Biogeochemical controls on ammonium accumulation in the surface layer of the Southern Ocean

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Abstract. The production and removal of ammonium (NH₄⁺) are essential upper-ocean nitrogen cycle pathways, yet in the Southern Ocean where NH₄⁺ has been observed to accumulate in surface waters, its mixed-layer cycling remains poorly understood. For surface seawater samples collected between Cape Town and the Marginal Ice Zone in winter 2017, we found that NH₄⁺ concentrations were 5-fold higher than is typical for summer and lower north than south of the Subantarctic Front (0.01–0.26 µM versus 0.19–0.70 µM). Our observations confirm that NH₄⁺ accumulates in the Southern Ocean’s winter mixed layer, particularly in polar waters. NH₄⁺ assimilation rates were highest near the Polar Front (12.9 ± 0.4 nM d⁻¹) and in the Subantarctic Zone (10.0 ± 1.5 nM d⁻¹), decreasing towards the Marginal Ice Zone (3.0 ± 0.8 nM d⁻¹) despite the high ambient NH₄⁺ concentrations in these southernmost waters, likely due to the low temperatures and limited light availability. By contrast, rates of NH₄⁺ oxidation were higher south than north of the Polar Front (16.0 ± 0.8 versus 11.1 ± 0.5 nM d⁻¹), perhaps due to the lower-light and higher-iron conditions characteristic of polar waters. NH₄⁺ concentrations were also measured along five transects of the Southern Ocean (Subtropical Zone to Marginal Ice Zone) spanning the 2018/19 annual cycle. These measurements reveal that mixed-layer NH₄⁺ accumulation south of the Subantarctic Front derives from sustained heterotrophic NH₄⁺ production in late summer through winter that, in net, outpaces NH₄⁺ removal by temperature-, light-, and iron-limited microorganisms. Our observations thus imply that the Southern Ocean becomes a biological source of CO₂ to the atmosphere in autumn and winter not only because nitrate drawdown is weak but also because the ambient conditions favour net heterotrophy and NH₄⁺ accumulation.

1 Introduction

The Southern Ocean impacts the Earth system through its role in global thermohaline circulation, which drives the exchange of heat and nutrients among ocean basins (Frölicher et al., 2015; Sarmiento et al., 2004). The Southern Ocean also plays an integral role in mediating climate, by transferring carbon to the deep ocean via its biological and solubility pumps (Sarmiento and Orr, 1991; Volk and Hoffert, 1985) and through the release of deep-ocean CO₂ to the atmosphere during deep-water ventilation (i.e. CO₂ leak; Broecker and Peng, 1992; Lauderdale et al., 2013; Sarmiento and Toggweiler, 1984). Upper Southern Ocean circulation is dominated by the eastward-flowing Antarctic Circumpolar Current (ACC) that consists of a series of broad circumpolar bands (“zones”) separated by oceanic fronts. These fronts can drive water mass formation (Ito et al., 2010) and nutrient upwelling that supports elevated productivity (Sokolov and Rintoul, 2007).

Concentrations of the essential macronutrients, nitrate (NO₃⁻) and phosphate (PO₄³⁻), are perennially high in Southern Ocean surface waters, in contrast to most of the global

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ocean. Assimilation of these nutrients, and thus primary productivity, is limited in the Southern Ocean by numerous overlapping factors, including temperature, light, micronutrient concentrations, and grazing pressure (e.g. Boyd et al., 2001; Martin et al., 1990; Reay et al., 2001; Smith and Lancelot, 2004). The strength of these limitations varies with sector (i.e. longitude), zone (i.e. latitude), and season, resulting in spatial and temporal variability in chlorophyll $a$, primary production, plankton community composition, and the nutrient uptake regime (Mdutyana et al., 2020; Mengesha et al., 1998; Shadwick et al., 2015; Thomalla et al., 2011). In addition to the seasonality of temperature and light, Southern Ocean ecosystems are influenced by seasonal changes in nutrient availability. In winter, deep mixing replenishes the nutrients required for phytoplankton growth but the low temperatures and light levels impede biological activity (Rintoul and Trull, 2001). Once the mixed layer shoals in spring and summer, phytoplankton consume the available nutrients until some form of limitation (usually iron; Nelson et al., 2001; Nicholson et al., 2019) sets in. This balance between wintertime nutrient recharge and summertime nutrient drawdown is central to the Southern Ocean’s role in setting atmospheric CO$_2$ (Sarmiento and Toggweiler, 1984).

The onset of iron limitation following the spring–early-summer bloom in the Southern Ocean drives phytoplankton to an increased reliance on recycled ammonium ($\text{NH}_4^+$; Timmermans et al., 1998), the assimilation of which has a far lower iron requirement than that of NO$_3^-$ (Price et al., 1994). The extent to which phytoplankton rely on NO$_3^-$ versus $\text{NH}_4^+$ as their primary N source has implications for Southern Ocean CO$_2$ removal since phytoplankton growth fuelled by subsurface NO$_3^-$ (“new production”) must be balanced on an annual basis by the export of sinking organic matter (“export production”; Dugdale and Goering, 1967), which drives CO$_2$ sequestration (i.e. the biological pump; Volk and Hoffert, 1985). By contrast, phytoplankton growth on $\text{NH}_4^+$ or other recycled N forms (“regenerated production”) yields no net removal of CO$_2$ to the deep ocean (Dugdale and Goering, 1967). Considerable research has focused on NO$_3^-$ cycling in the Southern Ocean mixed layer because of the importance of this nutrient for the biological pump (e.g. Francois et al., 1992; Johnson et al., 2017; Mdutyana et al., 2020; Primeau et al., 2013; Sarmiento and Toggweiler, 1984) and global ocean fertility (Fripiat et al., 2021; Sarmiento et al., 2004). By contrast, the cycling of regenerated N within the seasonally varying mixed layer – including the production of $\text{NH}_4^+$ and its removal by phytoplankton and nitrifiers – remains poorly understood.

$\text{NH}_4^+$ is produced in the euphotic zone as a by-product of heterotrophic metabolism (Herbert, 1999) and as a consequence of zooplankton grazing (Lehette et al., 2012; Steinberg and Saba, 2008), and it is removed by phytoplankton uptake (in euphotic waters) and nitrification (mainly in aphotic waters). Heterotrophic bacteria can also consume $\text{NH}_4^+$ (Kirchman, 1994) and have been hypothesised to do so at significant rates in the Southern Ocean mixed layer in winter (Cochlan, 2008; Mdutyana et al., 2020). The assimilation of $\text{NH}_4^+$ by phytoplankton requires relatively little energy (Dortch, 1990) such that $\text{NH}_4^+$ is usually consumed in the euphotic zone as rapidly as it is produced (Glibert, 1982; La Roche, 1983), resulting in very low surface $\text{NH}_4^+$ concentrations in the open ocean (<0.2 $\mu$M; Paulot et al., 2015). Additionally, $\text{NH}_4^+$ is often the preferred N source to small phytoplankton (Dortch, 1990), which typically dominate when iron and/or light are limiting (Deppeler and Davidson, 2017; Pearce et al., 2010; Tagliabue et al., 2014) since their higher cell surface-area-to-volume ratio renders them less vulnerable to diffusion limitation and/or light limitation (Finkel et al., 2004; Fujiki and Taguchi, 2002; Hudson and Morel, 1993; Mei et al., 2009).

In addition to the implications for size distribution, the dominant N source to phytoplankton is indicative of their potential for CO$_2$ removal, as per the new production paradigm (Dugdale and Goering, 1967). The N isotopic composition ($\delta^{15}$N, in ‰, vs. N$_2$ in air, equal to ($^{15}$N$/^{14}$N$_{\text{sample}}$/($^{15}$N$/^{14}$N$_{\text{air}}$ − 1) × 1000) of particulate organic N (PON, a proxy for phytoplankton biomass) can be used to infer the dominant N source to phytoplankton (Altabet, 1988; Fawcett et al., 2011, 2014; Lourey et al., 2003; Van Oostende et al., 2017) since the assimilation of subsurface NO$_3^-$ yields PON that is higher in $\delta^{15}$N than that fuelled by recycled $\text{NH}_4^+$ uptake (Treibergs et al., 2014). As such, measurements of bulk $\delta^{15}$N-PON can be used to infer the net N uptake regime.

Nitrification, the oxidation of $\text{NH}_4^+$ to nitrite (NO$_2^-$) and then NO$_3^-$ by chemoautotrophic bacteria and archaea, was historically considered unimportant in euphotic-zone waters due to the evidence for light inhibition of nitrifiers (Hooper and Terry, 1974; Horrigan and Springer, 1990; Olson, 1981) and the fact that they are outcompeted by phytoplankton for N$_2$ (Smith et al., 2014; Ward, 1985, 2005; Zakem et al., 2018). However, this view has been challenged in numerous ocean regions (Yool et al., 2007), including the Southern Ocean (Smart et al., 2015; Cavagna et al., 2015; Fripiat et al., 2015; Mdutyana et al., 2020). Wintertime upper-ocean $\text{NH}_4^+$ dynamics thus have implications for annual estimates of carbon export potential, insofar as NO$_3^-$ produced by nitrification in the winter mixed layer that is subsequently supplied to spring and summer phytoplankton communities constitutes a regenerated rather than a new N source on an annual basis (Mdutyana et al., 2020).

Surface concentrations of $\text{NH}_4^+$ are typically near zero in spring and early summer to mid-summer in the open Southern Ocean (Daly et al., 2001; Henley et al., 2020; Sambrotto and Mace, 2000; Savoye et al., 2004) due to assimilation by phytoplankton. In late summer, a peak in $\text{NH}_4^+$ concentration has been observed and attributed to enhanced bacterial and zooplankton activity following elevated phytoplankton growth (Becquevort et al., 2000; Dennett et al., 2001; Mengesha et al., 1998). The limited available observations suggest
that wintertime surface NH$_4^+$ concentrations are high (often > 1 µM), particularly south of the Subantarctic Front (SAF) (Bianchi et al., 1997; Henley et al., 2020; Philibert et al., 2015; Mdutyana et al., 2020; Weir et al., 2020). It thus appears that NH$_4^+$ is not depleted following the late-summer peak in its concentration, which indicates enhanced NH$_4^+$ regeneration, coincident with (but in excess of) NH$_2^+$ assimilation in winter and/or prior to this in late summer and/or autumn. Under these conditions, the Southern Ocean mixed layer may become net heterotrophic and thus a biological source of CO$_2$ to the atmosphere.

Here, we focus on NH$_4^+$ cycling in the Southern Ocean mixed layer, mainly in winter, which is a season assumed to be largely biologically dormant (Arrigo et al., 2008; Schaaffsma et al., 2018) and for which NH$_4^+$ cycle data are scarce. We confirm that NH$_4^+$ accumulates throughout the winter mixed layer south of the SAF, and examine the potential drivers thereof. Using NH$_4^+$ concentration data collected over a full annual cycle, we propose that these drivers include a contribution from the residual late-summer NH$_4^+$ pool, sustained NH$_4^+$ production in the autumn and winter, and limited wintertime NH$_4^+$ uptake and oxidation that nonetheless exceed the rate of in situ NH$_4^+$ production. Finally, from our temporally resolved NH$_4^+$ concentration data, we propose – for the first time – a measurement-based seasonal cycle for the mixed-layer NH$_4^+$ pool south of the SAF.

2 Methods

2.1 Cruise tracks and sample collection

Samples were collected for a series of analyses on the southward (S) and northward (N) legs of a winter cruise between Cape Town, South Africa, and the Marginal Ice Zone (MIZ) on board the R/V SA Agulhas II (VOY025; 28 June to 13 July 2017) (Fig. 1). Samples were also collected for NH$_4^+$ concentration analysis on three cruises on board the R/V SA Agulhas II during 2018–2019: early- and late-summer samples were collected during the SANAE 58 relief voyage (6 December 2018 to 15 March 2019; VOY035); winter samples were collected during the SCALE 2019 (http://scale.org.za/, last access: 17 January 2022) winter cruise to the MIZ (18 July to 12 August 2019; VOY039); and spring samples were collected during the SCALE 2019 spring cruise to the MIZ (12 October to 20 November 2019; VOY040) (Fig. S1).

Leg S of VOY025 in winter 2017 crossed the Atlantic sector and, due to logistical constraints, involved only surface underway collections, while leg N bordered the Atlantic and Indian sectors (30°E, WOCE IO6 line) and included eight conductivity–temperature–depth (CTD) hydrocast stations. Frontal positions were determined using the ship’s hull-mounted thermosalinograph, supported by temperature, salinity, and oxygen concentration data from CTD measurements made during leg N. The salinity and oxygen sensors were calibrated against seawater samples that were analysed for salinity using a Portasal 8410A salinometer and for dissolved oxygen by Winkler titration (Strickland and Parsons, 1972). Frontal positions were determined from sharp gradients in potential temperature, salinity, potential density, and oxygen concentrations (Belkin and Gordon, 1996; Lutjeharms and Valentine, 1984; Orsi et al., 1995). For leg N, the mixed-layer depth (MLD) was determined for each Niskin (up)cast as the depth between 10 and 400 m at which the Brunt–Väisälä frequency squared, $N^2$, reached a maximum (Carvalho et al., 2017).

During leg S, samples were collected every 4 h from the ship’s underway system (~7 m intake; “underway stations”), while samples on leg N were collected from surface Niskin bottles (~10 m, approximately 55 % light depth) mounted on the CTD rosette (“CTD stations”). NH$_4^+$ samples were also taken at 13 depths over the upper 500 m at the CTD stations. During the 2018–2019 cruises, NH$_4^+$ sam-

Figure 1. Winter 2017 cruise track overlaid on sea surface temperature (SST) measured by the hull-mounted thermosalinograph. The underway (leg S) and CTD (leg N) stations are indicated by white circles. Stations at which net primary production (NPP), nitrogen uptake, and ammonium oxidation experiments were conducted are denoted by red squares. The pink triangles indicate stations where only NPP experiments were conducted, and the green circles show stations where only ammonium oxidation was measured. Solid lines indicate the positions of the fronts, identified from measurements of temperature and salinity. Abbreviations for fronts: AF – Agulhas Front (~40.2°S); STF – Subtropical Front (~42.1°S); SAF – Subantarctic Front (~45.6°S); PF – Polar Front (~49.5°S); SACC – Southern Antarctic Circumpolar Current Front (~56.5°S); SB – Subantarctic Boundary (~58.5°S). Abbreviations for zones: STZ – Subtropical Zone; SAZ – Subantarctic Zone; PFZ – Polar Front Zone; OAZ – Open Antarctic Zone; PAZ – Polar Antarctic Zone; WG – Weddell Gyre; MIZ – Marginal Ice Zone. Together, the OAZ and PAZ constitute the Antarctic Zone (AZ). See Sect. S1 for detailed definitions of the fronts and zones. Figure produced using the package ggplot2 (Wickham, 2016).
ples were collected every 2 h from the ship’s underway system. At all stations, 40 mL of unfiltered seawater was collected for the analysis of NH$_4^+$ concentrations in duplicate 50 mL high-density polyethylene (HDPE) bottles that had been stored (“aged”) with an orthophthalaldehyde (OPA) working reagent. Unfiltered seawater was collected in duplicate 50 mL polypropylene centrifuge tubes for the analysis of NO$_3^-$, NO$_2^-$, and PO$_4^{3-}$ and in a single tube for urea. Immediately following collection, NH$_4^+$ and nutrient samples were frozen at $-20^\circ$C.

Duplicate size-fractionated chlorophyll a samples were collected by filtering seawater (500 mL) through 25 mm diameter glass fibre filters (0.3 and 2.7 µm; Sterlitech GF-75 and Grade D, respectively). Acetone (90 %) was added to foil-wrapped borosilicate tubes containing the filters and incubated at $-20^\circ$C for 24 h. Duplicate seawater samples (4 L) were also gently vacuum filtered through combusted 47 mm diameter, 0.3 µm GF-75 and 2.7 µm Grade-D filters for POC and PON concentrations and $\delta^{15}$N-PON. Filters were stored in combusted foil envelopes at $-80^\circ$C.

For microscopy, unfiltered seawater samples (250 mL) were collected during leg S in amber glass bottles and immediately fixed by the addition of 2.5 mL of Lugol’s iodine solution (2 % final concentration) and then stored at low room temperature in the dark until analysis. For flow cytometry, seawater samples were collected in triplicate 2 mL microcentrifuge tubes, fixed with glutaraldehyde (1 % final concentration), and stored at $-80^\circ$C until analysis (Marie et al., 2005).

Ten incubation experiments were conducted during leg S to measure net primary production (NPP). In addition, four NPP experiments were conducted during leg N using seawater collected from Niskin bottles fired at $\sim 10$ m. In all cases, pre-screened (200 µm mesh, to remove large grazers) seawater was collected in three 2 L polycarbonate bottles to which NaH$_{13}$CO$_3$ was added at $\sim 5$ % of the estimated ambient dissolved inorganic carbon (DIC) concentration. Enrichment of $^{13}$C was re-calculated post-cruise using measured DIC concentrations, and these enrichments were used in all NPP rate calculations. Bottles were incubated for 5 to 6.5 h in custom-built deck-board incubators shaded with neutral-density screens to mimic the 55 % light level and supplied with running surface seawater. Following incubation, each sample was divided (1 L per size fraction) and gently vacuum filtered through combusted 0.3 and 2.7 µm glass fibre filters that were stored in combusted foil at $-80^\circ$C until analysis.

N uptake (as NO$_3^-$, NH$_4^+$, and urea) and NH$_4^+$ oxidation experiments were conducted at five stations during leg S, with NH$_4^+$ oxidation measured at two additional stations at the ice edge (Fig. 1). On leg N, experiments were also conducted using seawater collected from $\sim 10$ m at the same four CTD stations as in the NPP experiments. Duplicate 1 L polycarbonate bottles were amended with $^{15}$N-labelled NO$_3^-$, NH$_4^+$, or urea at $\sim 10$ % of the ambient N concentration, estimated based on past wintertime measurements (Mdutyana et al., 2020) and, in the case of NH$_4^+$, coincident shipboard analyses. Enrichment of $^{15}$N was re-calculated post-cruise using the measured nutrient concentrations, and these enrichments were used in all rate calculations. Incubations and filtration were carried out as for NPP, although 500 mL was used per size fraction. For NH$_4^+$ oxidation, duplicate 250 mL black HDPE bottles were amended with 0.1 µM $^{15}$NH$_4^+$ and 0.1 µM $^{14}$NO$_3^-$ (the latter as a “trap” for the $^{15}$NO$_3^-$ produced by NH$_4^+$ oxidation; Ward, 2011). NH$_4^+$ oxidation bottles were incubated for 24 h under the same temperature conditions as the N uptake and NPP experiments. Subsamples (50 mL) were collected from each bottle immediately following tracer addition ($T_0$) and at the end of the experiments ($T_f$) and frozen at $-20^\circ$C until analysis.

2.2 Sample processing

2.2.1 Ammonium concentrations

On all cruises, NH$_4^+$ concentrations were measured shipboard using the fluorometric method of Holmes et al. (1999) and a Turner Designs Trilogy 7500-000 fluorometer equipped with a UV module. The detection limit, calculated as twice the pooled standard deviation (SD) of all standards, was 0.06 µM. To prevent possible influx and/or efflux of ammonia (NH$_3$) due to the temperature difference between surface waters and the shipboard laboratory, samples were frozen immediately upon collection, for a maximum of 24 h. OPA working reagent was added to the frozen samples prior to defrosting them for analysis. Samples were slowly warmed to room temperature in a water bath after OPA addition and incubated in the dark for 4 h once defrosted, and then each replicate was measured in triplicate. Standards and blanks were made daily using Type-1 Milli-Q water. Precision was $\pm 0.03$ µM for replicate samples and standards.

During VOY040 (spring 2019), we investigated the possibility that the ship’s underway system alters the seawater NH$_4^+$ concentrations (e.g. due to contamination or cell breakage). We collected surface samples from the underway and Niskin bottles concurrently and measured an average NH$_4^+$ concentration difference of $0.07 \pm 0.15$ µM ($n = 17$), with no noticeable trend of one method consistently yielding higher/lower concentrations. We thus have no reason to doubt NH$_4^+$ concentrations measured for seawater samples collected from the ship’s underway system.

2.2.2 Macronutrient concentrations

Following the winter 2017 cruise, duplicate seawater samples were analysed manually for NO$_3^-$ and PO$_4^{3-}$ (Bend-schneider and Robinson, 1952; Murphy and Riley, 1962) using a Thermo Scientific Genesyss 30 Visible spectrophotometer. The precision and detection limit were $\pm 0.05$ µM and 0.05 µM for NO$_3^-$ and $\pm 0.06$ µM and 0.05 µM for PO$_4^{3-}$.

The concentrations of NO$_3^-$ + NO$_2^-$ and Si(OH)$_4$ were measured using a Lachat QuikChem 8500 Series 2 flow in-
were calculated according to Dugdale and Wilkerson (1986).

\[ \rho = \frac{[PM] \times (\text{at. %}_{\text{meas}} - \text{at. %}_{\text{amb}})}{\frac{T}{\text{at. %}_{\text{init}} - \text{at. %}_{\text{amb}}}}, \quad (1) \]

\[ \text{at. %}_{\text{init}} = \frac{[(\text{at. %}_{\text{amb}}) + (\text{at. %}_{\text{tracer}} \times T)]}{\text{at. %}_{\text{tracer}}}. \quad (2) \]

Here, \( M \) is the species of interest (C, \( \text{NH}_4^+ \), \( \text{NO}_3^- \), or urea); \( \rho M \) is the uptake rate of that species (nM h\(^{-1}\), i.e. nmol C or N L\(^{-1}\) h\(^{-1}\)); \([PM]\) is the concentration of POC or PON (\( \mu \)M) on the filters; \([M]\) is the ambient concentration of DIC, \( \text{NH}_4^+ \), \( \text{NO}_3^- \), or urea at the time of sample collection; \([M_{\text{tracer}}]\) is the concentration of \( ^{15}\text{NaHCO}_3 \), \( ^{15}\text{NH}_4^+ \), \( ^{15}\text{NO}_3^- \), or \( ^{15}\text{N}\)-urea added to the incubation bottles; and \( T \) is the incubation period (hours). DIC concentrations were measured shipboard using a VINDTA 3C instrument and ranged from 2017 to 2130 \( \mu \)M (Bakker et al., 2016). The PM and \( \rho M \) of the picoplankton size class were calculated by subtracting the nanoplancton from the bulk measurements. Daily rates were computed by multiplying the hourly rates by the number of daylight hours, the latter calculated using the sampling latitude and day of the year (Forsythe et al., 1995).

The \( f \) ratio (Eppley and Peterson, 1979), used to estimate the fraction of NPP potentially available for export, was calculated as

\[ f \text{ ratio} = \frac{\rho \text{NO}_3^-}{\rho \text{N}_{\text{tot}}}, \quad (3) \]

where \( \rho \text{N}_{\text{tot}} = \rho \text{NH}_4^+ + \rho \text{NO}_3^- + \rho \text{Urea} \). Urea uptake was not measured at underway stations 50.7 and 55.5\(^\circ\) S (both in the Antarctic Zone); here, the \( f \) ratio was calculated omitting \( \rho \text{Urea} \). For the two Antarctic Zone stations at which urea uptake was measured, including \( \rho \text{Urea} \) decreased the \( f \) ratio by 8\%–25\% compared to that calculated using only \( \rho \text{NO}_3^- \) and \( \rho \text{NH}_4^+ \).

### 2.2.6 Ammonia oxidation rates

The azide method (McIlvin and Altabet, 2005) was used to convert \( \text{NO}_2^- \) produced by \( \text{NH}_4^+ \) oxidation to \( \text{N}_2\text{O} \) gas that was measured using a Delta V Plus IRMS with a custom-built purge-and-trap front end (McIlvin and Casciotti, 2011). This configuration yields a detection limit of 0.2 nmol N with a \( \delta^{15}\text{N} \) precision of \( \pm 0.1\%e \). The \( \delta^{15}\text{N} \) of \( \text{NO}_2^- \) was derived from \( ^{45}\text{N}_2\text{O}/^{44}\text{N}_2\text{O} \), and the rate of \( \text{NH}_4^+ \) oxidation (\( \text{NH}_4^+ \) ox; nM d\(^{-1}\)) was calculated following Peng et al. (2015) as

\[ \text{NH}_4^+ \text{ ox} = \frac{\Delta [15\text{NO}_2^-]}{[15\text{NH}_4^+] \times T}, \quad (4) \]

Here, \( \Delta [15\text{NO}_2^-] \) is the change in the concentration of \( ^{15}\text{NO}_2^- \) (nM) between the start and end of the incubation, calculated as the difference in the measured \( \delta^{15}\text{N} \) of \( \text{NO}_2^- \) between the
We note that isotope dilution (i.e. the dilution of $^{15}$NH$_4^+$ by co-occurring $^{14}$NH$_4^+$ regeneration) during the NH$_4^+$ uptake and oxidation experiments could potentially lead to an underestimation of the rates (Glibert, 1982; Mdutyana, 2021). For the NH$_4^+$ uptake experiments, their short duration (3 to 7.5 h) would have rendered the effect of regeneration minor (Mdutyana et al., 2020). Moreover, the $^{15}$NH$_4^+$ additions were high (100 nM) relative to both the ambient NH$_4^+$ concentrations north of the SAF and the $K_m$ values derived for NH$_4^+$ uptake and oxidation in the winter Southern Ocean (150–405 and 28–137 nM, respectively; Mdutyana, 2021), making a significant dilution effect unlikely (Lipschultz, 2008). Finally, at the stations south of the SAF, the ambient NH$_4^+$ concentrations were so high that even if the regeneration of $^{14}$NH$_4^+$ occurred at an elevated rate (e.g. 50 nM d$^{-1}$), as has been measured in the late-summer Southern Ocean when remineralisation is expected to be high; Goeyens et al., 1991), the $^{15}$N / $^{14}$N of the NH$_4^+$ pool would decrease by $< 1$ %–2 %. We thus consider the potential effect of isotope dilution to be minor.

A further consideration is possible stimulation of the NH$_4^+$ uptake and oxidation rates by $^{15}$NH$_4^+$ addition (Lipschultz, 2008). Given the $K_m$ values listed above and the high ambient NH$_4^+$ concentrations measured in the PFZ and AZ, a stimulation effect could be significant only at the stations north of the SAF where the NH$_4^+$ concentrations were 10–100 nM and, even then, to a lesser extent for NH$_4^+$ oxidation than NH$_4^+$ uptake given that ammonia oxidisers in the winter Southern Ocean become saturated at NH$_4^+$ concentrations of 100–200 nM (Mdutyana, 2021). The rates reported for the stations north of the SAF should therefore be considered “potential rates”. However, since our focus is mainly on explaining the accumulation of NH$_4^+$ south of the SAF, having potential rather than “true” rates for the STZ and SAZ does not affect our conclusions.

### 2.2.7 Plankton community composition

Microplankton groups ($> 15$ µm) were identified and counted in a subsample (20 mL) from each amber bottle using the Utermöhl technique (Utermöhl, 1958) and following the recommendations of Hasle (1978). Plankton groups and individual species were counted and identified using an inverted light microscope (Olympus CKX41) at 200× magnification. This level of magnification limited the cell sizes that could be reliably distinguished to $> 15$ µm. For each sample, at least 100 cells were enumerated to ensure a statistically valid count.

Pico- and nanoplankton cells ($< 15$ µm) were enumerated using an LSR II flow cytometer (BD Biosciences) equipped with blue, red, violet, and green lasers. Prior to analysis, 1 mL of sample was incubated with 1% (v/v) SYBR Green I (a DNA stain) at room temperature in the dark for 10 min (Marie et al., 1997). From light scatter and autofluorescence, the DNA-containing particles were identified as nano- and picoeukaryotes, and Synechococcus. Additionally, small heterotrophic prokaryotes (i.e. bacteria and possibly archaea, hereafter “bacteria”) were identified as DNA-containing particles with the lowest detectable autofluorescence (Marie et al., 1997; Gasol and del Giorgio, 2000) (see also Sect. S2). All particles lacking DNA were considered detritus. The populations of interest were gated using FlowJo 10.3 software (Tree Star, Inc.; https://www.flowjo.com/, last access: 17 January 2022).

In this study, we did not directly measure NH$_4^+$ regeneration (i.e. heterotrophy). Instead, we use the abundance of heterotrophic bacteria as a qualitative indicator of NH$_4^+$ regeneration potential, recognising that cell abundance does not imply activity. Additionally, we estimate the rate of NH$_4^+$ production from our concentration and rate data (see Sect. 3.3). The availability of organic matter to heterotrophs is inferred from the abundance of detritus.

### 2.3 Mixed-layer NH$_4^+$ residence time and NH$_4^+$ production rate estimates

The residence time of the mixed-layer NH$_4^+$ pool can be estimated using the measured ambient NH$_4^+$ concentrations and corresponding NH$_4^+$ removal rates as

$$\text{NH}_4^+ \text{ residence time} = \frac{[\text{NH}_4^+]}{\text{NH}_4^+ \text{ removal rate}}. \quad (5)$$

Here, $\text{NH}_4^+ \text{ residence time}$ is the time period (days) over which a given NH$_4^+$ concentration will be depleted assuming a constant NH$_4^+$ removal rate. We set NH$_4^+$ removal rate equal to $\rho_{\text{NH}_4^+} \text{ + } \rho_{\text{NH}_4^+ \text{ ox}}$ in winter and equal to $\rho_{\text{NH}_4^+}$ in late summer given the evidence for negligible mixed-layer NH$_4^+$ oxidation rates in this latter season (Bianchi et al., 1997; Mdutyana et al., 2020).

To determine the contribution of late-summer NH$_4^+$ production to the wintertime NH$_4^+$ pool (see Sect. 5.2), we define a rate of NH$_4^+$ concentration decline:

$$\text{NH}_4^+ \text{ rate of decline} = \text{NH}_4^+ \text{ production rate} - \text{NH}_4^+ \text{ removal rate}. \quad (6)$$

Here, $\text{NH}_4^+ \text{ production rate}$ is the NH$_4^+$ flux required to compensate for NH$_4^+$ removal over the late-summer-to-winter period, in order to yield the observed seasonal change in the ambient NH$_4^+$ concentration.

The rate of NH$_4^+$ concentration decline can also be defined as

$$\frac{\text{NH}_4^+ \text{ rate of decline}}{t} = \frac{[\text{NH}_4^+]_{\text{decline}}}{t}. \quad (7)$$

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where \([\text{NH}_4^+]_{\text{decline}}\) is the difference between the late-summer and winter \(\text{NH}_4^+\) concentrations and \(t\) is the time period (days) over which the \(\text{NH}_4^+\) concentration declines. Setting Eqs. (6) and (7) to be equal yields

\[
\text{NH}_4^+ \text{production rate} = \frac{[\text{NH}_4^+]_{\text{decline}}}{t} + \text{NH}_4^+ \text{removal rate},
\]

where \(\text{NH}_4^+ \text{removal rate} = \rho \text{NH}_4^+ + \text{NH}_4^+ \text{ox}\). Equations (7) and (8) assume that the elevated wintertime \(\text{NH}_4^+\) concentrations result from continuous \(\text{NH}_4^+\) production in excess of removal rather than from sporadic events of removal and/or production occurring between late summer and winter.

### 2.4 Statistical analyses

The correlations among latitude, N concentrations, NPP, N assimilation rates, and \(\text{NH}_4^+\) oxidation rates were investigated at the 5 % significance level using the Pearson correlation coefficient and the R packages stats (R Core Team, 2020) and corplot (Wei and Simko, 2017). Figures were made using the ggplot2 (Wickham, 2016), abind (Plate and Heiberger, 2019), ggpubr (Kassambara, 2019), lubridate (Grolemund and Wickham, 2011), MBA (Finley et al., 2017), metR (Campitelli, 2019), mgcv (Wood, 2017), oce (Kelley and Richards, 2022), and scales (Wickham and Seidel, 2020) R packages. Standard deviations were propagated using standard statistical practices.

### 3 Results

#### 3.1 Hydrography

Sea surface temperature (SST) decreased by \(\sim 17^\circ\text{C}\) between Cape Town (\(\sim 34^\circ\text{S}\)) and the edge of the MIZ (61.7° S), with similar gradients measured for legs S and N. During leg N, fairly deep MLDs were observed (124–212 m), which were similar to June and July climatological MLDs compiled from Argo float data for this region (Dong et al., 2018), which were similar to June and July climatological MLDs compiled from Argo float data for this region (Dong et al., 2018). While the focus of this study is the surface (i.e. upper \(\sim 10\) m), we report the MLDs here to show that sampling took place under typical winter conditions, with the deep MLDs evincing ongoing winter mixing and associated nutrient recharging. Where not specified, the trends discussed below refer to the surface data only. Latitudinal variations in each parameter are assessed by comparing the various Southern Ocean zones – the Subtropical Zone (STZ) north of the Subtropical Front (STF); the Subantarctic Zone (SAZ) between the STF and the Subantarctic Front (SAF), the Polar Frontal Zone (PFZ) between the SAF and the Polar Front (PF); and south of the PF, the Open Antarctic Zone and Polar Antarctic Zone (OAZ and PAZ), which are divided by the Southern Antarctic Circumpolar Current Front (SACCF) and collectively termed the Antarctic Zone (AZ); see Sect. S1 for detailed definitions of the fronts and zones and Figs. 1 and S1 for their positions at the time of sampling). For each parameter, the average \(\pm 1\) SD for each Southern Ocean zone is reported in Table 1.

#### 3.2 Macronutrient concentrations

In winter 2017, the surface and mixed-layer concentrations of \(\text{NH}_4^+\) ranged from below detection to 0.70 \(\mu\text{M}\) (Fig. 2a and b). Surface concentrations were higher in the PFZ, OAZ, and PAZ (0.42 \(\pm 0.01\), 0.52 \(\pm 0.01\), and 0.58 \(\pm 0.01\) \(\mu\text{M}\), respectively) than in the STZ and SAZ (0.08 \(\pm 0.03\) and 0.06 \(\pm 0.01\) \(\mu\text{M}\), respectively), with a sharp gradient observed at the SAF. South of the SAF, high \(\text{NH}_4^+\) concentrations persisted near-homogeneously throughout the mixed layer, with mixed-layer averages ranging from 0.65 \(\pm 0.01\) \(\mu\text{M}\) at station 58.5° S to 0.27 \(\pm 0.01\) \(\mu\text{M}\) at station 48.0° S and averaging 0.47 \(\pm 0.02\) \(\mu\text{M}\), with concentrations that were below detection north of the SAF (Fig. 2b). Below the mixed layer, \(\text{NH}_4^+\) concentrations decreased rapidly at all stations to values below detection by 200 m.

The concentrations of NO\(_3\) and PO\(_4^{3-}\) increased southwards from < 10 and < 1 \(\mu\text{M}\) in the STZ to > 20 and > 1.5 \(\mu\text{M}\) in the PFZ, OAZ, and PAZ (Fig. 2c and S3a), with the sharpest gradients occurring near the SAF. The concentrations of Si(\text{OH})\(_4\) increased rapidly across the PF, from an average of 3.2 \(\pm 1.1\) \(\mu\text{M}\) between 35.0 and 48.0° S to 45.6 \(\pm 0.6\) \(\mu\text{M}\) between 52.1 and 58.9° S (Fig. S3b). The NO\(_3\) concentrations were consistently low across the transect (0.16 \(\pm 0.02\) \(\mu\text{M}\); Fig. S3c), as were the concentrations of urea (0.20 \(\pm 0.04\) \(\mu\text{M}\); Table 1), with slightly lower urea concentrations observed in the SAZ than in the other zones.

#### 3.3 Chlorophyll a, POC, and PON

The highest bulk [chl a] was observed near the South African continental shelf, decreasing across the STF and remaining low thereafter (Fig. 3a). The proportion of chl a in the nano+ size class varied across the region but was > 50% at all stations, with higher (> > 80%) contributions near the fronts and at many OAZ and PAZ stations (Fig. 3b). The nano+ contribution was < 60% at only five stations (three in the SAZ, two in the OAZ).

The concentrations of bulk POC and PON were highest north of the STF and slightly higher in the OAZ than in the SAZ and PFZ (Fig. S4a and b). The contributions of the nano+ size fraction to POC and PON across the transect were 77.1 \(\pm 22.6\) % and 66.9 \(\pm 24.2\) %, respectively (Fig. S4c and d). The \(\delta^{15}\text{N}\)-PON decreased southwards from the STZ and SAZ (1.7 \(\pm 1.0\)%c) to the PFZ and OAZ (0.5 \(\pm 0.5\)%c; Fig. 4). Despite considerable differences among zones, the \(\delta^{15}\text{N}\)-PON was relatively homogeneous within each zone.
Table 1. Mean (±1 SD) of surface ocean POC, PON, chl a, and nutrient concentrations, cell abundances, and nutrient uptake rates measured in each zone of the Southern Ocean in winter 2017. Where no SD is given, only one sample was measured. The > 0.3 µM and > 2.7 µM size fractions are referred to as bulk and nano+, respectively. “% of nano+” refers to the average relative contribution of the nano+ size fraction to total chl a, POC, or PON, calculated for each station within a zone. The f ratio including ρUrea is shown only for zones where ρUrea was measured at all stations. “ND” indicates no data available. Abbreviations are as in Fig. 1.

|          | STZ | SAZ  | PFZ  | OAZ | PAZ |
|----------|-----|------|------|-----|-----|
| NH$_4^+$ (µM) | 0.08 ± 0.03 | 0.06 ± 0.01 | 0.42 ± 0.01 | 0.52 ± 0.01 | 0.58 ± 0.01 |
| PO$_4^{3-}$ (µM) | 0.44 ± 0.07 | 0.90 ± 0.06 | 1.59 ± 0.1 | 2.00 ± 0.13 | 1.99 ± 0.09 |
| NO$_3^-$ (µM) | 3.6 ± 0.2 | 10.5 ± 0.5 | 21.5 ± 0.2 | 26.7 ± 0.4 | 27.5 ± 0.4 |
| Si(OH)$_4$ (µM) | 2.6 ± 0.1 | 2.5 ± 1.8 | 6.6 ± 1.1 | 40.3 ± 0.5 | 45.0 ± 0.8 |
| NO$_2^-$ (µM) | 0.15 ± 0.02 | 0.13 ± 0.02 | 0.17 ± 0.02 | 0.19 ± 0.01 | 0.21 ± 0.02 |
| Urea (µM) | 0.23 ± 0.04 | 0.11 ± 0.04 | 0.26 ± 0.08 | 0.24 | 0.21 ± 0.03 |
| chl a (µg L$^{-1}$) | 77.5 ± 13.9 | 73.1 ± 19.0 | 69.8 ± 8.7 | 76.7 ± 11.3 | 80.1 ± 8.5 |
| POC (µM) | 4.4 ± 6.7 | 3.4 ± 0.4 | 3.2 ± 0.3 | 3.4 ± 0.5 | 3.5 ± 0.2 |
| POC (µM) | 0.50 ± 0.05 | 0.30 ± 0.04 | 0.24 ± 0.02 | 0.18 ± 0.02 | 0.17 ± 0.02 |
| POC (µM) | 0.15 ± 0.1 | 0.13 ± 0.07 | 0.11 ± 0.04 | 0.06 ± 0.03 | 0.04 ± 0.02 |
| chl a (µg L$^{-1}$) | 0.3 ± 0.1 | 0.3 ± 0.1 | 0.2 ± 0.3 | 0.2 ± 0.1 | 0.4 ± 0.0 |
| POC (µg L$^{-1}$) | 79.7 ± 24.6 | 79.6 ± 19.0 | 50.9 ± 33.2 | 77.2 ± 21.8 | ND |
| POC (µg L$^{-1}$) | 69.0 ± 31.9 | 67.1 ± 17.2 | 53.8 ± 24.1 | 67.0 ± 21.9 | 51.1 ± 24.7 |
| POC (µg L$^{-1}$) | 103.0 ± 22.1 | 102.5 ± 14.4 | 122.5 ± 11 | 234.1 ± 29.2 | 219.3 ± 1.0 |
| POC (µg L$^{-1}$) | 7.81 ± 0.69 | 6.90 ± 1.25 | 7.13 ± 0.71 | 6.72 ± 1.62 | 5.80 ± 3.75 |
| POC (µg L$^{-1}$) | 1.4 ± 0.9 | 1.2 ± 1.0 | 0.3 ± 0.5 | −1.3 ± 0.5 | −1.3 ± 0.4 |
| NPP (bulk) (nM d$^{-1}$) | 497.1 ± 42.4 | 277.5 ± 21.3 | 289.7 ± 19.2 | 85.3 ± 26.1 | 27.7 ± 0.2 |
| NPP (nano+) (nM d$^{-1}$) | 384.7 ± 29.7 | 178.2 ± 23.4 | 193.5 | 49.6 ± 5.0 | ND |
| ρNH$_4^+$ (nM d$^{-1}$) | 5.7 ± 0.8 | 8.9 ± 1.1 | 12.9 ± 0.4 | 4.8 ± 0.1 | 3.0 ± 0.8 |
| ρNH$_4^+$ (nM d$^{-1}$) | 4.0 ± 1.1 | 4.1 ± 1.2 | 4.2 ± 0.7 | 3.1 ± 0.4 | ND |
| ρNO$_3^-$ (nM d$^{-1}$) | 4.1 ± 0.4 | 11.5 ± 1.4 | 5.9 ± 1.4 | 3.6 ± 0.4 | 3.7 ± 1.8 |
| ρNO$_3^-$ (nM d$^{-1}$) | 3.4 ± 0.3 | 6.6 ± 0.4 | 4.3 ± 0.4 | 2.6 ± 0.8 | 2.7 ± 1.2 |
| ρUrea (nM d$^{-1}$) | 7.5 ± 0.6 | 6.9 ± 0.3 | 6.5 ± 1.0 | 2.1 ± 0.3 | 0.6 ± 0.01 |
| ρUrea (nM d$^{-1}$) | 4.9 ± 0.3 | 3.8 ± 0.2 | 4.0 ± 0.6 | 1.3 ± 0.2 | 0.7 ± 0.4 |
| ρUrea (nM d$^{-1}$) | 0.21 ± 0.31 | 0.43 ± 0.11 | 0.23 ± 0.18 | ND | 0.51 ± 0.53 |
| ρUrea (nM d$^{-1}$) | 0.43 ± 0.32 | 0.57 ± 0.12 | 0.31 ± 0.18 | 0.43 ± 0.16 | 0.55 ± 0.54 |
| NH$_4^+$ (nM d$^{-1}$) | 9.3 ± 0.5 | 12.9 ± 0.6 | 11.1 | 17.7 ± 0.6 | 14.3 ± 1.0 |
| Total microplankton (cells mL$^{-1}$) | 13 ± 11 | 5 ± 3 | 9 ± 3 | 6 ± 6 | 4 ± 2 |
| Centric diatoms (cells mL$^{-1}$) | < 1 | < 1 | < 1 | < 1 | 1 ± 2 |
| Pennate diatoms (cells mL$^{-1}$) | 2 ± 4 | < 1 | 2 ± 1 | 2 ± 3 | < 1 |
| Dinoflagellates (cells mL$^{-1}$) | 7 ± 6 | 4 ± 0 | 6 ± 2 | 3 ± 2 | 4 ± 0 |
| Microzooplankton (cells mL$^{-1}$) | 4 ± 3 | < 1 | 2 ± 2 | 1 ± 2 | < 1 |
| Nanoeukaryotes (cells mL$^{-1}$) | ND | 2.2 ± 1.4 × 10$^3$ | 1.5 ± 0.7 × 10$^3$ | 1.6 ± 0.7 × 10$^3$ | 1.4 × 10$^3$ |
| Picoeukaryotes (cells mL$^{-1}$) | ND | 4.5 ± 2.9 × 10$^3$ | 4.9 ± 3.7 × 10$^3$ | 1.5 ± 0.5 × 10$^3$ | 8 × 10$^2$ |
| Synechococcus (cells mL$^{-1}$) | ND | 3.8 ± 1.8 × 10$^3$ | 2.3 ± 1.1 × 10$^3$ | 1.4 ± 0.2 × 10$^3$ | 1 × 10$^3$ |
| Heterotrophic prokaryotes (cells mL$^{-1}$) | ND | 4.5 ± 3.2 × 10$^3$ | 2.3 ± 1.2 × 10$^3$ | 2.1 ± 2.3 × 10$^3$ | 3.2 × 10$^3$ |
| Detritus (particles mL$^{-1}$) | ND | 38.2 ± 14.9 × 10$^3$ | 63.8 ± 42.9 × 10$^3$ | 25.7 ± 18.6 × 10$^3$ | 2.57 × 10$^4$ |
3.4 Rates of net primary production, nitrogen uptake, and ammonium oxidation

Rates of bulk NPP were 2- to 6-fold higher in the SAZ and PFZ than has been reported previously for the Atlantic sector in winter (Mdutyana et al., 2020; Froneman et al., 1999) (Fig. 5a). By contrast, NPP was low in the OAZ, consistent with previous measurements (Kottmeier and Sullivan, 1987; Mdutyana et al., 2020). The relative contribution of the nano+ size class generally decreased southwards, from 85.4 % at 37.0°S to 24.4 % at 53.5°S, before increasing to >80 % near the SACCF.

The bulk NH$_4^+$ uptake rates ($\rho$NH$_4^+$) generally increased southwards from the STZ to the SAZ and PFZ and then decreased across the OAZ to reach a minimum at the southernmost station (Fig. 5b). In the nano+ size fraction, $\rho$NH$_4^+$ changed little latitudinally, although it was slightly lower in the OAZ than in the other zones. The contribution of nanoplankton to $\rho$NH$_4^+$ ranged from 32.8 % in the PFZ to 71.9 % in the STZ. The bulk NO$_3^-$ uptake rates ($\rho$NO$_3^-$) were also low in the STZ, while the highest $\rho$NO$_3^-$ was measured in the SAZ, with the rate then decreasing southwards. $\rho$NO$_3^-$ in the nano+ size class followed the same trend as total community $\rho$NO$_3^-$, with the nanoplankton accounting for 71.5 ± 0.3 % of bulk $\rho$NO$_3^-$ on average. The rates of bulk urea uptake ($\rho$Urea) were highest in the STZ, with the SAZ and the PFZ hosting similar rates, and the lowest rates were measured in the OAZ. $\rho$Urea for the nano+ size class followed a similar trend to bulk $\rho$Urea, and nanoplankton accounted for 51.8 % of $\rho$Urea in the SAZ, increasing to 66.1 % in the STZ. The uptake rates of the different N forms were not significantly correlated with one another or with the ambient N concentrations (Table S1).

Ammonium oxidation rates (NH$_4^+$ox) increased southwards, with higher NH$_4^+$ox in the OAZ and PAZ than in the STZ, SAZ, and PFZ (Fig. 5c). NH$_4^+$ox was generally comparable to previous wintertime measurements from the surface.

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Figure 5. (a) Surface rates of net primary production (NPP); rates of (b) ammonium ($\rho_{\text{NH}_4^+}$), (c) nitrate ($\rho_{\text{NO}_3^-}$), and (d) urea ($\rho_{\text{Urea}}$) uptake by the pico (light colours) and nano+ (dark colours) size fractions, with the full length of the bars indicating the bulk rates; and (e) rates of $\text{NH}_4^+$ oxidation. Error bars indicate ±1 standard deviation of duplicate experiments. The percentage of total NPP and N uptake attributable to the nano+ size fraction is written next to each bar in (a–d). NPP and $\text{NH}_4^+$ uptake were not measured for the nano+ size fraction at 58.5° S, and urea uptake was not measured at 50.7 and 55.5° S. Rates were not measured at the latitudes where no data are shown. In (b–e) the surface $\text{NH}_4^+$ concentration at each station is shown by the yellow circles. Leg N stations (at which samples were collected from Niskin bottles fired at 10 m) are indicated by black boxes surrounding the latitude. By contrast, samples were collected at the leg S stations (no square surrounding the latitude) from the ship’s underway system (~7 m). Fronts are indicated with arrows (labelled in e), and abbreviations are as in Fig. 1. Figure produced using the package ggplot2 (Wickham, 2016).

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of a bar with no stations at which no measurements were made while the absence of a bar with no " indicates that no cells were detected. Note that the abundances shown in (b) (top x axis) are > 2 orders of magnitude greater than those shown in (a). The "microplankton" shown in (a) are included on (b) (slim black bars) to illustrate the difference in abundance between the micro- and pico+nano populations. The frontal positions are indicated on (b), with abbreviations as in Fig. 1.

3.5 Plankton community composition

Microplankton abundance was low, with the highest cell counts recorded at stations 37.2 and 41.3° S in the STZ and no cells counted at 38.1° S (STZ) and 55.5° S (OAZ) (Fig. 6a). On average, microplankton abundance was higher in the STZ than in the SAZ, PFZ, and OAZ. The greatest diversity of microplankton groups was observed at 41.3° S in the STZ and at 50.0° S near the PF.

Centric diatoms (including *Planktoniella, Coscinodiscus,* and *Thalassiosira* species) were detected only at the southernmost station 58.9° S (3 cells mL⁻¹). Pennate diatoms (including *Pseudo-nitzschia, Pleurosigma,* and *Navicula* species) were more abundant in the STZ, PFZ, and OAZ, with negligible abundances in the SAZ. Higher pennate diatom abundances occurred near the PF (7 cells mL⁻¹), as has been observed in summer (e.g. Bracher et al., 1999). Dinoflagellates were identified at every station except 38.1° S and were most abundant in the STZ and PFZ. At all but three stations, small (≈ 15 µm) dinoflagellates were the most abundant group, although the larger *Protoperidinium* dinoflagellate species (mainly heterotrophic; Jeong and Latz, 1994) were almost as abundant in the PFZ and at 54.0° S. Microzooplankton (i.e. ciliates, 20–200 µM) were most abundant in the STZ and were also present in the PFZ at 46.1° S (3 cells mL⁻¹) and 48.9° S (3 cells mL⁻¹) and in the OAZ at 50.0° S (1 cell mL⁻¹) and 54.0° S (4 cells mL⁻¹). All other stations were characterised by negligible (< 1 cell mL⁻¹) microzooplankton abundances.

Nan- and picoeukaryotes, *Synechococcus,* and heterotrophic bacteria (collectively, "small cells") were roughly 10⁴ times more abundant than the microplankton (Fig. 6b). Notwithstanding a lack of data from the STZ, the highest small-cell abundances occurred in the SAZ near the SAF. Across the transect, picoeukaryotes were generally more abundant than all other phytoplankton groups (average contribution to total small cells of 12%–54% for picoeukaryotes, 7%–39% for nanoeukaryotes, and 15%–42% for *Synechococcus*). A similar trend has been observed for the Southern Ocean in spring (Detmer and Bathmann, 1997) and late summer (Fiala et al., 1998), in contrast to mid-summer observations showing nanoplankton dominance (e.g. Ishikawa et al., 2002; Weber and El-Sayed, 1987). Additionally, picoeukaryotes were 2 to 3 orders of magnitude more abundant in the SAZ and PFZ than in the OAZ. Nanoeukaryotes dominated near the PF at 50.0° S (39%) and in the southern OAZ at 55.5° S (36%), while *Synechococcus* dominated at 42.7 and 54.0° S (42% and 33%, respectively). In general, nanoeukaryote abundance was higher in the SAZ than in the PFZ and OAZ, as was that of *Synechococcus*.

The contribution of heterotrophic bacteria to the total small cells varied considerably (10%–62%), reaching a maximum south of the PF at 53.0 and 57.8° S (62% and 50%) and with higher abundances in the SAZ than in the PFZ and OAZ (Fig. 7). Additionally, heterotrophic bacterial abundances were 10-fold lower to 2-fold higher than the total pico- and nanophytoplankton cell counts. Detrital particles were most abundant near the southern edge of the SAF and were generally more abundant in the PFZ than in the SAZ and OAZ (Fig. S5).

3.6 2018–2019 cruises – ammonium concentrations

In early summer, surface NH₄⁺ concentrations were uniformly low across the transect (average of 0.11 ± 0.09 µM; Fig. 8a). South of the SAF, NH₄⁺ increased to an average concentration of 0.81 ± 0.92 µM by late summer (Fig. 8b). By winter 2019, the NH₄⁺ concentrations south of the SAF were ~ 40% lower than they had been in late summer (Fig. 8c), and they were similar to those observed in winter 2017 (0.50 ± 0.30 and 0.52 ± 0.11 µM, respectively), confirming that our 2017 observations are generally representative of the wintertime Southern Ocean. By early spring, the NH₄⁺ concentrations south of the SAF had declined to near or below detection (0.09 ± 0.08 µM; Fig. 8d) before rising again by late spring to an average value only slightly lower than that measured in winter (0.40 ± 0.74 µM; Fig. 8e). However, the late-spring NH₄⁺ concentrations were elevated only in
the PFZ (range of 0.11 ± 0.01 to 4.39 ± 0.03 µM, average of 0.77 ± 1.11 µM), as has been observed previously (Bathmann et al., 1997). Excluding the PFZ data yields a far lower late-spring average of 0.17 ± 0.11 µM south of the SAF, which we take as more broadly representative of this season.

### 3.7 Mixed-layer NH$_4^+$ residence time and NH$_4^+$ production rate estimates

The NH$_4^+$ residence time in winter 2017, computed using Eq. (5), ranged from 10 to 38 d (median of 21 d) south of the SAF and from 0 to 6 d (median of 2 d) north of the SAF. These values were estimated using wintertime measurements only and, as such, may not be representative of the transition from summer to winter. To refine our estimates, we used average $\rho$NH$_4^+$ and NH$_4^+$ concentration measurements. South of the SAF in late summer, $\rho$NH$_4^+$ = 50.6 ± 24.0 nM d$^{-1}$ and the NH$_4^+$ concentration was 0.81 ± 0.92 µM (Deary, 2020), which together yield an NH$_4^+$ residence time of 2 to 27 d (median of 5 d). The NH$_4^+$ residence time north of the SAF, calculated using $\rho$NH$_4^+$ = 20.7 ± 8.6 nM d$^{-1}$ and an NH$_4^+$ concentration of 0.16 ± 0.45 µM (Deary, 2020) was 1 to 17 d (median of 14 d).

The NH$_4^+$ production rate south of the SAF, calculated using Eq. (8) and an [NH$_4^+$]$_{\text{decline}}$ of 330 nM (i.e. the difference between late summer and winter 2019, 810 – 480 nM), of 141 d, and NH$_4^+$ removal rate of 50.6 ± 24.0 nM d$^{-1}$ (here, the average late-summer $\rho$NH$_4^+$ south of the SAF is used to approximate NH$_4^+$ removal rate), was 52.9 ± 25.0 nM d$^{-1}$. Similarly, north of the SAF (using an [NH$_4^+$]$_{\text{decline}}$ of 20 nM, i.e. 160 – 140 nM, and NH$_4^+$ removal rate of 20.7 ± 8.6 nM d$^{-1}$), the NH$_4^+$ production rate was 50.7 ± 9.3 nM d$^{-1}$. If we instead use the average NH$_4^+$ removal rate and NH$_4^+$ concentration measured in winter 2017 south (21.4 ± 0.6 nM d$^{-1}$ and 520 ± 110 nM) and north (18.4 ± 0.8 nM d$^{-1}$ and 80 ± 10 nM) of the SAF, the NH$_4^+$ production rate values were 23.4 ± 6.6 and 18.5 ± 6.6 nM d$^{-1}$, respectively. Using the range of NH$_4^+$ removal rate estimates and the average ambient NH$_4^+$ concentration measured south of the SAF in winter 2017 (16.7 to 31.2 nM d$^{-1}$ and 520 nM) and late summer 2019 (22.6 to 98.6 nM d$^{-1}$ and 810 nM), we calculate that over the late-summer-to-winter transition, the NH$_4^+$ production rate ranged from 18.8 to 100.9 nM d$^{-1}$ (compared to 6.3 to 28.8 nM d$^{-1}$ north of the SAF).

### 4 Discussion

#### 4.1 Drivers of NH$_4^+$ cycling in the surface layer of the Southern Ocean

Previous work has suggested that NH$_4^+$ accumulates in the Southern Ocean mixed layer following the late-summer increase in heterotrophy and then decreases into autumn as heterotrophic activity subsides, to be depleted by winter due to advective processes and biological removal (Koike et al., 1986; Serebrennikova and Fanning, 2004). However, our data show that NH$_4^+$ concentrations are elevated in the mixed layer in winter, particularly south of the SAF (Fig. 2). Similarly elevated winter surface-layer NH$_4^+$ has been observed previously in both the Atlantic and the Indian sectors, with concentrations typically increasing towards the south (Philibert et al., 2015; Mdutyana et al., 2020; Bianchi et al., 1997). Numerous overlapping processes are likely involved in setting the ambient NH$_4^+$ concentrations, as summarised in Fig. 9.

In this study, we directly measured the rates of NH$_4^+$ uptake and oxidation and estimated the rates of NH$_4^+$ production, along with qualitatively evaluating the role of heterotrophy from the relative abundance of heterotrophic bacteria, phytoplankton, and detritus. For the NH$_4^+$ cycle processes shown in Fig. 9 that are not quantified or inferred from our dataset, we consider their potential role in Southern Ocean NH$_4^+$ cycling based on findings reported in the literature.

The high NH$_4^+$ concentrations observed south of the SAF in winter may result from net NH$_4^+$ accumulation during late summer, autumn, and/or winter. The persistence of elevated NH$_4^+$ concentrations that are near-homogeneously distributed throughout the mixed layer is consistent with a residence time for the winter NH$_4^+$ reservoir in excess of the timescale for upper-ocean mixing. Indeed, we calculate a median residence time of 21 d south of the SAF, compared to 2 d north of the SAF. One implication of the long residence time computed for the polar zones is that the wintertime NH$_4^+$ pool likely reflects both ongoing processes and those that occurred earlier in the year. We posit that the elevated NH$_4^+$ concentrations south of the SAF may result from higher wintertime rates of NH$_4^+$ production than removal and/or from the gradual but incomplete depletion in winter of NH$_4^+$ produced mainly in late summer and autumn. We evaluate both possibilities throughout the discussion below.
Figure 8. Surface concentrations of NH$_4^+$ across the eastern Atlantic sector of the Southern Ocean measured between December 2018 and November 2019. Five unique transects (additional to the winter 2017 dataset presented in Fig. 2a) are shown: (a) early summer 2018, (b) late summer 2019, (c) winter 2019, (d) early spring 2019, and (e) late spring 2019. (f) The proposed seasonal cycle of NH$_4^+$ concentrations in the mixed layer south of the Subantarctic Front. The colour gradient in (f) shows the transition between late summer and late winter. Panels (a, b) cover a latitudinal extent of 30–70°S, while panels (c–e) cover 30–60°S due to the presence of sea ice. Abbreviations are as in Fig. 1, with AZ referring to the combined OAZ and PAZ. Figure produced using the package ggplot2 (Wickham, 2016).

Figure 9. Schematic of the possible mixed-layer NH$_4^+$ assimilation and production pathways. Bold text indicates components of the NH$_4^+$ cycle that were directly measured in this study (seawater concentrations of NH$_4^+$, NO$_2^-$, and urea; phytoplankton, bacterial, and microzooplankton cell abundances), and dotted lines indicate processes for which we have direct rate measurements (phytoplankton uptake of NH$_4^+$, oxidation of NH$_4^+$ to NO$_2^-$). Dashed-line boxes represent the atmosphere and sea ice, with all other processes occurring in the ocean. DON – dissolved organic nitrogen; NH$_3$(aq) – aqueous (seawater) ammonia; NH$_4$(g) – ammonium aerosols (including ammonium sulfate, ammonium bisulfate, and ammonium nitrate); NH$_3$(g) – ammonia gas.

4.1.1 Ammonium removal

Ammonium assimilation. Microbial growth is limited in the winter Southern Ocean (Arrigo et al., 2008; Smith et al., 2000; Takao et al., 2012), resulting in low cell abundances and nutrient uptake rates (Church et al., 2003; Iida and Odate, 2014; Mdutyana et al., 2020). However, while the concentrations of chl $a$ and rates of NPP were low across our transect, they were not negligible (Figs. 3a and 5a), consistent with previous reports for this season (Mordy et al., 1995; Pomeroy and Wiebe, 2001). Southern Ocean phytoplankton are adapted to survive suboptimal conditions; for example, numerous species achieve their maximum growth rates at temperatures that are considerably lower than the optimal growth temperatures of temperate and tropical species (2–9 °C versus 10–30 and 15–35 °C, respectively), with sharp declines in growth rates observed at temperatures outside this range (Boyd et al., 2013; Coello-Camba and Agusti, 2017; Fiala and Oriol, 1990). In addition, ice-free Southern Ocean waters typically extend to < 60°S in the eastern Atlantic and western Indian sectors in winter, so even though irradiance levels may not be optimal for phytoplankton growth, there is always some light available for photosynthesis. The hostile wintertime conditions of the open Southern Ocean do not, therefore, prevent ecosystem functioning, although the microbial dynamics and associated biogeochemical processes
differ from those occurring in summer (Smart et al., 2015; Mdutyana et al., 2020).

We measured fairly low surface NH$_4^+$ uptake rates (3.0–13.2 nM d$^{-1}$; Fig. 5b) compared to previous wintertime observations (range from 32–66 nM d$^{-1}$; Cota et al., 1992; Mdutyana et al., 2020; Philibert et al., 2015). Such low rates, if generally representative of winter, would limit mixed-layer NH$_4^+$ drawdown, especially south of the PF where ρNH$_4^+$ was particularly low. Recycled N (NH$_4^+$ + urea) nonetheless accounted for most of the N assimilated during winter, including in the AZ (Fig. 5b).

The available δ$^{15}$N-PON data suggest that the preferential reliance of phytoplankton on recycled N may have persisted from the late summer. In theory, PON generated in early summer through mid-summer from the assimilation of upwelled NO$_3^-$ (δ$^{15}$N-NO$_3^-$ of 5.2‰ in the AZ and 6.2‰ in the SAZ; Smart et al., 2015; Fripiat et al., 2019, 2021) will have δ$^{15}$N of ~0‰ in the AZ and 1‰~2‰ in the SAZ given the isotope effect of NO$_3^-$ assimilation and the degree of seasonal NO$_3^-$ drawdown (Sigman et al., 1999; Granger et al., 2004, 2010). Such δ$^{15}$N-PON values have indeed been measured in the early-summer and mid-summer Southern Ocean (Lourey et al., 2003; Smart et al., 2020; Soares et al., 2015). By late summer, δ$^{15}$N-PON has been observed to decline to between −5‰ and −1‰, with the lowest values occurring in the AZ (Lourey et al., 2003; Smart et al., 2020; Trull et al., 2008). Since the δ$^{15}$N of recycled N is expected to be low (<0‰; Checkley and Miller, 1989; Macko et al., 1986), the early-to-late-summer decline in δ$^{15}$N-PON implicates a switch from dominantly NO$_3^-$-supported to dominantly recycled N-supported phyttoplankton growth (Lourey et al., 2003). For the SAZ, the subsequent late-summer-to-winter rise in δ$^{15}$N-PON (i.e. from ca. −1‰ to 1‰~2.5‰; Fig. 4) has previously been attributed to PON decomposition by heterotrophic bacteria (Smart et al., 2020), during which $^{14}$N-NH$_4^+$ is preferentially remineralised, leaving the remaining PON enriched in $^{15}$N (Möbius, 2013). That NH$_4^+$ concentrations are not elevated in the SAZ mixed layer in winter (Fig. 2b) indicates that the remineralised NH$_4^+$ is rapidly re-assimilated by phytoplankton and/or oxidised to NO$_3^−$ in this zone. In the AZ, the much lower δ$^{15}$N-PON of $\sim$3‰ to $\sim$1‰ that we observe in winter surface waters requires the sustained assimilation of low-δ$^{15}$N N (i.e. recycled N) to offset a remineralisation-driven δ$^{15}$N rise akin to that of the SAZ. We conclude that Southern Ocean phytoplankton preferentially consume regenerated N from late summer until at least July (albeit at low rates in winter), particularly south of the PF.

The fact that NH$_4^+$ accumulated in the winter mixed layer despite being the preferred phyttoplankton N source in late summer through winter implies that low rates of NH$_4^+$ uptake contributed to its accumulation. Multiple factors may cause low rates of photoautotrophic NH$_4^+$ assimilation, including depleted NH$_4^+$ and micronutrient concentrations, light limitation, and low temperatures. North of the SAF, NH$_4^+$ concentrations below detection likely limited ρNH$_4^+$, as evidenced by the fact that in a series of experiments conducted on the same cruise, ρNH$_4^+$ increased with the addition of NH$_4^+$ at these stations (Mdutyana, 2021). By contrast, south of the SAF, NH$_4^+$ concentrations were similar to or higher than the half-saturation constant ($K_{\text{NH}_4}$) derived for NH$_4^+$ uptake in the winter Southern Ocean (0.2 to 0.4 µM; Mdutyana, 2021), suggesting that something other than NH$_4^+$ availability was limiting to phytoplankton at these latitudes.

Iron is not directly involved in NH$_4^+$ assimilation but is required for electron transport during photosynthesis and respiration, as well as for chlorophyll synthesis (Raven, 1988). While iron limitation is widespread across the Southern Ocean (Janssen et al., 2020; Pausch et al., 2019; Viljoen et al., 2019), iron availability appears to be higher in winter than during other seasons (Mtschali et al., 2019; Tagliabue et al., 2014) due to enhanced mixing, storms, and increased aeolian deposition (Coale et al., 2005; Honjo et al., 2000; Sedwick et al., 2008). The fact that ρNO$_3^-$ and ρNH$_4^+$ were generally similar across the transect (Fig. 5b) argues against a dominant role for iron in controlling ρNH$_4^+$ since NO$_3^-$ consumption has a far higher iron requirement than NH$_4^+$ assimilation (Morel et al., 1991).

In contrast to NH$_4^+$ and iron availability, light limitation is exacerbated in winter due to low insolation, increased cloud cover, and mixed layers that can be hundreds of metres deeper than the euphotic zone (Buongiorno Nardelli et al., 2017; Sallée et al., 2010). Light is thus often considered the dominant constraint on Southern Ocean primary productivity in this season (Thomalla et al., 2011; Llort et al., 2019; Wadley et al., 2014). However, since NH$_4^+$ assimilation by phytoplankton is fairly energetically inexpensive (Dortch, 1990), it should occur even under low-light conditions (recognising that light remains critical for coincident CO$_2$ fixation). Heterotrophic bacteria can also consume NH$_4^+$ (Kirchman, 1994), including in the dark, as they derive energy from organic carbon oxidation rather than light. At an ecosystem level, therefore, NH$_4^+$ assimilation may not be primarily limited by light, although this parameter clearly strongly controls the rate and distribution of NPP (Fig. 5a).

Previous observations suggest that temperature can influence NH$_4^+$ uptake, especially in winter (Gilbert, 1982; Reay et al., 2001). The negative effect of temperature appears to be enhanced under high-nutrient and low-light conditions, at least in the case of phytoplankton growth rates (Baird et al., 2001). Experiments conducted coincidentally with our sampling showed that the maximum rate of NH$_4^+$ uptake ($V_{\text{max}}$) achievable by the in situ community was strongly negatively correlated with temperature and latitude (Mdutyana, 2021), with the latter parameter representing the combined role of light, temperature, and possibly iron, the average concentration of which appears to increase from the SAZ to the AZ (Tagliabue et al., 2012). We conclude that these three drivers, along with NH$_4^+$ availability north of the SAF, may all play a role in controlling photoautotrophic NH$_4^+$ assimilation in
the winter Southern Ocean, with complex interactions among them that are difficult to disentangle.

In addition to physical and chemical limitations, microbial preference for other N species may impact NH$_4^+$ depletion. For example, the preferential uptake of urea and/or other dissolved organic N (DON) species by some organisms (e.g. picocyanobacteria, cyano- or heterotrophic bacteria) could cause a net decrease in the total NH$_4^+$ uptake rates. While urea has been shown to constitute a large fraction of the total N assimilated by Southern Ocean phytoplankton in summer and autumn (albeit mainly in the SAZ; Joubert et al., 2011; Thomailla et al., 2011), we measured fairly low urea (Fig. 5b), which is perhaps unsurprising given the low ambient urea concentrations (Table 1). The exceptions were stations 37.0 and 43.0°S where ρUrea was higher than ρNH$_4^+$, coincident with very low ambient NH$_4^+$ (0.10 µM and below detection, respectively) and relatively high urea concentrations (0.36 and 0.15 µM, respectively).

Community composition can also alter the N uptake regime. Small phytoplankton, such as the numerically dominant nano- and picocyanobacteria, are more likely to consume NH$_4^+$ and urea than NO$_3^-$ (Koike et al., 1986; Lee et al., 2012, 2013), especially under conditions of iron and light limitation (Sunda and Huntsman, 1997). Across our transect, reduced N (i.e. NH$_4^+$ + urea) uptake exceeded NO$_3^-$ uptake for both the total phytoplankton community (transect average of 12.0 ± 0.9 nM d$^{-1}$ for reduced N versus 5.8 ± 1.0 nM d$^{-1}$ for NO$_3^-$, f ratio of 0.36) and the pico size fraction (5.0 ± 1.2 nM d$^{-1}$ versus 1.9 ± 1.2 nM d$^{-1}$, f ratio of 0.27; Fig. 5b). That said, the NO$_3^-$ uptake rates were not negligible, including in the pico size fraction. In the PFZ and AZ, NO$_3^-$ uptake by the picoplankton was far more strongly correlated with the abundance of picocyanobacteria than *Synechococcus* ($r = 0.75$ and $r = 0.03$, respectively), consistent with observations of dominant reliance on NO$_3^-$ by picocyanobacteria and NH$_4^+$ by *Synechococcus* in other ocean regions (Fawcett et al., 2011, 2014; Painter et al., 2014). Additionally, *Synechococcus* abundance was strongly correlated with NH$_4^+$ concentration south of the SAF ($r = 0.65$). In the nano+ size class, NO$_3^-$ uptake was likely driven in the SAZ by dinoflagellates and nanoeukaryotes and in the PFZ and AZ by diatoms, which remain active in these zones in winter (Weir et al., 2020). By contrast, nanoeukaryotes, which have a higher per-cell nutrient requirement than the equally abundant picoeukaryotes, may have dominated NH$_4^+$ uptake in the PFZ and AZ given that higher nanoeukaryote abundances corresponded with lower NH$_4^+$ concentrations at a number of stations (e.g. stations 50.0, 51.1, and 55.5°S; Fig. 6b).

The low abundances of diatoms and dinoflagellates and absence of coccolithophores across our transect (Fig. 6a) are expected given the limitations imposed on nutrient uptake and CO$_2$ fixation by winter Southern Ocean conditions. The lower surface-area-to-volume ratio of large cells means that they rapidly experience diffusion limitation of NH$_4^+$ and micronutrient uptake and are more susceptible to light limitation (Finkel et al., 2004), resulting in their being outcompeted by smaller species for essential resources (Franck et al., 2005; Cavender-Bares et al., 1999). The near absence of centric diatoms is also best explained thus, particularly given their low surface-area-to-volume ratio compared to the more abundant pennate species (Kobayashi and Takahashi, 2002) that are more likely to consume NH$_4^+$ (Semeneh et al., 1998). Diatom success in winter may also be limited by enhanced mixing, as this group generally prefers stratified waters (Kopczynska et al., 2007).

In sum, NH$_4^+$ uptake rates were low across our transect but not negligible, indicating that phytoplankton activity in winter, which is dominated by smaller species, is a sink for NH$_4^+$. The hostile conditions of the winter Southern Ocean imposed limitations on NH$_4^+$ uptake that varied with latitude, with NH$_4^+$ concentrations controlling ρNH$_4^+$ north of the SAF, while light and temperature were important south of the SAF. Additionally, *Synechococcus*, nanoeukaryotes, and pennate diatoms likely dominated NH$_4^+$ assimilation, consistent with previous observations from the Southern Ocean and elsewhere (Klawonn et al., 2019; Semeneh et al., 1998).

**Ammonium oxidation.** Nitrification removes more mixed-layer NH$_4^+$ in winter than phytoplankton assimilation south of the PF, with NH$_4^+$ oxidation rates that were 2 to 5 times the co-occurring NH$_4^+$ uptake rates (Fig. 5c). The comparative success of ammonia oxidisers may be due to decreased competition with phytoplankton for NH$_4^+$, augmented by decreased photoinhibition (Wan et al., 2018; Lu et al., 2020), elevated NH$_4^+$ availability (Baer et al., 2014; Mdutyana et al., 2020; Mdutyana, 2021), and the apparently minor effect of temperature on NH$_4^+$ oxidation (Bianchi et al., 1997; Baer et al., 2014; Horak et al., 2013; Mdutyana, 2021). One implication of the dominance of NH$_4^+$ oxidation in winter is that in addition to the limitations on photoautotrophic NH$_4^+$ assimilation discussed above, low phytoplankton success in the AZ may result from nitrifiers outcompeting phytoplankton for scarce resources (e.g. trace elements required for enzyme functioning, such as iron and copper; Amin et al., 2013; Maldonado et al., 2006; Shafee et al., 2019) under conditions of low incident light and enhanced mixing.

The $K_m$ derived for NH$_4^+$ oxidation in the winter Southern Ocean has recently been reported to be low (0.03 to 0.14 µM), with ammonia oxidisers observed to become saturated at ambient NH$_4^+$ concentrations of ~0.1–0.2 µM (Mdutyana, 2021). This means that south of the SAF in winter 2017, ammonia oxidisers were not substrate limited (as implied by the lack of correlation between NH$_4^+$ oxid and NH$_4^+$ concentration; Table S1), which raises the question of why NH$_4^+$ oxidation did not occur at higher rates. The answer may indirectly involve temperature, in that psychrophilic organisms can be less responsive to high substrate concentrations at low temperatures (Baer et al., 2014). Another possibility is that NH$_4^+$ oxidation was iron-limited (Shiozaki et al., 2016; Shafee et al., 2019; Mdutyana, 2021). In any case, ammonia oxidisers were moderately successful across the surface Southern

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Ocean in winter, with low light, reduced competition with phytoplankton, and substrate repletion likely explaining the elevated NH$_4^+$ oxidation rates south of the PF compared to the stations to the north.

4.1.2 Ammonium production and other sources of ammonium

NH$_4^+$ production must have been sustained during the winter to maintain a mixed-layer NH$_4^+$ pool south of the SAF that was high in concentration relative to the early summer. Indeed, the residence time estimated for NH$_4^+$ in winter (10 to 38 d) is considerably shorter than the transition from late summer to winter (approximately 3 months), indicating that heterotrophic NH$_4^+$ production, which would have occurred coincidentally with NH$_4^+$ consumption, must have been ongoing in winter. We estimate the rate of this wintertime NH$_4^+$ production to be 23.4 ± 6.6 nM d$^{-1}$.

**Heterotrophic activity by bacteria.** Heterotrophic bacteria contribute significantly to NH$_4^+$ production in the Southern Ocean (Hewes et al., 1985; Koike et al., 1986; Tréguer and Jacques, 1992), including in winter (Rembauville et al., 2017). In our dataset, lower ratios of photosynthetic to heterotrophic cells were observed at stations with higher NH$_4^+$ concentrations (e.g. stations 48.9, 53.0, 54.0, and 57.8◦S; Fig. S5a), consistent with a role for the heterotrophic bacteria present at the time of sampling in generating the ambient NH$_4^+$ pool. The potential for ongoing heterotrophic activity can also be inferred from the high detrital particle counts along the transect (Fig. 7). However, since heterotrophic bacteria are likely more active in late summer and autumn when the temperature and the supply of labile PON are higher (Becquevort et al., 2000; Dennett et al., 2001; Pomeroy and Wiebe, 2001; Smart et al., 2020), we expect that the winter NH$_4^+$ pool includes NH$_4^+$ produced in late summer and autumn. A further consideration is assimilation of NH$_4^+$ by heterotrophic bacteria, reported to occur at elevated rates in the Southern Ocean mixed layer in winter (Mduyana et al., 2020; Sect. S3). If this process is a persistent feature of the winter Southern Ocean, it will decrease the net contribution of heterotrophic bacteria to NH$_4^+$ accumulation. We conclude that it is unlikely that the surface NH$_4^+$ pool measured in winter derived solely from wintertime bacterial NH$_4^+$ production given that yet higher NH$_4^+$ concentrations have been observed in late summer and autumn (Becquevort et al., 2000; Dennett et al., 2001), including in the present study (see Sect. 4.2 below).

**Heterotrophic activity by zooplankton.** While the microzooplankton enumerated in this study occurred at very low abundances, those that were present likely contributed to the NH$_4^+$ flux. For example, at stations 48.9 and 54.0◦S in the PFZ and AZ, respectively, both the ratios of photosynthetic to heterotrophic cells and the absolute abundances of heterotrophic bacteria were low, while the microzooplankton abundances and NH$_4^+$ concentrations were elevated compared to at nearby stations. The implication of these observations is that elevated microzooplankton abundances may help to explain high NH$_4^+$ concentrations in waters with low numbers of heterotrophic bacteria, although we note that this scenario occurred only at two stations. On balance, we posit that microzooplankton are less important for wintertime NH$_4^+$ production than heterotrophic bacteria given their low abundances in the surface layer (Fig. 6a; Atkinson et al., 2012). That said, it is possible that the contribution of microzooplankton (and/or macrozooplankton) to the NH$_4^+$ pool surpasses that of heterotrophic bacteria under certain conditions (Koike et al., 1986; Priddle et al., 1998), such as in (late) summer and near regions of frontal upwelling in response to elevated rates of phytoplankton biomass accumulation.

Above, we have assumed that NH$_4^+$ production is the direct result of heterotrophy. However, there are other possible mechanisms of NH$_4^+$ supply that should be considered. We briefly address some of these processes below, noting that for most, there are very few to no observations available from the Southern Ocean.

**DON cycling.** NH$_4^+$ can be released by heterotrophic bacteria that directly consume DON (e.g. urea; Billen, 1983; Tupas and Koike, 1990) and possibly also by ammonia oxidisers that convert DON to NH$_4^+$ intracellularly, through the equilibration of the intra- and extracellular NH$_4^+$ pools (Kitzinger et al., 2019). DON can also be converted to NH$_4^+$ through photodegradation by UV radiation (e.g. Aarnos et al., 2012). Bacterial decomposition of DON (rather than PON) to NH$_4^+$ is implicit in most estimates of ammonification, however, and cellular NH$_4^+$ efflux by ammonia oxidisers is likely extremely low given that they require NH$_4^+$ to fix CO$_2$. Additionally, the low light flux to the surface of the Southern Ocean in winter means that photodegradation will not yield a significant supply of NH$_4^+$. Thus, DON conversion to NH$_4^+$, through any mechanism, is probably negligible.

**External inputs of ammonium.** High surface ocean NH$_4^+$ concentrations may theoretically derive from external inputs of NH$_4^+$, such as N$_2$ fixation, NH$_4^+$ aerosol deposition, or sea-ice melt. N$_2$ fixation should be below detection in the winter Southern Ocean due to the cold temperatures, low-light and low-iron conditions, and high NO$_3^-$ concentrations (Jiang et al., 2018; Knapp et al., 2012; Kustka et al., 2003). NH$_4^+$ aerosols are unlikely to be abundant over regions of the Southern Ocean remote from islands and coastal Antarctica, particularly in winter when NH$_4^+$ aerosol concentrations have been shown to reach a minimum (Legrand et al., 1998; Xu et al., 2019). Moreover, the aerosols that are present over the open Southern Ocean will derive mainly from surface ocean NH$_4^+$ efflux; once re-deposited, this NH$_4^+$ does not constitute a new input to surface waters (Altieri et al., 2021). Finally, since our sampling took place before the sea ice reached its northernmost extent (Cavalieri and Parkinson, 2008), the dominant process would have been sea-ice formation rather than sea-ice melt, the latter an occasional source of NH$_4^+$ (Kattner et al., 2004; Zhou et al., 2014). In any case, we ob-
observed elevated NH$_4^+$ concentrations as far north as 46° S, ~1700 km beyond the influence of sea-ice melt.

4.2 Seasonal cycling of NH$_4^+$ in the Southern Ocean mixed layer south of the SAF

The NH$_4^+$ concentration data collected over the 2018/19 annual cycle provide context for interpreting our winter 2017 dataset, allowing us to address our hypothesis that NH$_4^+$ production in late summer and autumn contributes to the elevated NH$_4^+$ concentrations measured in winter.

The very low NH$_4^+$ concentrations observed in early summer (Fig. 8a) are consistent with high rates of phytoplankton NH$_4^+$ assimilation during the spring and early-summer growing period (Mdutyana et al., 2020; Savoye et al., 2004; Daly et al., 2001). By late summer, the NH$_4^+$ concentrations increased (Fig. 8b) presumably due to elevated heterotrophic activity (i.e. bacterial decomposition and zooplankton grazing) following the accumulation of algal biomass (Menge-sha et al., 1998; Le Moigne et al., 2013), coupled with iron limitation and/or silicate limitation of phytoplankton (Hiscock et al., 2003; Sosik and Olson, 2002) and enhanced grazing pressure (Becquevort et al., 2000). Mixed-layer NH$_4^+$ remained high between late summer and winter (Fig. 8b–c), likely due to sustained heterotrophic NH$_4^+$ production in excess of NH$_4^+$ removal. This notion is supported by estimates of the residence time of NH$_4^+$. We calculate that in summer, the in situ NH$_4^+$ pool would be depleted in 2 to 27 d (median of 5 d) without coincident NH$_4^+$ production. In addition, the net decline in NH$_4^+$ concentration of 0.31 ± 0.97 µM between late summer and winter requires an average NH$_4^+$ production rate of 52.8 ± 25.0 nM d$^{-1}$ given the observed NH$_4^+$ assimilation rates. This estimate is remarkably similar to the only measurements of NH$_4^+$ regeneration available for the Southern Ocean, measured near the Antarctic Peninsula in summer (average of 55 nM d$^{-1}$; Goeyens et al., 1991).

By early spring, the NH$_4^+$ concentrations had declined (Fig. 8d), implicating increased photosynthetic activity, and thus nutrient assimilation, following the alleviation of light limitation. We suggest that any NH$_4^+$ remaining in late winter would have been consumed in early spring prior to significant NO$_3^-$ drawdown because far less energy (i.e. light) is required for its assimilation (Dortch, 1990). The high NH$_4^+$ concentrations subsequently observed in late spring (mainly in the PFZ; Fig. 8e) can be explained by elevated heterotrophic activity in response to high levels of regional phytoplankton growth driven by the frontal upwelling of limiting nutrients (Becquevort et al., 2000; Mayzaud et al., 2002).

From our six transects of surface NH$_4^+$ concentrations across the Southern Ocean, we propose a seasonal cycle for mixed-layer NH$_4^+$ south of the SAF (Fig. 8f). Our proposal is consistent with previous characterisations of the early-summer-to-autumn evolution of Southern Ocean NH$_4^+$ concentrations (i.e. from below detection due to phytoplankton assimilation to elevated due to net heterotrophy). However, it contradicts the hypothesis that NH$_4^+$ will subsequently decline due to persistent but low rates of photosynthesis that yield insufficient biomass to support elevated heterotrophy in autumn, thus driving a coincident decrease in photosynthetic and heterotrophic activity (Koike et al., 1986; Serebrennikova and Fanning, 2004). Instead, our data evince a gradual decline in mixed-layer NH$_4^+$ concentrations from late summer through winter. This decline can be explained by heterotrophic NH$_4^+$ production outpacing NH$_4^+$ removal in late summer and/or autumn, with NH$_4^+$ regeneration decreasing during winter to lower rates than the combined rate of NH$_4^+$ assimilation and oxidation. By late spring, NH$_4^+$ reaches concentrations similar to those observed in early summer as the improved growing conditions (i.e. elevated light and iron availability; Ellwood et al., 2008; Mtshali et al., 2019) allow phytoplankton to rapidly consume any NH$_4^+$ remaining at the end of winter and subsequently produced in spring. An exception to this scenario is elevated, localised NH$_4^+$ production near fronts, such as we observed in late spring 2019, which likely resulted from biological activity supported by frontal upwelling of silicate- and iron-bearing Upper Circumpolar Deep Water (Prézelin et al., 2000).

5 Summary and implications

Our study of the upper Southern Ocean, focused on the infrequently sampled winter season, provides new insights into the internal cycling of N in the mixed layer of a globally important region. We attribute the elevated NH$_4^+$ concentrations that persist in the winter mixed layer south of the SAF to sustained heterotrophic NH$_4^+$ production in excess of NH$_4^+$ removal, driven by the temperature, light, and possibly iron limitation of phytoplankton and nitrifiers. We further suggest that heterotrophic bacteria are the main NH$_4^+$ producers in winter and that the contribution of external sources to the Southern Ocean’s mixed-layer NH$_4^+$ pool is negligible. From observations of surface NH$_4^+$ concentrations made between December 2018 and November 2019, we deduce that the elevated mixed-layer NH$_4^+$ concentrations measured in winter cannot be due solely to wintertime NH$_4^+$ production. Instead, we propose that NH$_4^+$ accumulates to its highest concentrations in late summer following the peak phytoplankton growing season, after which sustained heterotrophy throughout the autumn and winter prevents this NH$_4^+$ from being fully depleted until the early spring, even though the rate of NH$_4^+$ removal must exceed that of NH$_4^+$ production over this period. Measurements of heterotrophic NH$_4^+$ production rates are required to confirm the hypothesised seasonal cycle of NH$_4^+$ in the Southern Ocean mixed layer, and higher-spatial-resolution sampling of plankton community composition and N removal rates may help to explain local variability in NH$_4^+$ concentrations, particularly near the fronts.
In net, the Southern Ocean mixed layer is a biological source of CO$_2$ to the atmosphere in autumn and winter (Mongwe et al., 2018). The persistence of elevated NH$_4^+$ concentrations across the polar Southern Ocean between late summer and winter implies that this biological CO$_2$ production occurs not only because NO$_3^-$ drawdown is weak relative to NO$_2^-$ supply at this time (e.g. Gibson and Trull, 1999; Gray et al., 2018; Hauck et al., 2015; Mongwe et al., 2018; Shadwick et al., 2015) but also because the ambient conditions allow for NH$_4^+$ accumulation. There are additional implications of our observations. For example, NH$_4^+$ concentrations > 1 µM (and at times > 0.5 µM) have been reported to inhibit NO$_3^-$ assimilation, including in the Southern Ocean (Cochlan, 1986; Goeijens et al., 1995; Philibert et al., 2015; Reay et al., 2001). Inhibition of NO$_3^-$ assimilation due to the seasonal accumulation of NH$_4^+$ would constitute an inefficiency in the biological pump. However, we observed little evidence of this effect in winter 2017 – the southward decrease in $ρ$NO$_3^-$ was not stronger than that of $ρ$NH$_4^+$ despite the latitudinal increase in NH$_4^+$ concentration, and we observed no relationship between NH$_4^+$ concentration and the proportion of NO$_3^-$ to NO$_2^+$ + NH$_4^+$ uptake (i.e. the $f$ ratio; Table S1).

The implications of NH$_4^+$ cycling extend beyond the upper ocean to the atmosphere, since ammonium aerosols that influence Earth’s albedo (Tevlin and Murphy, 2019) are formed in the marine boundary layer from reactions of NH$_3$ gas with acidic species. In the remote Southern Ocean, marine NH$_3$ emissions, which are the largest natural contributors to NH$_3$ globally, are likely the dominant local source of NH$_3$ to the atmosphere (Paulot et al., 2015). Surface ocean NH$_4^+$ concentrations play a central role in determining the sign and magnitude of the air–sea NH$_4^+$ flux, along with wind speed, surface ocean temperature, and pH. Therefore, the biogeochemical pathways that underpin seasonal changes in surface ocean NH$_4^+$ concentrations represent an important control on the remote Southern Ocean air–sea NH$_3$ flux, with consequences for aerosol composition, cloud formation, and climate (Altieri et al., 2021).

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