The Spatiotemporal Variability of Snowpack and Snowmelt Water $^{18}$O and $^2$H Isotopes in a Subarctic Catchment

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Abstract This study provides a detailed characterization of spatiotemporal variations of stable water $^{18}$O and $^2$H isotopes in both snowpack and meltwater in a subarctic catchment. We performed extensive sampling and analysis of snowmelt water and meltwater isotopic compositions at 11 locations in 2019 and 2020 across three different landscape features: (a) forest hillslope, (b) mixed forest, and (c) open mires. The vertical isotope profiles in the snowpack’s layered stratigraphy presented a consistent pattern in all locations before snowmelt, and isotopic profiles homogenized during the peak melt period; represented by a 1–2‰ higher $^{18}$O value than prior to melting. Our data indicated that the liquid-ice fractionation was the prime reason that caused the depletion of heavy isotopes in initial meltwater samples prior to the peak melt period. The liquid-water isotopic composition was influenced by snowmelt rate, with higher fractionation during slow melting. The kinetic liquid-ice fractionation was evident only in close examination of meltwater le-excess values, not $^{18}$O values alone. Meltwater was isotopically heavier and more variable than the depth-integrated snowpack water; the weighted mean of meltwater $^{18}$O isotopic value was higher by 0.62–1.33‰ $\delta^{18}$O than the weighted mean of snowpack $^{18}$O isotopic values in forest hillslope and mixed forest areas, and 1.51–6.37‰ $\delta^2$H in open mires. Our results reveal close to 3.1‰ $\delta^{18}$O disparity between the meltwater and depth-integrated snowpack isotopic values prior to the peak melt period, suggesting that proper characterization of meltwater $\delta^{18}$O and $\delta^2$H values is vital for tracer-based ecohydrological studies and models.

Plain Language Summary Snowmelt water is an essential input for surface and subsurface flows, and soil water recharge in a catchment. Due to global changes in temperature and precipitation patterns, the role of meltwater in water resource management will be critical, particularly in high latitude and altitude regions. The conservative stable isotopes of water $^{18}$O and $^2$H have been proven to be valuable tools in hydrological investigations. In this study, we analyzed snowmelt water $^{18}$O and $^2$H isotopes and have presented their spatial and temporal variation. Here, we have also compared snowmelt water $\delta^{18}$O and $\delta^2$H values with the snowpack $\delta^{18}$O and $\delta^2$H values in diverse landscapes. This information is critical for tracking snowmelt water movement in different landscapes. Our results suggest that meltwater is isotopically heavier than the depth-integrated snowpack, and the range of variation of snowmelt water $\delta^{18}$O and $\delta^2$H values is wider than the range of the depth-integrated snowpack $\delta^{18}$O and $\delta^2$H values. The results also suggest that a low melt rate causes the lowering of meltwater $\delta^{18}$O and $\delta^2$H values through the interaction of liquid and solid content isotopically within the snowpack.

1. Introduction
Changes in air temperatures, precipitation patterns, and moisture conditions are producing unprecedented changes in snow accumulation and melt in northern regions (Arp et al., 2015; Irannezhad et al., 2022; Kiewiet et al., 2022; Rasouli et al., 2022; Rixen et al., 2022; Ruosteenoja et al., 2020; Vormoor et al., 2015). The changes in magnitude and timing of snow precipitation and snowmelt runoff have gradually resulted in a hydrological regime shift from a snowmelt to a rainfall-dominated system in cold climate regions (Berghuijs et al., 2014; Bintanja & Andry, 2017), with varying feedbacks at a catchment scale (Ala-aho et al., 2021; Meriö et al., 2019; Pi et al., 2021). Meltwater is vital for replenishing water storage and exporting solutes and nutrients (Ågren et al., 2010; Tomco et al., 2019). Thus, the characterization of snowmelt water processes and patterns is especially important for hydrological and ecosystem studies (Alstad et al., 1999; Young-Robertson et al., 2017).
Stable isotopes of oxygen and hydrogen ($^{18}$O and $^2$H) in water molecules have been widely exploited to aid greater understanding of the water cycle (Penna et al., 2018; Terzer-Wassmuth et al., 2021; Welker, 2000) and snow hydrology (Beria et al., 2018; Taylor et al., 2001). The range of snowmelt water $\delta^{18}$O and $\delta^2$H values can be different from the range of precipitation isotope values (Beria et al., 2018; Gat, 1996; Hooper & Shoemaker, 1986; McNamara et al., 1997; Rodhe, 1981, 1998). Snowmelt water $^{18}$O or $^2$H isotopes are often used as one of the end-members in isotope hydrograph separation methods for understanding streamflow generation processes and source water partitioning (Laudon et al., 2002; Pant et al., 2021; Penna & van Meerveld, 2019; Penna et al., 2014, 2016; Rodhe, 1981), in mixing models (Beria et al., 2020; Unnikrishna et al., 2002) and to help infer inter-transit time distributions of catchment storage (Ceperley et al., 2020; Fang et al., 2019; Lyon et al., 2010). The snowpack water $^{18}$O and $^2$H isotopes can be used to provide proxy records of water sources for plants (Alstad et al., 1999; Jespersen et al., 2018; Welker et al., 2005; Young-Robertson et al., 2017) and moisture transport mechanisms (Ala-aho et al., 2021; Pu, Kong, et al., 2020; Sinclair & Marshall, 2008).

Despite numerous applications of stable isotopes of water $^{18}$O and $^2$H in hydrology, different isotope fractionation processes in the snowpack complicate the use of $^{18}$O and $^2$H isotopes in snow-influenced regions (Laudon et al., 2004; Schilling et al., 2021; Tetzlaff et al., 2015). Freeze-thaw and condensation-sublimation processes result in frequent phase changes of the water (solid-liquid-vapor) and cause isotope fractionation in the snowpack (Evans et al., 2016; Hürkamp et al., 2019; Lee et al., 2010; O’Neil, 1968; Stichler et al., 1981; Zhou et al., 2008). Post-depositional snowpack processes lead to the spatial variability of snowpack and subsequent meltwater $^{18}$O and $^2$H isotopic composition (Ala-aho et al., 2021; Dietermann & Weiler, 2013; Hürkamp et al., 2019; Ohlanders et al., 2013; Siegenthaler & Oeschger, 1980; Sinclair & Marshall, 2008). The usual lapse rate of isotope values in the snowpack ($-0.6\%$ to about $-1.0\%$ of $\delta^{18}$O per 100 m (Niewodnizki et al., 1981)) is higher than the typically reported lapse rate for precipitation isotopes ($-0.2\%$ of $\delta^{18}$O per 100 m (Siegenthaler & Oeschger, 1980)). This might be due to the wind drift and fractionation caused by the accumulation of snow. When the accumulated snow starts melting, the temporal variability of meltwater in $^{18}$O and $^2$H, with a gradual growth in isotopic enrichment of snowmelt $^{18}$O and $^2$H isotopes, is often observed (Ala-aho et al., 2017; Laudon et al., 2002; Taylor et al., 2001). Moreover, the redistribution of snow cover caused by snow falling from the canopy also alters the snowpack $\delta^{18}$O and $\delta^2$H values in forest areas (von Freyberg et al., 2020; Koeniger et al., 2008; Vystavna et al., 2021).

As a result, characterizing the spatial and temporal variation of snowpack and meltwater $^{18}$O and $^2$H isotopic composition for their use in hydrology research, in snow-influenced regions, remains challenging (Beria et al., 2018; Penna et al., 2018). Studies that address the spatial and temporal variability of snowpack and snowmelt water $^{18}$O and $^2$H isotopes on a catchment scale (Marttila et al., 2021) are rare because snow and meltwater samples in high latitude and altitude regions are difficult to obtain (Pu, Wang et al., 2020; Tetzlaff et al., 2018). Due to the limited number of samples available for isotopic analysis, accurately capturing of the gradual $^{18}$O and $^2$H enrichment of melting snowpack and meltwater can be undermined (Taylor, Feng, Williams et al., 2002; Taylor et al., 2001). However, understanding the temporal variability of snowpack and meltwater $^{18}$O and $^2$H isotopes is essential for assessing the liquid-ice fractionation that leads to the progressive enrichment of the meltwater (Ham et al., 2019; Lee et al., 2009, 2010) and affects the isotope based hydrological analysis, such as the estimation of young water fractions (Ceperley et al., 2020).

Limited knowledge is available on comparisons between snowpack and meltwater isotopes from field studies. Isotopic fractionation of early meltwater can be different from peak and post-peak meltwater isotopic composition (Schmieder et al., 2016) due to the slow sublimation caused by the relatively low ambient temperature at the onset of melting (Evans et al., 2016; Gustafson et al., 2010) and different rates of liquid-ice isotopic fractionation (Feng et al., 2002; Lee et al., 2010; Taylor, Feng, Renshaw et al., 2002). Thus, the process understanding of earlier meltwater isotopic composition as compared to peak and post-peak meltwater isotopic composition is also limited. To our knowledge, the available field studies have focused either on (a) the high-resolution temporal evolution of snowmelt water $^{18}$O and $^2$H isotopes but only based on a few snowmelt lysimeters (Pu, Wang, et al., 2020; Rücker, Zappa, et al., 2019), thus ignoring the extensive spatial isotopic variation, or (b) on numerous lysimeters in the landscape, but ignoring the temporal variability in the landscape (Laudon et al., 2002; Rücker, Boss, et al., 2019). Ala-aho et al. (2017) simulated the spatial and temporal variability in snowmelt water $^{18}$O and $^2$H isotopes. However, high-resolution field data sets of concurrent snowpack and snowmelt water $^{18}$O and $^2$H to validate such simulations are inadequate and thus, little is known about the simultaneous isotope changes and fractionation processes of the snowpack and snowmelt water in field conditions (Carroll et al., 2022; Lee et al., 2010).
To address this knowledge gap, we investigated extensive spatiotemporal snowpack and meltwater $^{18}$O and $^2$H isotopes to answer the following questions:

1. How do snowmelt water and depth-integrated snowpack $\delta^{18}$O values compare in space and time?
2. How much do snowmelt water $\delta^{18}$O values vary over the course of the snowmelt period?
3. What processes or environmental conditions are responsible for the observed $\delta^{18}$O fractionation in snowmelt water in space and time?

To answer these questions, we examine $^{18}$O and $^2$H isotope data sets of daily snowmelt water, bi-daily (2–3 days) depth-integrated snowpack, layerd-snowpack stratigraphy during snow accumulation and different melting phases, and event-based precipitation over one snow season in a subarctic research catchment.

2. Study Area, Materials, and Methods

2.1. Study Area

The Pallas catchment is located in northern Finland (68.0°N, 24.2°E) (Figure 1). It has an area of 4.42 km$^2$ and the elevation ranges from 268 to 364 m a.s.l. The infrastructure for monitoring meteorological variables has a history of over 80 years (Marttila et al., 2021), while intensive hydrological measurements have been possible since 2008. Research activities at the Pallas site are growing, particularly since 2014, when ecohydrological investigations started being conducted. Annual mean surface air temperature has been 0.4°C since 2003 and precipitation has been 639 mm since 2008; 42% of this precipitation falls as snow. The average maximum snow depth has been 105 cm since 2008, measured by an automatic ultrasound snow sensor at the Kenttärova station. The quantity of snowfall, the thickness of snow cover, and the timing of snow melting were noticeably different in 2019 and 2020, with 45% and 48% of total precipitation observed as snowfall in these years. The maximum snow depths were 100 cm (observed on 23 March 2019) and 130 cm (observed on 13 April 2020). More information on the study area is provided in Supporting Information S1 (Text S1).

2.2. Sampling Design, Field Campaigns and Data Collection

A snow survey track of 2.2 km was established in the Pallas catchment that spans from a hillslope at the Kenttärova station, and extends through open mires in a downstream direction, to the catchment’s outlet stream at the Kenttärova station (as shown in Figure 1). The Snow Depth ($S_0$) was recorded in 46 locations (every 50 m) along the snow survey transect, and the Snow Density ($S_d$) was measured in 11 locations (approximately every 200 m) near the snowmelt lysimeter locations (Figure 1). A graduated plastic snow tube of 10.3 cm diameter, weighing 1.65 kg, was mostly used for measuring snow densities, but occasionally other tubes of different sizes were also used depending on the snowpack depth. The tube was weighed using a hanging weighing scale. The Snow Water Equivalent ($SW_E$) was computed from the measured $S_0$ and $S_d$. The data sets presented in this study were collected in 2019 and 2020. Daily resolution meteorological parameters were obtained from the database of the Finnish Meteorological Institute (FMI) for the Kenttärova station (ID: 101987).

2.2.1. Snowmelt Water $^{18}$O and $^2$H Isotope Sampling and Melt Efflux From Snowmelt Lysimeters

We designed, deployed, and installed a system of snowmelt lysimeters at 11 locations in the Pallas catchment (Figure 1 and Figure S1 in Supporting Information S1) across different landscape features: (a) along the catchment's transect on a forest hillslope containing mainly evergreen spruce trees (L1, L2, L3, L4, and L5), (b) in a mixed forest area containing evergreen and deciduous forest (L7, L8, L9, and L11), and (c) on peatlands in open mires (L6 and L10). The design of the snowmelt lysimeter is described in detail in Supporting Information S1 (Text S2). Snowmelt water samples were collected daily in plastic bags at 200 m intervals along the catchment’s transect as outflows from snowmelt lysimeters. Snowmelt efflux (mm/d) was computed from the weight of the meltwater sample divided by the density of water and cross-sectional area of the snowmelt lysimeter during the elapsed time between the two consecutive sampling rounds. The collected samples were analyzed for snowmelt water isotopic composition.

2.2.2. Stratigraphy Snow and Bulk Snow Sampling From Snowpack

Snowpits were excavated before the snowmelt period from the 2 to 6 April 2020 at 11 locations along the snow survey transect, adjacent to the locations of lysimeters (Figure 1). Snowpits at three locations were also excavated...
These three sampling sites were located (a) near L1 on a hillslope at the Kentarova station, (b) near L7 in the middle of the transect, and (c) located close to the catchment's outlet stream. From the snowpits, snow samples at a fixed vertical interval of 5 cm were taken to analyze their isotopic composition. Vertical depth-integrated bulk snow samples were also extracted using a plastic tube of 3.5 cm diameter at each lysimeter location (i.e., at 200 m intervals). The depth-integrated snowpack samples were collected biweekly during the winter seasons of 2019 and 2020. Such samples were occasionally collected during the snowmelt period in 2019 but were sampled every two to three days in 2020. The snow density and snow water equivalent from snow depths and weights were computed for each lysimeter's location. Details on precipitation sampling are provided in Marttila et al. (2021).
For vertically layered-snowpack isotope stratigraphy, the sampling depth was normalized to a scale of 0–1 by dividing it by the total snow depth monitored at each snowpit location in order to make the snowpit isotope data more readily comparable. For example, if the total snowpit depth is 100 cm and the 5 cm snow layer sample from 0 to 5 cm depth from the base (soil-snow interface) is analyzed, then the normalized depth for that particular layer will be 2.5/100, that is, equal to 0.025. The normalized depth for the given precipitation amount was obtained by dividing the cumulative precipitation amounts by the total precipitation amount that had fallen in the snow accumulation phase (i.e., until the day when maximum snow depth was observed).

### 2.3. Isotope Analysis

The snowpack samples were stored in airtight sealed plastic bags in-situ, which were later transferred to the GeoLab at the University of Oulu and stored in either 15 or 50 ml plastic tubes after completely melting at room temperature. Eventually, all samples (including snowmelt water) were refrigerated at 4°C before being analyzed for their isotopic composition. Samples were analyzed using a laser-based Cavity Ring-Down Spectroscopy (CRDS) Picarro system, consisting of an isotope analyzer (L2140-i), autosampler (A0325), and vaporizer (A0211) unit. More information is available in Supporting Information S1 (Text S3). Isotope ratios (values) of all samples were determined relative to the VSMOW (Vienna Standard Mean Ocean Water) international standard, and expressed as per mil, that is, ‰ (Gonfiantini, 1978). The secondary parameter line conditioned excess, that is, lc-excess (similar to deuterium excess, d-excess) was calculated from the equation of Local Meteoric Water Line, that is, \( \delta^2H - 7.68 \times \delta^{18}O \), where the slope is similar to the Global Meteoric Water Line (GMWL), that is, \( \delta^2H - 8 \times \delta^{18}O \) (Dansgaard, 1964). The slope of LMWL is based on snowfall precipitation that occurred in 2019 and 2020. The line-conditioned excess (lc-excess) values were calculated using the following equation (Landwehr & Coplen, 2004),

\[
\text{lc-excess} = \delta^2H - a \times \delta^{18}O - b
\]

where \( a \) is the slope of LMWL and \( b \) is the intercept of LMWL.

In the results and discussion sections, we have focused on lc-excess instead of d-excess, as this represents the deviation of each sample’s \( \delta^2H \) values from the LMWL. Deviation from a LMWL suggests non-equilibrium kinetic fractionation. If a sample lies on a LMWL, then the lc-excess will be \( \sim 0 \). Positive values show that the sample lies above the LMWL, and negative values show that the sample lies below the LMWL (Landwehr & Coplen, 2004). We used the flow-weighted mean for snowmelt water \( \delta^{18}O \) and \( \delta^2H \) values and the density-weighted mean for depth-integrated snowpack isotope values. Hereafter, the depth-integrated snowpack isotopes are described as snowpack isotopes unless otherwise specified, such as when described as layered-snowpack isotopes in order to avoid repetition of the “depth-integrated” term. We computed the flow-weighted mean of meltwater \( \delta_m^{18}O \) values using the following equation,

\[
\delta_m^{18}O = \frac{\sum_{i=1}^{n} F_{vol,i} \cdot \delta_i^{18}O_i}{\sum_{i=1}^{n} F_{vol,i}}
\]

where \( \delta_m^{18}O_i \) is the weighted means of snowmelt \( \delta^{18}O \) values, \( F_{vol,i} \) is the snowmelt efflux, \( i \) shows the \( i \)th number and \( n \) is the total number of samples. All isotope values are in per mil units (%e).

Similarly, we computed the density-weighted mean of snowpack isotope values by using the equation,

\[
\delta_s^{18}O = \frac{\sum_{i=1}^{n} S_d \cdot \delta_i^{18}O_i}{\sum_{i=1}^{n} S_d}
\]

where \( \delta_s^{18}O_i \) is the weighted means of snowpack \( \delta^{18}O \) values, and \( S_d \) is the snowpack density.
More information on how weighted means were computed for each landscape feature is available in Supporting Information S1 (Text S4).

3. Results

3.1. Comparison of Snowmelt Water and Snowpack $^{18}$O and $^2$H Isotopic Composition

We compared the snowmelt water $^{18}$O and $^2$H isotopes with snowpack $^{18}$O and $^2$H isotopes at a catchment scale (as mentioned previously, snowpack will refer to the depth-integrated snowpack hereafter). We found that the flow-weighted mean of snowmelt water $\delta^{18}$O and $\delta^2$H values was consistently higher than the density-weighted mean of snowpack $\delta^{18}$O and $\delta^2$H values in both years (see the weighted mean as a yellow circle and the standard error as red error bars in boxplots of Figure 2 and Figure S2 in Supporting Information S1 and Table S1). The weighted mean of snowmelt water $\delta^{18}$O values was higher in comparison to snowpack isotope values in both years, for example, it was higher by 1.98 ± 0.06 and 0.46 ± 0.24‰ $\delta^{18}$O in the forest hillslope and 2.33 ± 0.22 and 0.77 ± 0.27‰ $\delta^{18}$O in the mixed forest in 2019 and 2020, respectively. The highest variation was found in open mires, that is, the weighted mean of snowmelt water $\delta^{18}$O values was higher by 6.19 ± 1.64 and 1.51 ± 0.21‰ $\delta^{18}$O than the snowpack $\delta^{18}$O values. The most enriched weighted mean of snowmelt water $^{18}$O isotopes was observed in 2019 (Figure S2 in Supporting Information S1). The weighted mean and variation in ranges of snowpack $\delta^{18}$O and $\delta^2$H values were more consistent than snowmelt water $\delta^{18}$O and $\delta^2$H values, except for in open mires. Figure 2 and Table S1 provide a comparison between the $\delta^{18}$O and $\delta^2$H values of snowmelt water and snowpack. Figures for the 2019 data are provided in Supporting Information S1.

Figure 2. Dual isotope plots of snowmelt water (circles) and snowpack (triangles) for the year 2020, representing $\delta^{18}$O and $\delta^2$H values in different landscape features: (i) forest hillslope, (ii) mixed forest, and (iii) mire, along with boxplots, where weighted mean values are represented by yellow circles and error bars by red lines. The GMWL and LMWL are represented by solid and dotted black lines, respectively.
The LMWL $\delta^2H = 7.68^18O + 2.49$ was determined based on the snowfall $\delta^18O$ and $\delta^2H$ values only (excluding rainfall isotope values) from 2019 to the winter of 2020, as shown in Figure 2 and Figure S2 in Supporting Information S1. The dual-isotope plots show that most values for snowmelt water and snowpack $^18O$ and $^2H$ isotopes lie close to the LMWL, $\delta^2H = 7.68^18O + 2.49$ in 2019 and 2020.

The time-series plots of snowpack and snowmelt water $\delta^18O$ values (Figure 3) show that the temporal variation in the snowmelt water $\delta^18O$ values was greater compared to the snowpack $\delta^18O$ values. Furthermore, the comparison of snowmelt water and snowpack $\delta^18O$ values indicates that the snowmelt water $\delta^18O$ values were spatially more variable than the snowpack $\delta^18O$ values within the same landscape feature on most dates (see error bars in Figure 3). The spatial variation of snowpack $\delta^18O$ values was highest in open mires. Figure 3 shows that the snowmelt water $^18O$ isotopes were significantly depleted relative to the snowpack $^18O$ isotopes between 21 April and 21 May 2020, the difference being close to the equilibrium difference of 3.1‰ $\delta^18O$ between ice and liquid water (O’Neil, 1968). From 22 May to 6 June 2020, the snowmelt water $\delta^18O$ values were close to, and slightly higher than the snowpack $\delta^18O$ values. Around 21 May, the snowmelt water $\delta^18O$ values increased by 3.64 ± 0.01‰ $\delta^18O$ in the forest hillslope, 3.09 ± 0.08‰ $\delta^18O$ in the mixed forest, and 1.29 ± 0.33‰ $\delta^18O$ in open mires. More comparisons between snowmelt water and snowpack $\delta^18O$ values for both years (2019 and 2020) are provided in Tables S2, S3, and S4 and Text S5 in Supporting Information S1. Figure 3b shows that lc-excess values for snowpack were close to 0 in all three landscape features, while lc-excess values for snowmelt water exhibited relatively greater variation. Similar plots for 2019 isotope data sets are provided in Figure S3 in Supporting Information S1.

### 3.2. Snowmelt Water $^18O$ Isotopes Influenced by Snowmelt Rate

The snowmelt rate or snowmelt efflux magnitude was low between 21st April and 21st May in 2020, however, the meltwater volume was enough to sample for later isotope analysis (as shown in Figure 4). The snowmelt rate started to increase from mid-May, with a step-change to high melt rate around 22 May 2020. The snowmelt rate is primarily dependent on catchment topography, land cover, and meteorological conditions. Daily average air temperature was consistently above zero from around 22 May (Figure 4). The accelerated melt can also be observed from the rapidly declining snow depth (Figure 4). The snowmelt was relatively delayed and the timing of snowmelt was greater in 2020 compared with that of 2019 as there was a higher maximum snow depth in 2020 and the melting rate was steady (Figure 4 and Figure S4 in Supporting Information S1). The maximum snowmelt efflux was ~27 mm/day and ~59 mm/day in 2019 and 2020, respectively (Table S5 in Supporting Information S1).

More information on snowmelt efflux is provided in Supporting Information S1 (Text S6). Snowmelt efflux was spatially variable from day-to-day, even within the same landscape features (Figure 4). In contrast, snowmelt water $\delta^18O$ values were relatively less spatially variable (Figure 5).

Generally, the snowmelt with lower magnitude efflux had lower $\delta^18O$ values, and higher magnitude efflux had relatively higher isotope values (Figures 3 and 5). Therefore, the time-series snowmelt water $^18O$ isotope plots in three spatial locations: (a) forest hillslope, (b) mixed forest, and (c) open mires show that the initial isotope values were lower, representing the depletion of heavier isotopes in early snowmelt samples (Figure 5). The isotope values in the early snowmelt season remained relatively consistent, with a clear shift toward enriched values that correspond to the rapid increase in snowmelt efflux (22 May 2020 onwards). The isotope values covaried temporally in different landscape features (Figure 5). The range of variation in isotope values became smaller during the post peak melting period (see values after May 22 in Figure 5). The high-resolution snowmelt water $^18O$ isotope data also revealed that the initial isotope values were not the lowest values; in fact, the lowest values were observed a few days after the onset of the snowmelt period (see snowmelt water $\delta^18O$ values from May 5 to May 11 in Figure 5, where samples showed the lowest isotope values). The time-series snowpack isotope plots for 2019 and 2020 are provided in Figure S6 in Supporting Information S1.

The lc-excess values of snowmelt water were greater than or close to 0 in 2020 (as shown in the lower panel in Figure 5), and negative values were observed during and after the peak melt period only. Similarly to $\delta^18O$ values, there was considerable co-variation in lc-excess values between lysimeters. The highest lc-excess values were observed just prior to the peak melt period. In 2019, the lc-excess values of snowmelt water did not clearly co-vary between lysimeters (Figure S5 bottom panel in Supporting Information S1).
The comparison of the $\delta^{18}O$ values and lc-excess values of snowmelt water and snowpack shows how the $\delta^{18}O$ values are concentrated into two distinct clusters (represented by ellipses), with different snowmelt rates and air temperatures (Figure 6). In lc-excess plots, values are more dispersed. Whereas generally, the days with low efflux and near-zero air temperatures represent more positive lc-excess values in forest areas. The snowmelt $\delta^{18}O$
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and lc-excess values in particular exhibit a non-linear relationship with the environmental variables (see Figures S8 and S9 for 2019 and Figures S10 and S11 for 2020 in Supporting Information S1). The results for the year 2019 are shown in Figure S7 in Supporting Information S1.

To confirm the presence of two separate clusters statistically, that is, those represented by two separate ellipses for the dense forest areas, we used a \( k \)-means unsupervised clustering machine learning algorithm to observe any possible clustering of snowmelt water and snowpack \( ^{18} \)O isotopic comparison (we chose a \( k \)-value of 2). The algorithm is widely used for finding homogeneous groups (clusters) of compared values in data sets (Likas et al., 2003; Shi et al., 2010). We used an R-stats package to employ a \( k \)-means clustering algorithm in order to find the possible grouping of compared values of snowmelt water and snowpack. The package uses an algorithm presented by Hartigan and Wong (1979). The two clusters (represented by ellipses) are separated from each other in the forest and mixed forest areas but are slightly overlapped in open mires. When the two ellipses for the snowmelt water and snowpack \( ^{18} \)O isotopic relationship were analyzed visually (in Figure 6a) and quantitatively with the aid of other hydrometeorological data (i.e., flow (mm/d) and air temperature (°C)), one ellipse belongs to the isotopes prior to the peak melting, while the other belongs to peak and post-peak melting. This shows the different relationships between snowmelt water and snowpack \( \delta^{18} \)O values prior to peak melting, as well as during and post-peak melting.

**Figure 4.** Plots of different hydrometeorological variables for the year 2020, where (a) shows air temperature (orange line) and precipitation (bars/columns), (b) shows snow depth at Kenttärova station (yellow area) and snow depth for different landscape features along the snow survey transect (points), and (c) shows snowmelt efflux and snowmelt water temperature.
3.3. Spatiotemporal Variability of Isotope Values in Layered Snowpack Stratigraphy

The spatial variability and temporal evolution of isotope values in snowpack stratigraphy are presented in Figures 7 and 8. Before the snowmelt season in early April, consistent patterns were observed in both the snow isotope stratigraphy and density in all 11 locations across the catchment (Figure 7a). The base of the snowpack consisted of moderate-density snow, enriched in heavy isotopes. Maximum snow densities were observed at the normalized snow depth of 0.2–0.4 from the base of the snowpack. The snow layers at such depths were depleted in $^{18}$O isotopes. The surface layers represented the low snow densities typical for any recent snowfall events. Isotope values at the surface were also more negative in most profiles, suggesting a snowfall event that was depleted in $^{18}$O isotopes.

The snowpit sampling at different stages of snowmelt shows a homogenization of both snow isotope stratigraphy and density in the snowpack vertical profiles over the course of the snowmelt period (Figure 7b). Densities reached a fairly constant value ($\sim$400–450 kg/m$^3$) across the snowpack vertical profiles. As vertical variability in the isotope values started to reduce (i.e., the isotope values became more consistent) from the beginning to the end of peak melting, their average isotope values (represented by the dotted line in Figure 7) became higher, showing the enrichment of the snowpack in $^{18}$O isotopes as the snowmelt period progressed. What is notable is that the depleted $^{18}$O isotope surface layer that had been present in pits 1 and 11 before the snowmelt (Figure 7a) had mostly disappeared during the snowmelt.

Snowfall isotope values are plotted as step lines in Figure 8, corresponding to the normalized cumulative precipitation amounts on a scale of 0–1 and ranged from $-30.62$ to $-7.72$‰. The enriched early season snowfall events with high $\delta^{18}$O values can be observed at the base of the snowfall profiles (Figure 8). The most depleted snowfall events with low $\delta^{18}$O values can be seen between 0.1 and 0.4 normalized depths, and gradually enriched snowfall events between 0.5 and 0.9 normalized depths but with reduced variabilities. The top of the snowpack isotope profiles (0.9–1.0) represents the most depleted values. The patterns of isotope values generally remained...
Figure 6. The relationship between (a) snowmelt water δ¹⁸O values (‰) and (b) snowmelt water lc-excess (‰) with snowpack δ¹⁸O values (‰) and snowpack lc-excess (‰) for the year 2020, where the size of the circles represents the corresponding amount of measured snowmelt efflux (mm/d), and the color scale represents the average of the current and previous day's recorded air temperature. The ellipses around points (circles) in subplot (a) represent the distinction between the two dominant groups, estimated using a k-means clustering algorithm.
similar at the top of the hillslope, the middle of the catchment's transect and near the outlet stream’s location over the course of the snowmelt period, thus representing spatial coherence of vertical isotope profiles. During the maximum snow accumulation period (early April), the vertical isotope profiles exhibited the maximum isotopic variation, while at the post-peak melt period, less variation and more consistency was observed in the vertical isotope profiles. The ranges of layered-snowpack isotope stratigraphy, densities and $\Delta S\text{WE}S$ at different times are provided in Tables S6 and S7 in Supporting Information S1.

4. Discussion

4.1. Comparison of Spatial Snowmelt Water and Snowpack $^{18}$O and $^2$H Isotope Composition in Different Landscapes

We compared spatiotemporal changes in snowpack and snowmelt water $\delta^{18}$O and $\delta^2$H values for different landscape features, including forests and adjacent mires. We found that the snowmelt efflux was isotopically heavier compared to the snowpack across the Pallas catchment in Northern Finland. The snowmelt water $\delta^{18}$O and $\delta^2$H values also exhibited greater variations as compared to the snowpack $\delta^{18}$O and $\delta^2$H values within the divergent landscape features (see the ranges in Table S1 and the comparison of isotope variation in Figure 3). In 2019, the weighted mean of the snowpack isotope values in the forest hillslope was higher compared to that in open mires; the difference being $1.84 \pm 0.23\%e\delta^{18}$O and $11.28 \pm 0.86\%e\delta^2$H. It was slightly lower in 2020; the difference being $-0.03 \pm 0.05\%e\delta^{18}$O and $-0.14 \pm 0.34\%e\delta^2$H. A study by von Freyberg et al. (2020) found that the snowpack isotope values were higher in the forest compared with the open grasslands by $2.3\%e\delta^{18}$O and $13.4\%e\delta^2$H, likely due to isotope enrichment caused by canopy interception. However, unlike von Freyberg et al. (2020), our study was not designed to study the influence of interception, and we did not investigate the isotopic composition of throughfall to explain the influence of canopy interception. Koeniger et al. (2008) inferred from snow and streamflow isotope results that snow and streamflow were isotopically heavier in dense forest and partially cut forest than in the clear-cut forest. Our findings did not reveal a notable difference in the 2020 data, as presented by Koeniger et al. (2008). This might be attributed to the lack of a direct analysis of snowmelt water $^{18}$O and $^2$H isotopes, instead inferring from streamflow isotope that the snowmelt water might be isotopically heavier in the dense forest and partially cut forest than the snowmelt water $^{18}$O and $^2$H isotopes in the clear-cut forest. It
is important to note here that our landscape consists of open mires (wetlands). Snowmelt water sampling from snowmelt lysimeters was challenging, particularly during the spring flood episodes, where the snowmelt water in our lysimeters may have mixed with the water in the waterlogged mires. The old snowmelt water may affect the snowpack in mires through capillary rise, and the observed higher snowmelt water $\delta^{18}O$ and $\delta^2H$ values in mires may be partly attributed to its mixing with pre-event mire water.

We did not find elevation effects on changing isotope values in the snowpack and snowmelt water as the elevation range of the Pallas catchment was less than 100 m at the point where the measurements were made (Table S8). The elevation effects are typically significant in mountainous regions (Niewodnizański et al., 1981).

4.2. Temporal Evolution of Snowmelt Water and Snowpack $\delta^{18}O$ Isotope Values

Our results depict how the temporal variability of snowmelt water $\delta^{18}O$ values was much higher than the spatial isotopic variability along the 2.2 km catchment’s transect (Figure 5). Laudon et al. (2007) have used three replicates of snowmelt lysimeters in closed canopy, open canopy, and open field, and found that the spatial isotopic variability was less pronounced than the temporal variability between the three sites. The low (depleted) $\delta^{18}O$
values in earlier snowmelt water relative to the snowpack $\delta^{18}O$ values were an indication of the lighter water molecules melting out the snowpack (Feng et al., 2002; Taylor et al., 2001). Snowpack samples represented relatively constant isotope values compared with snowmelt water $\delta^{18}O$ values. These samples, however, also exhibited a subtle, gradual enrichment over the melt season (Figure 3). The change in snowmelt water $\delta^{18}O$ values, from the initial melt to the last melt events, was 7.25 and $4.17\%e$ $\delta^{18}O$ in the forest hillslope and 6.03 and $5.91\%e$ $\delta^{18}O$ in the mixed forest areas in 2019 and 2020, respectively (Figure 3), which is in agreement with snowmelt water $^{18}O$ and $^2H$ enrichment in previous laboratory (Feng et al., 2002; Taylor et al., 2001; Taylor, Feng, Renshaw, et al., 2002), and field studies (Laudon et al., 2002; Lee et al., 2010; Taylor et al., 2001). The temporal evolution from low (depleted) to high (enriched) isotope values during peak melt was mostly observed in forest and mixed forest areas, but was not as significant in open mires. Perhaps the trend in open mires could have been clearer, but sampling challenges in open mires might have affected the isotope values.

The statistical analysis of temporal trends shows that the trend of the snowmelt water $\delta^{18}O$ values was non-linear, while the trend of the snowpack $\delta^{18}O$ values was relatively linear in forest and mixed forest areas. However, if the residuals of the linearity assumptions are observed, there is an increasing non-linear trend of snowpack $\delta^{18}O$ values during the end phase of the snowmelt period in the two landscape features in 2020. These trends can be visually observed in Figures 3 and 5, Figures S3, and S5 in Supporting Information S1. More details and statistics on these trends are presented in Text S7 and Table S9 in Supporting Information S1.

Importantly, we discovered that snowmelt water $\delta^{18}O$ values covary temporally at several locations within the forest (represented by green and blue bubbles in Figure 5); however, isotope values do not strongly covary between forest areas and open mires. This indicates that a single sampling location may yield representative snowmelt water $^{18}O$ and $^2H$ isotope data for a given landscape unit or elevation band. The weighted mean of snowmelt water $\delta^{18}O$ and $\delta^2H$ values may be sufficient for simple and linear isotope-based mixing models to give a rough estimation of contributing endmembers (sources), but care must be taken when considering areas with both openings and dense forest. Open areas should be treated differently from dense forests in source water mixing models. The pattern of daily snowmelt water $\delta^{18}O$ values in different landscapes (represented by different colors in Figure 5 and Figure S5 in Supporting Information S1) was more consistent in 2020 compared to 2019. This means that the snowmelt isotope values may vary from season to season.

### 4.3. Snowmelt Rates Influence the Temporal Changes in Snowmelt Water $\delta^{18}O$ Isotope Values

Theoretical work simulating liquid-ice isotopic fractionation during snowmelt has suggested that low snowmelt rates may produce detectable differences in isotope values between early and late meltwater (Feng et al., 2002; Lee et al., 2009; Taylor et al., 2001; Taylor, Feng, Renshaw, et al., 2002). The decreasing slope of $\delta^2H$ versus $\delta^{18}O$ regression of meltwater can also be attributed to the liquid-ice isotopic fractionation, discussed by Lee et al. (2010). Results from our field data of snowpack and snowmelt isotope samples indicate that the relationship between snowmelt water and snowpack $^{18}O$ isotopes was different prior to peak melting, when melt efflux was smaller, compared with during peak melting, when melt efflux was greater. This is evident in Figure 6, where the two clusters separated by ellipses confirmed that the relationship between the two was different prior to the peak, and during and post peak melting. When environmental variables, such as air temperature, and snowmelt efflux were analyzed in conjunction with the relationship between snowpack and snowmelt water $\delta^{18}O$ values, we found a non-linear relationship between air temperature and snowmelt water $\delta^{18}O$ values, as shown in Figure 6 and Figure S11 in Supporting Information S1. Environmental variables influencing snowmelt dynamics, such as surface air temperature, net energy influxes, and other meteorological variables, are critical in assessing snowmelt water and snowpack spatiotemporal $^{18}O$ isotopic evolution. Snowmelt water $^{18}O$ and $^2H$ isotope dynamics can be different between the prior peak and post peak melt periods. This will have a strong impact on accurate evaluation of the isotope-based hydrograph separation (Taylor, Feng, Williams, et al., 2002).

Based on the clustering of two pools of snowmelt water $^{18}O$ isotopes in Figure 6, we selected the peak melt period as 21 May and computed regression slopes for the period prior to peak melt, and for the post peak melt period (Figure S12 in Supporting Information S1). All slopes are lower than the slope of the LMWL equation (i.e., 7.68). Slopes prior to the peak melt period (i.e., 6.43) and post peak melt period (i.e., 6.42) are similar in mixed forest areas, while the slope for the peak melt period (i.e., 5.73) is smaller/lower than the pre peak melt period (i.e., 7.17) in open mires. Conversely, the slope of the pre peak melt period (i.e., 4.91) is smaller than the post peak melt period (i.e., 6.44) in the forest hillslope. This smaller slope does not indicate evaporation.
effects as early snowmelt samples are plotted above the LMWL (Figure 2). The smaller slope may be caused by the kinetic liquid-ice fractionation which occurs during low melt episodes, indicated by higher lc-excess values in Figures 3 and 4.

Here, it should be noted that our study was not designed to investigate the sublimation and condensation processes that occur at snow and air interface and where the role of vapor isotopes, snow surface temperature, air temperature, wind speed, wind direction and turbulent humidity flux at different depths in the atmosphere is key to understanding the exchange of isotopes between snow surface and atmosphere through sublimation and condensation processes (Steen-Larsen et al., 2014, 2017; Wahl et al., 2021).

4.4. Earlier Snowmelt Water’s Depletion in ¹⁸O Isotopes

It has been well documented in the literature that the first pulse of snowmelt water exhibits low δ¹⁸O and δ²H values, which become progressively higher over the course of the snowmelt period (Langman et al., 2022; Laudon et al., 2002; Lee et al., 2010; Stichler, 1987; Taylor, Feng, Renshaw, et al., 2002; Taylor et al., 2001). The reason for this is often due to the occurrence of liquid-ice fractionation, causing melting of depleted water first and leaving behind enriched snowpack. This snowpack further enriches during the later stages of snowmelt due to isotopic fractionation by evaporation/sublimation (von Freyberg et al., 2020; Laudon et al., 2002; Lee et al., 2010; Taylor et al., 2001). In our study, snowmelt water and snowpack ¹⁸O isotope data sets in forest hillslope and mixed forest revealed three important processes in isotopic evolution (Figures 3 and 5): (a) the offset or clear distinction between snowmelt water and snowpack δ¹⁸O values during an earlier stage of snowmelt with low melt rates (Figure 3a); (b) the variability in lc-excess values during the earlier stage of the snowmelt with low melt rates (Figures 3b and 5); and (c) the lack of difference between snowmelt water and snowpack δ¹⁸O values in the late stage of snowmelt (Figure 3a).

We observed an apparent difference between the snowmelt water, and snowpack δ¹⁸O values during the initial melt or prior to the peak melt period. The offset between the two states of water was close to the offset of 3.1‰ δ¹⁸O between ice and liquid water in an equilibrium state (O’Neil, 1968). Meltwater samples at some locations were so depleted in heavy isotopes that their values were even lower (see Figure 2 from 6 May to 11 May 2020) than the most depleted (lowest) values observed in the vertical layered snowpack isotope profiles (vertical layered-snowpack isotope profiles in plot (a) in Figure 8). It should also be noted that most depleted values in Figure 5 are not the first melt values; they are instead the isotope values from 6 May to 11 May, coinciding with the slightly elevated melt rates during the early melt season (Figure 6). The first values at the start of the snowmelt period were slightly higher in ¹⁸O and ²H values. This was in line with the findings of Evans et al. (2016) and Lee et al. (2010), who found that the melting of top layers may directly contribute to the final observed snowmelt water δ¹⁸O values at the snowpack’s base. This means that the percolation of meltwater might not reach the liquid-ice equilibrium at the very beginning of the snowmelt period due to the entire surface layer melting and reaching the snowpack’s base through preferential flow paths or lateral flow (Evans et al., 2016). After the first samples and during the following melt episodes, the low volume of percolating meltwater prior to the peak melt period might have interacted with the solid-state of the snowpack. This may have caused isotopic exchange between the two states, wherein the lighter isotopes were preferentially present in the liquid state. Figure 3a shows how the difference between snowmelt water and snowpack δ¹⁸O values was not consistently 3.1‰ δ¹⁸O; small fluctuations were observed. This means that the liquid-ice interaction fluctuated between the equilibrium and non-equilibrium states. The ranges of snowmelt water δ¹⁸O values and vertical snowpack isotope profiles are provided in Tables S6 and S7 in Supporting Information S1.

The lc-excess values of snowmelt water prior to the peak melt period varied between values of 0 and 10 (Figures 3 and 5). In the dual-isotope space (i.e., δ²H vs. δ¹⁸O), this is represented by a slope lower than the LWML (Figure S12 in Supporting Information S1). The lc-excess values closer to 0 were indication that the equilibrium state during liquid-ice fractionation had been achieved. Larger lc-excess values were an indication of a move away from the equilibrium state, where kinetic fractionation remained a dominant process. The lc-excess was highest from 6–11 May prior to the peak melt period, reaching a value of 10‰. This was still at the melting phase with low melt efflux (Figure 5) when the vertical isotopic stratigraphy had not become significantly consistent and homogenized, compared to the isotopic stratigraphy during the peak and post peak melt period (plot (b) in Figure 7 and plots (c) and (d) in Figure 8). We suggest that the high lc-excess values prior to the peak melt period (with snowpack-snowmelt difference in δ¹⁸O being close to 3.1‰) indicated the time-variant kinetic liquid-ice
fractionation. The snowmelt water samples represent liquid-ice kinetic fractionation processes, while such fractionation effects were not indicated by the snowpack samples (Figure 3). Thus, composite snowpack samples alone may not clearly explain kinetic fractionation processes, as the exact reasons and processes governing the variation of ic-excess values are not clear. The co-variability in the snowmelt ic-excess values at the catchment scale do, however, point toward environmental conditions and snowmelt rates being key drivers of change in ic-excess values. Future studies focusing on high-resolution (preferably sub-daily resolution or dial cycles) numerical modeling of liquid-ice fractionation of $^2$H and $^{18}$O isotopes may provide answers for the temporal variation of the kinetic liquid-ice fractionation process.

The difference between the snowmelt water and snowpack isotope values, both $\delta^{18}$O and ic-excess values, started to disappear at the onset of the peak melt period, when the temperature started to rise, and the input energy fluxes to the snowpack started to mount. This magnified the snowmelt rate and the speed at which meltwater percolated through the snowpack. The isotope values at different snow depths tended to be homogenized at the initiation of the peak melt period. The relatively homogenized and consistent isotope values at the onset of the peak melt period have been observed in Figures 7 and 8. The consistency of the snowpack isotopic composition at different depths during the post peak melt period might be an indication of the diffuse movement of meltwater within the snowpack, caused by the widespread mass movement of percolated meltwater (Evans et al., 2016). The minor interaction between the percolated meltwater and the solid ice might have precluded isotopic exchange during liquid-ice equilibration or liquid-ice fractionation, as has been suggested before (Feng et al., 2002; Lee et al., 2009, 2010; Taylor, Feng, Renshaw, et al., 2002; Taylor et al., 2001). Thus, the resultant snowmelt water and snowpack showed similar isotopic compositions.

The depleted (lowest) value of early snowmelt water may occur either due to (a) melting of only fresh depleted snowfall events from the surface layers that percolate down, via by-passing flow mechanism (i.e., flowing through preferential flow paths or lateral flow) (Evans et al., 2016); (b) significant liquid-ice fractionation of percolating meltwater through the snowpack (Feng et al., 2002; Lee et al., 2009, 2010; Taylor, Feng, Renshaw, et al., 2002; Taylor et al., 2001); or (c) a combination of the two processes (Pu, Wang, et al., 2020). The consistency of the $\sim$3.1‰ offset between the snowpack and the snowmelt water throughout the early melt period (Figure 3) does not indicate that individual surface layers melted from the top and bypassed the snowpack through preferential flow or lateral flow. Therefore, the preferential flow of the surface layer meltwater is not likely to be the dominant reason for the temporal isotopic evolution seen in snowmelt water, though the vertical variability in the snow profile may influence temporal isotopic variability (Pu, Wang, et al., 2020). Based on our data, we suggest that liquid-ice fractionation during low melt rates was the primary reason for early depleted values, whereas increasing melt rates and vertical homogenization in the snow profiles might have resulted in less liquid-ice fractionation, meaning enriched snowmelt water resembled the snowpack isotopically.

5. Conclusions

This study provided a detailed characterization of $\delta^{18}$O and $\delta^2$H isotope values in (a) vertical snowpack layers displaying snowpack stratigraphy, (b) depth-integrated snowpack, and (c) snowmelt water for three divergent landscape features in space and time. We used 11 snowmelt lysimeters for different landscape features that were complemented by snowmelt efflux, measured at 200 m intervals along the catchment’s transect, and showed how snowpack and snowmelt water $\delta^{18}$O and $\delta^2$H values varied from the winter to spring seasons. Earlier snowmelt water was isotopically lighter compared to the depth-integrated snowpack, while the peak snowmelt water was isotopically heavier compared to the depth-integrated snowpack. The earlier snowmelt water at some locations had even lower $\delta^{18}$O values compared to the vertical isotope profiles of layered-snowpack stratigraphy. The temporal trend of the snowpack isotope values from low (depleted) to high (enriched) values was relatively linear during the snowmelt period in 2020, while the temporal trend of the snowmelt water $\delta^{18}$O values was highly non-linear. Our findings support the idea that estimating the spatiotemporal evolution of snowmelt water $\delta^{18}$O values based on snowpack $\delta^{18}$O values is not feasible without a detailed process understanding of liquid-ice isotope fractionation. Furthermore, our data indicates that liquid-ice fractionation processes (kinetic and equilibrium) occurring within the snowpack controlled the isotopic composition of the subsequent snowmelt water, and such processes were governed by snowmelt rate. Liquid-ice isotopic fractionation was apparent during periods of low melt, but became insignificant during high melt rates.
The analysis of extensive data sets of snowpack and snowmelt water $^{18}$O and $^2$H isotopes yielded new insights and baseline information for future isotope hydrology research in cold regions. We encourage the use of snowmelt water $^{18}$O and $^2$H isotopes as tools for ecohydrological and other tracer-based hydrological investigations in snow-influenced regions, provided that (a) snowmelt water is collected in-situ, or (b) liquid-ice fractionation processes during the snow season are adequately understood when estimating snowmelt $^{18}$O and $^2$H isotopes. Accounting for differences in landscape features is needed to encompass the temporal and spatial variability of snowmelt $^{18}$O and $^2$H isotopes.

Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

Data Availability Statement

The meteorological and daily snow depth data at the Kenttarova station, used in this study, were downloaded from the Finnish Meteorological Institute's database (https://en ilmaiteenlaitos.fi/download-observations#!/). Isotope data, used in this study, are available on a web repository accessible from https://doi.org/10.23729/c2d0ade1-cb3d-4dea-a48b-dbb5654bb662. The isotope data are available on the web repository accessed from https://agupubs.onlinelibrary.wiley.com/doi/10.1029/2022WR033101 by University Of Oulu. For full terms of use, visit https://agupubs.onlinelibrary.wiley.com/doi/terms.

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References


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