A high-resolution time series of oxygen isotopes from the Kolyma River: Implications for the seasonal dynamics of discharge and basin-scale water use

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[1] Intensification of the Arctic hydrologic cycle and permafrost melt is expected as concentrations of atmospheric greenhouse gases increase. Quantifying hydrologic cycle change is difficult in remote northern regions; however, monitoring the stable isotopic composition of water runoff from Arctic rivers provides a means to investigate integrated basin-scale changes. We measured river water and precipitation $\delta^{18}O$ and δD to partition the river flow into snow and rain components in the Kolyma River basin. On an annual basis, we found water leaving the basin through the river consisted of 60% snow and 40% rain. This is compared with annual precipitation inputs to the watershed of 47% snow and 53% rain. Despite the presence of continuous permafrost, and fully frozen soils in the spring, our analysis showed not all spring snowmelt runs off into the river immediately. Instead, a substantial portion is retained and leaves the basin as growing season evapotranspiration. **Citation:** Welp, L. R., J. T. Randerson, J. C. Finlay, S. P. Davydov, G. M. Zimova, A. I. Davydova, and S. A. Zimov (2005), A high-resolution time series of oxygen isotopes from the Kolyma River: Implications for the seasonal dynamics of discharge and basin-scale water use, *Geophys. Res. Lett.*, *32*, L14401, doi:10.1029/2005GL022857.

1. Introduction

[2] Total annual discharge from the six largest Eurasian rivers has increased significantly from 1936 to 1999 [*Peterson et al.*, 2002]. If this trend in freshwater input continues, North Atlantic deepwater formation may be disrupted, with potentially serious consequences for Earth's climate [*Clark et al.*, 2002]. In addition, hydrologic changes alter the delivery of carbon and nutrients from land to the Arctic Ocean [*Dittmar and Kattner*, 2003]. Increasing rates of permafrost melt and loss of forests to fire may contribute to a small part of the trend in river discharge, but cannot account for the entire magnitude of change. Increased precipitation is the most likely cause [*McClelland et al.*, 2004]. However, measuring precipitation over large Arctic watersheds has been challenging [*Serreze et al.*, 2003]. Stable water isotope measurements of major Arctic rivers have potential to provide insights about mechanisms responsible for observed discharge trends.

[3] The isotopic composition of precipitation is largely controlled by water vapor source, the formation temperature of precipitation, and relative fraction of water vapor removed from the atmosphere [*Gat*, 1996]. Most IAEA stations analyzed for stable isotopes exhibit seasonal cycles similar to air temperature variations [*Gat and Gonfiantini*, 1981], with enrichment of heavy isotopes during warm summer months and depletion in cold winter months. Previous work on Siberian water isotopes have exploited temperature dependence to reconstruct paleoclimate temperatures using stable water isotopes preserved in ice wedges [*Vasil'chuk*, 1992]. In Canada, modern lake stable isotopes have been used to partition regional evapotranspiration fluxes [*Gibson and Edwards*, 2002] and river stable isotopes were used to estimate the basin-scale transpiration flux of the Mississippi basin [*Lee and Veizer*, 2003].

[4] Here, we present a time series of $\delta^{18}O$ in the northeast Siberian Kolyma River. The Kolyma is the seventh largest Arctic river and the only river that showed no increase in discharge in the study by *Peterson et al.* [2002]. It is also the largest Arctic river completely underlain by continuous permafrost [*McClelland et al.*, 2004; *Brabets et al.*, 2000]. Permafrost influences watershed hydrology through shallow and seasonally varying active layer storage capacity [*McNamara et al.*, 1997]. Consequently, Arctic river transport of water, carbon, and nutrients is strongly seasonal [*Dittmar and Kattner*, 2003]. Our objectives were to (1) measure a baseline in the Kolyma against which to compare future effects of Arctic climate change and (2) use stable water isotopes and a two end-member mixing model to determine seasonal export of snow and rain from the basin.

2. Methods

2.1. Study Area

[5] The Kolyma drains an area of 650,000 km^2 with a mean annual discharge of 132 km^3 yr^{-1} [*McClelland et al.*, 2004]. Our sampling locations were near the town of Cherskii (69°N, 161°E), 150 km from the mouth of the Kolyma and ~50 km south of the northern tree line (Figure 1). Discharge was measured at the Kolymskoye gauge station, an additional 160 km upstream of Cherskii. The Anniui River confluence is located between the sampling and gauging stations and contributes ~4% of

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Figure 1. Water sampling was conducted at Cherskii, near the mouth of the Kolyma River, and the discharge gauging station was located at Kolymskoye. (Map source: Cornell University Interactive Mapping Tool).

the river flow. Cherskii has a strongly continental climate with warm summers (June–August average temperature of 12°C), cold winters (average January temperature is −35°C) and low annual precipitation (185 mm yr^{-1}). The Kolyma watershed is dominated by deciduous larch taiga and tundra vegetation, and ~10% of land area is covered by lakes.

2.2. Water Sampling and Isotopic Analysis

[6] We sampled the Kolyma from September 2002 to April 2004, with monthly collections during winter and more frequent collections during the ice-free season. We collected samples just below the river surface during ice-free months (under the ice during winter) in 4–25 mL glass vials with inverted cones inside screw top caps. The vials were sealed with parafilm, and kept cool during storage. We sampled individual rain events in Cherskii using pan collectors (emptied immediately following the event to minimize evaporation). Volunteers surveyed snow on 8–19 April 2003, at 23 locations within 300 km of Cherskii. By sampling snow immediately prior to spring thaw, we attempted to include post-deposition isotopic effects within the snow pack. Surface and sub-surface samples were allowed to thaw in 125 mL plastic bottles and transferred to 25 mL glass vials.

[7] We analyzed waters for $\delta^{18}O$ by CO_2 equilibration with a GasBenchII connected to a Thermo Finnigan DeltaPlus isotope ratio mass spectrometer at UC Irvine. We sent a subset to the Center for Stable Isotope Biogeochemistry at UC Berkeley for δD analysis using a Thermo Finnigan MAT H/Device. All isotope results in this paper are presented on the VSMOW scale. Analytical uncertainty was 0.1‰ for $\delta^{18}O$ and 1.6‰ for δD.

2.3. Partitioning Approach

[8] We calculated contributions of snow and rain inputs to the river for each sampling period using a two end-member mixing model:

$$Q(t) = Q(t)*f_{\mathrm{snow}}(t) + Q(t)*f_{\mathrm{rain}}(t) \tag{1}$$

where Q is total river discharge, f_{snow} and f_{rain} are snow and rain fractions of Q, and t is time.

$$\delta_{\mathrm{river}}(t) = f_{\mathrm{snow}}(t)*\delta_{\mathrm{snow}} + f_{\mathrm{rain}}(t)*\delta_{\mathrm{rain}} \tag{2}$$

δ_{river} is measured river $\delta^{18}O$, and estimates of the mean $\delta^{18}O$ of snow and rain end-members are denoted as δ_{snow} and δ_{rain}. We characterized δ_{snow} by taking the arithmetic mean of the spring snow survey and determined δ_{rain} in two steps. First, we weighted the $\delta^{18}O$ of each individual rain event by the amount of precipitation measured at the Cherskii airport within each month to determine a weighted mean. Second, each monthly mean (May through September) was weighted by the total precipitation for that month, minimizing bias from over or under-sampling rain events during some months with respect to others.

[9] We made a few assumptions about watershed processes other than mixing of snow and rain to use the partitioning method described above. The water budget of a river basin can be defined as,

$$Q = P - ET - \Delta S \tag{3}$$

where P is precipitation, ET is evapotranspiration, and ΔS is the change in groundwater storage. We assumed ΔS for one year was negligible because there is only one dam (hydroelectric) that finished filling in 1990 [*McClelland et al.*, 2004]. This simplified Equation 3 to:

$$Q = P - ET \tag{4}$$

ET is composed of evaporation (E), transpiration, and interception. Transpiration and interception do not modify the isotopic composition of surface water, and therefore, interception will be included in transpiration throughout this paper. E, however, enriches the residual water in heavy isotopes. If E was a large part of ET in this watershed, and evaporating elements (e.g., lakes) were closely linked to river flow, then δ_{river} cannot be assumed to be a simple mixture of snow and rain end-members.

[10] To test the isotopic influence of E on the Kolyma, we measured δD on a subset of precipitation and all river samples. Negative d-excess (d-excess = δD − 8*$\delta^{18}O$) values have been used as an elimination criterion for precipitation samples suspected of E influence prior to sample collection [*Kurita et al.*, 2004]. Isotopic equations derived to quantify the E/P ratio for lakes have been used to estimate evaporation from river systems [*Gibson and Edwards*, 2002; *Lee and Veizer*, 2003] (described in the auxiliary material).[1]

[11] Finally, we estimated the contributions of snow and rain to the annual water budget using a mass balance approach.

$$Q * f_{snow} = P * f^{P}_{snow} - ET * f^{ET}_{snow} \tag{5}$$

Knowing Q, f_{snow}, P and f^{P}_{snow}, one can solve for the snow fraction of ET (f^{ET}_{snow}). A similar equation can be written for the rain fraction.

2.4. Data Sources

[12] We obtained daily discharge records of the Kolymskoye gauge station on the Kolyma from ArcticRIMS (Rapid Integrated Monitoring System) (http://rims.

[1] Auxiliary material is available at ftp://ftp.agu.org/apend/gl/2005GL022857.

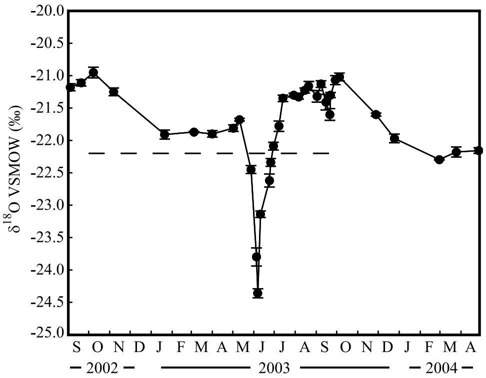

Figure 2. δ^{18}O of the Kolyma sampled at Cherskii, from September 2002 through April 2004. Error bars represent the analytical standard deviation of each sample. The Q-weighted mean for the one-year period from October 2002 to September 2003 was −22.2‰, marked by the dashed line.

unh.edu/). We also utilized several basin-scale NCEP products from Arctic-RIMS, including rescaled P and P-ET estimates by *Serreze et al.* [2003] and elevation adjusted surface temperature [*Oelke et al.*, 2003]. Serreze's P product was a rescaling of NCEP reanalysis data using measurements from a sparse precipitation gauge station network in the Arctic. We then subtracted P-ET from P to estimate the seasonal cycle of ET.

3. Results and Discussion

3.1. Stable Oxygen Isotope Observations

[13] We measured the Kolyma δ^{18}O from September 2002 through April 2004 (Figure 2). The Q-weighted annual mean δ^{18}O value for the Kolyma from October 2002 to September 2003 was −22.2‰. This compares remarkably well with the value of −22.4‰ mentioned in *Letolle et al.* [1993] used by *Ekwurzel et al.* [2001]. River

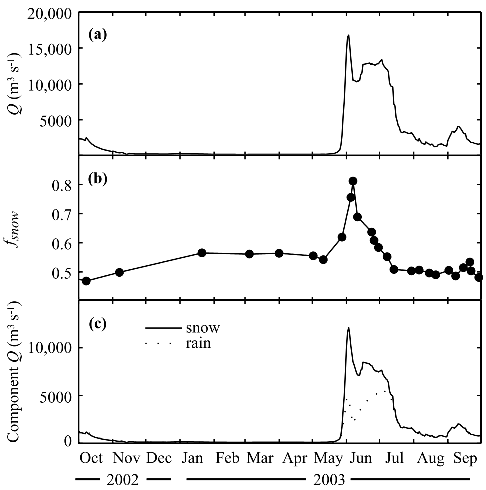

Figure 3. Kolyma (a) Q (m^3 s^{-1}) measured at Kolymskoye, (b) fraction of snow component calculated from δ^{18}O partitioning and (c) Q (m^3 s^{-1}) associated with snow (solid line) and rain components (dotted line). The fraction of snow peaked in early June at 82%, and the rain component peak was delayed until early July.

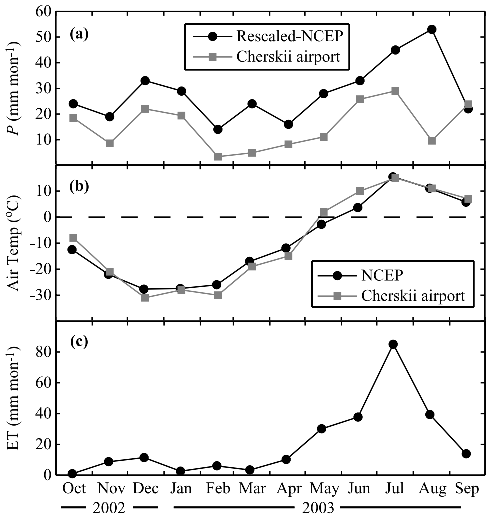

Figure 4. (a) Rescaled NCEP monthly mean P (mm mon^{-1}) [*Serreze et al.*, 2003] for the Kolyma watershed (circles) and measured at the Cherskii airport (squares). (b) Monthly mean surface air temperature from the elevation corrected NCEP reanalysis for the Kolyma watershed (circles) [*Oelke et al.*, 2003] and measured at the Cherskii airport (squares). (c) ET estimated from *Serreze et al.* [2003].

δ^{18}O rapidly became depleted (more negative) during the spring pulse of snowmelt in late May to early June. Minimum δ^{18}O (−24.4‰) occurred on 7 June 2003, near the seasonal maximum in Q, but delayed by 6 days (Figure 3a). The mean and standard deviation of sampled snow, −26.2 ± 5.0‰, explained the depletion during the spring thaw. From June through early July, Q remained high and river δ^{18}O rapidly increased due to summer rain inputs which had a weighted mean and standard deviation of −16.3 ± 3.8‰. Maximum δ^{18}O occurred during October of both 2002 and 2003 (with a value of −21.0 ± 0.1‰ in both years), and then slowly decreased over the baseflow months of October through January, stabilizing by February until the spring thaw in May. Rain, snow and river measurements are provided in the auxiliary material.[1]

[14] January through April baseflow δ^{18}O in 2003 was −21.9 ± 0.1‰ and was 0.3‰ lower in 2004 (−22.2 ± 0.1‰). The source of low winter Q (∼200 m^3 s^{-1}) is assumed to be deep groundwater and controlled release from the hydroelectric dam. The cause of interannual variability in river baseflow δ^{18}O is uncertain, but may reflect more snow input to the hydroelectric reservoir in the second year. Seasonal variations in Kolyma δ^{18}O are similar to Canadian river time series of measurements during ice-free conditions, and reconstructions of winter δ^{18}O from river ice cores [*Gibson and Prowse*, 2002].

3.2. Partitioning River Flow Into Snow and Rain Components

[15] We used mean snow and rain δ^{18}O end-members described above (−26.2 ± 5.0‰ and −16.3 ± 3.8‰) and measurements of the Kolyma from October 2002 to September 2003 (Figure 2) to partition snow and rain contributions to the river (Figure 3b). During the spring Q pulse, the snow component peaked at 82%. During

Table 1. Annual Water Budget of the Kolyma River Basin

Water Budget Component	$km^3\ yr^{-1}$	Snow	Rain	Error
River Discharge	74	60%	40%	10%
Precipitation Inputs	340	47%	53%	~15%
Evapotranspiration	266	43%	57%	~18%

baseflow conditions, snow and rain contributions to the river were both close to 50%. In Figure 3c we show the mass flux of snow and rain components. Snow dominated the river flow during spring, peaking in early June. In contrast, inputs from rain increased over the summer and peaked during early July. Summing these curves over the year, we found 60% of water leaving the Kolyma originally fell as snow and 40% as rain.

3.3. Annual Water Balance

[16] We compared our analysis of the annual snow and rain partitioning in river runoff to the proportions of annual snow and rain that fell on the watershed. Figure 4 shows monthly mean *P*, air temperature, and ET estimated for the Kolyma watershed, and measured air temperature and *P* at Cherskii for our study period. Cherskii airport air temperature was used to define the rain season as May through September (air temperatures above freezing) and the snow season as October through April. We estimated that 47% of annual *P* fell as snow and 53% as rain. We then used Equation 5 to partition the remaining loss pathway, ET, into 43% snow and 57% rain (Table 1).

[17] Water loss pathways in the Kolyma watershed are extremely seasonal. The months of May through September account for 90% of river *Q* and 83% of ET. Comparing fractions of snow and rain from incoming *P* and outgoing river *Q*, there is a higher proportion of snow in river *Q* than in *P*. This supports the assumption that much of the snowmelt immediately flows over the frozen soil surface into the river. However, it is apparent from ET partitioning, that snow contributes substantially to the summer ET flux. This is consistent with recent studies showing Siberian larch forests utilize snowmelt water during spring leafout, a period of enhanced transpiration [*Sugimoto et al.*, 2002], and local scale hydrologic experiments that show about half of snowmelt is lost as runoff, and the other half recharges soil water (S. P. Davidov, unpublished data, 2004).

3.4. Evaporation Influence Estimated From δD and $\delta^{18}O$

[18] Some precipitation samples we analyzed for δD (13 out of 44) had negative d-excess values, suggesting they may have been evaporatively enriched. If we discarded these samples, δ_{rain} decreased to $-16.6‰$ and the resulting river *Q* partitioning shifted by 2% (f_{snow} = 58% and f_{rain} = 42%). Using only precipitation samples with positive d-excess values, the local meteoric water line (LMWL) was $\delta D = 7.0 * \delta^{18}O - 11.7$, $r^2 = 0.99$, n = 31. None of the Kolyma samples had negative d-excess values, and the Kolyma waterline was $\delta D = 6.8 * \delta^{18}O - 19.1$, $r^2 = 0.98$, n = 30.

[19] The slopes of the LMWL and the river regression line were similar ($7.0 \pm 0.1‰$ compared to $6.8 \pm 0.2‰$), indicating a low *E*/*P* ratio for the watershed. Using the approach described in the auxiliary material, we estimated 6% of *P* for the basin was lost through *E*. Therefore, *E* does not appear to be a major component of ET in the Kolyma and our assumption that measured $\delta^{18}O$ in river samples represent a simple mixture of snow and rain (not enriched significantly by *E*) appears valid.

3.5. Partitioning Error and Sensitivity Analysis

[20] Sources of uncertainty in this study were (1) the assumption that precipitation collected near Cherskii was representative of the entire watershed and (2) there was no enrichment of snowmelt water above measured snowpack $\delta^{18}O$ [*Laudon et al.*, 2002]. We explored sensitivity to the choice of δ_{rain} and δ_{snow} by calculating the annual river partitioning using end-members $\pm 1‰$. Resulting f_{snow} values range from 50–70% with 6% standard deviation, which is within our estimated error of 10%. Also, statistically similar slopes for the river δD-$\delta^{18}O$ water line and the LMWL suggest there are not drastically different water vapor sources and precipitation conditions upstream.

4. Conclusions

[21] We determined the *Q*-weighted annual average of Kolyma $\delta^{18}O$ was $-22.2‰$ and during the same year, 60% of water that left the basin through the river originally fell as snow and 40% as rain. Our analysis suggests not all snowmelt leaves the watershed as spring runoff, instead, a substantial amount contributes to ET. We believe monitoring stable water isotopes of Arctic rivers will aid investigations of future climate change. Increases in river $\delta^{18}O$ are expected as a result of increasing mean annual temperature. However, a shift in the seasonality of precipitation, changes in surface vegetation (type, coverage, and water status) and increases in active layer depth as a result of permafrost melt may also contribute to changes in Arctic hydrology and the water isotope budget.

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