RESEARCH ARTICLE

Seasonal dynamics of carbonate chemistry, nutrients and CO$_2$ uptake in a sub-Arctic fjord

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Environmental change can have a significant impact on biogeochemical cycles at high latitudes and be particularly important in ecologically valuable fjord ecosystems. Seasonality in biogeochemical cycling in a sub-Arctic fjord of northern Norway (Kaldfjorden) was investigated from October 2016 to September 2018. Monthly changes in total inorganic carbon (C$_{total}$), alkalinity (A$_T$), major nutrients and calcium carbonate saturation (Ω) were driven by freshwater discharge, biological production and mixing with subsurface carbon-rich coastal water. Stable oxygen isotope ratios indicated that meteoric water (snow melt, river runoff, precipitation) had stratified and freshened surface waters, contributing to 81% of the monthly C$_{total}$ deficit in the surface layer. The timing and magnitude of freshwater inputs played an important role in Ω variability, reducing A$_T$ and C$_{total}$ by dilution. This dilution effect was strongly counteracted by the opposing effect of primary production that dominated surface water Ω seasonality. The spring phytoplankton bloom rapidly depleted nitrate and C$_{total}$ to drive highest Ω (~2.3) in surface waters. Calcification reduced A$_T$ and C$_{total}$, which accounted for 21% of the monthly decrease in Ω during a coccolithophore bloom. Freshwater runoff contributed C$_{total}$, A$_T$ and silicates of terrestrial origin to the fjord. Lowest surface water Ω (~1.6) resulted from organic matter remineralisation and mixing into subsurface water during winter and spring. Surface waters were undersaturated with respect to atmospheric CO$_2$, resulting in modest uptake of ~0.32 ± 0.03 mol C m$^{-2}$ yr$^{-1}$. Net community production estimated from carbon drawdown was 14 ± 2 g C m$^{-2}$ yr$^{-1}$ during the productive season. Kaldfjorden currently functions as an atmospheric CO$_2$ sink of 3.9 ± 0.3 g C m$^{-2}$ yr$^{-1}$. Time-series data are vital to better understand the processes and natural variability affecting biogeochemical cycling in dynamic coastal regions and thus better predict the impact of future changes on important fjord ecosystems.

Keywords: Carbonate chemistry; Ocean acidification; Nutrient cycling; Net community production; Sub-Arctic fjord; Norway

Introduction

Coastal oceans and marine shelves are regions of enhanced uptake of atmospheric carbon dioxide (CO$_2$) and disproportionately large fraction of primary productivity relative to their areal coverage (Thomas et al., 2004; Borges et al., 2005; Thomas et al., 2009; Cai, 2011). This biological productivity is fuelled by oceanic, atmospheric, and terrestrial nutrient inputs that drive biogeochemical exchanges between the coastal and open ocean and enhanced burial of organic and inorganic carbon (Thomas et al., 2004; Chen and Borges, 2009). Observations in high latitude fjord and coastal regions have shown these regions to be predominantly sinks for atmospheric CO$_2$, despite large regional variability (Omar et al., 2005; Else et al., 2008b; Signorini et al., 2013; Fransson et al., 2014; Evans et al., 2015; Omar et al., 2016; Yasunaka et al., 2016; Ericson et al., 2018). Mid- and low-latitude coastal regions were thought to be predominantly oceanic CO$_2$ sources, owing to higher water temperatures and organic matter inputs (Borges et al., 2005; Cai et al., 2006; Chavez et al., 2007; Chen and Borges, 2009). However, increased observations and the development of high-resolution regional models have enabled better resolution of carbon cycling in these regions and identified areas of annual CO$_2$ uptake in the coastal ocean (Takahashi et al., 2009; Laruelle et al., 2014; Bourgeois et al., 2016; Fennel et al., 2019). Seasonal dynamics are more variable and typically amplified, with temporal changes occurring faster, in the coastal ocean compared to open ocean environments. The processes controlling the biogeochemical cycling in coastal systems are difficult to determine without sustained seasonal measurements of the hydrographic, biogeochemical and meteorological variables (Borges et al., 2005; Bozec et al., 2006). Recent research efforts have significantly improved the understanding and quantification of carbon cycling...
in the coastal ocean, e.g., Fennel et al. (2019). However, seasonal baseline estimates of biogeochemical cycling and air–sea CO$_2$ exchange in some sub-Arctic and Arctic regions are limited. Greater spatial surveys would enable a better understanding of how climate change may affect carbon cycling at high latitudes.

The continental shelf and fjords of the Norwegian coast support a rich food web, cold-water coral reefs, large mammals (whales, seals), seabirds, valuable fish stocks and spawning grounds, and are used for recreation and aquaculture (Matthews and Sands, 1973; Erga and Heimdal, 1984; Salvanes and Noreide, 1993; Fosså et al., 2002; Asplin et al., 2014; Brattegard et al., 2011). The marine environment of the fjords is influenced by water mass circulation, tides, wind regimes, and freshwater inputs. Biogeochemical cycling and primary production exhibit strong seasonality that is controlled by variations in meteorology and hydrography (Eilertsen et al., 1984; Noji et al., 1992; Skardhamar and Svendsen 2005; Eilertsen and Frantzen, 2007; Eilertsen and Degerlund, 2010; Wiedmann et al., 2016). Steep topography leads to orographic steering of winds, which strongly control mixing and stratification in the fjord (Cottier et al., 2010; Myksvoll et al., 2012), in addition to effects from tides and currents. Persistent down-fjord (to the fjord opening) winds induce upwelling of nutrient-rich coastal water that can stimulate phytoplankton growth in the fjord (Skardhamar and Svendsen, 2005; Cottier et al., 2010). Up-fjord winds push surface waters coastwards and inhibit the upwelling of subsurface water and create a pressure gradient, which results in pulsed outflow of deep water (Skardhamar and Svendsen, 2005; Cottier et al., 2010). Katabatic winds descend from the surrounding mountainous regions into the fjord and cool the surface water and enhancing mixing of the water column (Spall et al., 2017). Wind-induced turbulent mixing resuspends detrital and lithogenic material in shallow areas and influences biogeochemical cycling and export fluxes in the water column (Noji et al., 1993; Keck and Wassmann, 1996; Reistad and Wassmann 1996).

Water column stratification is largely driven by salinity changes from freshwater input and warming/cooling in addition to currents, tides, and winds that drive vertical mixing with saltier subsurface water masses (Inall and Gillibrand, 2010; Cottier et al., 2010). These features influence particle transport, phytoplankton production and biogeochemical cycling (Klink et al., 1981; Svitski et al., 1987). The Norwegian Coastal Current transports fresher waters and supplies the fjords with nutrients and oxygen (Aure and Stigebrandt, 1989) to support enhanced primary production and a diversity of marine life (Erga and Heimdal, 1984; Salvanes and Noreide, 1993). From late autumn to winter, cooling, convective mixing and low light levels reduce phytoplankton activity and net respiration by heterotrophic organisms and organic matter remineralisation enriches the water column with inorganic carbon and nutrients (Noji et al., 1993; Eilertsen and Frantzen, 2007; Eilertsen and Degerlund, 2010). From spring, light availability, stratification, temperature and day length increase with increases in phytoplankton biomass and productivity (Reistad and Wassmann 1996; Eilertsen and Frantzen, 2007). Autotrophic activity utilises the winter stock of nutrients and drives biological carbon uptake during the growing season. Zooplankton abundance typically peaks during the spring bloom, declines throughout the summer as phytoplankton stocks diminish and often slightly increases during the smaller autumn blooms (Gronvik and Hopkins 1984; Michelsen et al. 2017).

High latitude surface waters are particularly sensitive to increases in atmospheric CO$_2$ and are likely to be the first areas to experience widespread ocean acidification; i.e., the lowering of pH and carbonate mineral saturation (Ω) states (Orr et al., 2005; Fabry et al., 2009; Doney et al., 2009). This sensitivity is due to the naturally low seawater carbonate ion concentrations as a result of greater freshwater inputs from melting sea ice, glacial meltwater, precipitation and river runoff (Chierici and Fransson, 2009; Fransson et al., 2013; Fransson et al., 2015). Compared to seawater, freshwater sources are low in total alkalinity ($A_t$), the natural buffer against acidity, and dilute the carbonate ion concentrations in seawater and decrease calcium carbonate (calcite or aragonite, CaCO$_3$) saturation. These processes enhance surface water acidification in coastal and seasonally ice-covered regions (Chierici and Fransson, 2009; Yamamoto-Kawai et al., 2009; Azetsu-Scott et al., 2014; Evans et al., 2014; Reisdorph and Mathis, 2014; Fransson et al., 2015). This enhancement occurs at potentially faster rates than the decreased pH and carbonate concentrations due to anthropogenic CO$_2$ uptake by the ocean; the impact on fjords will depend on the potential for atmospheric CO$_2$ uptake, as driven by biological production, and their geochemical buffering capacity to ocean acidification. Ocean acidification impacts the growth, metabolic processes and life cycles of marine organisms, especially those that precipitate CaCO$_3$ to form shells and skeletons (Orr et al., 2005; Fabry et al., 2008). When the calcium carbonate saturation decreases below the equilibrium threshold ($\Omega = 1$) for carbonate precipitation and dissolution in seawater, the potential for CaCO$_3$ to dissolve increases. The rate of acidification in the open ocean of the Norwegian Sea has been well documented with long-term decreases in pH and aragonite saturation state (Laufset et al., 2015; Jones et al., 2019). Few studies have been carried out to investigate the seasonal biogeochemical cycling and CO$_2$ uptake in northern Norwegian fjords, in contrast to the numerous studies from other northern fjord systems such as Svalbard (Omar et al., 2005; Fransson et al., 2014; Fransson et al., 2015; Ericson et al., 2019a; 2019b) and Greenland (Rysgaard et al., 2012; Meire et al., 2015).

This study presents the first time-series measurements of carbonate chemistry (dissolved inorganic carbon, CO$_2$ and total alkalinity, $A_t$), macronutrients (nitrate + nitrite, phosphate, silicate) and δ$^{18}$O covering a full annual cycle in a sub-Arctic fjord, Kalfjorden. The data provide baseline hydrographic and biogeochemical measurements in the full water column during spring, summer, autumn and winter and emphasise the importance of time-series sampling to unravel the processes controlling the seasonal variability of carbon cycling in this region of the coastal ocean. The main objectives of this work were to...
(1) investigate the seasonal cycling in carbonate chemistry in the context of physical (freshwater inputs and water mass mixing) and biogeochemical (photosynthesis/respiration, remineralisation, calcification, air–sea fluxes) forcing; (2) estimate net community production (NCP) and annual air–sea CO_2 exchange; and (3) determine the current ocean acidification state.

**Methods**

**Study area**

Kaldfjorden (69.75°N, 18.68°E) is an ice-free fjord, 15 km long and about 2 km wide, on the western part of Kvaløya, Troms county, northern Norway (Figure 1). The fjord has a north–south orientation with a typical structure of U-shaped valley bounded by steep, glacially carved sides and is connected to the north Atlantic Ocean across the Norwegian shelf. Kaldfjorden has a partial sill between 75 and 135 m at the mouth and two basins (150–220 m) separated by a ridge. The seafloor shallows towards the inner part of the fjord. The Kaldfjorden marine environment hosts pelagic and benthic calcifiers, plays a role in seasonal migration of herring and supports aquaculture production of Atlantic salmon (Register of Aquaculture Permissions, 2018). The region experiences the polar night from the end of November to the end of January and 24-hour daylight from the end of May until the end of July. The growing season in northern Norway is typically between the end of March and October/November, with peaks in phytoplankton biomass during the spring bloom in mid/late April and a smaller autumn bloom can occur by the end of August or early September (Eilertsen and Frantzen, 2007).

The hydrography of Kaldfjorden is influenced by the circulation of water masses of coastal and Atlantic origin, mixed with local freshwater. The North Atlantic Current carries warm and saline Atlantic Water (S > 35; 5 < T ≤ 10°C) northwards along the Norwegian continental slope (Skarðhamar and Svendsen, 2005). The Norwegian Coastal Current carries colder and less saline Norwegian Coastal Water (S < 35; 4 < T ≤ 12°C) along the continental shelf (Saetre, 2007), which is freshened by riverine inputs (Nordby et al., 1999; Skarðhamar and Svendsen, 2005; Albretsen et al., 2012). Winter Mode Water is formed from cooling and convective mixing of the local fjord water during wintertime. Several small streams transport freshwater (terrestrial snow and ice melt, precipitation) into the fjord.

**Meteorological observations**

Time-series measurements of precipitation, air temperature and wind speeds were recorded at an hourly resolution by the Norwegian Meteorological Institute (www.eKlima.met.no) at the Tromsø observation site (Figure 2). Wind speeds recorded at Tromsø are a better proxy for conditions in Kaldfjorden as the orientation is very similar for both sites and orographic effects will therefore be similar; wind data at other proximal sites at Maasvik and Hekkingen fyr showed more dramatic orographic effects. Data are freely available and were retrieved on 13 September 2018.
Hydrographic measurements

Hydrographic measurements and water samples were obtained along three transects across the outer (T1), middle (T2) and inner (T3) parts of Kaldfjorden from small motorboats or larger research vessels (Table 1). Vertical conductivity-temperature-depth (CTD) profiles were obtained east–west along each of the transects at near-monthly resolution from November 2016 to July 2018. On small boats, a handheld CTD (SAIV SD208) was used. On research vessels, onboard Seabird Electronics SBE911+ were used. Onboard RV Johan Ruud and on the small boats, water sampling was carried out using a single Niskin bottle mounted above the CTD. The bottle was lowered several times per station to each sampling depth. On the larger research vessels, a CTD rosette with 12 Niskin bottles attached was available, and all samples were taken on the upcast of the CTD profile. Table 1 summarizes details of the sampling, vessels and CTD sensors that were used during this study. The SAIV CTD #1321 used from September 2017 until September 2018 was calibrated in summer 2017. When possible, the SAIV CTD was deployed simultaneously with the SBE CTD frame for calibration and intercomparison of the sensors. Castaway CTD sensors were used as further indicators for potential differences between the SAIV CTDs. All Seabird CTD sensors are factory-calibrated annually and the conductivity cells are further calibrated against salinity samples throughout the year. All CTD data were averaged into 1-dbar pressure bins, and the upcasts of the CTD were used for analyses due to more stable vertical speed during recovery. Salinity data are reported on the Practical Salinity Scale.

Freshwater fractions (FW) were determined using the in situ salinity ($S$) relative to the mean salinity ($S_{\text{ref}}$) of Norwegian Coastal Water (Equation 1), as measured below 200 m at the outer fjord (T1) during the study period:

$$FW = \frac{S_{\text{ref}} - S}{S_{\text{ref}}}$$

Water sampling and analysis

Biogeochemical water samples were taken from the central hydrographic station on each transect (T1, T2, T3) from 3–6 depths in the water column, typically at the surface (0–5 m), 25 m, 50 m, 70 m, 100 m, 150 m, and bottom, depending on the water column depth, which was 230 m (T1), 175 m (T2) and 115 m (T3). Samples for carbonate
Table 1: Sampling event date, CTD type, water sampling details and vessels used. DOI: https://doi.org/10.1525/elementa.438.11

| Date (dd.mm.yyyy) | Vessel | CTD | Bottle/rosette | Water samples taken |
|-------------------|--------|-----|----------------|---------------------|
| 30.09 and 10.10.2016 | M/V Chinga | SAIV SD208 #1192 | bottle | C/A | Nutrients | δ¹⁸O |
| 11 and 14.11.2016 | M/V Chinga | SAIV SD204 #866 | bottle | x | x | x |
| 30.11 and 08.12.2016 | M/V Chinga | SAIV SD208 #1141 | bottle | x | x | x |
| 02.04.2017 | R/V Helmer Hanssen | SBE911plus | rosette | – | – | – |
| 05 and 08.09.2017 | M/V Chinga | SAIV SD208 #1321 | bottle | x | x | x |
| 05 and 06.10.2017 | M/V Chinga | SAIV SD208 #1321 | bottle | x | x | x |
| 31.10.2017 | R/V Helmer Hanssen | SBE911plus | rosette | x | x | x |
| 30.11–02.12.2017 | R/V Johan Hjort | SBE911plus | rosette | x | x | x |
| 22–25.01.2018 | M/V Dytiscus | SAIV SD208 #1321 | bottle | x | x | x |
| 19.02.2018 | M/V Dytiscus | SAIV SD208 #1321 | bottle | – | – | – |
| 13 and 14.03.2018 | R/V Johan Ruud | SAIV SD208 #1321 | bottle | x | x | x |
| 04–06.04.2018 | R/V Helmer Hanssen | SAIV SD208 #1321 | rosette | x | x | x |
| 22 and 23.05.2018 | M/V Dytiscus | SAIV SD208 #1321 | bottle | x | x | x |
| 08.06.2018 | M/V Dytiscus | SAIV SD208 #1321 | bottle | x | x | x |
| 19 and 25.07.2018 | R/V Kronprins Haakon, F/V Kjell Arne | SAIV SD208 #1321 | bottle/rosette | x | x | x |
| 06.09.2018 | S/V Verona | SAIV SD208 #1321 | bottle | x | x | x |

*aWater samples not taken.

chemistry were drawn from the Niskin bottle via a silicon tube into 250-mL borosilicate bottles, preserved with saturated mercuric chloride (60 µL) and stored in the dark at 4°C. Analyses for total dissolved inorganic carbon (C\textsubscript{T}) and total alkalinity (A\textsubscript{T}) were carried out at the Institute of Marine Research, Bergen, Norway, within 6 months. Following methods outlined in Dickson et al. (2007), C\textsubscript{T} was determined using gas extraction of acidified (8.5% H\textsubscript{2}PO\textsubscript{4}) samples followed by coulometric titration and photometric detection (Johnson et al., 1987) using a Versatile Instrument for the Determination of Titration carbonate (VINDTA 3D, Marianda, Germany). The determination of A\textsubscript{T} was carried out by potentiometric titration with 0.1 M hydrochloric acid in a semi-open cell using a Versatile Instrument for the Determination of Titration Alkalinity (VINDTA 3S, Marianda, Germany). The average standard deviation for C\textsubscript{T} and A\textsubscript{T} determined from replicate sample analyses, was within ±2 µmol kg\textsuperscript{-1}. Measurements were calibrated against Certified Reference Materials (CRM, provided by A. G. Dickson, Scripps Institution of Oceanography, USA).

Samples for macronutrients nitrate + nitrite (NO\textsubscript{3} + NO\textsubscript{2}), nitrite (NO\textsubscript{2}), phosphate (PO\textsubscript{4}) and silicic acid (Si[OH\textsubscript{4}]) were collected from the Niskin bottles into 20-mL vials, preserved with chloroform and stored at 4°C. Analysis was carried out at the Institute of Marine Research, Bergen, Norway, using a Flow Solution IV analyzer from O.I. Analytical, United States, following Grasshoff et al. (2009). The analyser was calibrated using reference seawater from Ocean Scientific International Ltd., United Kingdom. The semi-conservative tracer N* ([NO\textsubscript{3} + NO\textsubscript{2} – 16PO\textsubscript{4}]; Gruber and Sarmiento, 1997) was used to identify anomalies of nitrate + nitrite relative to phosphate, compared to global averages. As such, N* indicates periods of nitrogen replenishment or loss, as negative values suggest nitrogen deficits due to denitrification and positive values suggest nitrogen excess due to nitrogen fixation. However, these changes also include signals of advection of different water masses with different nutrient signatures.

Samples for determination of the ratio of the stable oxygen isotope of seawater (δ\textsuperscript{18}O) were transferred into 20-mL vials and stored in the dark at 4°C until analysis using a Thermo Fisher Scientific Delta V Advantage mass spectrometer with Gasbench II. Data were standardised relative to Vienna Standard Mean Ocean Water (VSMOW) for δ\textsuperscript{18}O (%) with a reproducibility of replicate analyses of ±0.04‰.
**Carbomate system determinations**

Calcium carbonate saturation and surface water fCO₂

Calcium carbonate (CaCO₃) saturation state (Ω) for the biomineral aragonite and the surface water fugacity of CO₂ (fCO₂) were determined from Cᵦ and Aᵦ and in situ temperature, salinity, pressure and macronutrient concentrations using the CO₂ system program CO2SYS (Lewis and Wallace, 1998; van Heuven, 2011). The carbonic acid dissociation constants (pKᵦ and pK₅) of Mehrbach et al. (1973) as refit by Dickson and Millero (1987) were selected, as they have shown good agreement between measured and calculated values in Arctic waters (Chen et al., 2015; Wooley et al., 2017) and were selected for similar studies in sub-Arctic/Arctic regions (Chierici et al., 2019; Ericson et al., 2019). Ω is used as an indicator for changes in carbonate chemistry in relation to ocean acidification. When Ω < 1, waters are undersaturated with respect to CaCO₃ and thus minerals are sensitive to dissolution.

Variability in surface water fCO₂ can be partitioned into temperature and biological signals by applying a temperature normalisation to the average surface water temperature (Tₐve; 7.51 ± 3.17°C, n = 49) for all seasons (Takahashi et al., 2002):

\[
fCO₂(T) = fCO₂(T_{ave}) \cdot \exp(0.0423 \cdot (T_{ave} - T_{obs}))
\]

where fCO₂(T) (Equation 2) is normalised fCO₂ and T_{obs} is the temperature corresponding to the surface water fCO₂. The remaining fCO₂ variability is attributed to changes in Cᵦ (assuming constant Aᵦ) due to biological processes. This approach does not account for variations in Cᵦ from other processes including Aᵦ riverine inputs, sediment fluxes and air–sea CO₂ exchange; these variations are incorporated into the biological signal.

Air–sea CO₂ fluxes

The potential for air–sea CO₂ exchange is determined by the difference between CO₂ in the sea and overlying air. Fluxes of CO₂ (Equation 3) were calculated from the quasi-monthly air–sea gradient in fCO₂ (ΔfCO₂), the solubility coefficient of CO₂ (Kᵦ) from Weiss (1974) and the gas transfer coefficient (K) which is a function of wind speed (Wanninkhof, 2014):

\[
CO₂ flux = k \cdot K₀ \cdot \Delta fCO₂
\]

The uncertainty in the parameterization of the gas transfer velocity k was taken as 20% (Wanninkhof, 2014). The ΔfCO₂ is the difference between the calculated fCO₂ in surface seawater and the daily mean air fCO₂ value as determined from atmospheric CO₂ values. Atmospheric CO₂ data were obtained as mixing ratios (xCO₂) at hourly resolution at the Sannaltunturi observation site in Pallas, northern Finland (67.9736°N, 24.1158°E), operated by the Finnish Meteorological Institute (FMI). Data were accessed via the World Data Centre for Greenhouse Gases on 5 October 2019. Daily atmospheric CO₂ concentrations were averaged into monthly means and converted to partial pressures, using the mean monthly air pressures and the seawater vapor pressure as determined from seawater salinity and temperature of the surface waters (Ambrose and Lawson, 1972; Millero and Leung, 1976). The average fCO₂ in dry air was 397 ± 9 ppm (n = 46) during the study period (30 September 2016 to 06 September 2018). Negative values of ΔfCO₂ and CO₂ flux indicate surface water CO₂ undersaturation and uptake of atmospheric CO₂. The uncertainty in the atmospheric CO₂ concentration is estimated as ±6 ppm as the maximum standard deviation of monthly means from daily concentrations, and the CO₂ flux uncertainty is estimated as ±0.1 mmol m⁻² day⁻¹. Wind speed data were averaged into monthly means and corrected to 10 m above sea level (Hartman and Hammond, 1985). The uncertainty in the wind speed data is estimated as ±10 m s⁻¹ as the root mean square error of one standard deviation per monthly mean from hourly observations. Air–sea CO₂ exchange per month was estimated from the daily CO₂ flux calculated for each (quasi-monthly) sampling event multiplied by the number in the respective sampling month.

Marine carbonate system

The temporal evolution of Cᵦ in the upper 50 m was determined from the quasi-monthly changes of Cᵦ (Equation 4) during the full annual cycle in 2017–2018. The 0–50 m depth range was selected to encompass the seasonal mixed layer, and changes were determined by integrating from the surface to 50-m depth. The total change in Cᵦ (ΔCₐₙₑₓₑₑ) is determined from the main processes that influence seawater Cᵦ: salinity changes (ΔCₐₙₑₓₑₑ), mixing with subsurface water (ΔCₐₙₑₓₑₑ), photosynthesis/respiration (ΔCₐₙₑₓₑₑ), air–sea CO₂ exchange (ΔCₐₙₑₓₑₑ) and calcium carbonate formation/dissolution (ΔCₐₙₑₓₑₑ):

\[
\Delta Cₐₙₑₓₑₑ = \Delta Cₐₙₑₓₑₑ + \Delta Cₐₙₑₓₑₑ + \Delta Cₐₙₑₓₑₑ + \Delta Cₐₙₑₓₑₑ + \Delta Cₐₙₑₓₑₑ
\]

where ΔCₐₙₑₓₑₑ is determined from the difference between the total monthly Cᵦ change and the change in salinity-normalised Cᵦ. Salinity normalisation using the traditional method (Equation 5) removes effects of dilution/enrichment, where variable (X) measured at in situ salinity (S) was salinity-normalised (Xₛₑₓₑₑ) to the subsurface Norwegian Coastal Water salinity reference (34.17; as described above) following Fris et al. (2003):

\[
Xₛₑₓₑₑ = \frac{X}{S} \cdot S_{ref}
\]

From the Cᵦ salinity and Cᵦδ¹⁸O relationships, values of Cᵦ at zero salinity were estimated at -696 µmol kg⁻¹ and 651 µmol kg⁻¹, respectively (see section on Seasonality in freshwater and deep-water effects). The negative and positive estimates yield different interpretations of the freshwater endmember and the resultant Cᵦ values, which are dependent upon the salinity normalisation method used (Fris et al., 2003). Therefore, in consideration of the significant difference in Cᵦ endmember estimates, with respect to the sign of the value, and following the normalisation of inorganic nutrient data (with Equation 5), the Cᵦ data were normalised using the traditional method, further
discussed below (see section on Uncertainty assessment). $\Delta C_{T-sal}$ integrates the signal from salinity changes due to freshwater inputs ($C_{t}$ and $A_{t}$) and advection of different water masses. $\Delta C_{mix}$ was estimated from monthly changes in the mixed layer ($C_{t}$) and the subsurface Norwegian Coastal Water ($2137 \pm 13$ $\mu$mol kg$^{-1}$, $n = 15$). Here a deepening of the mixed layer (increased potential density) infers vertical mixing and increased $C_{t}$, adapted from Chierici et al. (2011). $\Delta C_{mix} = 0$ if there is no change or a shallowing of the mixed layer and integrates the signal from potential density increases (increased $C_{t}$ and $A_{t}$) due to vertical mixing between sampling events.

Monthly changes in $C_{t}$ due to photosynthetic fixation of $C_{t}$ and production of organic carbon ($\Delta C_{T-frac}$) were determined by (1) using monthly changes in $C_{t}$-salinity normalised nitrate and the C/N Redfield ratio of 6.6 (Redfield et al., 1963) to estimate $C_{t}$ (uptake ($\Delta C_{T-bio}$) and by (2) residual difference between the total $C_{t}$ change and the sum of all other contributing factors ($\Delta C_{T-sal}$). These estimates yield the net community production (NCP), which describes the net primary production minus heterotrophic respiration. $\Delta C_{T-bio}$ is negative when respiration exceeds photosynthesis, thus the reverse transformation recycles organic carbon back into its inorganic form. $\Delta C_{T-sal}$ is determined from the air–sea CO$_2$ flux estimated at the time of each sampling event (quasi-monthly), multiplied by the number of days per respective sampling month. Changes in the $\Delta$CO$_2$ and wind speed are assumed to be linear (or the net result between positive and negative fluctuations) between each quasi-monthly sampling event. Negative fluxes indicate CO$_2$ undersaturation in surface water and atmospheric CO$_2$ uptake and thus yield a positive $\Delta C_{T-frac}$, i.e., input of $C_{t}$ to the surface water. $\Delta C_{T-CaCO_3}$ accounts for changes in $A_{t}$ that influence $C_{t}$, as outlined below.

Following the approach for $C_{t}$, monthly $A_{t}$ changes were determined (Equation 6) from contributions due to salinity changes ($\Delta A_{t-sal}$), mixing ($\Delta A_{t-mix}$), a minor contribution from photosynthesis/respiration ($\Delta A_{t-bio}$) and calcium carbonate (CaCO$_3$) formation/dissolution ($\Delta A_{t-CaCO_3}$), which likely includes terrestrial/benthic fluxes:

$$\Delta A_{T-total} = \Delta A_{T-sal} + \Delta A_{T-mix} + \Delta A_{T-bio} + \Delta A_{T-CaCO_3}$$

Changes in $A_{t}$ due to salinity variations ($\Delta A_{t-sal}$) were determined using the salinity normalisation method that accounts for a non-zero freshwater endmember (Friis et al., 2003), with $A_{t}$ of 337 $\mu$mol kg$^{-1}$ determined from linear regression analysis with $S = 0$ from the $A_{t}$-salinity relationship (presented in the section on Seasonality in freshwater and deep-water effects). $\Delta A_{t-mix}$ was estimated as described for $\Delta C_{mix}$ using an average $A_{t}$ in the subsurface Norwegian Coastal Water ($2282 \pm 10$ $\mu$mol kg$^{-1}$). Changes in $A_{t}$ associated with the uptake and release of nitrate during photosynthesis/respiration ($\Delta A_{t-bio}$) can be estimated as one unit of NO$_3$ uptake increases $A_{t}$ by one unit, therefore $\Delta A_{T-bio} = -0.15 \Delta A_{T-bio}$ (Brewer and Goldman, 1976). Changes in $A_{t}$ due to CaCO$_3$ formation/dissolution ($\Delta A_{T-CaCO_3}$) are estimated by considering the potential alkalinity ($A_{p}$); the sum of salinity-normalised $A_{t}$ and NO$_3$ (Brewer and Goldman, 1976). Thus, $\Delta A_{T-CaCO_3} = \frac{1}{2}(\Delta A_{T-CaCO_3})$, which accounts for carbonate mineral precipitation and dissolution (Zeebe and Wolf-Gladrow, 2001). The $\Delta A_{T-CaCO_3}$ term is also likely to include any terrestrial and sediment/benthic carbonate fluxes.

Monthly $\Delta C_{t}$ and $\Delta A_{t}$ were used to determine the corresponding changes in $\Omega$ for aragonite ($\Delta \Omega$ aragonite). For each process (salinity changes, mixing, photosynthesis/respiration, air–sea CO$_2$ exchange, calcium carbonate formation/dissolution) the associated $\Delta C_{t}$ and $\Delta A_{t}$ was added to the $C_{t}$ and $A_{t}$ of the previous sampling event with in situ temperature, salinity and macronutrient concentrations in the surface layer. The CO2SYS program was then used to calculate the perturbed $\Omega$ to yield monthly estimates of changes in surface water $\Omega$ from the key contributing processes: $\Delta \Omega_{sal}$, $\Delta \Omega_{mix}$, $\Delta \Omega_{bio}$, $\Omega_{CaCO_3}$ and $\Delta \Omega_{CaCO_3}$.

**Uncertainty assessment**

Uncertainties in the determined effects of the physical and biogeochemical processes on $\Delta C_{t}$, $\Delta A_{t}$ and $\Delta \Omega$ were estimated as follows. Errors associated with monthly $\Delta C_{T-total}$ and $\Delta A_{T-total}$ were estimated to be $\pm 0.2$ mol m$^{-2}$ based on analytical precision of $C_{t}$ and $A_{t}$ ($\pm 2$ $\mu$mol kg$^{-1}$). Uncertainties in $\Delta C_{T-sal}$ and $\Delta A_{T-sal}$ were estimated by consideration of the different normalisation methods that were applied (see section on Marine carbonate system). For the $A_{t}$ endmember of 337 $\mu$mol kg$^{-1}$, the difference between salinity normalised and measured $A_{t}$ ranged between $-23$ $\mu$mol kg$^{-1}$ and $131$ $\mu$mol kg$^{-1}$. For the $1340$ $\mu$mol kg$^{-1}$ endmember, the differences varied from $-11$ $\mu$mol kg$^{-1}$ to $58$ $\mu$mol kg$^{-1}$. Thus, the salinity-derived (lower) endmember value yields a greater correction to the $A_{t}$ values.

For $C_{t}$ and the traditional normalisation method, the difference between $C_{T-sal}$ and $C_{T}$ ranged between $-25$ $\mu$mol kg$^{-1}$ and $143$ $\mu$mol kg$^{-1}$. Using the $651$ $\mu$mol kg$^{-1}$ endmember and the non-zero freshwater endmember normalisation, the differences were from $-17$ $\mu$mol kg$^{-1}$ to $95$ $\mu$mol kg$^{-1}$. The upper bound of the $\Delta A_{T-sal}$ uncertainty can be considered as the maximum difference between the $A_{T-sal}$ values from each endmember, i.e., $131–58$ $\mu$mol kg$^{-1}$, which is $73$ $\mu$mol kg$^{-1}$. Similarly, the uncertainty in the $C_{T-sal}$ is estimated as $48$ $\mu$mol kg$^{-1}$. Therefore, the upper bound of the $\Delta A_{T-sal}$ and $\Delta C_{T-sal}$ uncertainty (maximum difference between $\Delta A_{T-sal}$ and $\Delta C_{T-sal}$ for the two normalisation methods used for each) is $1.75$ mol m$^{-2}$ month$^{-1}$ and $1.14$ mol m$^{-2}$ month$^{-1}$, respectively. Using the traditional normalisation technique likely over-corrects $C_{T-sal}$ as explained in Friis et al. (2003); however, the uncertainty is less compared to that estimated for the $A_{T-sal}$ methods, and therefore the standard normalisation method is considered suitable for the inorganic carbon (and nutrient) data in this study. Uncertainties in $\Delta C_{T-sal}$ and $\Delta A_{T-sal}$ estimated to be $\pm 0.4$ mol m$^{-2}$ based on analytical precision of $C_{t}$ and $A_{t}$ ($\pm 2$ $\mu$mol kg$^{-1}$). Uncertainties in $\Delta C_{T-bio}$ and $\Delta A_{T-bio}$ were estimated as $\pm 0.2$ mol m$^{-2}$ from the analytical precision of NO$_3$ of $3\%$ and the uncertainty in the $C/N$ ratio, which was set to $\pm 1$ $\mu$mol kg$^{-1}$ to account for variations in the ratio from 6.6 (Redfield et al., 1963) to 6.7 (Frigstad et al., 2014). The uncertainties in $\Delta A_{T-CaCO_3}$ and $\Delta C_{T-CaCO_3}$ were estimated as
±0.2 and ±0.1 mol m⁻², respectively, from the analytical precision of NO₃⁻ (±3%) and Aₕ (±2 µmol kg⁻¹).

Uncertainties in the calculated surface water fCO₂ (µCO₂) result from uncertainties in Cₐ, Aₕ, salinity, temperature, K1 and K2 that were added to each value of each property and used as inputs in CO2SYS to yield an upper bound error as ±9 µatm (±9 ppm). Based on a max value for k of 2.3, the associated error (20%; Wanninkhof, 2014) for k is ±0.46, thereby the uncertainty for ΔC₂ flux is estimated as ±0.01 mol m⁻². Uncertainties in ΔC₂ bioC and ΔA₂ bioC are estimated as ±1.1 mol m⁻² from the sum of all associated uncertainties for each contributing term. Following the same approach, errors associated with ΔΩ misc were estimated to be ±0.08 based on uncertainties of ΔΩ ± 0.04 from the input parameters run through CO2SYS. Using the associated errors for each ΔC₂ and ΔA₂ term in CO2SYS, uncertainties for ΔΩ were determined as ΔΩ misc ± 0.12, ΔΩ bio C ± 0.08, ΔΩ flux ± 0.02, ΔΩ CaCO₃ ± 0.09 and ΔC₂ bioC ± 0.43.

Results

Meteorology

Air pressure was stable during the summer and more variable for the rest of the year. Air temperatures were warmest (>20°C) in June, July and August and coldest (<−10°C) from January to March (Figure 2). Average daily precipitation was 2.7 ± 4.7 mm (n = 706), with intense events (10–32 mm day⁻¹) occurring sporadically throughout the year. Wind data measured every 6 hours showed that the average and the most frequent wind direction were 175 ± 90° (n = 2826) and 194°, respectively, which shows the impact of orographic effects at the Tromso site. The prevailing southerly winds in Tromso indicate that up-fjord (from the inner to outer part of the fjord) winds likely prevailed in Kaldfjorden. Wind speeds were on average 3.3 ± 2.1 m s⁻¹ (n = 2826), ranging from calm periods to strong gusts (up to 12.9 m s⁻¹). High pressure systems at the end of September and early October 2017 and February was accompanied by elevated temperatures and low rainfall. Winter storms (in December and January) were characterised by increased wind speeds, warmer air, and precipitation. June 2018 was a notably wet month relatively to the rest of the year. July 2018 had the warmest air temperatures (25–28°C) and very low precipitation as a result of higher and stable air pressure. Surface water temperatures were at the seasonal maximum during this period.

Hydrography

The water column was strongly stratified across Kaldfjorden from June to October (Figure 3). Highest potential temperatures (θ up to 12.27°C) were found in the upper 50 m from June to October 2017 and 2018 (Figure 4). Lower salinity (S ~ 32) water occupied the upper 20 m during this period (Figure 5). Subsequent cooling and convective mixing eroded the stratification in November and the water column was well mixed from December until May. During periods of weak stratification, the effects of tides are likely greater across the region (Skarðhamar and Svendsen, 2005). The fjord water resembled a modified variety of Norwegian Coastal Water (S < 34.8, 4 < θ ≤ 12°C; Nordby et al., 1999) with local effects of cooling, warming and freshening (Figure 6). The pycnocline persisted from May until November with modified Norwegian Coastal Water as the warmer and saltier watermass below 100 m. In the outer part of the fjord (T1), subsurface coastal water

Figure 3: Time series of seawater potential density anomaly from October 2016 to September 2018. Measurements of seawater potential density anomaly (kg m⁻³; color scale bar) from CTD deployments at the central hydrographic station of the outer (T1), middle (T2), and inner (T3) transects. Sampling events are indicated by black triangles. DOI: https://doi.org/10.1525/elementa.438.f3
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(below 200 m depth) had mean $S$ of 34.17 ± 0.24 and $θ$ of 5.35 ± 1.64°C ($n = 16$). Lowest salinity surface water ($S$ of 31.85), thus highest freshwater fraction (Figure 7a,i,q), was found in summer in the inner fjord. Temperature maxima occurred in the surface layer in July 2018. The stable oxygen isotope ($δ^{18}O$) varied between −0.26‰ and 0.76‰ in the water column (Figure 7b,j,r). Lower (isotopically depleted) $δ^{18}O$ is a signal of meteoric water input and was found in the fresher, stratified surface layer from May to November. Higher $δ^{18}O$ (isotopically enriched) values are indicative of coastal water (with $δ^{18}O$ 0.44 ± 0.21‰) in the full water column from December until April.

**Macronutrients**

Water column concentrations of nitrate ($NO_3$), phosphate ($PO_4$) and silicic acid ($Si(OH)_4$) had ranges of 0–10.69 µmol kg$^{-1}$, 0–0.97 µmol kg$^{-1}$ and 0–12.53 µmol kg$^{-1}$, respectively (Figure 7). Lowest concentrations and

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**Figure 4:** Time series of seawater temperature from October 2016 to September 2018. Seawater potential temperature (°C, color scale bar) from CTD deployments at the central hydrographic station of the outer (T1), middle (T2), and inner (T3) transects. Sampling events are indicated by black triangles. DOI: https://doi.org/10.1525/elementa.438.f4

**Figure 5:** Time series of seawater salinity from October 2016 to September 2018. Measurements of seawater salinity (color scale bar) from CTD deployments at the central hydrographic station of the outer (T1), middle (T2), and inner (T3) transects. Sampling events are indicated by black triangles. DOI: https://doi.org/10.1525/elementa.438.f5
Figure 6: Potential temperature-salinity plots from October 2016 to September 2018. Potential temperature (θ, °C) and salinity characteristics, overlying contours of potential density (kg m⁻³) anomaly, for all CTD casts at the central hydrographic station of the outer (T1), middle (T2), and inner (T3) transects per sampling event. DOI: https://doi.org/10.1525/elementa.438.f6
Figure 7: Hydrographic and biogeochemical depth profiles. Depth profiles in the full water column of (a,i,q) freshwater fraction (%), (b,j,r) δ_18O ‰, (c,k,s) nitrate (NO_3, µmol kg^-1), (d,l,t) phosphate (PO_4, µmol kg^-1), (e,m,u) silicic acid (Si(OH)_4, µmol kg^-1), (f,n,v) C_T (µmol kg^-1), (g,o,w) A_T (µmol kg^-1), (h,p,x) aragonite saturation state (Ω) per sampling month (color bar) at the central hydrographic station of the outer (T1), middle (T2), and inner (T3) transects. DOI: https://doi.org/10.1525/elementa.438.f7
episodic depletion of all macronutrients occurred in the surface layer between June and October across the fjord. Concentrations typically increased with depth but were spatially variable; enriched NO$_3$ and Si(OH)$_4$ was found at 100-m depth in the inner fjord in December (Figure 7s,u). The modified Norwegian Coastal Water (below 200-m depth at T1) had average (n = 16) concentrations of NO$_3$, PO$_4$ and Si(OH)$_4$ of 3.79 ± 3.55 µmol kg$^{-1}$, 0.35 ± 0.31 µmol kg$^{-1}$ and 4.49 ± 1.19 µmol kg$^{-1}$, respectively, that re-supplied the upper layers during vertical mixing.

Carbonate chemistry

The distribution of C$_T$ and A$_T$ showed lower values in the upper 50 m from June to November across the fjord. Surface water C$_T$ was lowest (<2050 µmol kg$^{-1}$) in summer (Figure 7f,n,v) and autumn, and A$_T$ was lowest (<2210 µmol kg$^{-1}$) from spring to autumn (Figure 7g,o,w). Reductions in sea surface C$_T$ and A$_T$ occurred from April, reaching lowest values of 1958 µmol kg$^{-1}$ and 2136 µmol kg$^{-1}$, respectively, in June 2018. Concentrations of C$_T$ increased with depth to >2105 µmol kg$^{-1}$ below 100 m. The modified Norwegian Coastal Water (below 200 m depth at T1) was characterised by high A$_T$ and C$_T$ of 2257–2296 µmol kg$^{-1}$ and 2108–2154 µmol kg$^{-1}$, respectively, throughout the study period. Highest C$_T$ of −2170 µmol kg$^{-1}$ was found close to the seafloor in November in the inner fjord. Aragonite saturation (Ω) was highest (2.20–2.33) in the upper 50 m from June to November across the fjord (Figure 7h,p,x). Values decreased with depth to low Ω of 1.34–1.66 below 150 m and lowest near the seafloor in the inner fjord.

Surface water seasonality and air–sea CO$_2$ exchange

Surface waters (0–5 m) were relatively warm (9.95–12.27°C) and fresh (<33.10) with isotopically light δ$^{18}$O (between −0.1‰ and −0.2‰) from July to October (Figure 8a–c). NO$_3$, PO$_4$ and Si(OH)$_4$ were rapidly reduced in April, with NO$_3$ and PO$_4$ nearly totally depleted from
May to October (Figure 8d–f). Lowest C\textsubscript{T, sal} (2059–2065 µmol kg\textsuperscript{-1}), controlled by biological drawdown, and A\textsubscript{T, sal} (2250 µmol kg\textsuperscript{-1}), from a likely calcification signal, occurred in September and October (Figure 9b–c). Depleted Si(OH)\textsubscript{4} also occurred at this time. Surface water fCO\textsubscript{2} was lowest (270–294 µatm) and strongly undersaturated (ΔfCO\textsubscript{2, sea-air} of −131 µatm) in April and May (Figure 9d–e). Greatest CO\textsubscript{2} uptake (2.7 mmol m\textsuperscript{-2} day\textsuperscript{-1}) occurred when ΔfCO\textsubscript{2, sea-air} was large (−129 µatm) and monthly wind speeds exceeded the yearly average (3.2 m s\textsuperscript{-1}) in May. Ω increased to highest saturation states (2.26–2.33) in September (Figure 9f). Sea surface CO\textsubscript{2} is influenced by temperature (+1°C raises fCO\textsubscript{2} by about 10 µatm; Takahashi et al., 1993). For a seasonal surface water increase of 10.1°C, increases in fCO\textsubscript{2} up to 100 µatm could be expected. Surface water fCO\textsubscript{2} was found to increase by ~60 µatm, thus the thermodynamic effects were compensated by photosynthetic CO\textsubscript{2} uptake. When the effects of temperature were removed, variations in fCO\textsubscript{2} showed similar changes compared with fCO\textsubscript{2} (Figure 9d) but with a larger seasonal amplitude, suggesting that biological processes dominated over thermodynamic control of fCO\textsubscript{2} to drive strong seasonality in surface water fCO\textsubscript{2} across the fjord (Figure 10).

The lowest salinity and highest freshwater fractions (5–6%) in June 2018 were accompanied by increases in Si(OH)\textsubscript{4} (0.8 µmol kg\textsuperscript{-1}) and C\textsubscript{T, sal} (~25 µmol kg\textsuperscript{-1}), indicating terrestrial sources of (remineralised) organic matter and weathered minerals entered the fjord during intense precipitation and runoff events. Cold (2.17–3.76°C) and saline (33.58–33.87) surface waters in March and April had enriched δ\textsuperscript{18}O due to vertical mixing with subsurface waters.
coastal water. Respiration, remineralisation and vertical mixing of the water column increased $C_{T\text{sal}}$, $NO_3\text{sal}$, $PO_4\text{sal}$ and $Si(OH)_4\text{sal}$ with a potential contributions from silicate dissolution, from January to March. The maximum absolute difference between measured and salinity-normalised concentrations for $NO_3$, $PO_4$ and $Si(OH)_4$ was 0.13 $\mu$mol kg$^{-1}$, 0.01 $\mu$mol kg$^{-1}$ and 0.15 $\mu$mol kg$^{-1}$, respectively, and relatively minor compared to 143 $\mu$mol kg$^{-1}$ for $C_T$ and 131 $\mu$mol kg$^{-1}$ for $A_T$. Increases in $A_T\text{sal}$ by $\sim$40 $\mu$mol kg$^{-1}$ occurred from early autumn to November. Surface water $\Omega$ was less variable during winter/early spring and lowest values (1.59) in March. $fCO_2$ was highest (381–388 $\mu$atm) between September and February, and $fCO_2\text{ T}$ peaked in January–March, which coincided with the period of convective mixing and colder temperatures. Surface waters were undersaturated with respect to atmospheric CO$_2$ with an average $\Delta fCO_2\text{sea-air}$ of $-58 \pm 33$ $\mu$atm ($n = 46$), driving air–sea CO$_2$ exchange of $-0.86 \pm 0.63$ mmol m$^{-2}$ day$^{-1}$ ($n = 46$; atmospheric CO$_2$ uptake). On an annual basis, the surface layer in Kalfjorden is estimated to be a sink for atmospheric CO$_2$ of 0.32 $\pm$ 0.03 mol C m$^{-2}$ yr$^{-1}$ ($n = 12$).

**Figure 10: Seasonality in fCO$_2$ driven by biological processes.** Surface water temperature-normalised $fCO_2$ ($fCO_2\text{T, $\mu$atm}$) as a function of potential temperature ($\theta$, °C) per season (by color) with key biological processes marked. The $fCO_2$ in air ($fCO_2\text{air, $\mu$atm}$) average (solid black line) and standard deviation (dashed lines) during the 2017–2018 annual cycle is indicated. The $fCO_2$ data below the $fCO_2\text{air}$ lines indicate undersaturation with respect to the atmosphere and increased potential for CO$_2$ uptake. DOI: https://doi.org/10.1525/elementa.438.f10

**Seasonal mixed layer carbonate chemistry dynamics**

The largest total monthly change in $C_T$ ($\Delta C_T\text{total}$) in the upper 50 m was $-4.1$ mol m$^{-2}$ month$^{-1}$, in the inner fjord, driven by changes in salinity $\Delta C_T\text{sal}$ of $-3.3$ mol m$^{-2}$ month$^{-1}$ (accounting for 81% of $\Delta C_T\text{total}$) due to freshwater inputs in June (Figure 11). $\Delta C_T\text{mix}$ showed little variation in its range of 0–0.1 mol m$^{-2}$ month$^{-1}$, constituting up to 5% of $\Delta C_T\text{total}$. The upper percentages of $\Delta C_T\text{mix}$ contributions coincided with the timing of erosion of water column stratification and increased $C_T$ in the upper layers from mixing with subsurface Norwegian Coastal Water, notably from December to April. $\Delta C_T\text{bio}$ ($\Delta C_T\text{bio C}$; determined from the residual difference between $\Delta C_T\text{total}$ and all other factors) ranged between $-0.4$ mol m$^{-2}$ month$^{-1}$ and $-0.8$ mol m$^{-2}$ month$^{-1}$ during the spring bloom and summer productive period from April to June. Biological production accounted for 20–30% of $\Delta C_T\text{total}$ and coincided with maximum $\Delta C_T\text{bio}$ of $-1.2$ mol m$^{-1}$ month$^{-1}$ from the rapid $NO_3$ drawdown in April. A second peak in biological $C_T$ drawdown occurred in August with $\Delta C_T\text{bio}$ up to $-0.5$ mol m$^{-2}$ month$^{-1}$ as a signal of late summer production and a possible autumn bloom. Greatest increases in $\Delta C_T\text{total}$ of
1.4–1.9 mol m\(^{-2}\) month\(^{-1}\) were driven by \(\Delta C_{\text{T \, sal}}\) of 1.1 mol m\(^{-2}\) month\(^{-1}\) (57% of \(\Delta C_{\text{T \, total}}\)) and \(\Delta C_{\text{T \, bio \, C}}\) of 0.4–0.7 mol m\(^{-2}\) month\(^{-1}\) (37% of \(\Delta C_{\text{T \, total}}\)) from September to October. These increases were a result of reduction in freshwater fluxes and increased surface layer \(C_t\) from respiration and remineralisation (as indicated by higher nutrient concentrations) in post-bloom conditions. \(\Delta C_{\text{T \, CaCO}_3}\) was lowest at −0.3 mol m\(^{-2}\) month\(^{-1}\), which accounted for up to 8% of the reduction in \(\Delta C_{\text{T \, total}}\) due to the presence of calcifying phytoplankton. Highest \(\Delta C_{\text{T \, CaCO}_3}\) of 0.3 mol m\(^{-2}\) month\(^{-1}\) represented up to 15% of \(\Delta C_{\text{T \, total}}\) at the time of intense freshwater runoff, indicating terrestrial inputs of \(A_t\), and a decaying coccolithophore bloom in September. Excess \(A_t\) (positive \(\Delta C_{\text{T \, CaCO}_3}\)) could also result from particulate inorganic carbon, e.g., \(\text{CaCO}_3\) shells in the water column that are captured by sampling and dissolve upon analysis. \(\Delta C_{\text{T \, flux}}\) range was 0–0.1 mol m\(^{-2}\) month\(^{-1}\) from the increased \(C_t\) in the upper layers due to uptake of atmospheric \(\text{CO}_2\), which was highest in May and June (up to 4% of \(\Delta C_{\text{T \, total}}\)). Monthly change in \(\Omega\) for aragonite (\(\Delta \Omega_{\text{total}}\)) ranged between −0.43 and 0.37 across Kaldfjorden (Figure 12). Greatest increases in \(\Delta \Omega\) were driven by biological production (decreased \(C_t\), slight increase in \(A_t\)) as \(\Delta \Omega_{\text{bio \, C}}\) varied between 0.25 and 0.40 from April to June. Biological production dominated the other processes and accounted for up to 99% of the monthly change in \(\Delta \Omega\). Increases in \(\Delta \Omega_{\text{bio \, C}}\) up to 0.20 in August and September coincide with biological \(C_t\) drawdown in a late summer/autumn bloom. Respiration (increased \(C_t\), slight decrease in \(A_t\)) in post bloom and winter conditions dominated the monthly decreases in \(\Delta \Omega_{\text{total}}\) with \(\Delta \Omega_{\text{bio \, C}}\) of −0.50. Lowest (negative) \(\Delta \Omega_{\text{CaCO}_3}\) corresponded to lowest (negative) \(\Delta \Omega_{\text{CaCO}_3}\) of −0.09, and 21% of \(\Delta \Omega_{\text{total}}\) as a result of calcification (greater decreases in \(A_t\) relative to \(C_t\)) in spring and summer. Lowest (negative) monthly

Figure 11: Temporal evolution of \(C_t\) in the upper layer during the 2017–2018 annual cycle. Monthly changes in total dissolved inorganic carbon (\(\Delta C_{\text{T \, total}}\), mol m\(^{-2}\) month\(^{-1}\)) in the upper layer (0–50 m) of Kaldfjorden are determined from contributions from salinity changes (\(\Delta C_{\text{T \, sal}}\)), mixing (\(\Delta C_{\text{T \, mix}}\)), photosynthesis/respiration (\(\Delta C_{\text{T \, bio}}\)), calcium carbonate formation/dissolution (\(\Delta C_{\text{T \, CaCO}_3}\)) and air–sea \(\text{CO}_2\) exchange (\(\Delta C_{\text{T \, flux}}\)) per sampling month during the 2017–2018 annual cycle at the central hydrographic station of the outer (T1), middle (T2), and inner (T3) transects. \(\Delta C_{\text{T \, bio}}\) was estimated by (1) using monthly changes in salinity-normalised nitrate and the C/N Redfield ratio of 6.6 (Redfield et al., 1963) to estimate \(C_t\) uptake (\(\Delta C_{\text{T \, bio \, N}}\)) and by (2) residual difference between the total \(C_t\) change and the sum of all other contributing factors (\(\Delta C_{\text{T \, bio \, C}}\)). DOI: https://doi.org/10.1525/elementa.438.f11
changes in $\Delta C_{\text{sal}}$ corresponded to highest (positive) $\Delta \Omega_{\text{sal}}$ up to 0.11 to reveal the net effect of $C_T$ dilution offset the effect of decreases in $A_T$ from freshwater inputs. Minor changes in $\Delta \Omega$ resulted from mixing with subsurface Norwegian Coastal Water, which was largest in winter and early spring with $\Delta \Omega_{\text{mix}}$ of −0.04 (representing up to 10% of $\Delta \Omega_{\text{total}}$ in December). $\Delta \Omega_{\text{flux}}$ ranged between −0.03 in December and January and 0.05 in August from changes in monthly $CO_2$ uptake and subsequent increased $C_T$ in the surface layer.

**Discussion**

**Seasonality in freshwater and deep-water effects**

Higher freshwater fractions (4–7%) and a shift towards depleted $\delta^{18}O$ (−0.26‰) from June to September coincided with warmer air temperatures and higher precipitation. The effects of increased precipitation and river runoff dominated the greatest monthly change in $C_T$ of −4.1 mol C m$^{-2}$, representing reductions of $C_T$ from dilution effects that accounted for 81% of monthly $C_T$ deficits in the inner part of the fjord in June. Summertime values of $C_T$ and $A_T$ in the surface layer of Kaldfjorden were similar to those reported for Svalbard fjords (Fransson et al., 2015; Ericson et al., 2019b) but generally higher than those of other high-latitude coastal and fjord systems (Table 2). The differences are largely due to greater meteoric water inputs in the other high-latitude regions compared to Kaldfjorden, which include glacial meltwater and contributions from sea-ice melt that result in higher dilution effects that lower $A_T$ (and $C_T$). In combination with strong biological carbon uptake, the fjord and coastal regions at higher latitudes and inland seas have nominally lower surface water $A_T$ and $C_T$ compared to Kaldfjorden.

The importance of freshwater in controlling the carbonate system and nutrient dynamics in Kaldfjorden is evident from the relationship between salinity and $A_T$ and $C_T$ during a full annual cycle (Figure 13a). The strong correlation between $A_T$ and salinity yielded $A_T = 56.95 + 337$ ($r^2 = 0.84$, $s_e = 55$ µmol kg$^{-1}$, $p < 0.0001, n = 234$), which shows evidence of a freshwater (zero salinity) end-member for $A_T$ of 337 ± 55 µmol kg$^{-1}$. For $C_T$ and salinity, the relationship yielded $C_T = 82.95 - 696$ ($r^2 = 0.76$;
Table 2: Summer $C_T$ and $A_T$ in surface waters of northern high-latitude coastal and fjord systems and inland seas. DOI: https://doi.org/10.1525/elementa.438.t2

| Region                        | $C_T$ range ($\mu$mol kg$^{-1}$) | $A_T$ range ($\mu$mol kg$^{-1}$) | Reference                  |
|-------------------------------|---------------------------------|---------------------------------|-----------------------------|
| Baltic Sea                    | 1200–2100                       | -                               | Thomas and Schneider (1999) |
| Glacier Bay, Alaska           | 1273–2044                       | 1412–2137                       | Reisdorph and Mathis (2014) |
| Puget Sound, Pacific Northwest, USA | 1431–2038                     | 1510–2101                       | Pelletier et al. (2018)     |
| Cumberland Sound, Canadian Arctic | 1779–1978                      | 1922–2147                       | Turk et al. (2016)          |
| Tempelfjorden, Svalbard       | 1960–2080                       | 2130–2260                       | Fransson et al. (2015)      |
| Adventfjorden, Svalbard       | 2050–2060                       | 2060–2150                       | Ericson et al. (2019b)      |
| Kaldsfjorden, northern Norway | 1958–2129                       | 2136–2273                       | this study                  |

*a* Not available.

Figure 13: Key carbonate chemistry, nutrients and salinity seasonal cycles and relationships. Trends in water column (a) $C_T$ ($\mu$mol kg$^{-1}$), $A_T$ ($\mu$mol kg$^{-1}$) and salinity, (b) silicic acid ($\text{Si(OH)}_4$, $\mu$mol kg$^{-1}$) and salinity, (c) $C_T$($\mu$mol kg$^{-1}$) and $A_T$($\mu$mol kg$^{-1}$), (d) $C_T$ ($\mu$mol kg$^{-1}$) and nitrate (NO$_3$, $\mu$mol kg$^{-1}$), (e) nitrate (NO$_3$, $\mu$mol kg$^{-1}$) and phosphate (PO$_4$, $\mu$mol kg$^{-1}$), (f) silicic acid ($\text{Si(OH)}_4$, $\mu$mol kg$^{-1}$) and nitrate (NO$_3$, $\mu$mol kg$^{-1}$) per sampling month (color bar) at the central hydrographic station of the outer (T1), middle (T2), and inner (T3) transects. Carbon and nutrient uptake/regeneration ratios (C/N, N/P, Si/N) determined from the linear regression trend of all seasonal data are shown in their respective plots. DOI: https://doi.org/10.1525/elementa.438.f13

$se = 102 \ \mu$mol kg$^{-1}$; $p < 0.0001$; $n = 235$), which indicates a deficit in $C_T$ in freshwater, as reported for freshwater $A_T$ endmember estimates from salinity relationships by Turk et al. (2016). A second approach in estimating the freshwater $A_T$ endmember is using the relationship with $\delta^{18}O$, as a comparison. Firstly, the meteoric $\delta^{18}O$ signature of freshwater was estimated as $-10.1‰$ from the relationship between $\delta^{18}O$ and salinity ($r^2 = 0.43$, $se = 0.79‰$, $p < 0.0001$, $n = 226$). As no direct measurements of freshwater endmembers are currently available for Kaldsfjorden,
the δ¹⁸O (−10.1‰) value determined here indicates that the freshwater sources are predominantly of meteoric origin (snow melt, precipitation and river runoff). The estimated δ¹⁸O value in Kaldfjorden is similar to that of the Hudson Bay rivers (latitude −60°N) and falls at the higher end (isotopically heavier) of previously reported ranges of δ¹⁸O in meteoric water endmembers in other high latitude fjord and coastal systems (Table 3). The spatial variability of the freshwater δ¹⁸O signature is due to the fact that precipitation becomes increasingly light isotopically at higher latitudes, in addition to localised variations such as influences of glaciers in sub-Arctic regions, as suggested by Turk et al. (2016).

When applied to the A₁-δ¹⁸O linear regression $A_1 = 88.78^{18}O + 2236$ ($r^2 = 0.44, \text{se} = 6.7 \, \text{mol kg}^{-1} \, \%_{\text{oo}}$; $p < 0.0001, n = 223$), the $A_1$ endmember is estimated at $1340 \pm 9 \, \text{mol kg}^{-1}$. The range of the estimated freshwater $A_1$ endmember (337–1340 μmol kg⁻¹) for Kaldfjorden falls within the range of Arctic rivers (Cooper et al., 2008) and for the Svalbard fjords (Fransson et al., 2015; Ericson et al., 2019a, 2019b), which are influenced by glacial runoff with a watershed containing carbonate and silicate bedrock. Values were higher than those for meteoric endmembers dominated by glacial meltwater in Greenland (Meire et al., 2015) and for Cumberland Sound (Turk et al., 2016). The associated freshwater source of $A_1$ likely originates from terrestrial inputs, e.g., weathered minerals of surrounding rock. The regression analysis showed that freshwater in Kaldfjorden had a diluting impact on surface water $A_2$. The freshwater-derived $A_2$ would act to slightly decrease dilution effects; however, the overall effect from freshwater inputs is a lowering of $A_3$ and $C_3$ in the surface layer. Variations in the $A_3$ content of meteoric water have been previously attributed to river drainage over carbonate and silicate-rich rocks that subsequently become enriched with minerals and transported into the fjord and coastal waters (Hjalmarsson et al., 2008; Azetsu-Scott et al., 2014; Fransson et al., 2015; Ericson et al., 2019a).

Similarly to $A_2$, $C_2$ decreased with increasing freshwater inputs in the upper layer of the fjord. From linear regression analysis with the local δ¹⁸O endmember (−10.1‰), $A_2 = 140.18^{18}O + 2066$ ($r^2 = 0.47, \text{se} = 9.9 \, \text{μmol kg}^{-1} \, \%_{\text{oo}}$; $p < 0.0001, n = 224$), the freshwater $C_2$ endmember was estimated as $651 \pm 13 \, \text{μmol kg}^{-1}$. In contrast to the endmember estimates for $A_2$, the $C_2$ estimates are similar in magnitude but opposite in sign. These findings suggest that the freshwater salinity and δ¹⁸O system could be influenced by contrasting processes of inorganic carbon removal and enrichment, respectively, depending upon which estimation method was selected. The δ¹⁸O-based endmember is similar to the estimated $C_1$ endmember in Tempelfjorden, Svalbard (Ericson et al., 2019a). This similarity leads to the hypothesis that freshwater runoff could contain a source of $C_1$ derived from atmospheric CO₂ uptake and terrestrial organic matter, which was subsequently re-mineralised upon transport to and release into the fjord. Increased surface water Si(OH)₄ and $C_{B_{\text{sal}}}$ of −25 μmol kg⁻¹ that was linked to persistent precipitation, most notably at the inner part of the fjord, indicated a supply of dissolved silica, perhaps from weathered silicate minerals, within freshwater runoff (Figure 13b). The enhanced surface water $C_{B_{\text{sal}}}$ associated with freshwater fluxes enriched the $C_1$ pool of the fjord and led to reductions in Ω. However, any additional $C_1$ was likely assimilated during biological production during the summer and thus constitutes a more transient signal following prolonged precipitation as observed in June 2018. Enhanced $A_3$ and silicates in freshwater delivered to Kaldfjorden could

Table 3: Literature values, ranges or mean ± standard deviation (n value) of measured and estimated δ¹⁸O, C₂ and A₁ in meteoric water endmembers in northern high-latitude fjord and coastal systems. DOI: https://doi.org/10.1525/elementa.438.13

| Region                  | δ¹⁸O (%) | $C_2$ (μmol kg⁻¹) | $A_1$ (μmol kg⁻¹) | Reference                  |
|-------------------------|----------|-------------------|-------------------|----------------------------|
| Godthåbsfjord, west Greenland | na²     | 80 ± 17           | 50 ± 20           | Meire et al. (2015)         |
| Siberian rivers; North American Arctic rivers | −20.5, −14.9 | na                | 800–1900          | Cooper et al. (2008)        |
| Cumberland Sound, Canadian Arctic | −19.2 ± 0.8 (40) | 247               | 174               | Turk et al. (2016)          |
| Hudson Bay rivers, Canada | −16.8, −10.8 | na                | 226–1870          | Burt et al. (2016)          |
| Adventfjorden, Svalbard | na       | 339 ± 7           | 294 ± 3; 418      | Ericson et al. (2018, 2019b) |
| Tempelfjorden, Svalbard | −16, −4.9 | 508 ± 52 (n = 36), 661 ± 45 (n = 27) | 526–1142; 355 ± 24 (n = 36), 601 ± 42 (n = 27) | Fransson et al. (2015); Ericson et al. (2019a) |
| Kongsfjorden, Svalbard | −15.9    | na                | 890               | Maclachlan et al. (2007); Fransson et al. (2016) |
| Kaldfjorden, northern Norway | −10.1 ± 0.8 (n = 226) | 651 ± 13 (n = 224) | 337 ± 55 (n = 234), 1340 ± 9 (n = 223) | this study |

²Not available in the cited reference.

N value not stated in the cited reference.
constitute a minor buffer in the surface layer against CO₂ increases and provide an additional source of silicate to silicous plankton species, e.g., diatoms. Future warming and increased precipitation and runoff in the sub-Arctic will result in higher freshwater inputs but, in regions of calcareous and silicous bedrock, a terrestrial supply of dissolved minerals could act to slightly counteract dilution effects on seawater A₂.

The seasonality in A₄ and C₄ in the modified Norwegian Coastal Water below 200 m at the mouth of the fjord varied from lower values in winter and higher values in summer and autumn as evidence of convective mixing of the surface layer and subsurface coastal water when stratification becomes eroded (Figure 7f–g). Deep vertical mixing in winter and spring, as indicated by a shift in δ¹⁸O from higher (-0.6‰) to lower (-0.1‰) isotopic values, homogenised the water column and enabled the coastal water source of macronutrients to be entrained into the upper layers and the freshwater-influenced surface to permeate deeper (Figure 7a–b). As such, the A₄ and C₄ signal from the productive surface layers was dispersed into the water column and lowered A₄ and C₄. Following spring and summer biological production, enriched C₄ and macronutrients were found at 150–250 m depth, with highest concentrations in November and December. This is due to enrichment from organic matter sinking out of productive surface waters and being remineralised in the subsurface and deep water. Seasonal variability in deep water A₄ and Si(OH)₄ closely followed variations in salinity and δ¹⁸O. Higher A₄ in November and December likely includes a contribution from sediment resuspension, superimposed onto the relatively elevated A₄ signal from Norwegian Coastal Water (with δ¹⁸O of -0.8). Increased concentrations of Si(OH)₄ arise from remineralisation of silicates, e.g., from diatoms and sediment resuspension, during wind-induced mixing.

The impact of these competing processes that increase and decrease C₄ and A₄, largely removed net seasonality from Ω variations below 150 m, yielding a narrow range of Ω (1.5–1.7) throughout the year. A notable increase in concentrations of all macronutrients and C₄ was found close to the sea floor at the inner part of the fjord in December 2017, which resulted in the lowest Ω of 1.3 during the time series. This low Ω could result from remineralisation, sediment fluxes or perhaps a manmade signal, for example water discharge in the shallower part of the fjord closest to land. This signal appeared to be localised and showed impacts to carbonate chemistry in areas exposed to greater land-water interactions, i.e., higher C₄ inputs reduced Ω. Removing the effects of salinity changes on C₄ and A₄ (Figure 13c) reveals further seasonal dynamics driven by biological production, calcification and atmospheric CO₂ uptake (discussed in the following section).

Seasonal biological processes and NCP

The importance of biological processes on the carbonate chemistry in Kaldfjorden is reflected in the seasonal NCP and substantial C₄ drawdown during spring and summer (Figure 11). With greater light availability, phytoplankton biomass increased in early spring and developed into a bloom of diatoms and Phaeocystis with chlorophyll a concentrations up to 10 mg m⁻² in April (Persson, 2018). Weak stratification and nutrient replenishment from subsurface waters and recycling in the upper water column fuelled biological production. Peaks in particulate matter fluxes and chlorophyll a fluxes, out of the surface layer during the spring bloom, have been observed in a neighbouring fjord, Balsfjorden in Tromsø (Eilertsen et al., 1981; Eilertsen and Degerlund, 2010). Coupled to strong biological carbon uptake and conversion to organic carbon, subsequent export of organic matter would contribute to reductions in water column C₄. Some organic matter may settle out of the water column and become buried in the sediments, driving seasonal A₄ depletion in the upper layer of the fjord during April. The abundance of zooplankton, e.g., copepods, was low, and export of particulate organic carbon likely continued prior to the development of strong stratification by late spring that would impede export of particulate matter to depth (Wassmann et al., 1991; Walker, 2018). By June and July, stronger stratification likely inhibited export of organic matter across the pycnocline, and thus respiration and recycling in the upper layers contributed to reduced monthly change in ∆C₂bio and net respiration signals relative to April.

NCP estimates were determined from the residual difference between the total monthly change in depth-integrated C₄ and the sum of all other contributing factors (ΔC₂bio+NCP) and from the total monthly change in depth-integrated NO₃ corrected for salinity changes (ΔC₂bio+N₂P₂C₄). During the productive months (growing season) of April–August, NCP was 12 ± 2 g C m⁻² (n = 5), which is about twice as large as the NCP estimate of 6 ± 2 g C m⁻² (n = 5). The differences in NCPc and NCPn for both seasonal and annual estimates arise from the sensitivity of the nitrate-based estimates to the C/N ratio selected. NCP is often computed using the Redfield ratio of C:N:P 106:16:1 (Redfield, 1963), which is most suitable when nitrate and phosphate are not depleted and C₄, NO₃, and PO₄ are assimilated and regenerated following the Redfield proportions (Arrigo, 2005). The relationship between C₄ and NO₃ (Figure 13d) yielded an average C/N of 7.5 (r² = 0.26, se = 0.83 µmol kg⁻¹, p < 0.001, n = 233) with a tendency towards lower C/N uptake ratios in the spring as a result of rapid NO₃ consumption as C₄ remained high, followed by a shift to higher C/N uptake ratios in the summer and early autumn upon intense drawdown of C₄ during high NCP. These C/N uptake ratios are higher than the Redfield C/N ratio (6.6) and consistent with previous observations of Frigstad et al. (2014) of C/N ratios that are higher relative to Redfield stoichiometry in northern high latitude regions, such as 6.7–7.0 in the Norwegian Sea region.

The strong de-coupling of C₄ and NO₃ at near-total depletion of NO₃ persisted during the growing season, suggesting that rapid recycling of nutrients, production of nitrogen-poor organic matter and/or other sources of nitrogen, such as ammonia, could be important factors (Kähler and Koeve, 2001). Furthermore, the C/N uptake ratio is likely to vary due to phytoplankton species composition (Sambrotto et al., 1993) and availability of dissolved...
iron (Takeda, 1998). As such, carbon-based estimates often exceed nitrogen-based estimates of NCP (Bozec et al. 2006; Tremblay et al. 2008; Ericson et al., 2019b). In addition, changes in light, temperature, salinity and availability of micro-nutrients are all likely to play a role. Therefore, using the traditional Redfield utilisation/replenishment ratio would not constrain the depletion in C relative to NO\textsubscript{3} in Kalfjorden, and N-based NCP estimates would not capture the extent of biological carbon uptake. Lower C/N in the winter and spring coincided with the smallest biologically driven monthly C deficits and was dominated by enrichment from respiration and remineralisation. Breakdown in water column stratification, mixing and diminishing light led to a steep decline in chlorophyll \textalpha concentrations (0.03–0.12 mg m\textsuperscript{-3}) and zooplankton abundance from October to February (Walker, 2018). The closer coupling of C and N at higher concentrations is evidence of export and remineralisation of organic matter in (sub-)surface waters, with likely contributions from sediment resuspension following episodic high winds (Walker, 2018) during winter.

The relationships between NO\textsubscript{3} and PO\textsubscript{4} (N/P; Figure 13e) and Si(OH)\textsubscript{4} and NO\textsubscript{3} (Si/N; Figure 13f) inform about the differences in supply and consumption of macronutrients. The average N/P was 11.6 (r\textsuperscript{2} = 0.83, se = 0.34 \mu mol kg\textsuperscript{-1}, p ≪ 0.001, n = 240) and shows close coupling of inorganic nitrogen and phosphorus. The N/P values were slightly lower in winter and higher in autumn. The temporal trends in N* (Figure 14a) and lower N/P observed during winter and early spring result from the

![Figure 14: Monthly variability in N* and potential alkalinity.](https://doi.org/10.1525/elementa.438.f14)
rapid removal of NO$_3^-$ and PO$_4^{3-}$ in the spring bloom. NCP increased from April as primary production exceeded respiration and caused a rapid reduction in C$_4$ in the surface layer. Reductions in N$_2$ from May show a shift in the system as nitrogen is depleted by ongoing biological production combined with denitrification and/or advective losses and remineralisation in subsurface water of organic matter with lower N/P ratios. The average seasonal Si/N was 0.3 (r$^2 = 0.28$, se = 0.03 μmol kg$^{-1}$, $p < 0.001$, n = 240) and shows variability and weak coupling with an excess of Si(OH)$_4$ relative to NO$_3^-$, at low NO$_3^-$ concentrations. Surface water NO$_3^-$ was depleted from May to October, while Si(OH)$_4$ and C$_4$ were continually consumed due to biological production by a diatom community. In addition, any iron limitation in the fjord would result in phytoplankton assimilating less NO$_3^-$ relative to Si(OH)$_4$, thus contributing to higher Si/N (Takeda, 1998), as observed in autumn.

The time lag in Si(OH)$_4$ drawdown of about one month, accompanied by lower N/P, reflects the species succession and prevalence of diatoms in the phytoplankton assemblage. The lower N/P uptake ratios coinciding with near depleted NO$_3^-$ concentrations suggests that intense nitrogen recycling occurred in the mixed layer and that other sources of nitrogen, such as ammonium, could be important to sustain phytoplankton production. The variable NO$_3^-$ (and PO$_4^{3-}$) concentrations below the mixed layer integrate the effects of uptake by phytoplankton in surface waters and regeneration from remineralisation of sinking organic matter and mixing with underlying waters. Due to the shallow seafloor towards the inner fjord and deep convective mixing, a substantial fraction of organically fixed carbon and nutrients (and exported silicones) is likely respired and returned to the water column. This mechanism would resupply the surface layers with NO$_3^-$ and possibly iron, to maintain biological production and induce high seasonality of C$_4$ and likely lower the potential for organic carbon burial in the sediments (Smith et al., 2015) in the inner fjord.

Surface waters were undersaturated with respect to atmospheric CO$_2$, but oceanic CO$_2$ uptake of 3.9 ± 0.3 g C m$^{-2}$ yr$^{-1}$, estimated for the full annual cycle, would resupply the upper layers with C$_4$. However, a net monthly decrease in the deficit in C$_4$ is observed due to the dominating effects of freshwater inputs and NCP. The growing season NCP$_{est}$ estimate for Kaldfjorden (14 ± 2 g C m$^{-2}$) is smaller compared to carbon-based NCP estimates of 49 g C m$^{-2}$ for the Amundsen Gulf region (Shadwick et al., 2011) and 65–85 g C m$^{-2}$ in Adventfjorden in Svalbard (Ericson et al., 2019a). The seasonal NCP$_{est}$ is slightly less than the equivalent NCP$_{up}$ estimates for the Nordic and Barents seas and Canadian Archipelago of 30–40 g C m$^{-2}$ (Codispoti et al., 2013). The annual NCP$_{est}$ estimates for Kaldfjorden (5 ± 2 g C m$^{-2}$ yr$^{-1}$) are modest in comparison to the carbon-based NCP estimates of 34 g C m$^{-2}$ yr$^{-1}$ in Adventfjorden in Svalbard (Ericson et al., 2019b) and 108 g C m$^{-2}$ yr$^{-1}$ in the Fram Strait–Svalbard region (Vaquer-Sunyer et al., 2013). These variations show enhanced seasonal and annual NCP in the higher latitude Arctic regions, likely due to higher productivity in sea-ice-influenced areas, enhanced oceanic exchanges for nutrient resupply, and greater export of organic matter from the productive surface layer to subsurface waters.

Accounting for freshwater effects, mixing and NCP, the remaining seasonal variations in surface water C$_4$ can be attributed to air–sea CO$_2$ exchange and residual changes that integrate variations in A$_4$, such as calcium carbonate formation/dissolution. To investigate this scenario, potential alkalinity (A$_4^+$) was evaluated to show that the water column average was 2285 ± 12 μmol kg$^{-1}$ with largest variations across the fjord from July to October (Figure 14b). Decreases in surface water A$_4^+$ of ~50 μmol kg$^{-1}$ are indicative of CaCO$_3$ formation during biogenic calcification, which was most pronounced from July to September. This period encompassed the summer phytoplankton bloom, when bloom-forming coccolithophores disturb the optical properties of surface waters and can be detected by signals in light reflectance by remote sensing (Tyrrell et al., 1999). Coccolithophores are widely abundant in the global oceans and influence seawater carbonate chemistry through the synthesis of CaCO$_3$ shells. The CaCO$_3$ or particulate inorganic carbon, that is exported out of the surface can either dissolve upon transport to deeper water layers or become buried in sediments (Rost and Riebesell, 2004). MODIS-Aqua satellite-corrected reflectance available from NASA Worldview shows development of a coccolithophore bloom in Kaldfjorden during this study (Figure 15). Images taken at the time of each sampling event (8 June, 25 July, 6 September), or those closest in time that are not obscured by cloud cover, and the 1st of each month show the bloom extended across Kaldfjorden by 26 July until 26 August. No reflectance in clear water was seen by the next sampling event on 6 September.

These observations support the proposed mechanism of removal of A$_4$ by calcification (and concomitant reduction in monthly ΔC$_{\text{total}}$), which would equate to a drawdown in C$_4$ of ~25 μmol kg$^{-1}$. The ΔC$_{\text{total}}$ accounts for about 20% of the ΔC$_{\text{alk}}$ and ΔC$_{\text{alk}}$ made a minor contribution (up to 8%) to ΔC$_{\text{total}}$, which shows the importance of calcifying phytoplankton in carbon cycling in the region. In the following autumn months, increases in A$_4^+$ up to 40 μmol kg$^{-1}$ (positive ΔC$_{\text{alk}}$) indicated A$_4^+$ inputs from terrestrial weathering and perhaps from CaCO$_3$ dissolution in a decaying coccolithophore bloom, likely from sampling events that captured particulate inorganic carbon (CaCO$_3$) shells in the water column.

**Ocean acidification state and CO$_2$ uptake**

The water column across Kaldfjorden remained saturated with respect to aragonite (and calcite, not shown) during the study period. The annual Ω range in Kaldfjorden compares very well (1.3–2.5) to that reported for fjords of western Norway (Omar et al., 2016). Seasonality in surface water Ω from minima in winter/spring to maximum in late summer was driven by changes in C$_4$ from the production (photosynthesis) and decay (respiration/remineralisation) of organic matter. Over the growing season, biological production reduced C$_4$ by 100 μmol kg$^{-1}$ in surface waters, leading to concomitant increases in Ω to 2.26–2.33 from March to September. Increases in ΔΩ$_{\text{bio-C}}$ (and ΔΩ$_{\text{bio-N}}$) of 0.25–0.40 between April to June dominated the monthly
changes $\Omega$ to show the importance of biological carbon uptake on the surface water acidification state (Figure 12). Removal of $A_i$ and subsequent lowering of $\Omega$ during calcification constituted a minor competing effect ($\Delta \Omega_{\text{calc}}$ of $-0.09$) that slightly counteracted the biologically-driven increases in $\Omega$ from $C_i$ drawdown in summer coccolithophore blooms. The effect of freshwater dilution reducing $C_i$ largely offset the parallel effect of $A_i$ reduction, resulting in increases in $\Delta \Omega_{\text{bio}}$ up to 0.11 across the fjord during the period of increasing freshwater fractions (April to June). Photosynthetically driven increases in $\Omega$ that counteract effects of freshwater dilution have been reported in other high latitude coast and fjord regions (Chierici and Fransson, 2009; Chierici et al., 2011; Franson et al., 2015, 2016; Ericson et al., 2019a). The decrease in $\Omega$ to $\sim1.9$ from May to June 2018 coincided with an episodic drop in salinity during a period of precipitation and peak freshwater fractions. Concomitant decreases in $\Delta \Omega_{\text{bio}}$ and $\Delta \Omega_{\text{bio}}$ suggested that suppression of $\Omega$ had occurred through increased $C_i$ from the degradation of organic matter and atmospheric CO$_2$ uptake in the freshwater flowing into the fjord, as previously reported (Anderson et al., 2009; Shadwick et al., 2011; Evans et al., 2014; Meire et al., 2015; Ericson et al., 2018). In addition, changes in water temperature affect $\Omega$ as colder waters have lower carbonate saturation through thermodynamic controls; increases in $\Omega$ by 1% arise due to warming of 1°C (Mucci, 1983). Coldest surface waters (2.17°C) with $\Omega$ of 1.59 in March warmed by 10.1°C to seasonal maximum (12.27°C) with $\Omega$ of 2.10 in July; thus, a thermodynamic increase in $\Omega$ of 0.16 could be expected and contribute to positive $\Delta \Omega_{\text{bio}}$ during this time. Saturation states steadily decreased during the autumn, winter and early spring in post-bloom conditions driven by organic matter remineralisation and net respiration ($\Delta \Omega_{\text{bio}}$ of $-0.50$) with effects of seasonal cooling. Impacts of mixing with subsurface carbon-rich Norwegian Coastal Water and air–sea CO$_2$ exchange played minor roles in monthly changes in $\Omega$ in surface waters.

Throughout each season, surface waters remained undersaturated with respect to the atmosphere and showed that Kaldfjorden was a sink for atmospheric CO$_2$ on an annual basis. Previous studies have also reported net atmospheric CO$_2$ sinks for sub-Arctic coast and fjord systems (Omar et al., 2016; Tynan et al., 2016; Yasunaka et al., 2016). The mean atmospheric CO$_2$ uptake of 0.86 ± 0.63 mmol m$^{-2}$ day$^{-1}$ is similar to that of 0.73 ± 0.40 mmol m$^{-2}$ day$^{-1}$ estimated for a marginal Arctic coastal environment of Hudson Bay, Canada (Else et al., 2008b). Wind speeds in Kaldfjorden were damped due to orographic steering, compared to measurements off the shelf (Nordby et al., 1999); thus, atmospheric CO$_2$ uptake in Kaldfjorden of 2.7 mmol m$^{-2}$ day$^{-1}$ is weak compared with the CO$_2$ influx of >15 mmol m$^{-2}$ day$^{-1}$ for the Norwegian Sea (Yasunaka et al., 2016) and the non-ice-covered Arctic shelf seas that generally absorb CO$_2$ at between 1 and 15 mmol m$^{-2}$ day$^{-1}$ (Omar et al., 2005; Cai et al., 2006; Else et al., 2008a, 2013; Ericson et al., 2018; Chierici et al., 2019). Enhanced CO$_2$ uptake in higher latitude waters results from substantial blooms, greater biological productivity and subsequent export of organic matter to deeper waters, coupled to strong winds enhancing oceanic CO$_2$ uptake. These results highlight the large spatial variability of the oceanic CO$_2$ sink and emphasize the need to resolve key processes on regional scales.

**Fjords in the future**

Strong seasonality in hydrography, carbonate chemistry, and macronutrients in the marine environment of Kaldfjorden, northern Norway, was driven by freshwater...
inputs, biological production, and mixing with subsurface coastal water. High latitude fjord and coastal environments may be particularly sensitive to future changes in ocean chemistry and the effects of ocean acidification, compared to the open ocean, due to the greater influence and spatial-temporal variations of freshwater sources and terrestrial influences. Fjords are dynamic ecosystems that naturally experience large ranges in carbonate chemistry and thus may exhibit a degree of resilience to future changes but may also be vulnerable to extreme values. Freshwater inputs play a key role in monthly changes in the acidification state (Ω) of surface waters through dilution effects on C$_T$ and A$_T$. Increases in air and seawater temperatures and greater freshwater fluxes will have consequences for the seasonal stratification and mixing of the water column, as well as phytoplankton species composition, bloom development, and biogeochemical cycling. Freshwater entering the fjord changes the seawater chemistry so that A$_T$ carbonate ion concentrations and pH decrease due to dilution. In addition to continued uptake of anthropogenic CO$_2$, these processes reduce the buffering capacity of the fjord water and increase the vulnerability of surface waters to acidification. Increased acidification is expected to have adverse effects on marine life, such as the pelagic calcifiers, coccolithophores, which contribute to monthly C$_T$ deficits during summer blooms and form an integral part of the food web (Andersson et al., 2015). Terrestrial inputs of organic matter and weathered minerals, such as carbonates and silicates from the surrounding bedrock (Fransson et al., 2015), enhance C$_T$ and provide a minor source of A$_T$ to surface waters. Effects of dilution from freshwater were strongly counteracted by primary production, with intense C$_T$ and nitrate drawdown from the period of the spring bloom to early autumn dominating the Ω seasonality. Estimates of NCP were modest but in accordance with other high latitude fjord and coastal regions: Kaldfjorden is an annual sink for atmospheric CO$_2$. High latitude fjords have been regarded as regions of high organic carbon sequestration, provided strong stratification does not prevail during the productive season, and if elevated fluxes of organic matter from productive surface waters and surrounding terrestrial sources can be exported and stored in fjord sediments (Smith et al., 2015). Changes in pH may also influence other chemical processes such as bioavailability of metals and toxins (Millero et al., 2009; Breitbarth et al., 2010).

The time-series data presented here emphasise the need for year-round sampling to better understand the natural variability in the marine environment. Addressing some of the remaining uncertainties could include resolving the characteristics (C$_T$, A$_T$, nutrients, δ$^{18}$O, salinity, organic matter) of the oceanic (coastal) and freshwater end-members to elucidate the main sources of inorganic and organic carbon, nutrients and minerals to Kaldfjorden. Phytoplankton community composition and net primary production, as well as drawdown of carbon and uptake of atmospheric CO$_2$, will ultimately depend on how the fjord marine ecosystems will respond to climatically induced changes. Observations of the biogeochemical dynamics on seasonal timescales further our understanding of carbon and nutrient cycling in these important marine systems and serve as benchmarks against which future changes can be compared and evaluated.

**Data Accessibility Statement**

The data collected during this study are publicly available at the Norwegian Marine Data Centre with the following reference: Angelika Renner (2020) Hydrography in Kaldfjorden, Troms, Norway. DOI: 10.21335/NMDC-1539462099.

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**Competing interests**

The authors have no competing interests to declare.

**Author contributions**

- Contributed to conception and design: AR, IW, MC, MB
- Contributed to acquisition of data: AR, EJ, MB, HHL
- Contributed to analysis and interpretation of data: EJ, HHL, AR, IW, MC
- Drafted and/or revised the article: EJ, AR, IW, MC
- Approved the submitted version for publication: EJ, AR, IW, MC, HHL, MB

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