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Publication Date:
July 2005

Series:
Faculty Publications

Permalink:
http://escholarship.org/uc/item/3h67587t

DOI:
https://doi.org/10.1029/2005GL022857

Keywords:
Climate change, Composition, Evapotranspiration, Isotopes, Natural water geochemistry, Oxygen, Permafrost, Precipitation (chemical), Rivers, Runoff, Time series analysis, Water, Arctic rivers, Basin-scale water use, Discharge water use, Oxygen isotopes, Seasonal dynamics, Snowmelt, Water runoff, Hydrology, climate change, hydrological regime, oxygen isotope, river discharge, Eastern Hemisphere, Eurasia, Kolyma River, Russian Federation, World

Abstract:
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 $\deltaD$ 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.

Supporting material:

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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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Received 28 February 2005; revised 6 June 2005; accepted 10 June 2005; published 20 July 2005.

[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 $\delta D$ 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 evaportranspiration. 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.

[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-endmember 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...
δ_river is measured river δ^{18}O, and estimates of the mean δ^{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 δ^{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,

\[ Q = P - ET - \Delta S \]  

where \( P \) is precipitation, \( ET \) is evapotranspiration, and \( \Delta S \) is the change in groundwater storage. We assumed \( \Delta S \) 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:

\[ Q = P - ET \]  

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 \( \delta_{river} \) 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*δ^{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.

\[ Q \cdot f_{snow} = P \cdot f_{snow}^P - ET \cdot f_{snow}^{ET} \]  

Knowing \( Q \), \( f_{snow} \), \( P \) and \( f_{snow}^P \), one can solve for the snow fraction of \( ET \) (\( f_{snow}^{ET} \)). 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 Arctic-RIMS (Rapid Integrated Monitoring System) (http://rims.

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\(^1\)Auxiliary material is available at ftp://ftp.agu.org/apend/gl/2005GL022857.
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 $\delta^{18}O$ from September 2002 through April 2004 (Figure 2). The $Q$-weighted annual mean $\delta^{18}O$ value for the Kolyma from October 2002 to September 2003 was $-22.2\%_{\circ}$. This compares remarkably well with the value of $-22.4\%_{\circ}$ mentioned in Letolle et al. [1993] used by Ekwurzel et al. [2001]. River $\delta^{18}O$ rapidly became depleted (more negative) during the spring pulse of snowmelt in late May to early June. Minimum $\delta^{18}O$ ($-24.4\%_{\circ}$) 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, $-26.2 \pm 5.0\%_{\circ}$, explained the depletion during the spring thaw. From June through early July, $Q$ remained high and river $\delta^{18}O$ rapidly increased due to summer rain inputs which had a weighted mean and standard deviation of $-16.3 \pm 3.8\%_{\circ}$. Maximum $\delta^{18}O$ occurred during October of both 2002 and 2003 (with a value of $-21.0 \pm 0.1\%_{\circ}$ 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 $\delta^{18}O$ in 2003 was $-21.9 \pm 0.1\%_{\circ}$ and was $0.3\%_{\circ}$ lower in 2004 ($-22.2 \pm 0.1\%_{\circ}$). The source of low winter $Q$ ($\sim 200 \text{ m}^3 \text{s}^{-1}$) is assumed to be deep groundwater and controlled release from the hydroelectric dam. The cause of interannual variability in river baseflow $\delta^{18}O$ is uncertain, but may reflect more snow input to the hydroelectric reservoir in the second year. Seasonal variations in Kolyma $\delta^{18}O$ are similar to Canadian river time series of measurements during ice-free conditions, and reconstructions of winter $\delta^{18}O$ 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 $\delta^{18}O$ end-members described above ($-26.2 \pm 5.0\%_{\circ}$ and $-16.3 \pm 3.8\%_{\circ}$) 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 pulse, the snow component peaked at $82\%$. During...
not appear to be a major component of ET in the Kolyma and our assumption that measured δ18O 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 δ18O [Laudon et al., 2002]. We explored sensitivity to the choice of δsnow and δrain by calculating the annual river partitioning using end-members ±1‰. Resulting fsnow 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-δ18O 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 δ18O 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 δ18O 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.

[22] Acknowledgments. This work was supported by an NSF RAISE grant OPP-0097439. We thank H. Kristenson and S. Hayden for field sampling and J. Neff for logistical assistance. We also thank X. Xu and S. Trumbore at UCI for the use of their stable isotope facilities. LRW received support from the EPA’s STAR fellowship program.

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