CLIMATOLOGICAL AND HYDROLOGICAL INFLUENCES ON STABLE HYDROGEN AND OXYGEN ISOTOPOES OF ACTIVE LAYER WATERS, LEVINSON-LESSING LAKE AREA, TAYMYR PENINSULA

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Abstract

Stable isotope data (δ18O, δD) of precipitation and active layer waters were collected in the Levinson-Lessing Lake catchment, Taymyr, Siberia during 1994 and 1995. They are used here in conjunction with hydrological and microclimate data to provide information on water-forming processes in the active layer. Summer precipitation defines a local meteoric water line different from the global meteoric water line, suggesting that the source of summer precipitation has been exposed to evaporative isotopic enrichment. The stable isotope composition of active layer waters indicate that they are mostly fed by summer rain. Small isotopic variabilities between sites in the catchment indicate differences in microclimatic and hydrologic characteristics. Furthermore, the isotope ratios have been successfully applied to differentiate between sources of active layer waters (snow and ground ice melt, rain water). Further studies are required to develop universally applicable models that describe isotopic behaviour in the active layer in these areas.

Introduction

Information on water sources and pathways is often required to accurately assess active layer hydrology. Natural abundance variations of the rare, heavy stable isotopes 18O and D in water have been shown to be useful hydrological tracers (Clark and Fritz, 1997) and have been applied in a variety of studies in the Arctic (Bursey et al., 1991; Cooper et al., 1993). Gibson et al. (1993, 1996) developed an isotopic method for calculation of lake evaporation in the continental Arctic in Canada. A good agreement was found between the evaporation rates calculated with the isotopic mass approach and evaporation rates calculated using standard methods (water and energy balance, aerodynamic profile, class A pan). Furthermore, any significant climatic variation results in changes of the 18O and D content of meteoric waters (Clark and Fritz, 1997). Paleowaters therefore reflect paleoenvironmental conditions. Within the multidisciplinary project on ‘Late Quaternary environmental history of Central Siberia’ a program of detailed hydrological investigations was undertaken to establish the 18O and D content of recent waters as a reference for the interpretation of paleowaters. This is of special interest in this region where paleo data archives such as ice cores are absent.

For the correct understanding of permafrost waters, information on how permafrost waters are formed is necessary. This included an isotopic survey of permafrost waters, hydrological inputs (snow, ground ice melt, rain), and physical characteristics, including position and slope aspect, ice content, vegetation, snow coverage and lithology. Few studies have examined isotopic characteristics of permafrost waters. One notable exception is a study on δ18O in active layer ground waters in differing geomorphological and lithological locations by Mikhalev (1989) in the Kolyma Lowland. He concluded that the main source of active layer groundwater is the late summer and fall precipitation which would be reflected in the isotopic composition of active layer ground ice.

This study examines the spatial and temporal variation of stable isotopes in permafrost affected waters in the Levinson-Lake catchment collected during 1994 and 1995. The study was designed to answer several important questions: (1) What are the recent, seasonal isotopic variations in the local precipitation? (2) What are the sources of active layer waters and is there a seasonal change in their isotopic composition? (3) Based on our current knowledge, is it useful to apply stable isotopes for arctic hydrological processes and paleoenvironmental interpretation in this permafrost setting?

Methods

This paper reports results from field work carried out in 1994 and 1995 in the Levinson-Lessing lake catch-
ment, Taymyr Peninsula, Siberia, the northernmost continental area of the circumpolar Arctic (Figure 1). A varying topography (slopes and flat areas) and arctic tundra vegetation in the area facilitate the study of transport processes in the active layer under varying geological and geomorphological, but similar climatic conditions. Three transects with a sum of 16 sites were instrumented during the summer of 1994 in the Levinson-Lessing lake catchment (Figure 1). These slopes differ in slope aspect and inclination, soil parent material, vegetation and thaw depth of the active layer. At each site, triple wire time domain reflectometry (TDR) probes, PT 100 temperature probes, wells, piezometers and suction lysimeters were installed. Volumetric moisture content, bulk electrical conductivity, water level in wells and piezometers, electrical conductivity, and pH of ground and soil waters, were measured daily from July to late August 1994. Depth of thaw of the active layer was recorded at least once a week. Ground water samples were collected from wells and piezometers using PVC tubing and plastic syringes and from the vadose zone using suction lysimeters. Precipitation was sampled from each event. At least two samples (each 30 ml) were taken from each type of water: one was analyzed for pH and electrical conductivity in the field; the second one was kept cool for stable isotope and radionuclide analysis. Laboratory analysis of water samples was undertaken using the Epstein-Mayeda technique for oxygen isotopes (Epstein and Mayeda, 1953) with an analytical error of 0.2 ‰ and a new water-H2 gas equilibration technique for hydrogen isotopes with an error of 2 ‰. The stable isotope ratios \( r_{H}= D / ^{1}H \) and \( r_{O}= ^{18}O / ^{16}O \) are calculated relative to Vienna Standard Mean Ocean Water (V-SMOW) in permil:

\[
\delta = \frac{r_{\text{sample}} - r_{\text{v-SMOW}}}{r_{\text{v-SMOW}}} \times 10^3
\]  

[1]

Positive \( \delta^{18}O \) and \( \delta D \) values, therefore, indicate an enrichment of the heavy isotopes \(^{18}O \) and \( D \) compared to the SMOW, and negative values a depletion.

**Results and discussion**

**Precipitation**

With establishment of a Local Meteoric Water Line (LMWL), permafrost waters in this catchment can be interpreted with respect to recent meteoric water input. Figure 2 depicts the relationship between \( \delta^{18}O \) and \( \delta D \) of meteoric water samples collected in the Levinson-Lessing lake catchment in 1994 and 1995. Values for \( \delta^{18}O \) range between -32.4 and -9.1 ‰, \( \delta D \) values between -245.3 and -78.6 ‰ with an average of \( \delta^{18}O = -20.1 \) ‰ and \( \delta D = -154.5 \) ‰ (\( N = 42 \)). The average values agree well with the theoretical predicted values using the relationship between the mean annual air temperature (MAAT) and mean annual \( \delta^{18}O \) of precipitation by Dansgaard (1964). Given a MAAT of -10 to -15 °C for the Taymyr peninsula (Ershov, 1989), this theoretical equation produces \( \delta^{18}O \) values between -20.6 to -24.1 ‰ and \( \delta D \) values between -156 to -184 ‰.

Craig (1961), using a large number of precipitation samples from all latitudes, defined a linear relationship between \( \delta^{18}O \) and \( \delta D \), which is called the Global Meteoric Water Line (GMWL). The seasonal variations of precipitation correspond to variations of air temperature, i.e. the winter and fall snow is depleted in heavy isotopes compared to summer precipitation (Figure 2). Using the least squares fit, winter snow samples define a linear relationship of

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\text{Figure 1. Location of the three installed transects in the Levinson-Lessing lake catchment, Taymyr, Siberia.}
\]

\[
\text{Figure 2. Plot of } \delta^{18}O \text{ versus } \delta D \text{ of precipitation samples collected in 1994 and 1995. The LMWL (determined from summer rain samples) is presented relative to the GMWL.}
\]
\[ \delta D = (8.02 \pm 0.31) \delta^{18}O + (14.2 \pm 8.7); \]
\[ N = 15, \quad r^2 = 0.98 \]  

where \( N \) is the number of data points and \( r^2 \) is the coefficient of determination. The slope and intercept are within the uncertainties of the GMWL therefore suggesting the oceans as the source of precipitation. Summer rain, however, deviates from the GMWL and scatters around a local meteoric water line (LMWL) defined as:

\[ \delta D = (7.23 \pm 0.21) \delta^{18}O + (9.9 \pm 3.5); \]
\[ N = 27, \quad r^2 = 0.98 \]

The scattering of precipitation data along the LMWL (Figure 2) likely reflects seasonal differences in air mass history that depend on temperature during condensation of vapour. The smaller slope compared to the GMWL and the intercept of -9.9 suggest a different origin of summer air masses compared to the winter snow. The slope suggests that the summer air masses carry some moisture from terrestrial surface waters which are already enriched in heavy isotopes by evaporation (Gat, 1980; Clark and Fritz, 1997).

**CATCHMENT WATERS**

**DIFFERENCES IN STABLE ISOTOPE COMPOSITION BETWEEN SLOPES**

The three slopes where data were collected are different with respect to their microclimatic conditions. Slope 1 has a south-west exposure, but is located in the trough of a valley and therefore receives some shadow. Slope 2 has the same aspect, but little vegetation cover and is steeply inclined. Thaw depths are greatest on this slope and it is well drained. Slope 3, which faces north-east, has the lowest inclination, thickest vegetation cover, receives a considerable amount of shadow, and is poorly drained. As a consequence, the water table reaches the surface during most of the summer. Upslope of site 3a is a shattered block field with steep inclination.

Figure 3 shows the isotopic composition of active layer water samples (ground and soil water) collected from July to August 1994. These values are displayed in relation to the GMWL and LMWL. For all slopes, ground and soil waters fall between the GMWL and LMWL. Values for slope 1 and 2 fall in the same range: -19 \( \leq \delta^{18}O \leq -15 \%o \) and -140 \( \leq \delta D \leq -110 \%o \). In comparison, some water samples of slope 3 are more negative, ranging from -21 to -15 \%o and -160 to -120 \%o. A closer examination shows that these samples are from site 3a, the uppermost site on slope 3.

This suggests that site 3a receives water from an additional source, possibly from melting snow or basal ice. Basal ice, first defined by Woo et al. (1982), refers to ice that is formed by refreezing snow meltwater. For example, when snowmelt water infiltrates large rock fields, it refreezes at the base of the rockfield due to negative temperatures. In addition, basal ice can supply meltwater over the entire summer since it is protected from direct sunlight or strong air temperature variation. Supporting evidence was found in the field, where meltwater leaving the rock field was observed. Furthermore, snow was absent, suggesting basal ice
was an important source of active layer waters at site 3a.

The range of active layer stable isotope composition for the entire set of slope water samples ($-20.8 \leq \delta^{18}O \leq -14.7 \%$ and $-156 \leq \delta D \leq -109 \%$) relative to summer precipitation (Figure 2) indicates that active layer waters are mostly fed by rainwater or melting of active layer ground ice. In 1995, a few samples were collected from the upper centimetres of the frozen active layer. The extracted waters have the isotopic

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Figure 4. Soil water isotope composition at various depths in the active layer on slope 2. Isotope composition of ground water samples is shown with crosses (+).

Figure 5. Plot of soil temperature versus $\delta^{18}O$ from soil water at the same depths. Within each slope, one symbol represents one site.
composition of $\delta^{18}$O=-17.5‰, $\delta^{18}$D=-129‰ (N=1) on slope 2 and $\delta^{18}$O=-17.9‰, $\delta^{18}$D=-132‰ (N=2) on slope 3 which is essentially unchanged relative to the 1994 active layer waters. As little difference is observed between samples collected in late fall and the early spring, it is inferred that fall and winter processes negligibly affected the active layer during this period. Furthermore, the similarity between isotopic composition of the frozen active layer and summer precipitation makes an evaluation of the individual active layer water budget components (active layer ice, precipitation, lateral flow) more difficult.

Therefore it can be assumed that processes during the fall and winter essentially leave the isotope composition of the active layer unchanged. This similarity in isotopic values between the frozen active layer and summer precipitation makes the isotopic quantification of these active layer water budget components more difficult.

Some waters show an offset to the left of the LMWL (1b, 2d) which is difficult to interpret and requires further investigations using additional geochemical data (e.g., pH, anions, cations).

**INTRA-SLOPE VARIABILITY**

An examination of slope 2 waters reveals site-specific differences in active layer waters (Figure 4). Site 2a is located furthest upslope and 2d at the base of the slope (Figure 1). Site 2x and 2c are located at the same height on the slope, approximately 50-80 metres apart from each other.

With the exception of waters from 2x, all samples fall in the range of -110 to -135‰ and -15 to -18‰ (Figure 4), a range consistent with summer rain and thawing active layer ice. The waters of site 2x are isotopically depleted compared to other sites on slope 2 and range between -125 to -145‰ $\delta^{18}$D and -17 to -20‰ $\delta^{18}$O suggesting snowmelt water as an additional source of active layer water. Field observation supports this trend in the isotopic data. Site 2x received meltwater from a small snowpatch until the latter half of July while the rest of the slope was already snow-free.

**SOIL TEMPERATURE AND $\delta^{18}$O**

Figure 5 depicts the relationship between soil temperatures in the active layer and the $\delta^{18}$O of the soil water at that depth. For most sites, $\delta^{18}$O values become heavier with increasing soil temperatures, a trend which is most pronounced for sites on slope 2. Sites that do not show this trend are located in the upper soil profiles (1a-9 cm, 3a-9 cm, 2x). These samples reflect the infiltration of rainwater (1a) or mixing of active layer waters with snow (2x) or basal ice contribution (3a). At site 1a, samples from the unsaturated zone could only be extracted after precipitation events, since the soil was usually very dry (the maximum water content at this site was 15%).

The positive relationship between $\delta^{18}$O and soil temperatures supports the observation that at higher soil temperatures, evaporation is enhanced and fractionation increases. Another explanation is that, concomitant with progress of the summer and warming of the active layer, precipitation $\delta^{18}$O values become more enriched. The latter hypothesis cannot be supported by precipitation $\delta^{18}$O values since they do not follow a seasonally increasing trend.

Using the $\delta^{18}$O values as a ‘microthermometer’ for these slopes, slope 3 has the coldest active layer which corresponds to lowest isotope values. Slope 2 is warmest, with highest $\delta^{18}$O values. Furthermore, the slope of the $\delta^{18}$O-soil temperature relationship indicates the magnitude and type of vapour loss. Although soil temperatures of slopes 1 and 2 extend over a similar range, the inclination of the $\delta^{18}$O - soil temperature relationship on slope 2 is greatest. This can be explained by distinguishing between evaporation and evapotranspiration dominated fractionation. As stated earlier, slope 2 has almost no vegetation cover compared to slope 1. Therefore, evaporation on slope 2 will cause an isotopic fractionation. On slopes 1 and 3, however, a considerable part of evaporation may proceed as evapotranspiration which has no fractionation effect on the isotopic composition (Gat, 1995).

**Summary and conclusions**

The active layer waters of this site are of meteoric origin, fed mostly by summer precipitation. This is in agreement with results presented by Mikhaalev (1989) who found that the main source of active layer waters in the Kolyma region is summer and fall precipitation. In contrast, Mackay (1983) found that the isotopic composition of active layer waters of the Tuktoyaktuk Peninsula, Canada, is composed of snowmelt water and rainfall. This underlines the importance of a good understanding of the study site, so that local effects (such as runoff from a late lying snowpack) can be excluded when isotope signatures are discussed.

At our sites, the active layer is fed mainly by summer rain, therefore the isotopic signatures reflect the late summer air temperatures. This is useful for the paleoclimatic interpretation of isotopic composition of intrasedimental ice preserved in the active layer in this setting. However, these results are contrary to Nikolayev and Mikhailen (1995) who found the best correlation...
between $\delta^{18}O$ of active layer intrasedimental ice and average air temperature of the winter season, although the active layer was also fed by late summer precipitation. The role of active layer isotopes as paleothermometers is not yet fully understood. Future process studies should therefore investigate the relationship between stable isotopes in active layer waters, formation of intrasedimental ice and air temperatures on a long-term basis.

The interpretation of stable isotope signatures on the basis of physical parameters of the active layer (soil water content and temperature) allows the identification of microclimatic differences between slopes. Moreover, the isotopes have been applied to discern different sources of waters within one slope (such as basal ice at site 3a). This stresses the fact that on a small scale, i.e. within one catchment, isotopes are sensitive indicators for microclimatic and hydrological processes affecting the active layer. However, independent hydrological measurement should be carried out in support of stable isotope measurements.

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