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Biogeochemical Processes in the Active Layer and Permafrost of a High Arctic Fjord Valley

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Warming of ground is causing microbial decomposition of previously frozen sedimentary organic carbon in Arctic permafrost. However, the heterogeneity of the permafrost landscape and its hydrological processes result in different biogeochemical processes across relatively small scales, with implications for predicting the timing and magnitude of permafrost carbon emissions. The biogeochemical processes of iron- and sulfate-reduction produce carbon dioxide and suppress methanogenesis. Hence, in this study, the biogeochemical processes occurring in the active layer and permafrost of a high Arctic fjord valley in Svalbard are identified from the geochemical and stable isotope analysis of aqueous and particulate fractions in sediment cores collected from ice-wedge polygons with contrasting water content. In the drier polygons, only a small concentration of organic carbon (<5.40 dry weight%) has accumulated. Sediment cores from these drier polygons have aqueous and solid phase chemistries that imply sulfide oxidation coupled to carbonate and silicate dissolution, leading to high concentrations of aqueous iron and sulfate in the pore water profiles. These results are corroborated by $\delta^{34}\text{S}$ and $\delta^{18}\text{O}$ values of sulfate in active layer pore waters, which indicate the oxidative weathering of sedimentary pyrite utilising either oxygen or ferric iron as oxidising agents. Conversely, in the sediments of the consistently water-saturated polygons, which contain a high content of organic carbon (up to 45 dry weight%), the formation of pyrite and siderite occurred via the reduction of iron and sulfate. $\delta^{34}\text{S}$ and $\delta^{18}\text{O}$ values of sulfate in active layer pore waters from these water-saturated polygons display a strong positive correlation ($R^2 = 0.98$), supporting the importance of sulfate reduction in removing sulfate from the pore water. The significant contrast in the dominant biogeochemical processes between the water-saturated and drier polygons indicates that small-scale hydrological variability between polygons induces large differences in the concentration of organic carbon and in the cycling of iron and sulfur, with ramifications for the decomposition pathway of organic carbon in permafrost environments.

Keywords: permafrost, biogeochemistry, iron-sulfur, carbon, Svalbard

INTRODUCTION

Permafrost regions account for 24% of the land area of the Northern Hemisphere and store immense quantities (1330 to 1580 Pg) of organic carbon (Tarnocai et al., 2009; Hugelius et al., 2014). Permafrost ecosystems are currently a net sink of carbon, due to the drawdown of carbon dioxide exceeding emissions of methane and carbon dioxide (Schaefer et al., 2011; Kirschke et al., 2013; Parmentier et al., 2013). As rapid climate change occurs in the high Arctic, rising permafrost temperatures and progressively deeper active layers are exposing previously frozen soil organic carbon to microbial decomposition (Romanovsky et al., 2010; Koven et al., 2011; Schuur et al., 2015). This releases carbon dioxide and methane to the atmosphere and is predicted to cause permafrost ecosystems to become a net carbon source by the mid-2020s (Lee et al., 2012; Elberling et al., 2013; Schuur et al., 2015).

To predict the impact of permafrost thaw on biogeochemical cycling, it is necessary to understand the spatial heterogeneity of biogeochemical processes in Arctic soils. The distribution of stored organic carbon varies across the permafrost landscape, and with depth (Kuhry et al., 2010). Organic carbon accumulation and storage in permafrost is maximised in regions where waterlogged conditions dominate, as under these conditions, net primary production exceeds decomposition (Kolka et al., 2015). Hydrology exerts a powerful influence on the type of gaseous products released from decomposing permafrost organic carbon. For example, drained soil allows previously frozen organic carbon to decompose rapidly under oxic conditions, producing carbon dioxide (Liljedahl et al., 2012; Elberling et al., 2013), whereas water saturation limits the oxygen supply and enhances anaerobic respiration and methane production (Turetsky et al., 2008; Lipson et al., 2012; Olefeldt et al., 2013). However, under anaerobic conditions, alternative electron acceptors such as nitrate, manganese, iron and sulfate may be used preferentially in the microbial respiration of organic carbon, and inhibit the production of methane (Kristjansson and Schönheit, 1983; Dise and Verry, 2001). The use of alternative electron acceptors increases carbon dioxide production relative to methane, and hence the coupling between hydrology and the availability of alternative electron acceptors plays a crucial role in determining the magnitude of permafrost carbon emissions.

Hodson et al. (2016) conducted hydrological monitoring that demonstrated how lowland runoff draining the reactive, fine-grained sediments of Svalbard's Central Tertiary Basin (i.e., Adventdalen, the site of the present study) acquires reactive iron from pyrite oxidation. Pyrite oxidation exerted a major control upon the composition of this runoff, as evidenced by the presence of high sulfate and cation concentrations (Hodson et al., 2016). This process has also been documented in several nearby catchments (Fardalen, Bolterdalen and Longyeardalen; Yde et al., 2008; Rutter et al., 2011; Hindshaw et al., 2016). Therefore, sediments and water in Adventdalen and the surrounding area have an abundance of alternative electron acceptors, such as ferric iron and sulfate, for the microbial oxidation of organic carbon. The abundance of ferric iron and sulfate in the waters draining through Adventdalen is important in the context of

greenhouse gas emissions because Fe(III)- and sulfate-reduction are thermodynamically favourable, and therefore competitive, relative to methanogenesis. The availability of Fe(III) has been reported to decrease methanogenesis in the permafrost landscape of Barrow, Alaska (Lipson et al., 2012; Miller et al., 2015) and the availability of sulfate can decrease methane production in wetlands (e.g., Pester et al., 2012). Hodson et al. (2016) used the decrease in concentrations of sulfate and ^{54}Fe (i.e., dissolved and colloidal iron that passed through a 0.45 μm filter) at low flow to infer removal by sulfate reduction and precipitation of pyrite (FeS_2), iron monosulfide (FeS) and possibly elemental sulfur in ground waters draining through an alluvial sediment fan in Endalen (a tributary to Adventdalen). However, the processes removing ^{54}Fe and sulfate from these waters were hypothesised to be limited by a lack of organic matter (e.g., Raiswell and Canfield, 2012). A scarcity of organic matter has also been observed to limit the removal of ^{54}Fe and sulfate from pore water in fjord sediment cores in Svalbard (Wehrmann et al., 2014). From these findings, it is clearly necessary to understand if the accumulation of organic carbon in such environments results in enhanced removal of iron and sulfate from solution.

The objective of this study is to determine how the vertical and lateral heterogeneity of organic carbon accumulation in a high Arctic fjord valley influences the biogeochemical processes in the active layer and shallow permafrost sediments and pore water. This study presents and discusses geochemical analyses of pore water and sediment sampled from cores collected from the active layer and shallow permafrost of ice-wedge polygonal terrain in a high Arctic fjord valley.

MATERIALS AND METHODS

Field Site Description

Adventdalen (78°19'N, 15°93'E) is a 12 km \times 4 km valley, oriented NW-SE, in central Svalbard. The van Mijenfjord and Adventdalen Groups, which contain sedimentary rocks (sandstones, shales, and carbonates), comprise the lithology of the Adventdalen catchment. The sandstones and shales contain iron-bearing mineral phases, such as pyrite, siderite and glauconite (Dallmann et al., 1999; Riber, 2009; Svinth, 2013). There is also a small amount of iron in biotite and chlorite (chamosite; Hodson et al., 2016).

Glacial retreat and postglacial isostatic rebound influence landscape evolution (Milne and Shennan, 2007). Consequently, permafrost aggradation in many high Arctic fjord valleys has involved the freezing of reactive, fine-grained sediment originally deposited on the fjord floor via glacio-marine sedimentation and delta progradation throughout deglaciation (Haldorsen et al., 2010; Gilbert et al., 2018). During the Last Glacial Maximum, Adventdalen was filled with an erosive ice stream; this caused permafrost to thaw (Humlum et al., 2003; Humlum, 2005; Landvik et al., 2005). After retreat of the ice sheet, the glacial isostatic rebound of Svalbard meant that relative sea level fell, and a Gilbert-type delta prograded into Adventfjorden (Lønne and Nemeč, 2004). In Adventdalen, the deeper permafrost is

epigenetic, as it formed after the progradation of the delta (Gilbert et al., 2018). The shallower syngenetic permafrost in the aeolian terraces of Adventdalen aggraded concurrently with aeolian sedimentation (Gilbert et al., 2018). Although permafrost only aggraded since deglaciation, it is continuous in Svalbard, where it reaches a thickness of 80–100 m near the coast (Brown et al., 1997; Humlum, 2005). The seasonally thawed layer, or “active layer,” can be 0.4–6 m thick in Svalbard (Christiansen, 2005). In Adventdalen, the active layer thickness is around 1 m (Christiansen et al., 2010). This study focuses on the zone 0–2 m depth, which involves sampling the active layer and shallow permafrost.

The climate of Svalbard is polar tundra (Kottek et al., 2006) and in Longyearbyen between 1981 and 2010, the mean annual air temperature (MAAT) was -4.6°C (Førland et al., 2011). Between 1989 and 2011, MAAT increased by 1.25°C per decade (Førland et al., 2011) and in 2018, the MAAT was -1.8°C (eklima.met.no). The mean annual ground temperature (MAGT) ranges from -5.6°C to -3.2°C in the Adventdalen area (Christiansen et al., 2010) and the rising air temperatures have caused a recent increase in permafrost temperatures (Isaksen et al., 2019). The vegetation cover in the valley varies between 1.3 g m^{-2} and 27.2 g m^{-2} (moss) and 2.8 g m^{-2} and 9.6 g m^{-2} (vascular plants), with the amount of moss dependent on the soil moisture (Sjögersten et al., 2006).

The study sites (Figures 1A–C) were selected with the aim of sampling the heterogeneity of permafrost environments in the fjord valley. Table 1 lists the sample locations with details of the sampling methods. The sites are covered with Late Holocene loess (aeolian) deposits overlying alluvial and deltaic deposits (Cable et al., 2017; Gilbert et al., 2018). The aeolian deposits and alluvial fans dominate the syngenetic permafrost aggradation, whilst the permafrost beneath constitutes the epigenetic permafrost aggradation (following downward freezing; Gilbert et al., 2018; Toft-Hornum et al., In Review). The study sites are situated in areas of low-centred ice-wedge polygons, which are common in Adventdalen (Sørbel and Tolgensbakk, 2002). Ice Wedge North (Figure 1B) is a water-saturated wetland, mainly fed by springs, whereas Ice Wedge South (Figure 1C) is mainly precipitation-fed and only sporadically inundated by local snowmelt and rainfall.

Coring and Core Subdivision

Coring was undertaken before the onset of thaw, when air temperatures were below 0°C . Sediment cores of the frozen active layer and shallow permafrost were extracted in segments (5 cm diameter; between 5 and 50 cm length) to a depth of 2 m using motorised hand drilling equipment (a Stihl BT 130 drilling engine with a cylindrical drill head and rods). The core segments were extruded into sterile Whirl Pak® bags, which were sealed and frozen during transport to the University of Sheffield, where they were stored at -18°C . Cores were subdivided by sawing into 2 cm depth slices while frozen. The freshly cut surfaces were scraped with a scalpel, and the outer 2 cm removed with a hollow brass tube (3 cm diameter), to prevent contamination. The sawblade, scalpel and brass tube were cleaned with 70% isopropanol between slices.

Pore Water Extractions

A pore water extraction method was adapted from Spence et al. (2005). Vials containing samples were transferred to a Coy Vinyl Anaerobic chamber with a N_2 atmosphere (0 ppm oxygen). Each vial was weighed to determine the sample mass. Nitrogen-sparged de-ionised water (Milli-Q) was added to fill each vial. The vials were reweighed to determine the mass and volume of water added. A 3 ml volume of water was subsequently removed from the top of the vial to create a headspace. The vials were crimp-capped, inverted and stored for 5 days at 4°C whilst submerged in water (to prevent gas diffusion across the septa). This storage time enabled the de-ionised water to equilibrate with the sediment pore water (e.g., Spence et al., 2005). 7 days after first saturation of the sample, the vials were centrifuged at 7750 rpm for 5 min and transferred back to the anaerobic chamber. The equilibrated supernatant was filtered (0.22 μm nylon syringe filter) for chemical analysis and the sediment remaining in the vials was weighed after drying at 105°C for 24 h (Kokelj and Burn, 2003; Ernakovich et al., 2017). The vials were then reweighed to determine the initial moisture content of the samples.

Major ions (Ca^{2+} , Mg^{2+} , Na^+ , K^+ , Cl^- , NO_3^- , and SO_4^{2-}) were determined by ion chromatography (Dionex ion chromatograph, DX 90; limit of detection or “LOD” = 0.02 mg l^{-1} for the lowest, undiluted analysis; precision $<5\%$ for the mid-range standards). Trace metals (Fe, Mn) were determined on a 5 ml sample acidified with 50 μl reagent grade HNO_3 (Fisher Scientific Trace Metal Grade), using Inductively Coupled Plasma Mass Spectrometry or ICPMS (PerkinElmer Elan DRC II, MA, United States). The precision errors for repeat analyses of mid-range standards were $<5\%$, and the detection limits were $1.0\text{ }\mu\text{g l}^{-1}$. Analyte concentrations were corrected for blank analyses and corrected for the dilution during the pore water extraction.

Solid Phase Analyses

Carbon and nitrogen elemental abundance were determined by drying sediment samples at 105°C , acidifying in 6 M HCl, rinsing, drying, homogenising, weighing between 25 and 50 mg of each sample into a tin capsule and analysing on an Elementar vario EL cube (Animal and Plant Sciences Department, University of Sheffield). Elemental concentrations were checked with acetanilide standards ($\text{C}_8\text{H}_9\text{NO}$; Merck; $n = 28$, $C = 70.96 \pm 0.67\%$, $N = 10.34 \pm 0.10$), with 2 blanks and 2 acetanilide standards run every 15 samples.

Acid-volatile sulfur (AVS) and chromium-reducible sulfur (CRS) were determined at the University of Leeds. A two-step distillation method was applied to freeze-dried and milled sediment samples, first using 6 M HCl and then boiling 3 M CrCl_2 solution (Canfield et al., 1986; Fossing and Barker Jørgensen, 1989). In each extraction, H_2S was precipitated as Ag_2S , filtered, dried, and sulfide was determined gravimetrically. The stoichiometry of the phase was used to convert the mass to weight percent (FeS for AVS; FeS_2 for CRS).

Different operationally defined iron mineral phases were targeted with a four-step sequential extraction procedure applied to 100 mg freeze-dried and milled sediment samples. To extract amorphous and nanoparticulate iron (oxyhydr)oxide phases

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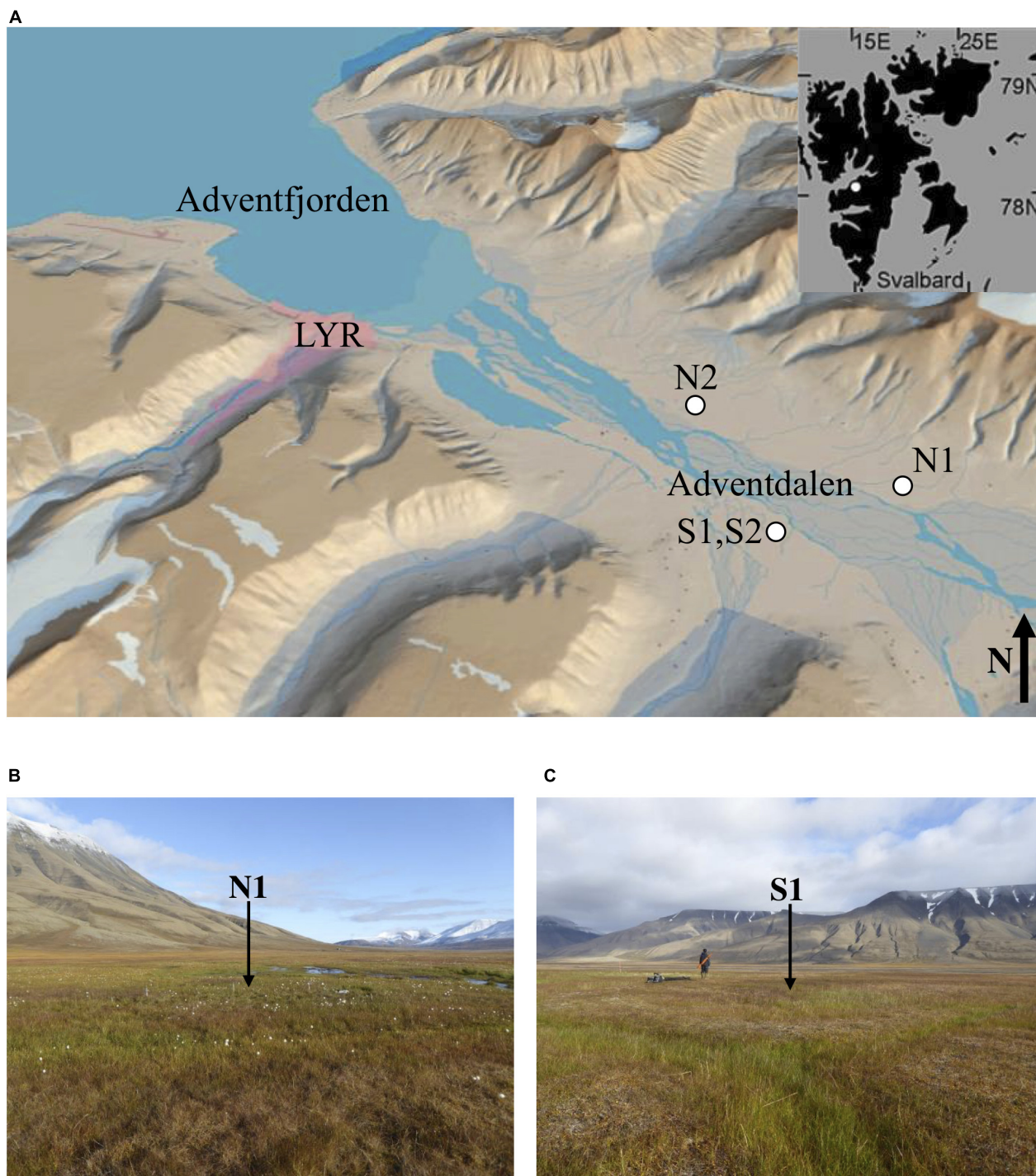


FIGURE 1 | (A) The Svalbard archipelago (inset) and Adventdalen, with a white circle indicating each site where ice-wedge polygons were sampled, **(B)** overlooking polygon N1, facing east, and **(C)** overlooking polygon S1, facing north. Photo credit: E.L. Jones.

($\text{Fe}_{\text{ascorbate}}$), each sample was shaken for 24 h with 10 ml ascorbate solution (50 g l^{-1} sodium citrate, 50 g l^{-1} sodium bicarbonate and 10 g l^{-1} of ascorbic acid; buffered at pH 7.5; Raiswell et al., 2008). To target iron bound in carbonates ($\text{Fe}_{\text{acetate}}$), the residual sample was shaken for 48 h at 50°C with 10 ml

sodium acetate solution (1 M sodium acetate solution buffered with acetic acid to pH 4.5; Poulton and Canfield, 2005). To target crystalline iron (oxyhydr)oxides ($\text{Fe}_{\text{dithionite}}$), the residual sample was then shaken for 2 h with 10 ml dithionite solution (50 g l^{-1} sodium dithionite buffered to pH 4.8 with acetic acid

TABLE 1 | Sampling locations, in UTM zone 33X.

Site Name	Site Code	Latitude (UTM)	Longitude (UTM)	Sample Type	Sampling Date	Shipping Date
Ice Wedge South	S1			1 core	15.06.15	28.09.15
		8679400	521010	8 water samples	26.08.17 & 27.08.17	29.09.17
	S2a	8679343	521042	1 core	12.02.17	19.06.17
	S2b	8679343	521042	1 core	14.02.17	19.06.17
Ice Wedge North	N1	8680446	522541	1 core	02.05.16	10.05.16
	N2	8681819	519780	1 core	15.04.14	29.04.14
	N1w	8680446	522541	9 water samples	31.08.17 & 01.09.17	29.09.17

The sample types are cores of the active layer and shallow permafrost and waters from the active layer extracted via Macro Rhizon samplers. All samples were extracted from the polygon centre.

and sodium citrate; Poulton and Canfield, 2005). Lastly, for magnetite ($\text{Fe}_{\text{oxalate}}$), the residual sample was shaken with 10 ml ammonium oxalate solution (0.2 M ammonium oxalate/0.17 M oxalic acid buffered with ammonium hydroxide to pH 3.2) for 6 h (Poulton and Canfield, 2005). These extractions were performed at room temperature (except for Fe_{carb} at 50°C). After centrifugation (4000 rpm for 4 min at 21°C), the supernatant from each extraction was stored at 4°C until analysis by atomic absorption spectroscopy (AAS). Recent research has shown that mineralogical associations based on sequential chemical extractions need to be treated with caution (Oonk et al., 2017; Slotznick et al., 2020; Hepburn et al., 2020). For instance, the efficacy of the $\text{Fe}_{\text{acetate}}$ stage in extracting iron bound in carbonates ranges from 3 to 85% (Oonk et al., 2017; Hepburn et al., 2020). The solubility of the targeted minerals depends upon their grain size, crystallinity and mineralogical association (Slotznick et al., 2020). Therefore, this study draws only tentative links between the extraction steps and the specific minerals represented by each step.

In situ Pore Water Sampling and Analyses

In late summer 2017, pore waters from the S1 polygon at Ice Wedge South and the N1 polygon at Ice Wedge North were sampled *in situ* using MacroRhizon soil moisture samplers (Van Walt Ltd.). Water samples for the analysis of $\delta^{18}\text{O}\text{-SO}_4$ and $\delta^{34}\text{S}\text{-SO}_4$ were collected to fill 50 ml centrifuge tubes and were stored at 4°C until processing and analysis at Lancaster Environment Centre, Lancaster University, United Kingdom. Water samples for the analysis of $\delta^{18}\text{O}\text{-H}_2\text{O}$ were collected to fill Eppendorf tubes and were stored at 4°C until analysis at the University of East Anglia (UEA). Water samples for the analysis of $\delta^{18}\text{O}\text{-SO}_4$ and $\delta^{34}\text{S}\text{-SO}_4$ were loaded onto ion exchange resins (Supelite™ DAX-8 for removal of dissolved organic matter; Dowex 50 W-X8 for removal of cations; and Dowex AG2 × 8 for removal of anions). Anions were eluted from the Dowex® AG2 × 8 using aliquots of 1 M ultrapure HCl to a total volume of 1.5 ml. A 0.2 ml volume of 1 M BaCl_2 was added to the eluted sample, and the samples were left for 48 h at 4°C to allow BaSO_4 to precipitate. Each sample was rinsed three times with Milli-Q de-ionised water using centrifugation and re-suspension to remove any interfering products (e.g., chlorides). The samples were dried at 40°C.

The dry samples were weighed into tin or silver capsules for analysis of $\delta^{34}\text{S}\text{-SO}_4$ and $\delta^{18}\text{O}\text{-SO}_4$, respectively. Isotopic analysis was undertaken by Elemental Analyser (Elementar Pyrocube) linked to a continuous flow isotope ratio mass spectrometer (Isoprime 100 with dual inlet capability for injection of monitoring gases), following methods in Wynn et al. (2015). Combustion of BaSO_4 within tin capsules yielded SO_2 for determination of $\delta^{34}\text{S}$. Analytical conditions demanded the use of vanadium pentoxide as an oxidizing agent and a combustion temperature of 1120°C. Pyrolysis of BaSO_4 at 1450°C within silver capsules and in the presence of carbon black, yielded CO for the determination of $\delta^{18}\text{O}$. $\delta^{34}\text{S}$ values were corrected against Vienna Cañon Diablo Troilite (VCDT) using within run analyses of international standards NBS-127 and SO5. $\delta^{18}\text{O}$ values were corrected to Vienna Standard Mean Ocean Water (VSMOW) using within-run analyses of NBS-127 and SO6. Internal standard MLSG (a subglacial meltwater precipitate of BaSO_4) was used to monitor drift and precision within each run as well as external precision between analytical sequences. Within-run standard replication (1SD) was better than $\pm 0.3\text{‰}$ for both sulfur and oxygen isotope values.

Water samples for the analysis of $\delta^{18}\text{O}\text{-H}_2\text{O}$ were analysed at the University of East Anglia (UEA) with a Picarro 1102i analyser, by direct injection of 2.6 μl of water. Samples were measured together with two United States Geological Survey (USGS) standards: USGS 64444 and USGS 67400 and a UEA Norwich Tap Water (NTW) internal laboratory standard. Using the calibration line defined by the USGS standards, the true isotopic composition of the samples was calculated, relative to VSMOW.

Precipitation Correction

Pre-melt snowpack chemistry data from Svalbard were compiled (Hodgkins et al., 1997; Wynn et al., 2006; Tye et al., 2007; Yde et al., 2008) to calculate the mean X/Cl ratios (where X is a major anion or cation). The following ratios in the snowpack were established: $\text{Na}/\text{Cl} = 0.82$; $\text{K}/\text{Cl} = 0.02$; $\text{Mg}/\text{Cl} = 0.10$; $\text{Ca}/\text{Cl} = 0.08$; $\text{SO}_4/\text{Cl} = 0.11$. By assuming that chloride behaves conservatively and originates only from precipitation (*pd*), these ratios were used to calculate the fraction of each anion or cation in pore water (*pw*) derived from precipitation (fX_{pd} ; Equation 1):

$$fX_{pd} = \left(\frac{X}{\text{Cl}}\right)_{\text{snow}} \cdot \left(\frac{\text{Cl}}{X}\right)_{pw} \quad (1)$$

The residual (fX_{npd}) represents the fraction of each anion or cation in pore water derived from weathering (Equation 2):

$$fX_{npd} = 1 - fX_{pd} \quad (2)$$

The $\delta^{34}\text{S}\text{-SO}_4$ values were corrected for snow inputs, following Equation (3) (from Hindshaw et al., 2016) and using the mean snowpack $\delta^{34}\text{S}\text{-SO}_4$ value of 17.5‰ from the compilation of pre-melt snowpack chemistry from Svalbard.

$${}^{34}S_{npd} = \frac{({}^{34}S_{pw} - f \cdot {}^{34}S_{snow})}{(1 - f)} \quad (3)$$

Where f is the fraction of sulfate derived from snowmelt. The $\delta^{18}\text{O}\text{-SO}_4$ values were corrected in the same way as in Equation (3), using the mean snowpack $\delta^{18}\text{O}\text{-SO}_4$ value of 9.28‰ from the compilation of pre-melt snowpack chemistry from Svalbard.

Bedrock Sulfide $\delta^{34}\text{S}$ Sampling and Analyses

Bedrock samples (sandstones, siltstones, shales) from each geological formation in the Adventdalen Group and the Carolinefjellet Formation were obtained from cores 13/2013 (33X E522859 N8685197, Dirigenten), BH9/05 (Urdkollbreen, 33X E528365 N8647669) and outcrops in Adventdalen. Rock samples were cut into small blocks using a lapidary trim saw (Lortone, United States) and weathered surfaces removed using a grinding surface (Saphir 330, ATM, Germany). Rocks were washed with de-ionised water in an ultrasonic bath before being dried in an oven at 50°C. Dried rocks were pulverised in a steel pestle and mortar into small chips which were further crushed to a fine powder in an agate disc mill (Tema, United Kingdom). Crushing equipment was washed with water, dried with compressed air and cleaned with ethanol between samples to prevent contamination. Chromium-reducible sulfur (CRS) was extracted from ground rock using the method previously described (Section “Solid Phase Analyses”). The resultant Ag_2S precipitates were weighed into tin capsules for $\delta^{34}\text{S}$ analysis via combustion in an Elemental Pyrocube elemental analyser coupled to an Isoprime continuous flow mass spectrometer at the University of Leeds. Samples were combusted at 1150°C to SO_2 in the presence of pure oxygen (N5.0) into a stream of helium (CP grade). The SO_2 produced flowed through tungstic oxide packed into the combustion column to ensure quantitative conversion. Excess oxygen was removed by reaction with hot copper wires at 850°C and water was removed in a Sicapent trap. Duplicate aliquots of silver sulfide were prepared because of a small sulfur isotope memory effect incurred during SO_2 processing in the pyrocube. $\delta^{34}\text{S}$ values were corrected to the VCDT scale using international standard IAEA S-3 (−32.06‰) and an inter-lab chalcopyrite standard CP-1 (−4.56‰). Reproducibility of a within-run check standard (BaSO_4) was $\pm 0.24\text{‰}$ (1 SD).

Data Analyses

Data analyses were performed in Microsoft Excel and R Core Team (2017) and graphics were prepared with Excel or the ggplot2 R package. Prior to testing correlations

between variables, Shapiro Wilk’s method was used to test whether data were normally distributed. Many variables were not normally distributed and their inter-relationships were non-linear. Therefore, monotonous relationships between the variables were tested using Spearman correlation, which is appropriate for non-normally distributed data and is particularly robust for dealing with skewed distributions and outliers (du Prel et al., 2010).

RESULTS

Figure 2 shows that the Ice Wedge North cores had a significantly higher gravimetric water content (g water g^{-1} dry sediment) than the Ice Wedge South cores according to Welch’s t -test (Table 2). The mean gravimetric water content in the N1 core was 2.79 g g^{-1} (maximum was 9.14 g g^{-1}) and highest both at the surface and below 90 cm depth, whereas N2 had a mean gravimetric water content of 6.05 g g^{-1} , with no clear patterns with depth. In S1, however, the mean water content was 0.57 g g^{-1} , with peaks near the surface, and at 105, 141 and 169 cm, and a max water content of 1.32 g g^{-1} . The mean gravimetric water content of S2a and S2b was similar (0.58 and 0.45 g g^{-1} , respectively). Both these cores had an increased water content near the surface.

The concentration of chloride was higher in the cores from Ice Wedge South, compared to those from Ice Wedge North (Figures 2F–J). The concentration of chloride peaked close to the base of the active layer in cores from both sites (Figures 2F–J). There was also a peak in chloride concentration near the ground surface at N1 (Figure 2F). There was an accumulation of chloride in the underlying permafrost of cores N2, S2a and S2b (Figures 2G–J).

The sedimentary organic carbon content for the two sites also differed significantly according to Welch’s t -test (Table 2). Figure 2 and Table 2 show that Ice Wedge North had a higher organic carbon content than Ice Wedge South. The surface of N1 had a high organic carbon content (29.7 wt.%). In the permafrost, the organic carbon content was high, but variable (low at 30–70 cm depths, peaking at 100–140 cm depth). S1, however, had a more uniform and lower organic carbon content (mean of 2.38 wt.%), but with a peak near the surface (5.40 wt.%) and at 61 cm depth (4.65 wt.%).

The concentrations of $\text{Fe}_{(aq)}$ and sulfate in pore waters differed significantly between the two sites, according to Welch’s t -test (Table 2). Figures 3A,B show that the $\text{Fe}_{(aq)}$ in N1 and N2 was $<3.0 \text{ mmol L}^{-1}$. In N1, peaks in aqueous iron occurred at 31, 67, 93, and 155 cm depth. In N2, the peaks were broader and at depths of 15–73 cm, 109–139 cm, and 197 cm. Figures 3F,G show that while N2 had a sulfate concentration $<1.3 \text{ mmol L}^{-1}$, N1 had up to 4.7 mmol L^{-1} in the peak at the active layer base. N1 had a clear pattern in sulfate, increasing from the surface toward the base of the active layer, and then decreasing into the permafrost (with a small peak ~ 150 cm).

Figures 3C–E show that the concentration of $\text{Fe}_{(aq)}$ was low ($<1.5 \text{ mmol L}^{-1}$) in the active layer of all the Ice Wedge South cores. $\text{Fe}_{(aq)}$ in S1 peaked at 75 cm and 123 cm depth. The $\text{Fe}_{(aq)}$ concentration in the permafrost of S2a and S2b had less distinct

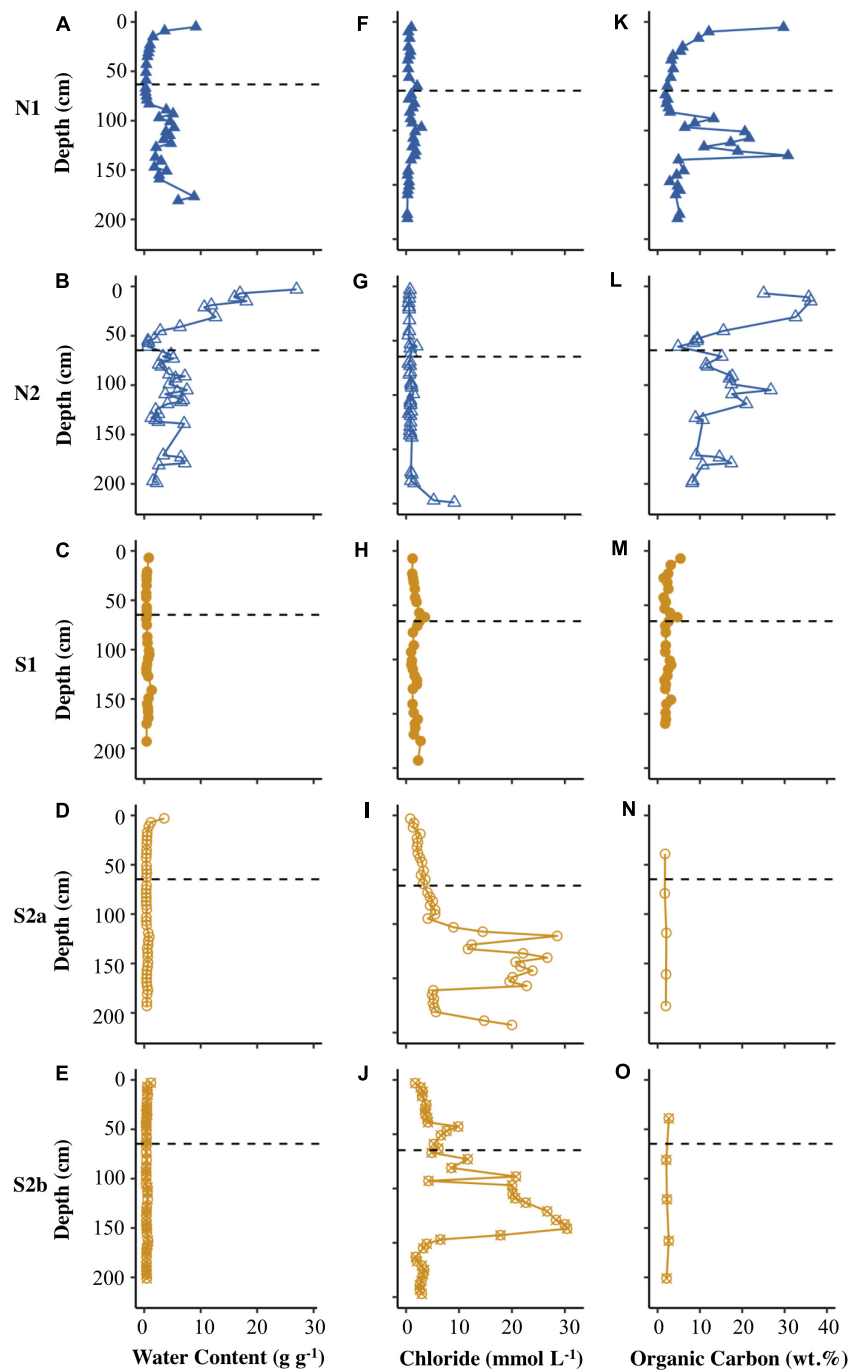


FIGURE 2 | Depth profiles of water content (A–E), chloride concentration (F–J), and concentration of organic carbon (K–O) for N1, N2, S1, S2a, and S2b. The horizontal dashed line on each plot represents the base of the active layer in 2017.

peaks, but increased with depth, reaching over 7.5 mmol L^{-1} . **Figures 3H–J** show that sulfate followed similar depth trends to $\text{Fe}_{(\text{aq})}$ for all three Ice Wedge South cores, reaching $>20 \text{ mmol L}^{-1}$ in core S2b. The sulfate concentration in the active layer was higher than the $\text{Fe}_{(\text{aq})}$ concentration.

Figures 4A–F show that N1 had only low concentrations of calcium and magnesium in pore water ($<6 \text{ mmol L}^{-1}$),

with the highest values of both cations in the uppermost core sample. N2 had similarly low concentrations of calcium and magnesium (**Figures 4B,G**). S1 displayed distinct peaks in all cations at the base of the active layer and $>1 \text{ m}$ depth in the permafrost (**Figures 4, 5C,H**). S2a and S2b show the greatest range in concentrations of calcium and magnesium (reaching $\sim 15 \text{ mmol L}^{-1}$), with higher concentrations in the permafrost

799 **TABLE 2** | Results of Welch's *t*-tests comparing organic carbon, gravimetric water content, Fe_(aq), sulfate, CH_{4(aq)}, CRS, Fe_{ascorbate}, Fe_{dithionite}, Fe_{acetate}, and Fe_{oxalate} 856
800 for Ice Wedge South and Ice Wedge North. 857

Variable	<i>t</i>	<i>df</i>	<i>p</i> -Value	Ice Wedge South			Ice Wedge North			Units of mean
				mean	<i>SD</i>	<i>n</i>	mean	<i>SD</i>	<i>n</i>	
Organic carbon	-7.9	60.8	***	2.30	0.86	30	12.5	9.94	60	Dry wt.%
Gravimetric water content	-7.4	72.5	***	0.53	0.34	113	4.6	4.68	73	g g ⁻¹
Fe _(aq)	8.7	127.8	***	2.92	2.76	113	0.6	0.60	73	mmol l ⁻¹
Sulfate	14.6	125.9	***	7.49	4.81	113	0.7	0.97	73	mmol l ⁻¹
CH _{4(aq)}	-10.3	70.1	***	1.24	2.13	113	98.0	79.0	71	μmol l ⁻¹
CRS	-3.1	6.2	*	0.01	0.01	9	0.1	0.04	7	Dry wt.%
Fe _{ascorbate}	-4.1	11.6	**	0.47	0.14	10	1.0	0.36	10	Dry wt.%
Fe _{dithionite}	7.5	17.7	***	0.44	0.09	10	0.1	0.10	10	Dry wt.%
Fe _{acetate}	-3.5	10.2	**	0.50	0.11	10	1.0	0.43	10	Dry wt.%
Fe _{oxalate}	5.1	17.4	***	0.76	0.17	10	0.3	0.21	10	Dry wt.%

814 *t* is the *t* statistic, *df* is the degrees of freedom, *p* is the significance level, *SD* is the standard deviation of the mean and *n* is the number of samples. Asterisks indicate 872
815 level of significance: **p* < 0.05; ***p* < 0.01; and ****p* < 0.001. 873

817 (Figures 4D,E,I,J). The concentration of potassium was generally 875
818 <2.5 mmol L⁻¹ in N2, S1, S2a and S2b (Figure 5), but reached 876
819 >20 mmol L⁻¹ in N2. The sodium concentration was <10 mmol 877
820 L⁻¹ in all the cores (Figures 5F–J). 878

822 The concentration of solid phase iron species was variable 879
823 in the N1 core, with concentrations: Fe_{ascorbate} < 1.6 wt.%, 880
824 Fe_{dithionite} < 0.35 wt.%. Fe_{acetate} was the dominant extracted iron 881
825 phase at N1, reaching a maximum of 1.57 wt.% (Table 3 and 882
826 Supplementary Figures S1A,C,E). Fe_{oxalate} at N1 was < 0.71 883
827 wt.% (Supplementary Figure S1G). N1 had the highest CRS at 884
828 15 cm depth and at 115 cm depth (Supplementary Figure S1K), 885
829 which tracked the organic carbon content. AVS was detected at 886
830 15 cm in N1, and also in all three samples measured between 114 887
831 and 156 cm (Supplementary Figure S1I). Compared with the 888
832 porewater Fe_(aq) and sulfate profiles, the concentration of solid 889
833 phase iron species was constant with depth in S1 (Supplementary 890
834 Figure S1B,D,E,H,J,L). Fe_{ascorbate} was <0.8 wt.%, Fe_{dithionite} was 891
835 <0.6 wt.%, Fe_{acetate} was <0.8 wt.%, Fe_{oxalate} was <1 wt.%, CRS 892
836 was <0.05 wt.%, and no AVS was detected in this core (Table 3). 893
837 Overall, N1 had less Fe_{dithionite} and Fe_{oxalate} than S1. N1 contained 894
838 an order of magnitude more CRS than S1 (Table 3). 895

839 Table 4 summarises the δ³⁴S-SO₄ results obtained from 896
840 samples of water within the active layer. Overall, water from 897
841 Ice Wedge North water was more enriched in ³⁴S than water 898
842 from Ice Wedge South. The Ice Wedge North δ³⁴S-SO₄ was 899
843 increasingly depleted in ³⁴S with increasing depth. At Ice Wedge 900
844 South, the water δ³⁴S-SO₄ was most enriched in ³⁴S at 30 cm 901
845 depth, whereas the δ³⁴S-SO₄ was more depleted in ³⁴S at depths 902
846 of 9 cm and 60 cm. 903

847 DISCUSSION 848

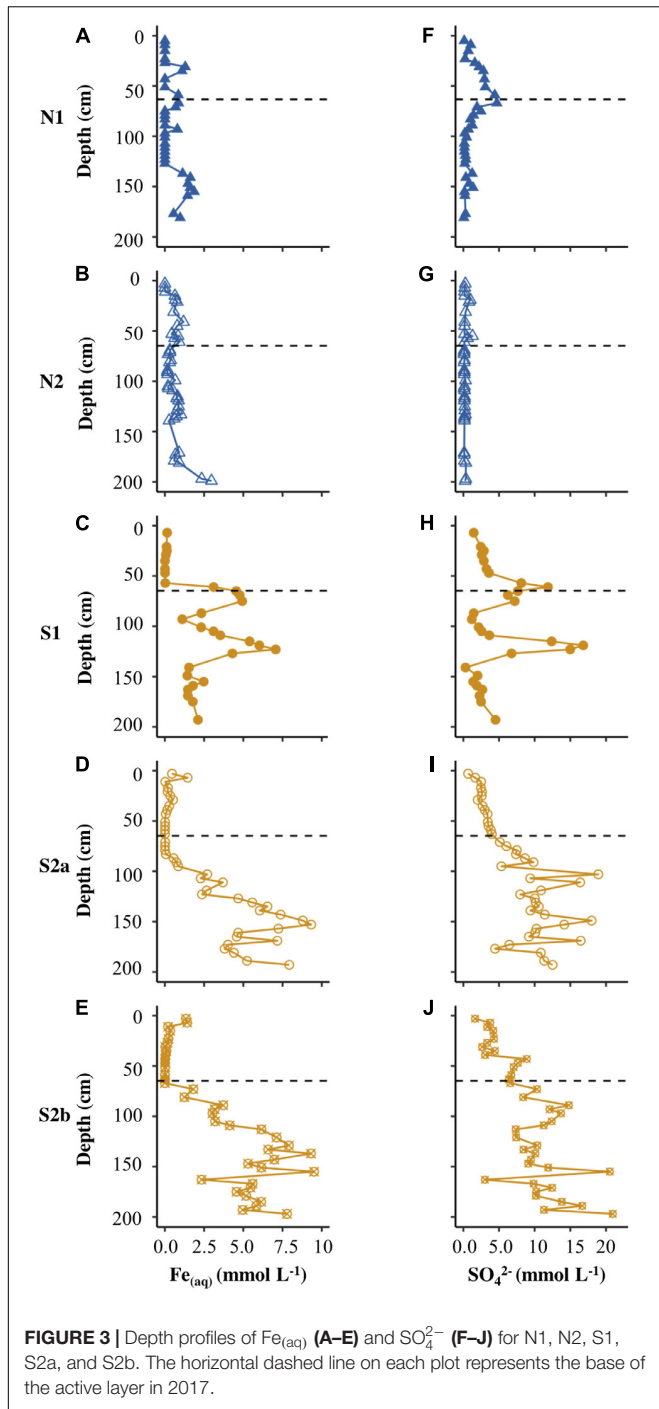
849 Since emergence from the sea during the Holocene epoch, 904
850 permafrost aggradation and ongoing organic carbon 905
851 accumulation have most likely caused significant changes 906
852 in the biogeochemical processes and mineral precipitation 907
853 reactions within this high Arctic floodplain. The two sites 908
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described above help understand these changes because their 875
876 contrasting hydrological regimes result in marked differences 877
878 in the accumulation and decomposition of permafrost organic 879
880 carbon, in spite of their proximity within the same valley. 881
882 Below, we describe how the quantity of organic carbon regulates 883
884 the consumption of alternative electron acceptors used for 885
886 microbial oxidation of organic carbon, thus causing a switch 887
888 in the water-saturated areas of the floodplain to iron- and 889
890 sulfate-reduction, with net iron and carbon storage via increased 891
892 CRS and Fe_{acetate} precipitation. In contrast, the drier areas 893
894 store far less organic carbon and instead remain dominated 895
896 by the biogeochemical signatures of pyrite oxidation. These 897
898 processes seem most likely during the earlier stages of floodplain 899
900 development, when they play a dominant role in the weathering 901
902 of fresh mineral surfaces, as expressed by the composition 903
904 of glacial meltwaters in the nearby Endalen, Bolterdalen and 905
906 Longyeardalen catchments (Yde et al., 2008; Rutter et al., 907
908 2011; Hodson et al., 2016, respectively). In the following 909
910 discussion, we therefore explore first of all the dominant 911
912 weathering processes that are linked to pyrite oxidation, 913
914 before examining the other processes associated with sulfur 915
916 and iron biogeochemistry that better explain the later stages 917
918 of biogeochemical evolution, once more organic carbon has 919
920 become available. 921

901 Dominant Weathering Reactions 902

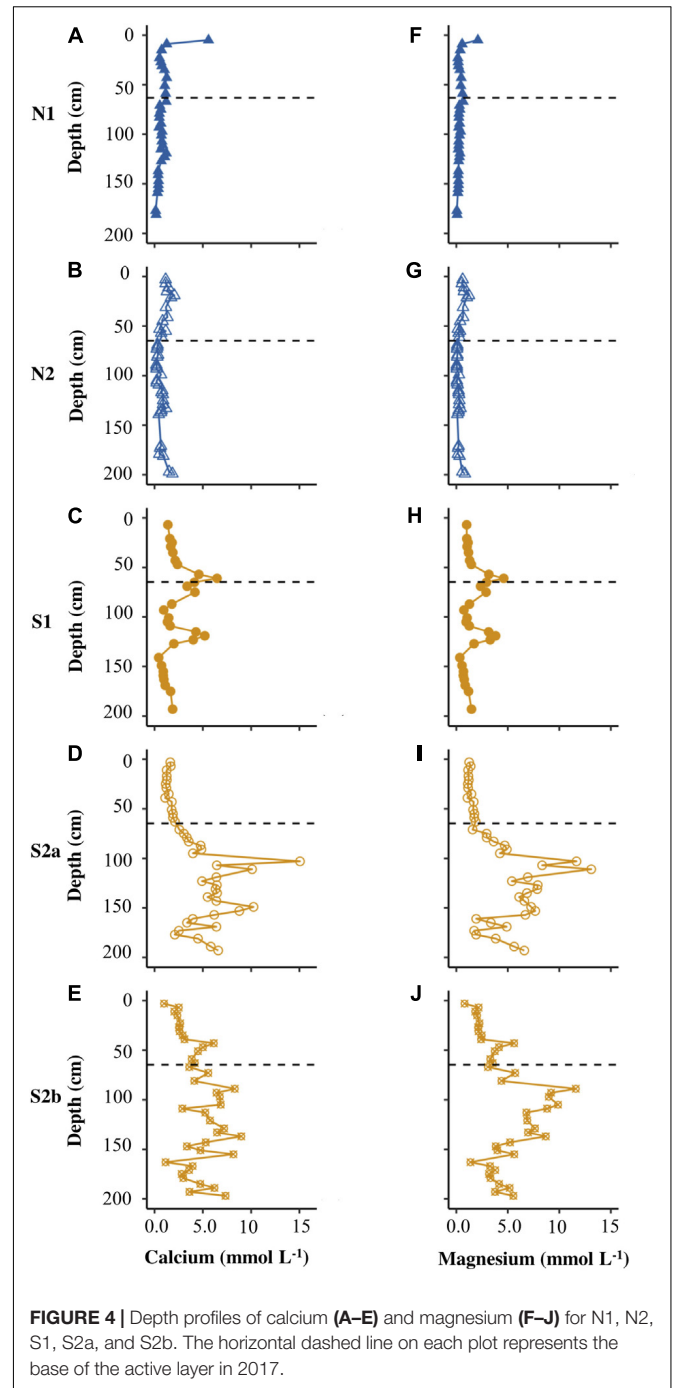
The pore water profiles in Figures 2–5 represent the following 903
904 attributes of both the active layer and the underlying permafrost: 905
906 (1) the *in situ* distribution of extractable or adsorbed solutes, 907
908 reactive mineral phases, and gases, and (2) the products 909
910 of additional rock-water-microorganism interactions following 911
912 thaw. Across both sites, except for N1 where potassium 913
914 dominated, calcium and magnesium were the dominant cations 915
916 (Figures 4, 5), indicating that carbonate dissolution prevails 917
918 over silicate dissolution, despite a low carbonate content in 919
920 the sediments (< 0.8 dry wt.% at Ice Wedge South) and low 921
922 volumetric carbonate contents (~0.3 to 1.8%, but up to 10.7%) in 923
924 rocks from the Todalen and Endalen endmembers (Svinth, 2013). 925

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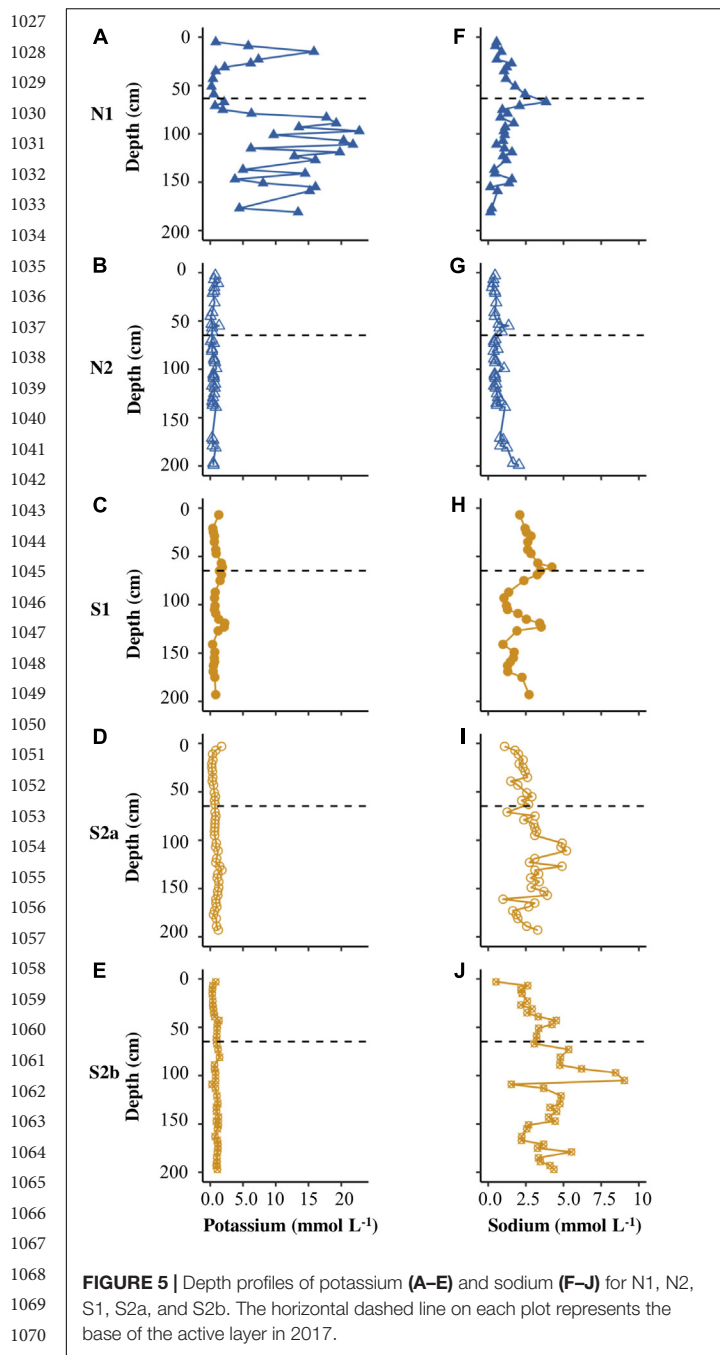
This contrasts with a study in a nearby unglaciated catchment (Fardalen) that found a relatively high proportion of silicate weathering, suggested to be the result of a combination of relatively rapid leaching of carbonate phases from the active layer and low rates of physical weathering in the sediments, failing to expose fresh carbonates to weathering (Hindshaw et al., 2016). However, carbonate weathering has previously been shown to control the water chemistry of both glacial and non-glacial watersheds, even where the bedrock is predominantly

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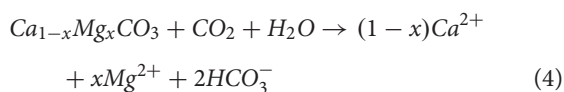
silicate, with only trace amounts of carbonate (Blum et al., 1998; Horton et al., 1999). This strongly suggests that although the low carbonate content coupled with the potential for active layer leaching might limit the importance of carbonate weathering in Adventdalen to some degree, the reactivity of the carbonate phases (compared to silicates) compensated for these factors.

The weathering processes responsible for the acquisition of Ca^{2+} and Mg^{2+} in the sediment pore water are most likely to be

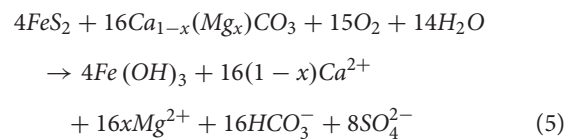


represented by the following reactions (after Tranter et al., 2002; Yde et al., 2008; Hindshaw et al., 2016):

- (1) Carbonate dissolution with carbonic acid (where x is equal to 1 or 0):



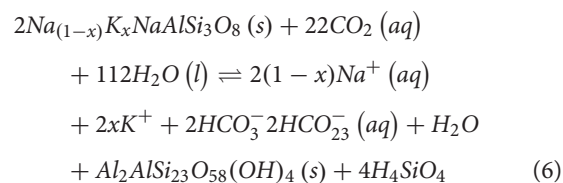
- (2) Sulfide oxidation coupled to carbonate weathering (where x is equal to 1 or 0):



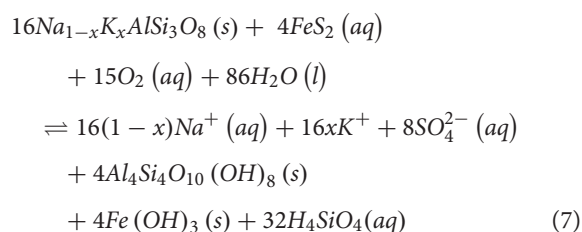
The covariance between ions is used here to identify the dominant weathering processes in the entire active layer and permafrost. Evidence for dolomite weathering as a source of both Ca^{2+} and Mg^{2+} was therefore provided by the strong positive correlation between these ions at Ice Wedge South ($p < 0.0001$, $\rho = 0.95$) and a regression slope close to unity (0.89). When corrected for precipitation inputs (rain and snow) of both cations, the slope remained unchanged (Figure 6), but the intercept decreased significantly toward zero (0.24 mmol L^{-1}). Therefore dolomite represents a credible, common source for both ions.

Although $(Ca + Mg)_{npd}$ was strongly correlated with SO_4^{2-npd} ($p < 0.0001$, $\rho = 0.81$), the regression slope was 0.73, which suggests that some of the sulfate was associated with other processes. Given the presence of silicates in the catchment, sulfide oxidation coupled to silicate dissolution may play a role in making up the deficit. Since silicates in the catchment are mainly present as Na- and K-feldspars, they may be represented by the formulae: $NaAlSi_3O_8$ and $KAlSi_3O_8$ in the following reactions (Tranter et al., 2002; Hindshaw et al., 2016):

- (1) Silicate dissolution with carbonic acid (where x is equal to 1 or 0):



- (2) Sulfide oxidation coupled to silicate dissolution (where x is equal to 1 or 0):



When silicate and carbonate dissolution are both driven by sulfide oxidation, the total base cation ($Ca^{2+} + Mg^{2+} + Na^+ + K^+$) ratio to SO_4^{2-} tends toward unity when precipitation inputs are insignificant (Fairchild et al., 1994; Tranter et al., 2002; Wadham et al., 2010). However, this characteristic signature can be overprinted or masked by gypsum dissolution, ion exchange reactions

TABLE 3 | Summary statistics calculated for the length of each core for the solid phase data of the cores S1 and N1 from Ice Wedge South and Ice Wedge North, including AVS, CRS, Fe_{ascorbate}, Fe_{dithionite}, Fe_{acetate}, Fe_{oxalate}, and organic carbon.

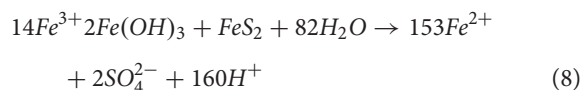
	Unit		Ice Wedge North		Ice Wedge South		
			N1	N2	S1	S2a	S2b
AVS	dry wt. %	mean	0.02	na	0	na	na
		(min-max)	(0.00–0.05)	na	(0–0)	na	na
		n	7	na	9	na	na
CRS	dry wt. %	mean	0.12	na	0.02	na	na
		(min-max)	(0.02–0.27)	na	(0.00–0.05)	na	na
		n	7	na	9	na	na
Fe _{ascorbate}	dry wt. %	mean	0.97	na	0.44	na	na
		(min-max)	(0.32–1.64)	na	(0.30–0.77)	na	na
		n	10	na	12	na	na
Fe _{dithionite}	dry wt. %	mean	0.13	na	0.44	na	na
		(min-max)	(0.05–0.32)	na	(0.24–0.52)	na	na
		n	10	na	12	na	na
Fe _{acetate}	dry wt. %	mean	1.00	na	0.53	na	na
		(min-max)	(0.38–1.57)	na	(0.39–0.73)	na	na
		n	10	na	12	na	na
Fe _{oxalate}	dry wt. %	mean	1.58	na	4.11	na	na
		(min-max)	(0.55–3.43)	na	(2.48–5.78)	na	na
		n	10	na	12	na	na
Organic Carbon	dry wt. %	mean	8.55	16.7	2.41	1.9	2.34
		(min-max)	(1.84–30.8)	(4.83–45.6)	(1.44–5.41)	(1.71–2.08)	(2.06–2.66)
		n	33	29	28	5	5

TABLE 4 | $\delta^{34}\text{S-SO}_4$ in water from the active layer in polygons S1 and N2.

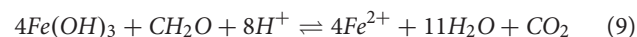
Site	Depth (cm)	$\delta^{34}\text{S-SO}_4$			
		mean	min	max	n
S1	9	–6.90	–6.95	–6.85	2
S1	30	–2.23	–3.74	–0.13	3
S1	60	–6.78	–8.37	–5.70	3
N1	9	15.1	3.33	25.3	3
N1	30	5.19	–0.98	11.4	2
N1	60	–2.83	–4.03	–2.16	3

or mineral precipitation reactions that remove base cations from solution. Of these, gypsum may be ignored, because it is absent from the bedrock (Svinth, 2013). Prior to correction for precipitation inputs, the total base cation ratio to sulfate at Ice Wedge South was 0.98 ($r^2 = 0.68$). The standard correction for precipitation inputs resulted in a lower regression slope of 0.76 ($r^2 = 0.70$; **Figure 7A**). Of the base cations, sodium was particularly affected by the above masking effects, with non-precipitation inputs appearing negative, similar to that which may be inferred from pore water data for Adventdalen sediment cores presented by Cable et al. (2017). This is indicative of ion exchange reactions or mineral precipitation (e.g., albite) and hence precludes the use of the non-precipitation ion ratios to estimate the importance of sulfide oxidation coupled to silicate dissolution.

Since the concentration of SO_4^{2-} derived from pyrite oxidation was high and developed within a sometimes-anoxic environment, alternative processes of pyrite oxidation to those shown by Equations 5 and 7 require identification. For instance, under acidic conditions, the $\text{Fe}(\text{OH})_3$ produced via Equations 5 and 7 dissociates to form Fe^{3+} , the reduction of which could result in a significant contribution of ferrous iron to the total cations in solution (Raiswell and Canfield, 2012; Hodson et al., 2016; Raiswell et al., 2018).



However, the reduction of iron (oxyhydr)oxide need not be achieved in combination with sulfide oxidation (Equation 9).



In contrast to Ice Wedge South, **Figure 7B** shows that the Ice Wedge North pore waters generally do not plot on the 1:1 line, and so there is no relationship between sulfate and total cations for these. Despite this, some samples from this site have a sulfate to chloride ratio greater than the snowpack sulfate to chloride ratio. This indicates that sulfide oxidation has enhanced the sulfate concentrations. The absence of a 1:1 relationship between total cations and sulfate at this site therefore strongly suggests that the sulfate produced by sulfide oxidation is removed in part by sulfate reduction during respiration of organic matter (Equation

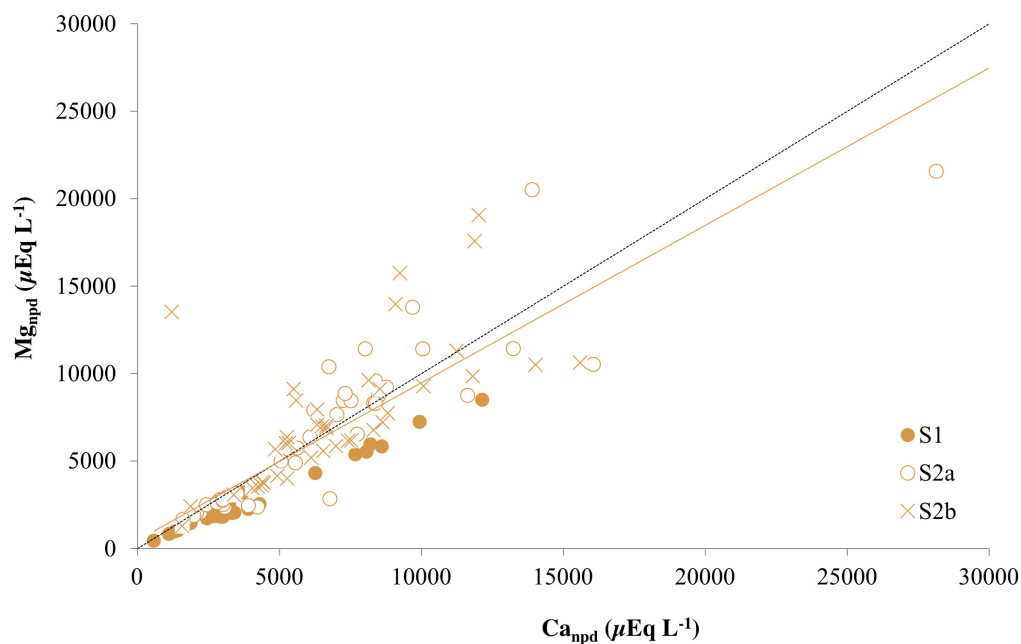
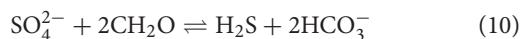


FIGURE 6 | Non-precipitation-derived magnesium (Mg_{npdp}) and non-precipitation-derived calcium (Ca_{npdp}) in the three cores from Ice Wedge South. The dashed black line corresponds to a 1:1 relationship between Ca_{npdp} and Mg_{npdp} that represents dolomite dissolution. The solid orange line is a regression for all points from the three cores from Ice Wedge South ($R^2 = 0.69$).

10; Wadham et al., 2004).



The major ion ratios show that sulfide oxidation coupled to carbonate dissolution can contribute substantial quantities of sulfate to the sediment pore water at Ice Wedge South. Here, the weathering mechanisms are therefore analogous to those reported in glacial catchments of the areas, which is intuitive when the source of the sediments is considered (i.e., the aeolian deposition of glaciofluvial sediments desiccated during early winter). The sulfuric acid produced by sulphide oxidation may further act as a weathering agent in this system, akin to in glacial catchments, potentially producing carbon dioxide during weathering of carbonates (e.g., Torres et al., 2017). In contrast, processes removing sulfate from the sediment pore water at Ice Wedge North preclude the use of major ion ratios to determine the significance of sulfide oxidation. In spite of this, sulphide oxidation is still very likely to occur because the provenance of the sediments is the same as that at Ice Wedge South.

Sources and Sinks of Sulfate

Sulfur and oxygen isotopic values of sulfate in waters provide compelling evidence for the identification of the sources and sinks of sulfate (e.g., Wynn et al., 2006, 2015; Turchyn et al., 2013; Hindshaw et al., 2016). Pre-melt snowpacks in Svalbard at Midtre Lovenbreen (Wynn et al., 2006) and near Ny Ålesund (Tye and Heaton, 2007) suggest a $\delta^{34}S$ range of +17 to +18‰ and a $\delta^{18}O$ range of +8.6 to +9.7‰ for the precipitation-derived sulfate contribution to the active layer pore water (i.e., SO_4^{2-pd}).

A study of dissolved organosulfur compounds in a raised peat bog showed that atmospheric sulfur in surface water sulfate is also taken up by plants (plant $\delta^{34}S$ was 0.1‰ and 4.2‰) and released when they decay, producing humic organosulfur with $\delta^{34}S$ values reflecting the precipitation-derived origin of the sulfate (Bottrell et al., 2010). Other sources of sulfate to the pore water include the oxidative weathering of pyrite (OWP), as in Equations 5, 7, and 8. The mass-weighted mean sulfide $\delta^{34}S$ values in the geological formations of the study region vary widely, from -40.6‰ in the Grumantbyen Formation to 1.78‰ in the Aspelintoppen Formation (Table 5). In the geological formations measured closest to the sites, the range in mass weighted mean sulfide $\delta^{34}S$ values is narrower, from -13.8‰ in the Carolinefjellet Formation to -2.01‰ in the Firkanten Formation. These nearby $\delta^{34}S$ values are therefore used to discriminate the rock-derived sulfate from the snowpack-derived sulfate.

The $\delta^{18}O$ of sulfate depends on the oxidation pathway; the oxygen atoms in the sulfate can originate either from atmospheric oxygen (+23.5‰) or from the surrounding water (-11‰ to -14‰ at our sites). In sulfate produced by OWP via Fe^{3+} (Equation 8), the oxygen atoms are derived solely from the surrounding water molecules. Experiments have demonstrated that there is no isotopic discrimination during the incorporation of oxygen atoms from water molecules into sulfate (Lloyd, 1968). In contrast, the incorporation of oxygen atoms from O_2 molecules into sulfate molecules during OWP via O_2 causes an isotopic fractionation of -8.7‰. Consequently, sulfate produced by OWP via O_2 is depleted in ^{18}O by -8.7‰ compared with atmospheric O_2 , which is strongly enriched in ^{18}O at +23.7‰ (Bottrell and Tranter, 2002). However, during OWP via O_2 ,

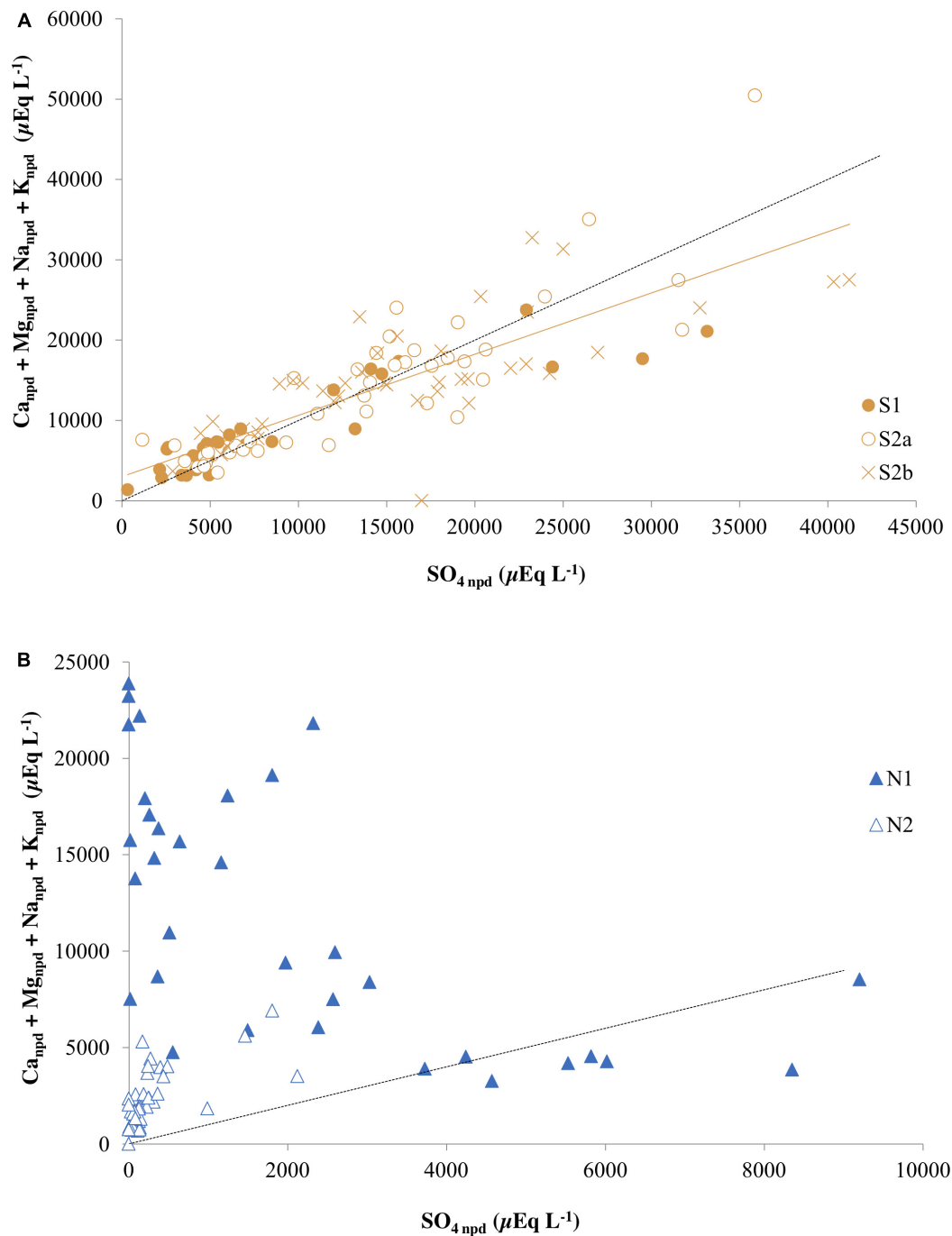


FIGURE 7 | Total non-precipitation-derived base cations and non-precipitation-derived sulfate in panel **(A)** the three cores from Ice Wedge South and **(B)** the two cores from Ice Wedge North. The dashed black line corresponds to a 1:1 relationship between $\text{SO}_{4\text{npd}}$ and $(\text{Ca}_{\text{npd}} + \text{Mg}_{\text{npd}} + \text{Na}_{\text{npd}} + \text{K}_{\text{npd}})$ that represents sulfide oxidation coupled to silicate dissolution. The solid orange line is a regression for all points from the three cores from Ice Wedge South ($R^2 = 0.70$).

there can be isotopic exchange between water and oxygen atoms in sulfoxy anions of intermediate valency, obscuring the isotopic signal of atmospheric oxygen (Balci et al., 2007). Hence, even in OWP via O_2 , three out of four oxygen atoms in the sulfate molecule could show an isotopic signal from water (Bottrell and Tranter, 2002).

To elucidate whether the sulfate in the samples from Adventdalen could derive from OWP only via O_2 , the approach of Bottrell and Tranter (2002) was applied to the $\delta^{18}\text{O}$ values of sulfate from the pore waters in the active layer. This conservative approach assumed that only the final oxygen atom incorporated into sulfate will still carry an isotopic signature indicative of its

TABLE 5 | The mean quantity of sulfide (wt.%) in bedrock, the numerical mean sulfide $\delta^{34}\text{S}$ (‰) in bedrock and the mass-weighted mean sulfide $\delta^{34}\text{S}$ (‰) in bedrock.

Formation	n	Mean S	Numerical mean $\delta^{34}\text{S}$	Mass weighted mean $\delta^{34}\text{S}$
		wt.%	‰	‰
Aspelintoppen	4	0.09	0.44	1.78
Battfjellet	3	0.06	1.56	1.49
Frysjoaodden (Gilsonryggen)	4	0.15	-4.98	1.34
Grumantbyen	3	0.10	-32.35	-40.60
Basilika	4	1.71	8.45	2.04
Firkanten (Endalen)	3	0.24	-6.22	-7.88
Firkanten (Todalen)	3	3.55	-0.76	-2.01
Carolinefjellet	3	0.15	-13.27	-13.79

TABLE 6 | Oxygen isotopes of weathering-derived sulfate ($\delta^{18}\text{O}\text{-SO}_4^{2-\text{npd}}$) and water ($\delta^{18}\text{O}\text{-H}_2\text{O}$) compared to the threshold for anoxic oxidation of pyrite ($\delta^{18}\text{O}_{\text{THRESH}}$).

Site	Depth	$\delta^{18}\text{O}\text{-SO}_4^{2-\text{npd}}$	$\delta^{18}\text{O}\text{-H}_2\text{O}$	$\delta^{18}\text{O}_{\text{THRESH}}$	OWP
	cm	‰	‰	‰	
S	9	-5.36	-11.4	-4.78	anoxic
S	9	-3.36	-11.3	-4.75	oxic
S	30	3.47	-12.3	-5.45	oxic
S	30	3.48	-12.3	-5.45	oxic
S	30	2.54	-12.4	-5.57	oxic
S	30	3.95	-12.1	-5.30	oxic
S	60	1.39	-12.0	-5.24	oxic
S	60	3.83	-13.1	-6.05	oxic
S	60	5.66	-12.4	-5.57	oxic
N	9	9.16	-13.0	-6.00	oxic
N	9	5.65	-13.0	-6.01	oxic
N	9	1.40	-13.0	-5.96	oxic
N	30	5.06	-13.7	-6.56	oxic
N	30	6.12	-13.5	-6.34	oxic
N	60	5.33	-14.1	-6.79	oxic
N	60	5.39	-14.1	-6.79	oxic
N	60	4.32	-13.8	-6.59	oxic
N	60	4.32	-13.5	-6.38	oxic

source (water or atmospheric oxygen). Only if there is less than 25% of the oxygen in a sulfate molecule derived from O_2 can it be certain that part of the sulfate was produced anoxically, by OWP via Fe^{3+} . Equation 11 uses the measured $\delta^{18}\text{O}\text{-H}_2\text{O}$ water isotopic compositions (Table 6) to calculate a threshold sulfate $\delta^{18}\text{O}$ ($\delta^{18}\text{O}_{\text{THRESH}}$) for the formation of sulfate with one oxygen atom from O_2 and three from water (Bottrell and Tranter, 2002).

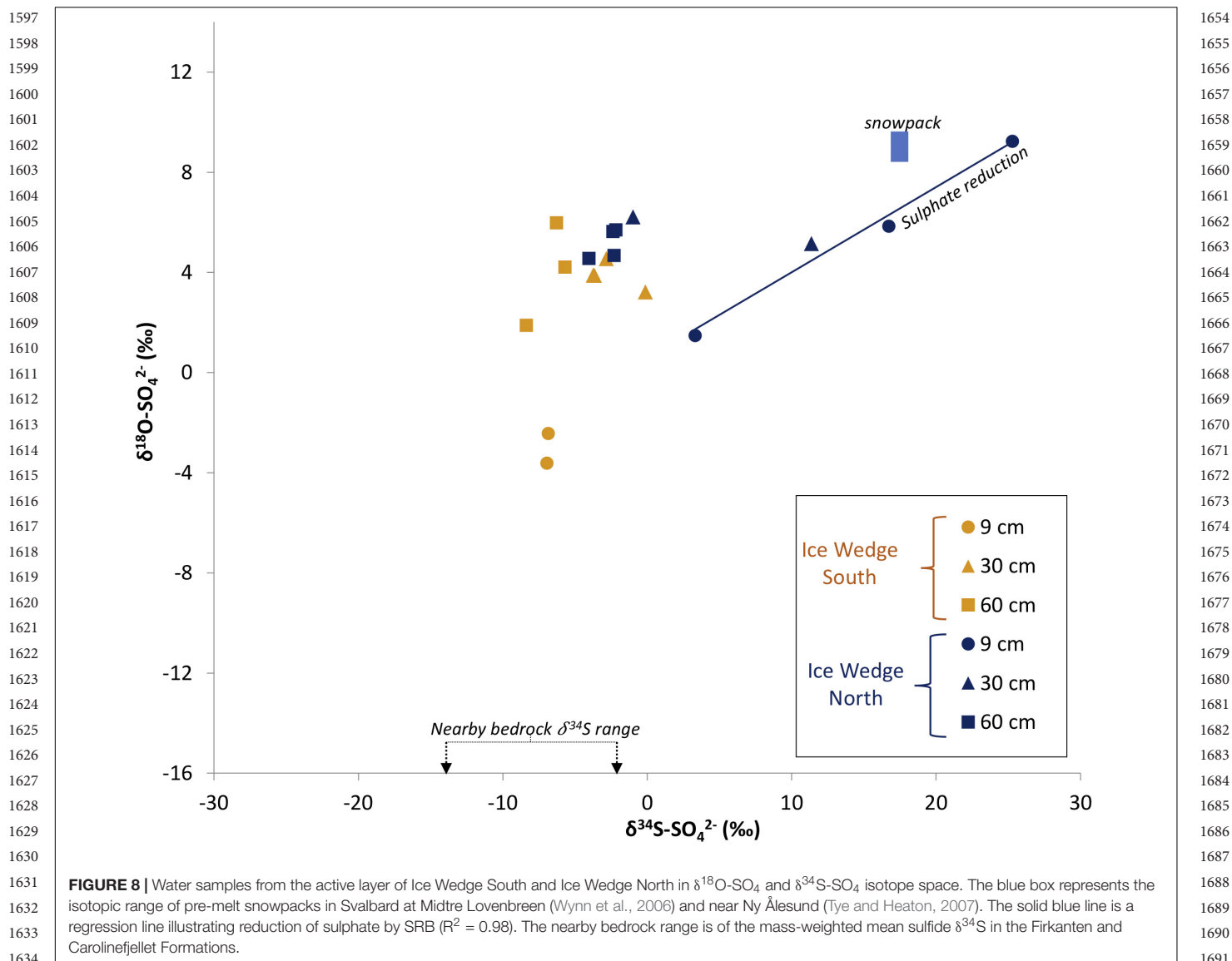
$$\delta^{18}\text{O}_{\text{THRESH}} = (23.7 - 8.7) \times 0.25 + 0.75 \times \delta^{18}\text{O}_{\text{THRESH}} \quad (11)$$

Comparing $\delta^{18}\text{O}\text{-SO}_4^{2-}$ and $\delta^{18}\text{O}_{\text{THRESH}}\text{-SO}_4^{2-}$ data, only one sample from the Adventdalen active layer pore water falls below

the threshold for OWP via Fe^{3+} (Table 6), indicating that $\text{SO}_4^{2-\text{npd}}$ in this sample originates from OWP via Fe^{3+} . In the rest of the samples, $\text{SO}_4^{2-\text{npd}}$ could have originated from OWP via Fe^{3+} , but the isotopic data do not require that and it is probable that $\text{SO}_4^{2-\text{npd}}$ in the remaining samples instead originated from OWP via O_2 . This is a surprising result, as these samples are from between 9 and 60 cm below the ground surface and it was anticipated that oxygen penetration would decrease with profile depth. However, it is possible that radial oxygen loss from the roots of wetland plants may have provided an oxygen source to this deeper pore water (e.g., Johnston et al., 2014). In addition, ice-wedge cracking and shallower cracking restricted to the active layer (O'Neill and Christiansen, 2018) provide a route for ingress of oxygenated rain and meltwater.

Figure 8 shows how plotting sulfate $\delta^{18}\text{O}$ and $\delta^{34}\text{S}$ in sulfur and oxygen isotopic space can provide a unique solution to elucidating the sulfate sources. The stoichiometric stage of pyrite oxidation results in sulfur isotopic fractionation between pyrite and sulfate ($\epsilon_{\text{SO}_4\text{-pyrite}}$) of -1.3‰ to -0.6‰ (Balci et al., 2007; Pisapia et al., 2007; Brunner et al., 2008). The mass-weighted mean sulfide $\delta^{34}\text{S}$ of the nearby Firkanten and Carolinefjellet Formations, combined with the sulfur isotopic fractionation during the stoichiometric stage of sulfide oxidation, indicates that any pore water sample with a $\delta^{34}\text{S}\text{-SO}_4^{2-\text{npd}}$ value between -15.1‰ and -2.6‰ is likely to have derived all of its sulfur from the stoichiometric oxidation of pyrite. Figure 8 shows that half of the pore water samples from Adventdalen contain $\text{SO}_4^{2-\text{npd}}$ within this $\delta^{34}\text{S}\text{-SO}_4^{2-\text{npd}}$ range, indicating that the stoichiometric oxidation of pyrite is an important process contributing sulfate to active layer pore water in Adventdalen. This corroborates the evidence from the $\delta^{18}\text{O}\text{-SO}_4^{2-}$ results presented earlier.

Although all samples except for the most ^{34}S -enriched sample have a $\delta^{34}\text{S}$ range between that of the bedrock and snowpack $\delta^{34}\text{S}$ values (Figure 8), the origin of sulfate in these samples in ^{34}S and ^{18}O cannot be explained purely as a mixing of sulfate derived from these two sources. There are two lines of evidence for this. Firstly, if mixing between (relatively ^{34}S -enriched) precipitation-derived sulfate and (relatively ^{34}S -depleted) weathering-derived sulfate were solely responsible for the $\delta^{34}\text{S}$ values of pore water sulfate in the active layer, a negative linear correlation between the concentration and $\delta^{34}\text{S}$ values of sulfate would result. There is no such negative linear correlation in the pore water samples from Adventdalen ($R^2 < 0.2$). Secondly, a negative correlation between $\delta^{34}\text{S}\text{-SO}_4^{2-}$ and the sulfate-to-chloride molar ratio ($\text{SO}_4^{2-}/\text{Cl}^-$) would result from mixing between weathering-derived sulfate (high $\text{SO}_4^{2-}/\text{Cl}^-$, relatively ^{34}S -depleted) and precipitation-derived sulfate (low $\text{SO}_4^{2-}/\text{Cl}^-$, relatively ^{34}S -enriched). In fact, there is a positive correlation between $\delta^{34}\text{S}\text{-SO}_4^{2-}$ and $\text{SO}_4^{2-}/\text{Cl}^-$ ($R^2 = 0.49$). Neither line of evidence supports mixing between snowmelt and pyrite oxidation as the sole reason for the $\delta^{34}\text{S}\text{-SO}_4^{2-}$ values observed in the pore water samples from Adventdalen. Hence, these statistical tests indicate that sulfate removal from the pore water by sulfate-reducing bacteria (SRB) may be a factor in enriching

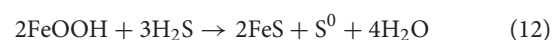


the remaining sulfate in these samples. SRB preferentially reduce the lighter isotopes of sulfur and oxygen, leading to isotopic enrichment of the residual sulfate. It is common for the product H_2S to be incorporated into iron sulfides or organic matter (Brown, 1985, 1986; Blodau et al., 2007). This causes both $\delta^{34}\text{S}$ and $\delta^{18}\text{O}$ values in the remaining sulfate to increase, producing a positive correlation between them (Mandernack et al., 2003). Since there is a positive correlation ($R^2 = 0.98$) between $\delta^{34}\text{S}$ and $\delta^{18}\text{O}$ in four samples from Ice Wedge North, and $\delta^{34}\text{S}$ in one sample exceeds the snowpack $\delta^{34}\text{S}$, it seems highly likely that SRB are active in the pore water of the active layer at Ice Wedge North.

Iron and Sulfur Mineral Precipitation

The $\delta^{34}\text{S}$ and $\delta^{18}\text{O}$ values of pore water sulfate in the active layer are indicative of sulfate reduction at Ice Wedge North. The sulfate concentration was relatively low at Ice Wedge North ($< 4.7 \text{ mmol l}^{-1}$; Figures 3F,G), supporting the isotopic

evidence for sulfate reduction. To produce a distinctive $\delta^{18}\text{O-}\delta^{34}\text{S}$ signature of sulfate reduction, the reduced sulfur must be sequestered in the solid phase (iron sulfide or carbon-bonded sulfur). The reactions of iron with hydrogen sulfide can be expressed with the following simplified equation scheme, where Equation 12 represents hydrogen sulfide reacting with iron oxides to form iron monosulfide (AVS) and sulfur, and Equation 13 represents the formation of the more stable pyrite (CRS) from the metastable iron monosulfide and sulfur (Raiswell and Canfield, 2012):



The AVS formed in Equation 12 consists mainly of the iron sulfide minerals mackinawite, greigite and amorphous FeS. These phases are usually only stable for short periods before their

1711 re-oxidation or further reduction to pyrite (Chesworth, 2008).
 1712 **Table 3** and **Supplementary Figure S1I** show that the AVS
 1713 concentrations in the Ice Wedge North sediments were low, but
 1714 detectable, in just over half the samples, reaching a maximum
 1715 of 0.05 dry weight percent (equivalent to $6.02 \mu\text{moles g}^{-1}$
 1716 dry sediment; 115 cm depth). The low concentrations of AVS
 1717 indicate that AVS is not a significant long-term store of the
 1718 products of iron and sulfate reduction. The CRS includes pyrite,
 1719 which is the most thermodynamically stable iron sulfide (Berner,
 1720 1967). Concentrations of CRS are higher than AVS, reaching
 1721 0.27 dry weight percent (equivalent to $22.4 \mu\text{moles g}^{-1}$ dry
 1722 sediment; 115 cm depth) at Ice Wedge North (**Table 3** and
 1723 **Supplementary Figure S1K**).

1724 The Ice Wedge North sediments contain abundant iron in
 1725 $\text{Fe}_{\text{ascorbate}}$, $\text{Fe}_{\text{dithionite}}$, $\text{Fe}_{\text{oxalate}}$, $\text{Fe}_{\text{acetate}}$, and CRS. Ferrous iron
 1726 in the porewaters is probably derived from the dissimilatory
 1727 reduction of iron (oxyhydr)oxides (Equation 9), as well as the
 1728 oxidation of allogenic and authigenic pyrite (Equation 8). In
 1729 addition to reaction with hydrogen sulfide to form AVS or
 1730 CRS, the dissolved ferrous iron reacts with bicarbonate ions to
 1731 form $\text{Fe}_{\text{acetate}}$ (iron bound in carbonates). Siderite (FeCO_3) is
 1732 an iron carbonate, and tends to occur in reducing, CO_2 -rich,
 1733 hydromorphic environments, such as peatlands (Chesworth,
 1734 2008). **Table 3** shows that $\text{Fe}_{\text{acetate}}$ reaches >1.5 dry weight%
 1735 in the sediments at Ice Wedge North, indicating that it is a more
 1736 significant sink of ferrous iron than CRS. As the precipitation of
 1737 $\text{Fe}_{\text{acetate}}$ dominates over the precipitation of CRS, it is possible
 1738 that an additional sink for the hydrogen sulfide is carbon-
 1739 bonded sulfur (CBS), which has previously been shown to be
 1740 an important sink for reduced sulfur in peat soils (Spratt and
 1741 Morgan, 1990; Blodau et al., 2007). Although CBS was not
 1742 measured in these cores, a strong positive correlation between
 1743 organic carbon content and CRS ($\rho = 0.9$; $p < 0.001$) and also
 1744 $\text{Fe}_{\text{acetate}}$ ($\rho = 0.67$; $p < 0.01$) indicates that where the organic
 1745 carbon content is high, sulfate reduction, CRS precipitation and
 1746 $\text{Fe}_{\text{acetate}}$ precipitation occur. Given the high concentration of
 1747 sedimentary organic carbon, it seems likely that CBS exists and
 1748 is forming at this location.

1749 In contrast to Ice Wedge North, the $\delta^{34}\text{S}$ and $\delta^{18}\text{O}$ values of
 1750 sulfate in pore water from the drier active layer of S1 indicate
 1751 primarily OWP via O_2 , and some OWP via Fe^{3+} . Evidence for
 1752 a mostly oxidised active layer at S1 is in the mostly low $\text{Fe}_{\text{(aq)}}$
 1753 concentration in the pore water from the active layer and the
 1754 low water table (summer 2017). CRS and AVS concentrations at
 1755 this site are low, corroborating the isotopic evidence that sulfate
 1756 reduction is negligible at this site. In addition, the concentration
 1757 of iron bound in carbonate ($\text{Fe}_{\text{acetate}}$) is lower than at Ice Wedge
 1758 North. Finally, **Figures 3C,H** show that aqueous iron and sulfate
 1759 co-vary at this site, which is indicative of pyrite oxidation (e.g.,
 1760 Hodson et al., 2016).

1761 The quantity of sedimentary organic carbon exerts a strong
 1762 control on biogeochemical processes and mineral precipitation
 1763 across both sites. Firstly, the organic carbon content is strongly
 1764 positively correlated with CRS ($\rho = 0.90$; $p < 0.001$), $\text{Fe}_{\text{acetate}}$
 1765 ($\rho = 0.67$; $p < 0.01$), and $\text{Fe}_{\text{ascorbate}}$ ($\rho = 0.75$; $p < 0.001$). In
 1766 addition, the sedimentary organic carbon content is negatively
 1767 correlated with pore water sulfate ($\rho = -0.68$; $p < 0.001$) and

1768 aqueous iron ($\rho = -0.61$; $p < 0.001$). This further supports
 1769 the mechanisms discussed above, whereby in organic carbon-
 1770 poor sediment, the oxidation of pyrite produces aqueous iron
 1771 and sulfate that are not reduced to form authigenic CRS
 1772 (primarily pyrite) and $\text{Fe}_{\text{acetate}}$ (primarily siderite). In contrast,
 1773 in organic carbon-rich sediment, the dissolved iron and sulfate
 1774 are reduced, forming $\text{Fe}_{\text{acetate}}$ and CRS. Finally, sedimentary
 1775 organic carbon content appears to influence the formation
 1776 of $\text{Fe}_{\text{dithionite}}$ (primarily crystalline iron (oxyhydr)oxides) and
 1777 $\text{Fe}_{\text{oxalate}}$ (primarily magnetite). Organic carbon content was
 1778 negatively correlated with $\text{Fe}_{\text{dithionite}}$ ($\rho = -0.85$; $p < 0.001$) and
 1779 $\text{Fe}_{\text{oxalate}}$ ($\rho = -0.89$; $p < 0.001$). A plausible explanation is a
 1780 combination of cycling redox conditions and dissolved oxygen
 1781 levels at Ice Wedge South, which increase the crystallinity of
 1782 ferrihydrite, coupled with an inhibition of the change from
 1783 $\text{Fe}_{\text{ascorbate}}$ (primarily poorly crystalline ferrihydrite) to $\text{Fe}_{\text{dithionite}}$
 1784 (primarily crystalline goethite) by organic compounds at Ice
 1785 Wedge North (e.g., Schwertmann and Murad, 1988; Thompson
 1786 et al., 2006; Amstaetter et al., 2012; Herndon et al., 2017).
 1787 These significant relationships demonstrate that the quantity of
 1788 sedimentary organic carbon exerts a landscape-scale control on
 1789 the active layer and permafrost biogeochemistry.

1790 Variations in Pore Water Geochemistry 1791 Due to Physical Processes

1792 The distribution of chemical species described above is partly
 1793 governed by a set of complex physical processes, including
 1794 hydrological inputs to the active layer (precipitation, advection
 1795 and ground ice melt; Throckmorton et al., 2016), hydrological
 1796 outputs from the active layer (evaporation, freezing and
 1797 advection) and the diffusion of ions from regions of high
 1798 concentration to regions of low concentration. Each of these
 1799 physical processes can influence the distribution of chemical
 1800 constituents in the active layer and their signatures commonly
 1801 overlap or mask one another. For instance, both evaporation
 1802 of water and ion freeze-out from the active layer concentrate
 1803 the pore water chemistry and enrich the $\delta^{18}\text{O}$ and δD of the
 1804 remaining water (Throckmorton et al., 2016). However, these
 1805 physical processes are secondary to the ion ratio interpretations
 1806 discussed earlier.

1807 The sediments in this study were dominantly fine-grained,
 1808 with the median grain size in the cores at Ice Wedge South
 1809 ranging from 34 to 60 μm . In fine-grained sediments, migration
 1810 of unfrozen water (and solutes) occurs along temperature-
 1811 induced pressure gradients toward colder ground (Kokelj and
 1812 Burn, 2003, 2005), and forms segregated ice lenses behind the
 1813 freezing front. During active layer freezeback, the migration is
 1814 upward toward the freezing front descending from the surface,
 1815 and downward toward the permafrost table during upward
 1816 freezing from the permafrost table (Cheng, 1983). This results
 1817 in the formation of ice lenses and the concentration of solutes
 1818 close to the ground surface, in the transient layer and at
 1819 the top of permafrost, whereas the middle of the active layer
 1820 becomes desiccated (Mackay, 1983). During thawing of the active
 1821 layer in summer, unfrozen water and solute migration occurs
 1822 downward into frozen ground below the advancing thaw front
 1823 (Cheng, 1983). As the thaw front reaches its maximum depth,
 1824

TABLE 7 | Results from Welch's *t*-test, which was used to test whether the concentration of chloride was significantly different for permafrost (PF) versus active layer (AL) in each core.

Core	Variable	<i>t</i>	<i>df</i>	<i>p</i>	AL			PF			Units for mean
					mean	<i>SD</i>	<i>n</i>	mean	<i>SD</i>	<i>n</i>	
S1	chloride	0.59	11.42	n.s.	1.84	0.77	9	1.67	0.52	20	mmol L ⁻¹
S2a	chloride	6.44	28.01	***	2.39	0.81	14	12.60	8.31	28	mmol L ⁻¹
S2b	chloride	3.52	31.62	**	4.66	2.20	14	11.81	10.29	28	mmol L ⁻¹
N1	chloride	1.22	22.15	n.s.	0.75	0.52	10	1.01	0.67	23	mmol L ⁻¹
N2	chloride	1.62	31.10	n.s.	0.76	0.40	13	1.35	1.78	27	mmol L ⁻¹

t is the *t* statistic, *df* is the degrees of freedom, *p* is the significance level, *SD* is the standard deviation of the mean and *n* is the number of samples. Asterisks indicate level of significance: **p* < 0.05; ***p* < 0.01; and ****p* < 0.001; n.s., not significant.

moisture and solutes can migrate into the top of permafrost. The downward migration into the top of permafrost in summer is greater than the upward migration out in winter when the temperature and pressure gradients are reversed, because the unfrozen water content and hydraulic conductivity are greatly reduced at lower temperature (Cheng, 1983). Consequently, there is a net annual downward migration that enriches the top of permafrost with segregated ice and solutes following repeated freeze-thaw cycles (Cheng, 1983; Kokelj and Burn, 2003, 2005).

There are multiple lines of evidence that indicate that these mechanisms contribute to redistributing solutes at the study sites. Segregated ice lenses were observed at the base of the active layer in cores from Ice Wedge South, and the top of permafrost was enriched with segregated ice at Ice Wedge North, indicated by the elevated water contents (Figures 2A,B). Assuming that chloride behaves conservatively, without participating in dissolution or precipitation reactions, its concentration profile in each core can be used to establish the net effect of moisture distribution by repeated freeze-thaw in the sediments (Jessen et al., 2014). The peak in the concentration of chloride close to the base of the active layer in cores from both sites coincides with the presence of ice lenses and is likely to have been caused by unfrozen water and solute migration during upward freezing of the active layer (Figures 2E,G,H,I). The peak in chloride concentration, coincident with an increase in water content near the ground surface at N1, may be indicative of upward moisture migration during active layer freezing.

The upper permafrost at Ice Wedge South formed syngenetically as the permafrost table rose in conjunction with sediment deposition (Gilbert et al., 2018), which means that the present-day permafrost is comprised of material formerly in the active layer. The accumulation of chloride in the permafrost is thus probably due to a combination of moisture migration into the top of permafrost at the end of summer (Cheng, 1983) and incorporation into permafrost of solutes from the base of the active layer during permafrost aggradation (Figures 2H–J). Cores S2a and S2b have significantly more concentrated chloride in the permafrost compared with the active layer (Table 7), and the chloride concentration in the active layer is greater than in the other cores. The formation of efflorescent salts on the ground surface at Ice Wedge South (Mora et al., 2015) further highlights the role of solute migration along potential gradients at the site.

The higher concentration of chloride in the cores from Ice Wedge South, compared to those from Ice Wedge

North, could be the result of diffusion of chloride from the underlying deltaic sediments. Nearby cores display an increase in chloride concentration from a mean of ~1 mmol L⁻¹ in the loess sediments to a mean of ~68 mmol L⁻¹ in the underlying deltaic sediments (Cable et al., 2017). Diffusion upward from these marine sediments may enhance the pore water chloride concentration at shallower depths in some locations, although the reasons for the location-specific diffusion are not clear. Additionally, it is likely that the topography of the sites has changed over time with the development of depositional landforms and ice-wedge polygons (Gilbert et al., 2018; O'Neill and Christiansen, 2018). Changes in topography are likely to have influenced patterns of water movement through the active layer, while variations in active layer thickness changed the amount of ground subject to advection over time. The data do not enable enhanced chloride concentration due to upward diffusion to be distinguished from enhanced chloride concentration due to low advection rates. However, greater rates of advection most likely occur at Ice Wedge North due to the spring-fed hydrologic regime and high water content.

In summary, unfrozen water and solute migration along potential gradients contribute to water content and solute variations with depth observed in the cores. The complex depositional and periglacial history at the sites makes further interpretation of chloride concentration patterns difficult. Ice Wedge South cores display evidence of chloride diffusion from underlying sediments and/or variations in rates of advection as the aeolian terrace aggraded. Despite the complexity introduced by these physical processes, the use of covariance between ions has enabled the dominant biogeochemical processes in Adventdalen to be elucidated.

CONCLUSION

This study highlights the importance of landscape evolution and demonstrates that permafrost aggradation and organic carbon accumulation have caused significant changes in the biogeochemical processes and mineral precipitation reactions within this high Arctic floodplain. The contrasting hydrological regimes of the study sites result in marked differences in the accumulation and decomposition of permafrost organic carbon. The drier areas of the floodplain store little organic carbon and

1939 are representative of the earlier stages of floodplain development.
 1940 These areas are dominated by the biogeochemical signatures of
 1941 pyrite oxidation and the weathering of fresh mineral surfaces,
 1942 similar to glacial meltwaters in nearby catchments. In contrast,
 1943 the water-saturated areas of the floodplain represent the later
 1944 stages of floodplain development, where the accumulation of
 1945 organic carbon causes a switch to iron- and sulfate-reduction,
 1946 with net iron and carbon storage via increased CRS (FeS₂) and
 1947 Fe_{acetate} (siderite) precipitation. In addition, contrasting chloride
 1948 concentrations demonstrate the geochemical contrast between
 1949 sediments of marine and aeolian origin. As air temperatures
 1950 continue to rise in the high Arctic and as thaw progresses deeper
 1951 into the permafrost, there are likely to be major changes in the
 1952 iron, sulfur and carbon cycling in this valley, depending on how
 1953 permafrost thaw impacts the geomorphology and hydrology of
 1954 the ice-wedge polygonal terrain.

1955 DATA AVAILABILITY STATEMENT

1956
 1957
 1958 Jones (2019). Permafrost and active layer biogeochemical data
 1959 from Adventdalen (2015–2017) (Version 1.0) in the UK Polar
 1960 Data Centre, Natural Environment Research Council, UK
 1961 Research & Innovation (<https://doi.org/10.5285/4c90d954-3db2-4084-9fe9-e050c839a6fe>).

1962 AUTHOR CONTRIBUTIONS

1963
 1964
 1965
 1966
 1967 EJ wrote the manuscript. EJ, AH, and HB collected the cores of
 1968 permafrost and water from the active layer. EJ and JR processed
 1969 and analysed the cores of permafrost. TD analysed the rock
 1970 cores. ST and PW facilitated the laboratory analyses of permafrost

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- 2287 **Conflict of Interest:** The authors declare that the research was conducted in the
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