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1 **Measurements of ^{14}C in ancient ice from Taylor Glacier, Antarctica constrain in situ**
2 **cosmogenic $^{14}\text{CH}_4$ and ^{14}CO production rates**

3
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26

27 **Abstract**

28

29 Carbon-14 (^{14}C) is incorporated into glacial ice by trapping of atmospheric gases as well
30 as direct near-surface in situ cosmogenic production. ^{14}C of trapped methane ($^{14}\text{CH}_4$) is a
31 powerful tracer for past CH_4 emissions from "old" carbon sources such as permafrost and
32 marine CH_4 clathrates. ^{14}C in trapped carbon dioxide ($^{14}\text{CO}_2$) can be used for absolute
33 dating of ice cores. In situ produced cosmogenic ^{14}C in carbon monoxide (^{14}CO) can
34 potentially be used to reconstruct the past cosmic ray flux and past solar activity.
35 Unfortunately, the trapped atmospheric and in situ cosmogenic components of ^{14}C in
36 glacial ice are difficult to disentangle and a thorough understanding of the in situ
37 cosmogenic component is needed in order to extract useful information from ice core ^{14}C .
38 We analyzed very large (≈ 1000 kg) ice samples in the 2.26 – 19.53 m depth range from
39 the ablation zone of Taylor Glacier, Antarctica, to study in situ cosmogenic production of
40 $^{14}\text{CH}_4$ and ^{14}CO . All sampled ice is > 50 ka in age, allowing for the assumption that most
41 of the measured ^{14}C originates from recent in situ cosmogenic production as ancient ice is
42 brought to the surface via ablation. Our results place the first constraints on cosmogenic
43 $^{14}\text{CH}_4$ production rates and improve on prior estimates of ^{14}CO production rates in ice.
44 We find a constant $^{14}\text{CH}_4 / ^{14}\text{CO}$ production ratio (0.0076 ± 0.0003) for samples deeper
45 than 3 m, which allows the use of ^{14}CO for correcting the $^{14}\text{CH}_4$ signals for the in situ
46 cosmogenic component. Our results also provide the first unambiguous confirmation of
47 ^{14}C production by fast muons in a natural setting (ice or rock) and suggest that the ^{14}C
48 production rates in ice commonly used in the literature may be too high.

49

50 **1. Introduction**

51

52 ***1.1 The potential and challenges of ^{14}C in polar ice***

53

54 The carbon-14 (^{14}C) content of trace gases in glacial ice is a potentially useful tracer in
55 several paleoclimatic and geochemical applications. ^{14}C of methane ($^{14}\text{CH}_4$) in ancient air
56 is an indicator of the "fossil" (^{14}C – depleted) fraction of the atmospheric methane budget
57 (Petrenko et al., 2009). Ice core $^{14}\text{CH}_4$ measurements can identify changes in emissions
58 from sources such as marine methane clathrates and thawing permafrost at past times of
59 global warming. Marine methane clathrates and permafrost have been proposed as
60 potentially important atmospheric methane sources both for past warming events and for
61 the future as a result of anthropogenic global warming (e.g., Kennett et al., 2000;
62 O'Connor et al., 2010; Walter et al., 2007a; Walter et al., 2007b). ^{14}C of CO_2 ($^{14}\text{CO}_2$) in
63 ice cores is potentially useful as a dating tool (Andree et al., 1984; Van de Wal et al.,
64 1990) and could serve to improve these valuable records of past environmental change
65 with absolute age constraints.

66

67 In addition to being trapped with C-containing atmospheric gases, ^{14}C is also produced in
68 situ in glacial ice by cosmic rays (e.g., Lal et al., 1987). A significant fraction of this in
69 situ cosmogenic ^{14}C appears to form carbon monoxide (^{14}CO) (e.g., Lal et al., 2000; van
70 der Kemp et al., 2002), and the in situ cosmogenic ^{14}CO is expected to be dominant over
71 the trapped atmospheric ^{14}CO content (e.g., van Roijen et al., 1995). This potentially
72 allows ^{14}CO in ice cores to be used as a tracer for the past cosmic ray flux. The solar
73 magnetic field modulates the flux of cosmic rays (e.g., Masarik and Beer, 1999). The
74 cosmic ray flux history reconstructed from ^{14}CO in polar ice cores could therefore be a
75 useful indicator of past solar activity and potentially even solar irradiance (e.g.,
76 Steinhilber et al., 2009). Such reconstructions would require ice coring sites with well-
77 constrained past snow accumulation rates and a thorough understanding of ^{14}C retention
78 in the firn column (firn is the thick compacted snow layer at the top of an ice sheet; see
79 Section 1.2).

80

81 At high geomagnetic latitudes where polar ice cores are drilled, in situ ^{14}C production
82 rates are insensitive to variations in the geomagnetic field strength (Lal et al., 2005). This
83 gives ice core in situ ^{14}C an important advantage over a tracer like meteoric ^{10}Be , which
84 has been widely used to study past solar activity (e.g., Muscheler et al., 2007). Tracers
85 like meteoric ^{10}Be , which are produced globally in the atmosphere, also have the added
86 complication of variability in transport and deposition (e.g., Field et al., 2006). ^{14}C of
87 atmospheric CO_2 , which has also been used to study past cosmic ray flux and solar
88 activity (e.g., Beer et al., 1988; Knudsen et al., 2009) suffers from uncertainties
89 associated with temporal variations in the global carbon cycle. None of these
90 complications are present for in situ cosmogenic ^{14}C in ice.

91

92 While it is clear that both the trapped atmospheric and in situ cosmogenic ^{14}C
93 components in glacial ice are highly promising as geochemical tracers, they are present in
94 a combined form and impossible to separate analytically. The in situ component of ^{14}C
95 must be thoroughly understood before useful information can be extracted from ice core
96 ^{14}C measurements. This understanding must include cosmogenic production rates of ^{14}C
97 in ice and their depth dependence, the efficiency of retention of produced ^{14}C in glacial
98 firn, the partitioning of in situ cosmogenic ^{14}C between $^{14}\text{CO}_2$, ^{14}CO and $^{14}\text{CH}_4$ and the
99 relative magnitude of the cosmogenic ^{14}C signal as compared to the trapped atmospheric
100 signal. Presently, our understanding of these processes is, at best, incomplete.

101

102

103 ***1.2 Polar firn and the firn-ice transition***

104

105 Processes in polar firn influence the retention of in situ cosmogenic ^{14}C that is produced
106 near the surface of ice sheets, and are relevant to some aspects of discussion of our
107 results. We therefore provide a basic introduction to polar firn and the firn-ice transition
108 for readers not closely familiar with ice core studies; a more comprehensive introduction
109 can be found in Buizert et al. (2013). The top $\approx 50 - 120$ m of an ice sheet are composed
110 of compacted snow; this layer is termed the firn. The firn is sponge-like in its structure
111 and is comprised of the ice matrix and the porosity (air-filled space). At snow

112 accumulation zones, snow layers move downward with time through the firn column.
113 Densification of the layers accompanies this downward movement as the porosity volume
114 is reduced (Herron and Langway, 1980). The porosity is interconnected (“open porosity”)
115 for most of the firn thickness, allowing gases to move freely and exchange with the
116 atmosphere, primarily by molecular diffusion (e.g., Schwander et al., 1988). As
117 densification continues with increasing depth, a significant fraction of the air successively
118 becomes trapped in bubbles in the ice (“closed porosity”). Eventually, some continuous
119 impermeable ice layers form, impeding further gas diffusion and exchange with the free
120 atmosphere (e.g., Schwander et al., 1993). This depth horizon is known as the “lock-in
121 depth”, and is typically found $\approx 5 - 15$ m above the depth at which all of the air is fully
122 sealed into bubbles and the firn zone ends.

123

124

125 *1.3 Production, retention and partitioning of in situ cosmogenic ^{14}C in glacial ice*

126

127 ^{14}C is produced by secondary cosmic rays from ^{16}O directly in the ice lattice in relatively
128 shallow glacial ice in accumulation zones (e.g., Jull et al., 1994) as well as in ablation
129 zones (e.g., van der Kemp et al., 2002). Confirmed production mechanisms include
130 energetic neutrons (e.g., Lal et al., 1990) as well as negative muon capture (van der Kemp
131 et al., 2002, Lupker et al., 2015). Production via fast muon interactions with ^{16}O in quartz
132 has been demonstrated in the laboratory (Heisinger et al., 2002a) but not confirmed in a
133 natural setting. Production of ^{14}C by thermal neutron capture from ^{14}N in air trapped in
134 the ice has been shown to be insignificant (Buizert et al., 2012). Production rates decrease
135 exponentially with depth for the neutron mechanism:

$$136 \quad P_n(z) = P_n^0 e^{\left(\frac{-\rho z}{\Lambda_n}\right)} \quad (1)$$

137 where P_n^0 is the production rate at the surface (^{14}C atoms $\text{g}^{-1} \text{a}^{-1}$), z the depth in cm, Λ_n the
138 neutron absorption mean free path in g cm^{-2} , and ρ the density of ice (0.92 g cm^{-3}). For
139 the two muon mechanisms, the relationship of production rate with depth is more
140 complex and has been formulated by Heisinger et al. (2002a, 2002b) and summarized
141 recently by Lupker et al. (2015). Muon production rates decrease more slowly with depth
142 than a simple exponential function. To obtain the overall ^{14}C production rate, the rates

143 from the three individual mechanisms are summed. Surface production rates and depth
144 dependence for each mechanism are shown in Table 1 and Figure 1.

145

146 The hot ^{14}C atom that is produced from ^{16}O reacts to form predominantly $^{14}\text{CO}_2$ and ^{14}CO
147 (e.g., Lal et al., 2000; van de Wal et al., 2007) Non-negligible in situ cosmogenic
148 production of $^{14}\text{CH}_4$ was also proposed (Petrenko et al., 2009) and recently confirmed
149 (Petrenko et al., 2013). To date, there is no agreement among prior studies on the
150 partitioning of in situ cosmogenic ^{14}C between the different gas species. Multiple studies
151 have included measurements of both ^{14}CO and $^{14}\text{CO}_2$, allowing for the ^{14}CO fraction of
152 total produced ^{14}C to be examined. The results show surprisingly high variability, with
153 estimated ^{14}CO fractions spanning almost the full 0.0 – 1.0 range, sometimes even within
154 the same study (e.g., Lal et al., 2001; van de Wal et al., 2007). It may therefore be
155 possible that the ^{14}C partitioning is different for each of the three cosmogenic production
156 pathways.

157

158 An important additional uncertainty is related to the retention of produced in situ
159 cosmogenic ^{14}C in glacial firn. If gases containing in situ cosmogenic ^{14}C are able to
160 escape from the ice grains into the interconnected open porosity, they can be lost to the
161 atmosphere. Prior studies have disagreed on the efficiency of retention of in situ ^{14}C , with
162 measurements from some laboratories showing that a large fraction of ^{14}C is retained
163 (e.g., Lal, et al., 2000; Lal, et al., 2001), while measurements from other laboratories
164 showed very little or no retention of ^{14}C (e.g., de Jong et al., 2004; Petrenko et al., 2013;
165 Smith et al., 2000). As discussed in detail in Petrenko et al. (2013), the studies that found
166 high ^{14}C retention in firn were likely affected by procedural artifacts associated with the
167 use of an acidified melt extraction approach. Low or no retention of in situ cosmogenic
168 ^{14}C in firn is therefore more likely.

169

170

171 ***1.4 Goals of this study***

172

173 The study presented in this paper utilized an outcrop of ancient (≈ 52 ka) glacial ice in the
174 ablation zone of Taylor Glacier, Antarctica to examine the in situ cosmogenic ^{14}C content
175 in the upper 20 m of the glacier. The relatively old age of the ice has allowed for
176 essentially complete decay of all in situ cosmogenic or trapped atmospheric ^{14}C retained
177 from the accumulation zone. The ^{14}C signal in this ice is therefore due mainly to in situ
178 cosmogenic production that occurs while the ice is brought up to the surface via the
179 process of ablation. Further, because this is solid ice, all of the in situ cosmogenic ^{14}C is
180 quantitatively retained. Finally, the sublimation-driven ablation rates at the site have been
181 well-characterized using stake measurements (Bliss et al., 2011). The combination of
182 these factors allows for measurements at this site to be used to improve estimates of ^{14}C
183 production rates and partitioning between different gas species.

184

185 Specifically, the goals of the study were as follows:

- 186 a) To further confirm in situ cosmogenic production of $^{14}\text{CH}_4$ using a site where a
187 relatively large signal is expected.
- 188 b) To provide the best estimates to date for in situ cosmogenic production rates of ^{14}CO
189 and $^{14}\text{CH}_4$ in glacial ice.
- 190 c) To examine the in situ cosmogenic $^{14}\text{CH}_4 / ^{14}\text{CO}$ ratio and explore the possibility that
191 ^{14}CO can be used to quantify the in situ cosmogenic component of $^{14}\text{CH}_4$ in younger ice,
192 ultimately allowing for the paleoatmospheric $^{14}\text{CH}_4$ signal to be reconstructed.

193

194 While the method used in this study for gas extraction from the ice was not optimized for
195 quantifying $^{14}\text{CO}_2$, these measurements were still obtained and have allowed for some
196 constraints on total ^{14}C production rates.

197

198

199 **2. Ice sampling and analytical methods**

200

201 ***2.1 Site description***

202

203 Taylor Glacier is located in the McMurdo Dry Valleys region of Antarctica (Figure 2).
204 Taylor Glacier ice is sourced from Taylor Dome, where a deep ice coring project has
205 revealed the existence of > 100 ka old ice (Steig et al., 2000). The very dry and windy
206 conditions over the glacier, combined with relatively slow ice movement produce an
207 unusually long (≈ 80 km) ablation zone in the lower part of the glacier. Most of the
208 ablation is via sublimation, with only the lowermost ≈ 5 km of the glacier regularly
209 experiencing summer melt (Bliss et al., 2011). Several recent studies have examined the
210 ice dynamics and mass balance of Taylor Glacier (Kavanaugh and Cuffey, 2009;
211 Kavanaugh et al., 2009a; Kavanaugh et al., 2009b) as well as established that ice
212 spanning a large range of ages (Holocene to at least 130 ka) is present at the surface of
213 the glacier (Aciego et al., 2007; Buizert et al., 2014). The ice drilling site for this study
214 ($77^{\circ}45.699'S$, $161^{\circ}43.179'E$, 527 m asl; Figure 2) is located along the center flowline of
215 the glacier 14.0 km from the glacier terminus. The ice flow velocity at this site is ≈ 10 m /
216 year, with an ablation rate of 0.196 ± 0.020 m / year (Bliss et al., 2011; Buizert et al.,
217 2012).

218

219

220 ***2.2 Ice core drilling and field melt-extraction of trapped air***

221

222 24-cm diameter ice cores in the 1.50 – 20.25 m depth range were recovered using the
223 Blue Ice Drill from Ice Drilling Design and Operations in Madison, WI, USA (Kuhl et
224 al., 2014). All drill components were pre-cleaned with ethanol and deionized water prior
225 to drilling, and no drilling fluid was used. Seven complete samples were obtained, with
226 average mid-depths of 2.26, 3.77, 5.27, 6.77, 10.02, 15.01 and 19.53 m. Each sample
227 spanned a 1.5 m depth range. A very large amount of ice is needed for precise $^{14}\text{CH}_4$
228 determinations (≈ 1000 kg, Petrenko et al., 2008a); for this reason cores for each sample
229 were obtained from 21 parallel boreholes. These boreholes were located along the center
230 flow–line of the glacier with an average spacing of 0.87 m between the boreholes. In
231 preparation for on-site gas extraction, the ice cores were cleaned by scraping all surfaces
232 with electropolished stainless steel chisels.

233

234 Ice cores were melted on-site to extract the trapped gases using the large-volume clean
235 sampling apparatus (“melter”) described in Petrenko et al. (2008a) and Petrenko et al.
236 (2013). Modifications from the gas extraction procedure described in Petrenko et al.
237 (2013) were as follows. During the melter headspace flush–evacuation sequence (prior to
238 ice melting), ultrapure air ($[\text{CH}_4] < 10$ ppb, $[\text{CO}] < 10$ ppb, $[\text{CO}_2] < 1$ ppm) was used
239 instead of ultrapure N_2 and Ar, and the flush was done 3 times. After the final flush, the
240 ice melter headspace was evacuated for 15 min after the pressure reached a value 1 torr
241 above the expected vapor pressure over ice. Unlike in the Petrenko et al. (2013) firm
242 study, no carrier gas was needed with Taylor Glacier ice; instead, the air trapped in the
243 ice cores served as the carrier for in situ cosmogenic ^{14}C . The gas recirculation step
244 (needed to equilibrate the in situ cosmogenic ^{14}C between the melt water and headspace)
245 was 30 min long. On average, 386 kg of ice per melt–extraction was used, and air from 3
246 melt–extractions was combined into a single sample.

247

248 The 30 min duration of the recirculation step (to allow for equilibration of trace gases
249 between the water and the headspace) was established at the start of the field season as
250 follows. Melt water from a sample extraction was purged with ultrapure air, and the
251 headspace was evacuated. One of the standard gases was introduced into the headspace
252 (to a pressure mimicking a sample extraction), and the recirculation of gas through the
253 bubbler was started. At five-minute intervals, an aliquot of the gas was analyzed for CO_2
254 mole fraction ($[\text{CO}_2]$) using a LI-820 infrared CO_2 analyzer from LI-COR[®] Biosciences
255 (Figure S1a in the Electronic Annex). The time needed until $[\text{CO}_2]$ was no longer
256 appreciably changing was set as the procedural recirculation time. To further test that gas
257 equilibration time is the same when the gas is initially dissolved in the water (which may
258 be the case for in situ cosmogenic ^{14}C), the headspace was again evacuated, and ultrapure
259 air was introduced. The recirculation was started again, and the gradual evasion of CO_2
260 from the water into the headspace was monitored, confirming that 30 min was again
261 sufficient (Figure S1b).

262

263 In an attempt to obtain a better characterization of in situ cosmogenic $^{14}\text{CO}_2$, we
264 measured pH of the melt water following the gas extraction. This measurement,

265 combined with the known $[\text{CO}_2]$ in the headspace in principle allows for quantification of
266 the total amount of inorganic carbon in the system (CO_2 gas, dissolved CO_2 , bicarbonate
267 ion (HCO_3^-) and carbonate ion (CO_3^{2-})). Following the transfer of sample air from the
268 melter headspace into storage canisters, the melter headspace was vented with CO_2 -free
269 air to avoid interference of atmospheric CO_2 with the pH measurement. pH
270 measurements were made inside a box flushed with CO_2 -free air, with the melt water
271 plumbed directly into this box. An Orion™ ROSS Ultra™ Triode™ was used for the
272 measurement, in combination with a set of low ionic strength buffers for electrode
273 calibration as well as an ionic strength adjustor solution for pure water. Approximately 1
274 L of melt water was flushed through the tubing prior to taking the pH sample aliquots,
275 and the measurement was done in triplicate on separate aliquots of melt water.

276

277

278 ***2.3 Tests of melt-extraction procedural effects***

279

280 Two types of procedural tests on the large-volume gas melt–extraction system were
281 conducted to investigate effects of added extraneous carbon on measured $^{14}\text{CH}_4$ in the
282 samples. As previously described for this system (Petrenko et al., 2008a; Petrenko et al.,
283 2013), tests involving introducing standard gases over the melt water from sample ice
284 were conducted in the field. For this study, two different standard gases were used. Both
285 gases contained ≈ 490 ppb of CH_4 , to mimic the average CH_4 mole fraction ($[\text{CH}_4]$)
286 expected in the sampled ice. One of the gases contained CH_4 derived from a fossil source
287 and expected to be essentially free of ^{14}C (“ ^{14}C -dead standard”), while the other gas
288 (“ ^{14}C -modern standard”) contained modern atmospheric CH_4 (from Niwot Ridge, CO,
289 USA), with an expected ^{14}C activity of ≈ 130 percent modern carbon (pMC; as defined in
290 Stuiver and Polach (1977)). Both gases also contained CO_2 (≈ 215 ppm) and CO (≈ 85
291 ppb) to mimic the expected levels in the sample ice. The gases also contained ≈ 100 ppm
292 each of Kr and Xe, for use as tracers of gas partitioning between the melt water and the
293 headspace in the ice melter (Petrenko et al., 2013). The pH measurements described
294 above were also performed on the melt water following the field procedural tests.

295

296 Using two standard gases with distinct ^{14}C activities allows for complete
297 characterization of both the mass and ^{14}C activity of extraneous carbon added during
298 sample processing. The air samples derived from these field procedural tests undergo
299 treatment identical to that of air derived from the actual ice samples, allowing the tests to
300 characterize extraneous carbon from all steps of sample processing.

301

302 As described by Petrenko et al. (2013), the field procedural tests successfully mimic all
303 aspects of sample processing except for the ice melting step. The melting step results in
304 heating of the walls of the aluminum melt-extraction chamber (up to 50°C) and
305 potentially increased outgassing of CH_4 . For this reason, additional procedural tests were
306 conducted at SIO to specifically investigate the effects of ice melting. More than 2000 kg
307 of gas-free ice was obtained from a commercial supplier that provides clear ice blocks
308 for carving ice sculptures. Very slow uni-directional freezing of this ice over several days
309 allows for all gases to be excluded from the ice lattice. 16 sub-samples of this gas-free
310 ice were analyzed for $[\text{CH}_4]$ at OSU using the usual “blank” ice procedure (Mitchell et
311 al., 2013); these analyses revealed no differences between this ice and gas-free ice
312 routinely made at the OSU laboratory, confirming that no significant amount of CH_4 was
313 present. The standard gases described above were used to provide the air during
314 simulated melt extractions with this gas-free ice, with the gas being introduced just
315 before the start of the melting step. One complete test sample was collected using the ^{14}C -
316 dead standard, and another test sample using the ^{14}C -modern standard; as in the field,
317 each sample was comprised of 3 melt-extractions. To ensure the most direct assessment
318 of procedural effects from the ice-melting step, further over-water procedural tests were
319 conducted at SIO immediately following the gas-free ice tests.

320

321 All of the over-water procedural tests included a water purging step prior to each
322 simulated extraction, in which ultrapure air ($[\text{CH}_4] < 10$ ppb) was introduced into the
323 melter via a bubbler manifold at the bottom (Petrenko et al., 2013) and evacuated from
324 the headspace via the top. The progress of the purge was monitored by directing part of
325 the evacuated airstream through a cryogenic drying trap into the LI-820 CO_2 analyzer.
326 The purge was conducted until measured CO_2 reached the baseline value determined by

327 flowing the ultrapure air directly into transfer lines and pumps, bypassing the melter
328 itself. Figure S1c in the Electronic Annex shows the typical time evolution of CO₂ in air
329 being pumped from the melter headspace during a purge.

330

331 Because the amount of in situ cosmogenic ¹⁴C in sampled ice was expected to be very
332 large relative to any ¹⁴C addition from the melt-extraction procedure, no tests to
333 quantify such procedural ¹⁴C addition were conducted.

334

335

336 ***2.4 Laboratory analyses and sample processing***

337

338 Air samples and procedural tests were first analyzed at SIO for [CH₄] by a gas
339 chromatograph (GC) with a flame ionization detector (FID), as well as for CFC-11, CFC-
340 12 and SF₆ (to constrain inclusion of modern ambient air) by GC-mass spectrometry
341 (GC-MS) (Mühle et al., 2007). The procedural tests were further analyzed for δ(Xe/N₂)
342 and δ(Kr/N₂) (Thermo MAT 253 isotope ratio MS (IRMS)) at SIO. CO₂ for subsequent
343 ¹⁴C analyses was extracted at SIO from ≈ 1 standard liter (SL) of air for all samples, field
344 procedural tests and standard gases (Leucker, 1998).

345

346 [CO] and [CO₂] in the samples, procedural tests, and standard gases were analyzed at
347 NIWA using a GC setup as described by Lowe et al. (1991) with a reducing gas detector
348 and a methanizer-FID, respectively. δ¹³CH₄ in the samples was analyzed by a continuous-
349 flow method at NIWA using an IsoPrime IRMS (Ferretti et al., 2005), with a correction
350 applied for krypton interference following Schmitt et al. (2013). δ¹³CH₄ in the standard
351 gases was analyzed by dual-inlet on a MAT 252 IRMS as described in Petrenko et al.
352 (2008b).

353

354 The air in the samples, procedural tests and standard gases was then processed to oxidize
355 CH₄ to CO₂; this CO₂ was captured for further handling, as described in Petrenko et al.
356 (2008b) and Petrenko et al. (2009). Unfortunately, one of the ¹⁴CH₄ samples (5.27 m) was
357 lost at this stage due to a problem with flame-sealing of sample CO₂ into a pyrex tube.

358 ≈ 20 SL of air was retained in each sample canister for ^{14}C analyses, and this remaining
 359 air was diluted with a gas containing 10.23 ppm of ^{14}C -depleted CO in ultrapure air to
 360 increase carbon mass for ^{14}C measurements. $\delta^{13}\text{C}$ in this high-[CO] gas was
 361 determined at NIWA following Brenninkmeijer (1993). After the dilution, the samples
 362 were processed to oxidize CO to CO_2 ; this CO_2 was captured for further handling, as
 363 described in Petrenko et al. (2008b) and Petrenko et al. (2009).

364

365 The resulting pure CO_2 samples (derived from CO_2 , CO or CH_4) were converted to
 366 graphite by using an excess of H_2 over a Fe catalyst in the conventional graphitization
 367 lines at ANSTO (Hua et al., 2004). We used Sigma-Aldrich Fe_2O_3 (mesh size
 368 unspecified, 99.999% purity). Before the graphitization, the iron oxide was reduced to Fe
 369 by reaction with ≈ 750 mbar pure H_2 at 600°C for 1 hour. By-product water vapor
 370 produced during the graphitization was efficiently trapped using a cold finger system at -
 371 80°C (Yang et al., 2013). Typical graphitization efficiency was high: $\approx 90\%$ for 20 μgC
 372 samples and $\approx 100\%$ for 70-100 μgC samples. Unfortunately, one further $^{14}\text{CH}_4$ sample
 373 (10.02 m) was lost at this stage while extracting sample CO_2 from the sealed pyrex tube.
 374 Graphite derived from all samples, procedural tests and standard gases was subsequently
 375 measured for ^{14}C by AMS on the 10 MV ANTARES accelerator at ANSTO (Fink et al.,
 376 2004). The final sample sizes (determined prior to graphitization) were $\approx 55 \mu\text{g C}$ for
 377 $^{14}\text{CO}_2$, $\approx 20 \mu\text{g C}$ for $^{14}\text{CH}_4$ and $\approx 85 \mu\text{g C}$ for ^{14}CO .

378

379

380 **3. Results**

381

382 The in-situ cosmogenic ^{14}C content in ice for the gas species X can be calculated as (see
 383 the Electronic Annex for a derivation):

$$\begin{aligned}
 {}^{14}\text{C}_X = & \frac{pMC}{100} \times e^{-\lambda(y-1950)} \times \frac{\left(1 + \frac{\delta^{13}\text{C}_X}{1000}\right)^2}{0.975^2} \times \\
 & \times 1.1694 \times 10^{-12} \times [X] \times \text{Air} \times \frac{1}{1000} \times \frac{1}{22.4} \times N_A
 \end{aligned} \tag{2}$$

384

385

386 where $^{14}C_X$ is the number of ^{14}C -bearing molecules of gas X per gram of ice, pMC is the
387 ^{14}C activity of the gas after all corrections, λ is the ^{14}C decay constant ($1.216 \times 10^{-4} \text{ a}^{-1}$), y
388 is the year of measurement, $\delta^{13}C_X$ is the $\delta^{13}C$ of the gas, 0.975 is a factor arising from
389 ^{14}C activity normalization to $\delta^{13}C$ of -25 ‰ (part of definition of pMC (Stuiver and
390 Polach, 1977)), 1.1694×10^{-12} is the $^{14}C / (^{13}C + ^{12}C)$ ratio corresponding to the absolute
391 international standard activity (AISA) (Hippe and Lifton, 2014), $[X]$ is the gas molar
392 fraction in extracted air after all corrections, Air is the air content of the ice in cc STP g^{-1} ,
393 22.4 is the number of STP liters of gas per mole, and N_A is the Avogadro constant.

394

395 The ^{14}C activity as well as the gas molar fraction measured in the samples can be altered
396 by addition of extraneous carbon at various stages of processing. These procedural effects
397 must be quantified and corrected for before the true ^{14}C content of the ice can be
398 determined. Detailed discussion of the corrections for all gas species as well as of the
399 determination of the air content is presented in the Electronic Annex. Briefly, all reported
400 uncertainties are 1σ unless otherwise specified, and uncertainties for calculated
401 quantities are determined by standard error propagation techniques. Sample $[CH_4]$ and
402 $^{14}CH_4$ were corrected for the effects of a) gas dissolution in the melter, b) overall
403 procedural effects based on procedural test results, c) a small amount of additional CH_4
404 produced during the melting step and d) inclusion of a small amount of ambient air via
405 fractures in the shallowest samples. Measured $[CH_4]$ and $^{14}CH_4$ values as well as values
406 after all corrections and the calculated $^{14}CH_4$ content in the ice are presented in Table 2
407 and Figure 3.

408

409 Sample $[CO]$ and ^{14}CO were corrected for the effects of a) dilution with the high- $[CO]$,
410 ^{14}C -depleted gas, b) in situ cosmogenic production of ^{14}CO in the sample canisters
411 during transport and storage, c) CO added during the melt-extraction and sample storage
412 and d) gas dissolution in the melter. Measured ^{14}CO values and the ^{14}CO content in the
413 ice calculated after all corrections are shown in Table 3 and Figure 3.

414

415 We consider the $^{14}CO_2$ analyses and results somewhat speculative because of problems
416 and uncertainties inherent in the melt-extraction approach. However, these results do add

417 some useful information to the discussion and were therefore included. We were only
418 able to constrain the upper limit of $^{14}\text{CO}_2$ present in the ice; see the Electronic Annex for
419 a detailed discussion. Table 4 and Figure 3 show the measured $^{14}\text{CO}_2$ activities as well as
420 estimated maximum ice $^{14}\text{CO}_2$ content.

421

422

423 **4. Discussion**

424

425 *4.1 Comparison with prior ice cosmogenic ^{14}C data quality and measurement methods*

426

427 Our method is currently the only one capable of $^{14}\text{CH}_4$ measurements in ice, because it
428 accommodates the very large sample size requirement. In the one prior study of in situ
429 cosmogenic $^{14}\text{CH}_4$ (at Summit, Greenland; Petrenko et al., 2013) the average relative
430 error on $^{14}\text{CH}_4$ determinations was 13%, while in the present work it is 4%. Most of the
431 improvement has resulted from larger $^{14}\text{CH}_4$ content in Taylor Glacier ice compared to
432 Greenland Summit firn, with additional improvements from better determinations of
433 effective air content in the ice as well as of procedural ^{14}C effects.

434

435 The ^{14}CO determinations were also improved as compared to Petrenko et al. (2013); with
436 a reduction in relative error from 22% to 2.6%. In this case, most of the improvement
437 arises from better-constrained dilution with ^{14}C -depleted CO , with additional
438 improvement due to larger ^{14}CO content in the ice. In comparison with the best prior
439 work that determined ^{14}CO in ablation zone ice samples with similar ^{14}CO content (van
440 der Kemp et al., 2002), our errors are an order of magnitude lower. This is mainly due to
441 the much larger sample size used in our work.

442

443 Petrenko et al. (2013) provided a detailed discussion of the general advantages of our
444 approach for in situ cosmogenic $^{14}\text{CH}_4$ and ^{14}CO determinations as compared to most
445 prior work; we summarize these briefly here. First, our air extractions are performed in
446 the field immediately after core recovery, eliminating the uncertainties associated with
447 non-negligible additional ^{14}C production during ice core transport and storage. Second,

448 melt-extraction combined with water – headspace gas equilibration ensures complete
449 recovery of ^{14}C from ice grains, which may be a problem with dry-extraction methods.
450 Finally, our method includes very thorough characterization of procedural ^{14}C effects for
451 all parts of sample processing.

452

453 Our method performed poorly for $^{14}\text{CO}_2$ because melt-extraction techniques inherently
454 lead to artifacts in CO_2 (see Electronic Annex for detailed discussion). However, our
455 $^{14}\text{CO}_2$ results are likely more reliable than the majority of prior work that used the melt-
456 extraction approach (Jull et al., 1994; Lal et al., 1990; Lal et al., 1997; Lal et al., 2000;
457 Lal et al., 2001). As discussed in detail in Petrenko et al. (2013), the use of acid in those
458 prior studies combined with inadequate procedural blank characterization likely resulted
459 in addition of considerable extraneous ^{14}C . Of the in situ cosmogenic $^{14}\text{CO}_2$ studies
460 available to date, we consider the studies done by dry extraction (e.g., Smith et al., 2000;
461 van de Wal et al., 2007) to be more reliable than any melt-extraction approaches,
462 including our own. However, even results from those studies harbor unknown
463 uncertainties from $^{14}\text{CO}_2$ production during ice core transport and storage, as well as from
464 possibly incomplete release of in situ cosmogenic $^{14}\text{CO}_2$ from ice grains during the
465 extraction.

466

467

468 ***4.2 Possible contribution of ^{14}C from the accumulation zone***

469

470 As discussed in Sections 1.1 and 1.3, accumulation-zone ice can acquire ^{14}C via both
471 trapping of atmospheric gases and in situ cosmogenic production. To assess whether the
472 sampled ice has any remnant signal of ^{14}C from the accumulation zone, both the age of
473 the ice and the possible cumulative in situ cosmogenic ^{14}C production at the
474 accumulation site need to be constrained.

475

476 Prior work on Taylor Glacier used measurements of δD of ice to establish a preliminary
477 age scale for ice outcropping along the center flowline shown in Figure 2 (Aciego et al.,
478 2007). Further samples collected along the center flowline during the 2009-10 Antarctic

479 season added constraints based on $[\text{CH}_4]$ and the isotopic composition of atmospheric O_2
480 ($\delta^{18}\text{O}_{\text{atm}}$) (Baggenstos, 2015). The oldest unambiguously dated ice (≈ 52 ka) identified
481 based on the preliminary analysis of the 2009-10 results was at 14 km from the glacier
482 terminus (Figure 2); this was the location chosen for our in situ cosmogenic ^{14}C study.
483 The center flowline sample taken at 5 m depth at this location yielded an average $\delta^{18}\text{O}_{\text{atm}}$
484 value of 0.247 ‰ and a $[\text{CH}_4]$ value of 492 ppb (Figure 4). Such a $\delta^{18}\text{O}_{\text{atm}}$ value is
485 plausible for gas ages of at least 49 – 56 ka (Severinghaus et al., 2009). The only younger
486 period when such a $\delta^{18}\text{O}_{\text{atm}}$ value occurs is during the Preboreal between 10 – 11 ka, but
487 the Preboreal is associated with $[\text{CH}_4] > 600$ ppb. The $[\text{CH}_4]$ values contained in this
488 low- $\delta^{18}\text{O}_{\text{atm}}$ section of the TG profile (14.0 km: 492 ppb; 13.7 km: 522 ppb; 13.4 km:
489 557 ppb; Figure 4) are also indicative of a Dansgaard-Oeschger type $[\text{CH}_4]$ oscillation.
490 Such an oscillation is present in Antarctic ice cores at 50 - 55 ka (Brook et al., 2005;
491 WAIS Divide Project Members, 2015). The $[\text{CO}_2]$ value of 212 ppm measured in the
492 14.0 km sample is also consistent with a gas age of 52 – 55 ka (Ahn and Brook, 2008).
493 We cannot identify another time interval during the last glacial when the same
494 combination of values for these 3 tracers re-occurs. Finally, the age we infer is in
495 agreement with the age obtained through ^{81}Kr radiometric dating of ice at this location,
496 which suggested an age of 46 ± 21 ka (Buizert et al., 2014).

497

498 Sample age increases with depth at most sampled locations on Taylor Glacier
499 (Baggenstos, 2015), including at our ^{14}C sampling site. On Figure 4, the $[\text{CH}_4]$ trend
500 measured in the large-volume samples was visually matched to the high-resolution $[\text{CH}_4]$
501 record from the WAIS Divide ice core (WAIS Divide Project Members, 2015) to arrive
502 at sample gas age estimates. The 2.26 m (youngest) sample has an estimated age of 50.4
503 ka, while the 19.53 m (oldest) sample has an estimated age of 53.4 ka. After 50.4 ka of
504 decay, the remaining $^{14}\text{CO}_2$, ^{14}CO and $^{14}\text{CH}_4$ from trapped air are much smaller than the
505 respective uncertainties on our measurements and can be ignored. Given the large age of
506 the samples, we also assume that we can ignore the contribution from in situ cosmogenic
507 ^{14}C produced in the accumulation zone. We test this assumption in Section 4.4, finding it
508 to be valid.

509

510

511 *4.3 $^{14}\text{CH}_4$ and $^{14}\text{CH}_4 / ^{14}\text{CO}$ ratio*

512

513 The values for $^{14}\text{CH}_4$ content (Table 2) in the ice provide further confirmation that in situ
514 cosmogenic production of $^{14}\text{CH}_4$ takes place. This confirmation is stronger in the present
515 study than in the earlier Petrenko et al. (2013) work, with more samples and higher
516 measured $^{14}\text{CH}_4$.

517

518 Table 5 shows the $^{14}\text{CH}_4 / ^{14}\text{CO}$ ratio. As can be seen, this ratio appears to be effectively
519 constant for all except the shallowest (2.26 m) sample. As Table 5 also shows, expected
520 ^{14}C contribution from the neutron mechanism is relatively small or negligible for all
521 except the 2.26 m sample. ^{14}C fraction from negative muon capture is relatively constant
522 and represents the largest component at all depths, while the contribution from fast
523 muons grows steadily with increasing depth. The expected ratio between ^{14}C from
524 negative muons and ^{14}C from fast muons decreases from 1.9 at 3.77 m to 1.0 at 19.53.
525 The constant $^{14}\text{CH}_4 / ^{14}\text{CO}$ ratio in this entire depth range suggests that this ratio (0.0076
526 ± 0.0003) is the same for both muon ^{14}C production mechanisms.

527

528 The significantly higher $^{14}\text{CH}_4 / ^{14}\text{CO}$ ratio for the 2.26 m sample suggests that the ^{14}C
529 partitioning is different for the neutron production mechanism, resulting in relatively
530 more $^{14}\text{CH}_4$. Due to the relatively small overall ^{14}C contribution from neutrons, our data
531 leave the neutron $^{14}\text{CH}_4 / ^{14}\text{CO}$ production ratio poorly constrained (see Figure 3). We
532 note, however, that a relatively higher $^{14}\text{CH}_4 / ^{14}\text{CO}$ production ratio for neutrons than for
533 muons is qualitatively consistent with the value observed in shallow Greenland firn by
534 Petrenko et al. (2013) (0.021 ± 0.005 for Sample 1, which was considered to be the most
535 reliable of the three $^{14}\text{CH}_4$ determinations). For the firn samples in the Petrenko et al.
536 (2013) study, 95% of all in situ cosmogenic ^{14}C was expected to be from the neutron
537 production mechanism.

538

539 The constant $^{14}\text{CH}_4 / ^{14}\text{CO}$ ratio for both muon production mechanisms means that the in
540 situ cosmogenic component of $^{14}\text{CH}_4$ can be accurately determined based on measured

541 ^{14}CO in ice that does not contain significant neutron-produced ^{14}C . This cosmogenic
542 component could then be corrected for, allowing for accurate reconstructions of
543 paleoatmospheric $^{14}\text{CH}_4$. The trapped atmospheric component of ^{14}CO is negligibly small
544 in comparison to ^{14}CO from in situ cosmogenic production. For example, we calculate
545 that trapped air from the Younger Dryas - Preboreal climate transition at 11.5 ka would
546 contain only 0.3 ^{14}CO molecules per gram of ice, or about 100 times less than what is
547 observed due to ablation-zone in situ cosmogenic production in Taylor Glacier ice at 10
548 m depth.

549

550 In addition to ablation-zone production, ice with ages much younger than 50 ka would
551 also contain significant ^{14}C from in situ cosmogenic production in the accumulation zone.
552 However, this ^{14}C would begin accumulating starting only at the lock-in depth (see
553 Section 1.2) where the first impermeable ice layers begin to form, and gas exchange with
554 the atmosphere and the upper part of the firn effectively stops (e.g., Schwander et al.,
555 1993). The lack of in situ cosmogenic ^{14}C accumulation above the lock-in depth is
556 supported by the results of our recent firn ^{14}C study at Summit, Greenland (Petrenko et
557 al., 2013), as well as by results of most other studies (e.g., de Jong et al., 2004; van der
558 Kemp et al., 2000) that avoided the problematic melt-acidify air extraction approach. At
559 the lock-in depth (typically $> 25\text{m}$ solid ice equivalent), ^{14}C production from the shallow-
560 penetrating neutron mechanism is minimal or non-existent. It can therefore be assumed
561 that effectively all of the in situ cosmogenic ^{14}C in sampled ice at Taylor Glacier (if taken
562 below $\approx 5\text{ m}$ in ablating ice) is from the muon production mechanisms. The determined
563 $^{14}\text{CH}_4 / ^{14}\text{CO}$ ratio could then be applied to make the correction for in situ cosmogenic
564 $^{14}\text{CH}_4$.

565

566

567 ***4.4 Discussion of observed ^{14}C production rates***

568

569 Figure 3 shows the $^{14}\text{CH}_4$ and ^{14}CO content in sampled ice after all corrections, as well as
570 the upper limit of the $^{14}\text{CO}_2$ and total ^{14}C content. We use these results to estimate surface
571 ^{14}C production rates for each species by each mechanism. This is accomplished via a

572 least-squares curve fit to the data using an expression that describes cumulative ^{14}C as a
 573 function of depth. The expression used for cumulative ^{14}C is described in full detail in the
 574 Electronic Annex; here we provide a more abbreviated discussion. For the neutron
 575 mechanism, an exponential depth dependence of production rate is assumed (Eq. 1),
 576 using a value of 150 g cm^{-2} for the mean free path Λ_n (Lal et al., 1987; van de Wal et al.,
 577 2007). Integrated over depth, this yields the following for cumulative ^{14}C from neutrons
 578 (Lal et al., 1990):

$$579 \quad {}^{14}C_n(z) = \frac{P_n^0 e^{-\rho z/\Lambda_n}}{\frac{\rho A}{\Lambda_n} + \lambda} \quad (3)$$

580 where A is the ice ablation rate (in cm a^{-1}) and other terms are as previously defined.
 581

582 For the muon mechanisms, we use the depth dependence as formulated by Heisinger et
 583 al. (2002a, 2002b) and summarized recently by Lupker et al. (2015). For the purpose of
 584 curve fitting, we successfully approximate the depth dependence for negative muon
 585 capture with a 2-term exponential, and with a 3-term exponential for fast muons (Figure
 586 S2 in Electronic Annex). For negative muon capture, integrating over depth yields an
 587 expression for cumulative ^{14}C that has 2 exponential terms equivalent in form to Equation
 588 3.

589

590 The expression for cumulative ^{14}C from fast muons is slightly more complex because,
 591 unlike for neutrons and negative muon capture, there is non-negligible ^{14}C production
 592 during long-term ice transport at large depths. Ice flow modeling suggests that our
 593 sampled ice travels at a depth of $\approx 300 - 500 \text{ m}$ through much of the length of Taylor
 594 Glacier (Buizert et al., 2012). In this case, the cumulative ^{14}C from fast muon production
 595 (whose depth dependence is approximated by a 3-term exponential) is given by:

$$596 \quad {}^{14}C_{\mu f}(z) = D e^{\lambda z/A} + P_{\mu f}^0 \left(\frac{f_{\mu f 1} e^{-\rho z/\Lambda_{\mu f 1}}}{\frac{\rho A}{\Lambda_{\mu f 1}} + \lambda} + \frac{f_{\mu f 2} e^{-\rho z/\Lambda_{\mu f 2}}}{\frac{\rho A}{\Lambda_{\mu f 2}} + \lambda} + \frac{f_{\mu f 3} e^{-\rho z/\Lambda_{\mu f 3}}}{\frac{\rho A}{\Lambda_{\mu f 3}} + \lambda} \right) \quad (4)$$

597

598 The Λ 's represent the effective absorption mean free path for each exponential term and
 599 the f 's represent the fractional contribution of each term to the total production rate at the
 600 surface. D is a constant (in ^{14}C atoms / g) that is related to the ^{14}C content at a reference

601 depth z_{deep} . For the case of long-term transport at z_{deep} , it can be shown that D is directly
602 proportional to the surface production rate $P_{\mu f}^0$ (see Electronic Annex). In the resulting
603 expression for cumulative ^{14}C from all mechanisms as a function of depth, each of the 7
604 exponential terms is proportional to one of the surface production rates (P_n^0 , $P_{\mu-}^0$, or $P_{\mu f}^0$)
605 and the three production rates are the only free parameters that are determined by the
606 least-squares curve fit.

607

608 One major source of uncertainty in determining surface ^{14}C production rates from our
609 data is the choice of average depth for long-term ice transport between the accumulation
610 zone and our ablation site. This mainly affects $P_{\mu f}^0$. For the purposes of curve fitting, we
611 assume 400 m for the average depth of long-term transport. We incorporate the
612 uncertainty in this depth by also running scenarios where the average depth of transport is
613 as shallow as 300 m or as deep as 500 m ($P_{\mu f}$ is 3x greater at 300 m than at 500 m), and
614 including the full range of resulting $P_{\mu f}^0$ values in the reported error.

615

616 Another major source of uncertainty in determining surface ^{14}C production rates from our
617 data is the ice ablation rate. Ablation stake measurements yield $0.196 \pm 0.020 \text{ m a}^{-1}$ for
618 the sampled part of the glacier (Bliss et al., 2011; Buizert et al., 2012). However, the
619 ablation stakes represent at most 8 years of measurements (2002 – 2010), whereas longer-
620 term rates are needed for the cumulative ^{14}C calculations. If we consider that the ice rises
621 from a depth of $\approx 400 \text{ m}$, then the last ≈ 2000 years are in principle of interest, although
622 ablation rate variations become less important further back in time due to lower
623 production rates at greater depths.

624

625 Stronger winds and warmer temperatures result in faster ablation rates on Taylor Glacier,
626 with katabatic wind speed being the dominant variable for much of the glacier length
627 (Bliss et al., 2011). A weather record from nearby McMurdo Sound suggests that 2002 –
628 2010 was 0.4°C warmer than 1957 – 2003 (National Climatic Data Center, 2015). It is
629 likely that the Dry Valleys – McMurdo Sound area was $\approx 2^\circ\text{C}$ cooler and experienced
630 somewhat stronger katabatic winds during the Little Ice Age (Bertler et al., 2011; Rhodes

631 et al., 2012). It is also possible that, during the initial phase of its ascent from 400 m to
632 the surface, sampled ice was traveling in a zone with higher ablation rates of $\approx 0.25 \text{ m a}^{-1}$
633 (Buizert et al., 2012). To account for all these factors, we use a doubled uncertainty of \pm
634 0.04 m a^{-1} for the long-term ablation rate. We used curve fit scenarios that span the full
635 range of possible ablation rates to estimate the corresponding contribution to
636 uncertainties in surface production rates.

637

638 Our data provide better constraints for production rates for ^{14}CO than for $^{14}\text{CH}_4$, due to
639 the larger number of ^{14}CO data points. P_n^0 , $P_{\mu^-}^0$ and $P_{\mu f}^0$ for ^{14}CO were all determined by a
640 least-squares curve fit as discussed above (Figure 3a). The constant $^{14}\text{CH}_4/^{14}\text{CO}$
641 production ratio for the muon mechanisms discussed above provides an additional
642 constraint for $P_{\mu^-}^0$ and $P_{\mu f}^0$ for $^{14}\text{CH}_4$. For $^{14}\text{CH}_4$, we calculated negative muon capture
643 and fast muon production rates by multiplying the corresponding ^{14}CO production rates
644 by 0.0076 ± 0.0003 ; P_n^0 was the only completely free parameter in the $^{14}\text{CH}_4$ curve fit. All
645 the $^{14}\text{CH}_4$ and ^{14}CO data points are fitted within the 1σ uncertainties (Figure 3). The
646 curve fitting approach for $^{14}\text{CO}_2$ was the same as for ^{14}CO .

647

648 ^{14}C production from ^{16}O in quartz by fast muons has been shown in laboratory irradiation
649 experiments (Heisinger et al., 2002a). However, to the best of our knowledge, the fast
650 muon mechanism has never been unambiguously confirmed in a natural setting (ice or
651 rock). Recent ^{14}C results from a 15 m deep rock core at Leymon High, Spain, suggested
652 no significant ^{14}C production by the fast muon mechanism (Lupker et al., 2015).
653 However, there was low confidence in this particular conclusion, owing mainly to
654 analytical uncertainties that were as large as the ^{14}C signal for the deepest samples. We
655 specifically tested for the presence of ^{14}C from the fast muon mechanism by attempting to
656 fit our most complete data set (^{14}CO) with only the neutron and negative muon capture
657 mechanisms (Figure 3a). As can be seen, such a curve fit fails to match the observations,
658 confirming that the fast muon mechanism is significant. Our results thus provide the first
659 unambiguous confirmation of ^{14}C production from ^{16}O by fast muons in a natural setting.

660

661 Surface production rates for all mechanisms, scaled to sea level and high latitude using
662 the scaling of Lifton et al. (2014) are shown in Table 7. Our results yield useful
663 constraints for $P_{\mu f}^0$ (25% relative uncertainty) and $P_{\mu-}^0$ (29%) for ^{14}CO and $^{14}\text{CH}_4$. For
664 $P_{\mu f}^0$, the uncertainty in the depth of long-term ice transport provides the single largest
665 contribution to the error. For $P_{\mu-}^0$, the curve-fitting procedure itself (which takes into
666 account data uncertainties) and the ablation rate uncertainty are the largest sources of
667 error. Constraints on P_n^0 are weaker due to a relatively small contribution from this
668 mechanism (88% relative uncertainty for ^{14}CO and 47% for $^{14}\text{CH}_4$).

669

670 We place low confidence in the production rate estimates for $^{14}\text{CO}_2$ because of the large
671 amount of scatter in those data as well as the unknown and likely variable extraneous ^{14}C
672 component introduced by the melt-extraction. If we make the speculative assumption that
673 the ^{14}C content of added extraneous inorganic carbon (which affects $^{14}\text{CO}_2$) is
674 approximately uniform across all our samples, then we could conclude that $^{14}\text{CO}_2$ is the
675 dominant species produced by all mechanisms. This conclusion would be consistent with
676 a prior study that was done by dry-extraction and included procedural blank
677 characterization at another Antarctic ablation area with similar ice ^{14}C content (van der
678 Kemp, et al., 2002).

679

680 Given the large uncertainties in $^{14}\text{CO}_2$ determinations, we cannot place strong constraints
681 on the total ^{14}C production rates. Further, experiments involving laboratory irradiation of
682 ice suggest that non-negligible amounts of simple compounds other than $^{14}\text{CO}_2$, ^{14}CO and
683 $^{14}\text{CH}_4$ may also be formed (e.g., Rossler et al., 1984). Our measured set of species thus
684 may not be capable of fully characterizing the in situ cosmogenic ^{14}C . If we assume that
685 the fraction of in situ ^{14}C that forms other compounds is relatively small, our data (Figure
686 3d, Table 6) do suggest that the total ^{14}C production rates in ice commonly used in the
687 literature (Table 1, Table 7) are overestimated.

688

689 As we now have data-based estimates of ^{14}C production rates, we are able to test our
690 earlier assumption (Section 4.2) that in-situ cosmogenic ^{14}C inherited from the ice
691 accumulation zone is insignificant. A detailed discussion is presented the Electronic

692 Annex. For ^{14}CO , we find that, at the upper limit, inherited ^{14}C represents 5% or less of
693 measured values. This value is smaller than several other sources of uncertainty involved
694 in the interpretation (depth of long-term ice transport, long-term ablation rate, curve
695 fitting); neglecting this small possible contribution is therefore acceptable.

696

697

698 ***4.5 Comparison with the Scharffenbergbotnen ice ablation site***

699

700 We compare our results with ^{14}CO and $^{14}\text{CO}_2$ data from a prior study at the
701 Scharffenbergbotnen ice ablation area in Antarctica ($74^\circ34.67'\text{S}$, $11^\circ02.97'\text{W}$, 1173 m
702 asl) (van der Kemp et al., 2002). We consider these to be the most reliable prior
703 determinations of ^{14}CO and $^{14}\text{CO}_2$ in ablating ice, as a dry-extraction method was used
704 (avoiding the problematic melt-acidify approach of many other prior studies) and the
705 study included procedural blank characterization. To perform the comparison, we
706 attempted to fit the Scharffenbergbotnen data with the same approach we applied to our
707 data. However, we constrained the ^{14}CO and $^{14}\text{CO}_2$ surface production rates for each
708 mechanism to be within the uncertainties of those determined from the Taylor Glacier ^{14}C
709 data. The value for P_n^0 was scaled by a factor of 1.80 with respect to Taylor Glacier, the
710 value for $P_{\mu^-}^0$ by a factor of 1.45 and the value for $P_{\mu^+}^0$ with a factor of 1.18 (as per Lifton
711 et al. (2014)) to account for the higher elevation of Scharffenbergbotnen. Ablation rate
712 was adjusted to 0.16 m a^{-1} (van der Kemp et al., 2002).

713

714 Scharffenbergbotnen ice has a relatively younger age (van der Kemp et al., 2002) and is
715 expected to have a non-negligible ^{14}C component inherited from the accumulation zone.
716 Further, the van der Kemp et al. (2002) study did not include a correction for ^{14}C
717 production during storage and transport of the ice. To account for these ^{14}C components,
718 we treat D in Equation 4 as an additional free parameter.

719

720 Figure 5 shows the results of the curve fits. Scharffenbergbotnen ^{14}C results are
721 reproduced well with ^{14}C production rates determined at Taylor Glacier, even without
722 allowing for any uncertainty in the Scharffenbergbotnen ablation rate. This agreement is

723 encouraging and suggests that all in situ cosmogenic ^{14}C migrates from the ice lattice into
724 the air bubbles relatively rapidly, allowing for the dry-extraction method used in the van
725 der Kemp et al. (2002) study to quantitatively recover the ^{14}CO and $^{14}\text{CO}_2$.

726

727

728 **5. Conclusions and Future Outlook**

729

730 In this study, in situ cosmogenic production of $^{14}\text{CH}_4$ in ice has been corroborated, and
731 $^{14}\text{CH}_4$ production rates have been constrained in the 2.26 – 19.53 m depth range of the
732 Taylor Glacier samples. ^{14}CO measurements have improved substantially over prior
733 work, resulting in much less scatter in the data. As a result, ^{14}CO production rates have
734 also been successfully constrained at Taylor Glacier and appear consistent with
735 independent data from Scharffenbergbotnen. This agreement between results from melt-
736 extraction and dry extraction approaches suggests that in situ cosmogenic ^{14}CO is
737 released rapidly from the ice grains into the air bubbles in the ice. This study has also
738 provided the first confirmation of in situ cosmogenic ^{14}C production from ^{16}O by the fast
739 muon mechanism in a natural setting.

740

741 Taylor Glacier results revealed that the $^{14}\text{CH}_4 / ^{14}\text{CO}$ production ratio is the same (0.0076
742 ± 0.0003) for the two muon mechanisms and is independent of depth. A higher $^{14}\text{CH}_4 /$
743 ^{14}CO production ratio inferred for the neutron mechanism is qualitatively consistent with
744 earlier results from Summit, Greenland firn. The constant $^{14}\text{CH}_4 / ^{14}\text{CO}$ production ratio
745 for the muon mechanisms means that ^{14}CO can be used to correct $^{14}\text{CH}_4$ measurements
746 for the in situ cosmogenic component in ice with negligible neutron-produced $^{14}\text{CH}_4$.
747 This means that it should be possible to retrieve the paleoatmospheric $^{14}\text{CH}_4$ signal from
748 Taylor Glacier ice deeper than ≈ 6 m, allowing for reliable reconstructions of the fossil
749 fraction of the past methane budget, including testing the hypotheses of major
750 atmospheric methane releases from methane clathrates and permafrost.

751

752 This study did not produce reliable constraints on $^{14}\text{CO}_2$ production rates because of
753 complications associated with the melt-extraction approach. $^{14}\text{CO}_2$ production rates and
754 total ^{14}C production rates in ice thus remain to be experimentally verified.

755

756 Ongoing work involves the construction of a system capable of extracting CO_2 from \approx
757 1.5 kg of ice by sublimation. This approach would allow for complete $^{14}\text{CO}_2$ release from
758 the ice grains without the complications of a melt extraction. Ongoing work is also
759 attempting to extend the sampling at the ≈ 52 ka site at Taylor Glacier to a depth of 70 m
760 or more, to better characterize the two muon production mechanisms, as well as collect
761 samples from very shallow ice to improve constraints on ^{14}C production by neutrons.

762

763

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TABLES

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Mechanism	P^0 (^{14}C atoms $\text{g}^{-1} \text{a}^{-1}$), sea level and high latitude	P^0 (^{14}C atoms $\text{g}^{-1} \text{a}^{-1}$), calculated for Taylor Glacier (elev. 527 m)
Neutrons	20.0 ± 1.5	39.6 ± 3.0
Muon capture	4.75 ± 0.4	6.96 ± 0.59
Fast muons	0.74 ± 0.4	0.88 ± 0.47

778

779 Table 1. Estimates of surface total ^{14}C production rates in ice based on prior work for the
780 neutron, negative muon capture and fast muon mechanisms. Cosmic ray flux scaling is as
781 in Lifton et al. (2014). Production rate for neutrons is as determined by Young et al.
782 (2014) in West Greenland for quartz, and for muons as determined by Heisinger et al.,
783 (2002b) via laboratory irradiation of quartz; the rates are transferred from quartz to ice by
784 accounting for the number of O atoms per gram ice versus per gram quartz. We use a
785 value of 150 g cm^{-2} for the absorption mean free path for neutrons in ice (Λ_n) (Lal et al.,
786 1987; van de Wal et al., 2007), although we note that Nesterenok and Naidenov (2012)
787 estimated a lower Λ_n of $\approx 130 \text{ g cm}^{-2}$.

788

Sample mid-depth, m	Measured [CH ₄], ppb	[CH ₄] after all corrections, ppb	Measured ¹⁴ CH ₄ , pMC	¹⁴ CH ₄ after all corrections, pMC	Ice ¹⁴ CH ₄ content, molecules / g
2.26	468.8 ± 0.3	461.6 ± 4.9	33.2 ± 0.3	29.9 ± 1.6	0.418 ± 0.026
3.77	473.5 ± 0.2	476.1 ± 2.4	24.2 ± 0.3	22.8 ± 0.8	0.327 ± 0.013
5.27	485.8 ± 0.2	492.0 ± 6.1	sample lost		
6.77	493.0 ± 0.2	502.7 ± 2.0	17.4 ± 0.2	17.4 ± 0.5	0.273 ± 0.010
10.02	515.9 ± 0.4	525.4 ± 2.0	sample lost		
15.01	518.7 ± 0.2	528.6 ± 2.0	13.9 ± 0.2	13.7 ± 0.5	0.206 ± 0.008
19.53	531.4 ± 0.3	541.6 ± 2.0	13.0 ± 0.2	12.8 ± 0.5	0.182 ± 0.008

789

790 Table 2. Sample [CH₄] and ¹⁴CH₄ as measured and after all procedural corrections, as

791 well as calculated ice ¹⁴CH₄ content. All [CH₄] values are reported on the Tohoku

792 University scale (Cunnold et al., 2002). Note that the effect of ¹⁴C corrections is largest

793 for the shallowest 2 samples which are affected by ambient air inclusion via fractures in

794 the ice.

795

796

Sample mid-depth, m	Measured ^{14}CO , pMC	Ice ^{14}CO content, molecules / g
2.26	126.9 ± 0.6	45.2 ± 1.6
3.77	121.7 ± 0.6	41.9 ± 1.1
5.27	96.2 ± 0.5	37.5 ± 1.0
6.77	91.7 ± 0.5	36.4 ± 0.9
10.02	81.3 ± 0.4	31.4 ± 0.8
15.01	79.5 ± 0.4	26.9 ± 0.7
19.53	75.3 ± 0.5	23.9 ± 0.6

797

798

799 Table 3. ^{14}CO activities measured in the samples and final calculated ice ^{14}CO content.

800

Sample mid-depth, m	Measured $^{14}\text{CO}_2$, pMC	Upper limit of $^{14}\text{CO}_2$ in sampled ice, molecules / g
2.26	41.6 ± 0.4	397 ± 24
3.77	32.7 ± 0.3	296 ± 17
5.27	30.0 ± 0.4	286 ± 16
6.77	31.2 ± 0.3	305 ± 17
10.02	30.3 ± 0.4	293 ± 16
15.01	21.7 ± 0.3	180 ± 10
19.53	25.8 ± 0.3	220 ± 12

801

802

803 Table 4. Measured $^{14}\text{CO}_2$ activities in the samples, as well as estimated maximum $^{14}\text{CO}_2$

804 content in the ice. $^{14}\text{CO}_2$ activities have not been normalized for $\delta^{13}\text{C}$, as $\delta^{13}\text{C}$ was not

805 measured. Since other uncertainties associated with the $^{14}\text{CO}_2$ determination are

806 relatively large, the lack of $\delta^{13}\text{C}$ values results in a negligible increase in the overall

807 uncertainty ($\approx 0.25\%$ relative uncertainty increase).

808

Sample mid-depth, m	$^{14}\text{CH}_4 / ^{14}\text{CO}$ ratio	Expected fraction of ^{14}C from neutrons	Expected fraction of ^{14}C from muon capture	Expected fraction of ^{14}C from fast muons
2.26	0.0092 ± 0.0005	0.10	0.61	0.29
3.77	0.0078 ± 0.0003	0.05	0.63	0.33
5.27		0.02	0.63	0.35
6.77	0.0075 ± 0.0003	0.01	0.62	0.37
10.02		0.00	0.59	0.40
15.01	0.0077 ± 0.0003	0.00	0.55	0.45
19.53	0.0076 ± 0.0003	0.00	0.51	0.49

809

810

811 Table 5. Measured $^{14}\text{CH}_4 / ^{14}\text{CO}$ ratio for finalized ice ^{14}C content, and the fraction of

812 expected total ^{14}C attributable to each production mechanism. Expected fractions were

813 calculated using Taylor Glacier surface production rates as in Table 1 and depth-

814 production rate formulations as described in Section 4.4 and the Electronic Annex.

815

816

Sample mid-depth, m	Upper limit of total ^{14}C , atoms / g ice	Expected total ^{14}C , atoms / g ice	Upper limit Measured / Expected
2.26	443 ± 24	828 ± 136	0.53 ± 0.09
3.77	338 ± 17	727 ± 132	0.47 ± 0.09
5.27	324 ± 16	662 ± 128	0.49 ± 0.10
6.77	342 ± 17	614 ± 125	0.56 ± 0.12
10.02	324 ± 16	536 ± 118	0.60 ± 0.14
15.01	207 ± 10	448 ± 110	0.46 ± 0.12
19.53	244 ± 12	390 ± 104	0.62 ± 0.17

817

818

819 Table 6. Comparison of upper limit of total ^{14}C in sampled ice based on measurements

820 with expected values based on production rates from prior studies as in Table 1.

821

	^{14}CO	$^{14}\text{CH}_4$	$^{14}\text{CO}_2$	<i>Total</i>	Total, prior studies
Neutrons	0.69 ± 0.61	0.024 ± 0.011	13.6 ± 10.1	14.3 ± 10.1	20.0 ± 1.5
Muon Capture	0.24 ± 0.07	0.0018 ± 0.0005	1.5 ± 0.7	1.7 ± 0.7	4.75 ± 0.4
Fast Muons	0.053 ± 0.014	0.00040 ± 0.00010	0.51 ± 0.17	0.56 ± 0.17	0.74 ± 0.4

822

823 Table 7. In situ cosmogenic ^{14}C production rates by the individual mechanisms at the ice

824 surface as determined from Taylor Glacier data by curve fitting (Section 4.4 and Figure

825 3), scaled to sea level and high latitude. All production rates are in atoms of $^{14}\text{C g}^{-1} \text{ a}^{-1}$.

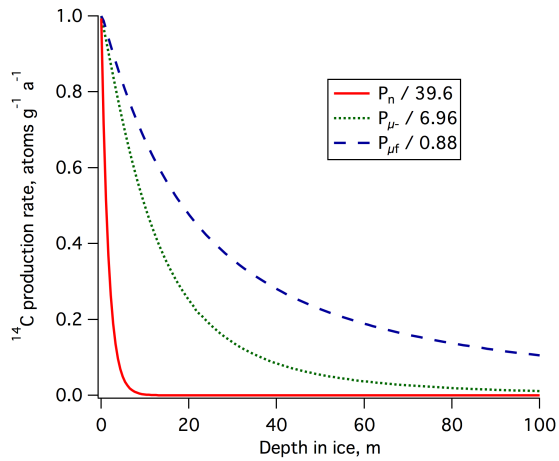
826 The values for $^{14}\text{CO}_2$ and total production rates are less certain, as discussed in the text.

827 Typical total ^{14}C production values given in prior studies (same as in Table 1) are also

828 shown for comparison.

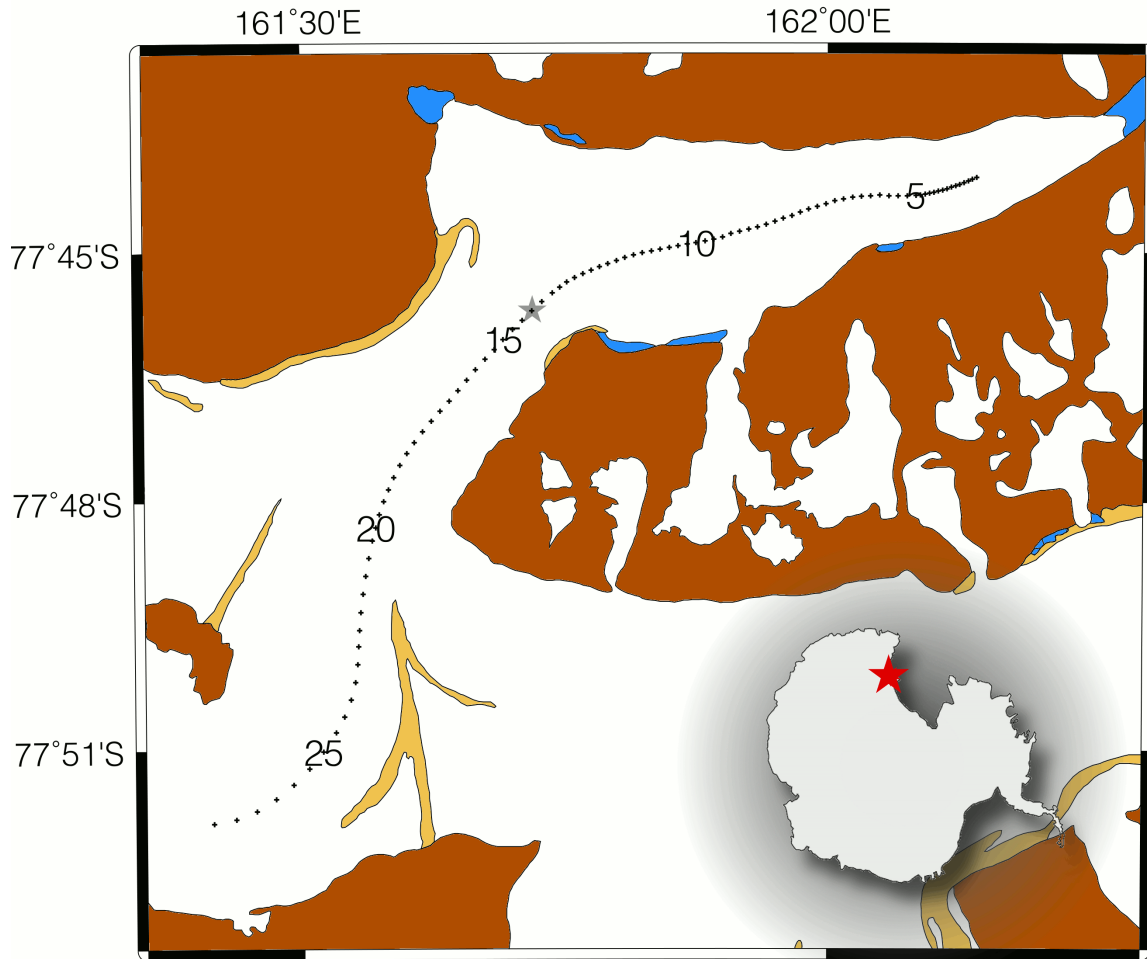
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FIGURES



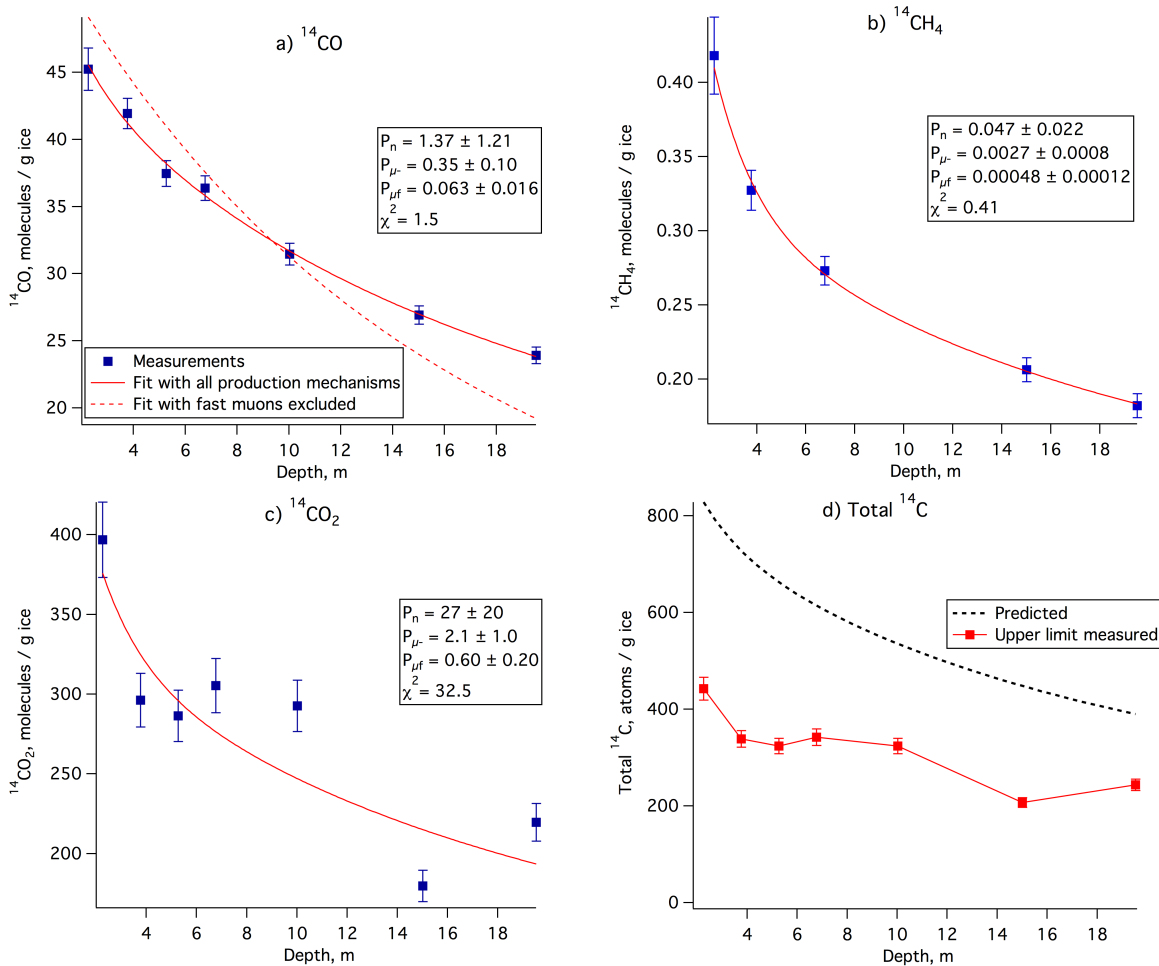
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Figure 1. Depth dependence of ^{14}C production rates in Taylor Glacier ice for the neutron (P_n), stopped negative muon (P_{μ^-}) and fast muon (P_{μ^+}) mechanisms. All production rates from Table 1 have been normalized to 1 at the surface for easier visual comparison of the depth dependence. Production by neutrons is negligible below about 10 m, while production by fast muons remains significant even at depths as large as 400 m.



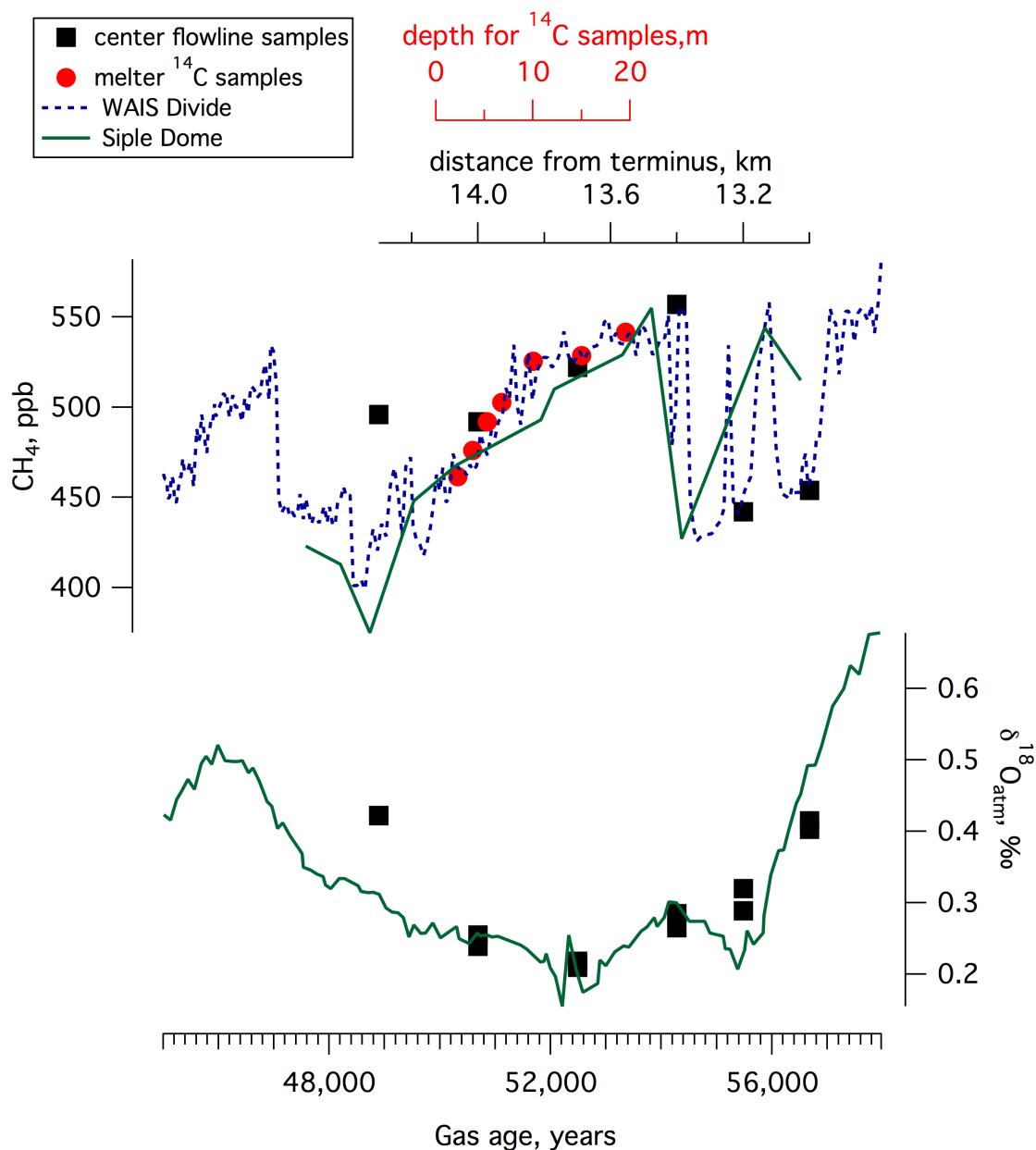
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Figure 2. Map of Taylor Glacier, showing locations of small ice samples taken along the center flowline for CH_4 and $\delta^{18}\text{O}_{\text{atm}}$ analyses during the 2009-10 season (black dots). Numbers denote distance in km from glacier terminus. Grey star shows the location of the ≈ 52 ka ice used in this study. Inset shows Taylor Glacier location in Antarctica.



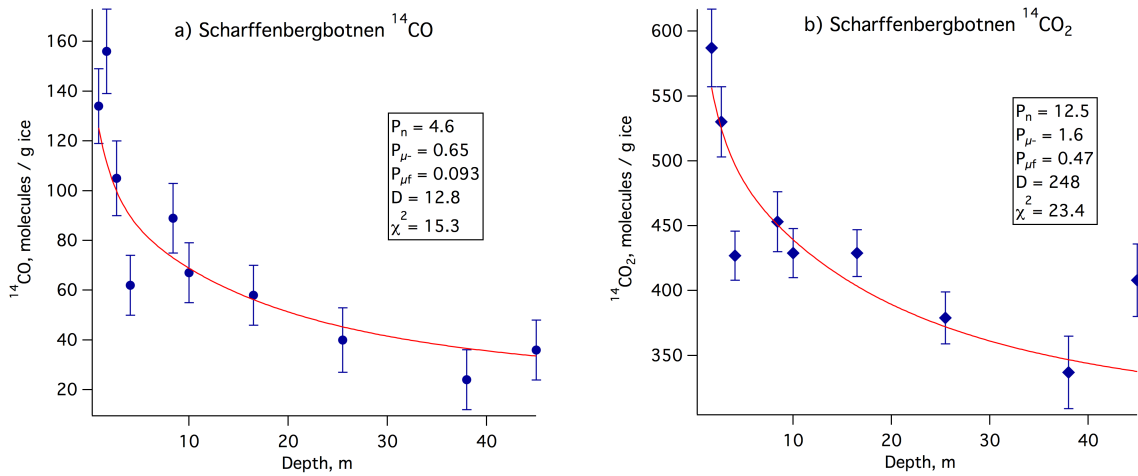
846

847 Figure 3. a) ^{14}CO content after all corrections (markers) and least-squares fit lines as
 848 described in the text. All production rate estimates are for Taylor Glacier site ice surface
 849 (^{14}C atoms $\text{g}^{-1} \text{a}^{-1}$). χ^2 value for the quality of “all mechanisms” fit is also shown. The
 850 best fit with fast muons excluded had $\chi^2 = 190$. Uncertainties for P_n , P_{μ^-} and $P_{\mu f}$
 851 incorporate contributions from the curve fit, ablation rate and depth of long-term
 852 transport. b) Similar to a), but for $^{14}\text{CH}_4$. Uncertainties for P_{μ^-} and $P_{\mu f}$ are propagated
 853 from corresponding uncertainties for ^{14}CO production rates and $^{14}\text{CH}_4 / ^{14}\text{CO}$ ratio. c)
 854 Similar to a) and b), but for upper limit of $^{14}\text{CO}_2$ content in the ice. d) Upper limit of total
 855 ^{14}C measured in the ice and predicted total ^{14}C . Predicted ^{14}C is determined by using
 856 cumulative ^{14}C formulations as described in Section 4.4 and the Electronic Annex, and
 857 surface production rates as in Table 1.



858
859

860 Figure 4. Data used to constrain the gas age of the large-volume samples used for ¹⁴C
861 analyses. Siple Dome [CH₄] is from Brook et al. (2005). WAIS Divide ice core [CH₄] is
862 from Buizert et al. (2015) and WAIS Divide Project Members (2015). Siple Dome ice
863 core δ¹⁸O_{atm} data are from Severinghaus et al. (2009). Note that neither the distance-age
864 relationship for the center flowline samples, nor the depth-age relationship for the ¹⁴C
865 samples is expected to be linear; this likely causes the mismatches in [CH₄] and δ¹⁸O_{atm}
866 outside the 50 – 55 ka age range of main interest.



867

868 Figure 5. Scharffenbergbotnen data from van der Kemp et al. (2002) and fit curves

869 generated using production rates constrained by Taylor Glacier data. Production rate

870 estimates shown are determined by the curve fit (within bounds allowed by Taylor

871 Glacier results) and are for Scharffenbergbotnen ice surface, in ^{14}C atoms $\text{g}^{-1} \text{a}^{-1}$.

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