measurements on a common age scale.

5 Conclusions

This paper provides the first high-resolution, multi-proxy characterization of “long snapshot” Allan Hills ice. We identify strong sub-meter-scale layering in 1.3–1.8 Ma ice near the bed of the ALHIC1901 core across multiple proxies. By integrating particle, chemistry, and gaseous measurements, we conclude that layering in the near-basal portion of this core is likely driven by integration of basal material, via the inter-layering of distinct ice units with different concentrations of basal material. Our findings point to a complex pattern of layering in the near-basal ice, with heterogeneous integration of basal material across multiple layers of distinct compositions. Our observations of discrete layers with varying levels of impurities also help to reconcile the presence of discrete packets with altered CO_2 values even tens of meters above the bed, sandwiched by otherwise pristine ice.



To achieve these results, we developed novel approaches to sampling, using 3D ECM layer orientations to coordinate sam-
pling of complex ice core stratigraphy. Large samples ($^{40}\text{Ar}_{\text{atm}}$ dating samples and CH_4 samples in section 228_4) were cut
at angles to match the ice stratigraphy. This has the benefit of minimizing the averaging which occurs within each sample,
enabling samples to fall more completely within certain layers, but comes at the cost of slower sample preparation. Our second
approach relies on dip-adjusting conventional straight sticks of rectangular samples. While this does result in more averaging
across layers, it is faster and similar to conventional ice core processing. As we have done here, we suggest that the first, more
complex method is most appropriate for large-volume samples of highly thinned ice, where the sample size approaches or
exceeds layer thicknesses in the ice. For small samples like water isotopes, CO_2 , chemistry, and particles, as well as for CFA
measurements, dip-adjusting a conventional sampling plan is the logical strategy.

Code and data availability. Data will be published on the US Antarctic Program Data Center and code will be posted with a stable Zenodo
DOI at publication. Both code and data are currently available on GitHub (https://github.com/liamkirkpatrick/alhic1901_deeplayering)



665 Appendix A: Water isotope standards

Table A1. University of Washington Δ^* IsoLab reference waters used for normalization to the VSMOW-SLAP scale.

Standard Name	δD (‰ vs. VSMOW)	$\delta^{18}O_{ice}$ (‰ vs. VSMOW)
WW	-268.30	-33.82
WGW	-318.82	-39.77
SPS	-365.01	-46.96
SPG	-435.31	-55.39

Appendix B: Impact of gas loss on $\delta\text{O}_2/\text{N}_2$

The slope of the gravitation-corrected $\delta\text{O}_2/\text{N}_2$ and $\delta\text{Ar}/\text{N}_2$ ($\delta_{\text{grav}} = \delta - \Delta m \cdot \delta^{15}\text{N}$) relationship represents fractionation due to gas loss, while vertical offsets (in the $\delta\text{O}_2/\text{N}_2$ dimension) represent real changes in $\delta\text{O}_2/\text{N}_2$ (Yan et al., 2021a). While the number of samples is small, the slope of the relationship in section 228_4 (blue points) is similar to that seen in pristine (i.e., samples not impacted by oxidation) measurements (yellow points) from Marks-Peterson et al. (2026). However, as shown in Fig. B1, measurements from this study exhibit much lower $\delta\text{O}_2/\text{N}_2$ than pristine measurements with similar $\delta\text{Ar}/\text{N}_2$. This result suggests that, while gas loss has occurred, the measurements in section 228_4 have depleted $\delta\text{O}_2/\text{N}_2$.

Notably, the measurements of $\delta\text{O}_2/\text{N}_2$ in this study are 5-13 ‰ less depleted than the three oxidation-impacted measurements in Marks-Peterson et al. (2026), while the CO_2 concentrations are similar. As shown in Fig. B1, $\delta\text{Ar}/\text{N}_2$ measurements indicate this is the result of gas loss, but is also indicative of a real change in $\delta\text{O}_2/\text{N}_2$. The number of data points in both the existing data ($n=3$) and this study ($n=8$) are small, but this may point to variation in an O_2 consumption process not related to CO_2 production within the deepest portions of the ALHIC1901 core.

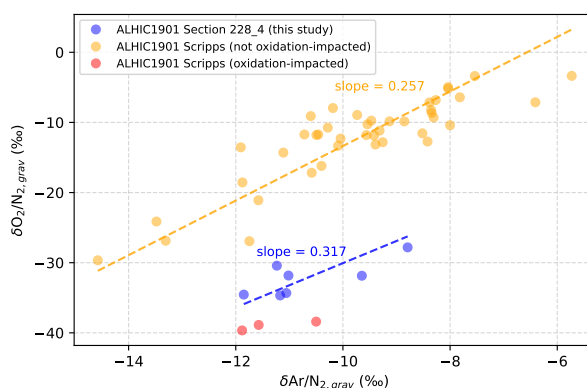


Figure B1. Overview of ALHIC1901 $\delta\text{O}_2/\text{N}_2$ and $\delta\text{Ar}/\text{N}_2$ data measured at Scripps from Marks-Peterson et al. (2026) and this study. The strong correlation between $\delta\text{O}_2/\text{N}_2$ and $\delta\text{Ar}/\text{N}_2$ seen in the yellow samples (not oxidation impacted) reflects fugitive O_2 and Ar loss during firn close-off, drilling and storage. The deep basal samples (red, blue dots) are vertically offset from the fugitive gas loss curve, suggesting additional oxygen consumption in the ice.

Appendix C: $\delta\text{O}_2/\text{N}_2$ and CO_2

There is evidence for a strong relationship between the consumption of O_2 (and resulting depletion of $\delta\text{O}_2/\text{N}_2$) and the production of CO_2 in Allan Hills data. As shown in Fig. C1, the $\delta\text{O}_2/\text{N}_2$ measurements proximal to elevated CO_2 measurements in the ALHIC1901 core have significantly more depleted $\delta\text{O}_2/\text{N}_2$ than those alongside pristine CO_2 measurements (Marks-Peterson et al., 2026). We calculate the average CO_2 in dip-adjusted space for each $\delta\text{O}_2/\text{N}_2$ sample in this study. The result is elevated CO_2 alongside depleted $\delta\text{O}_2/\text{N}_2$, consistent with other measurements from this core.

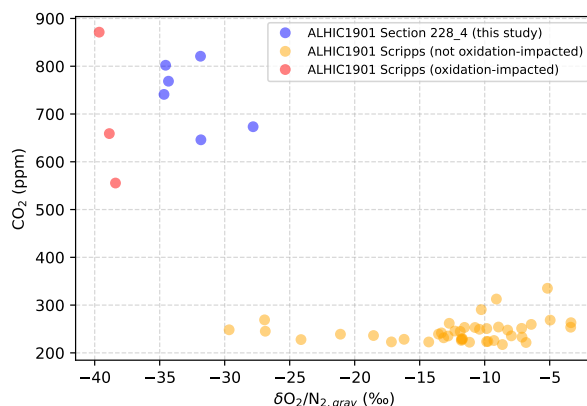


Figure C1. Here we plot pairs of $\delta\text{O}_2/\text{N}_2$ and CO_2 measurements from Marks-Peterson et al. (2026) alongside data from this study. Note that samples impacted by the O_2 oxidation process (red dots) show depleted $\delta\text{O}_2/\text{N}_2$ and elevated CO_2 , while samples which are not impacted (yellow dots) have much lower CO_2 and less depleted $\delta\text{O}_2/\text{N}_2$. Samples from this study (blue dots) appear consistent with other impacted measurements, exhibiting both depleted $\delta\text{O}_2/\text{N}_2$ and elevated CO_2 .

Appendix D: CO_2 and $\delta^{18}\text{O}_{\text{ice}}$

685 Here we explore the relationship between CO_2 and $\delta^{18}\text{O}_{\text{ice}}$ in both sections 228_4 and 230_4. We calculate the average $\delta^{18}\text{O}_{\text{ice}}$ value within the dip-adjusted range represented by each individual CO_2 value. Note that we find more elevated CO_2 with more depleted $\delta^{18}\text{O}_{\text{ice}}$ in section 228_4, but the opposite relationship in section 230_4.

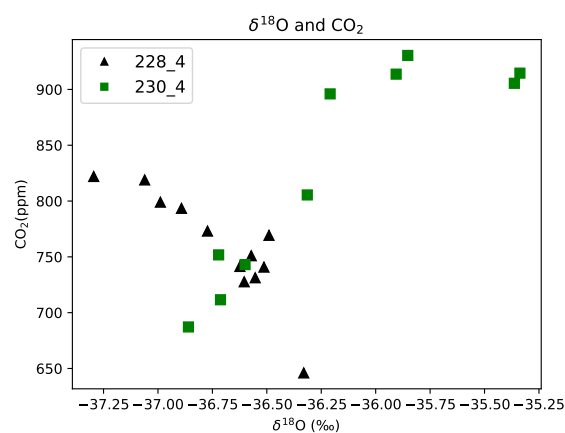


Figure D1. Sections 228_4 and 230_4 show opposite relationships between CO_2 and $\delta^{18}\text{O}_{\text{ice}}$; in section 228_4 elevated CO_2 is associated with more depleted $\delta^{18}\text{O}_{\text{ice}}$, while in section 230_4 higher CO_2 is associated with more enriched $\delta^{18}\text{O}_{\text{ice}}$.



Author contributions. Conceptualization: LRK, JIC, TJF. Methodology: LRK, JIC. Software: LRK. Visualization and formal analysis: LRK, JIC. Investigation (CFA, IC and ICP-MS prep): JIC, LT, UAJ, ECO. Investigation (Coulter Counter): JIC, LT, BGK. Investigation (ICP-MS):
690 MJH. Investigation (IC): EVK, DB, JCD. Investigation (water isotopes): LRK, LD, HLB, AJS, EJS. Investigation (GHG): AChoi, JMP, MK, CB, EB. Investigation (major gas isotopes and elemental ratios): RHS, CB. Investigation ($^{40}\text{Ar}_{\text{atm}}$ dating): VH, JH. Investigation (particles interpretation): ACarter, SA, AChoi, JIC, BGK. Investigation (physical properties): NS. Investigation (fieldwork): ACarter, SS, JMP, JH. Writing (original draft): LRK, JIC. Writing (review and editing): all authors.

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