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Key Points:
- Topography controls hydrology and thereby mediates stream thermal regimes
- Snowmelt thermally buffers streams to air temperature in steeper watersheds
- Thermal heterogeneity across river basins could be lost with reduced snowpack

Supporting Information:
- Figures S1–S5 and Table S1

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Watershed geomorphology and snowmelt control stream thermal sensitivity to air temperature

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Abstract How local geomorphic and hydrologic features mediate the sensitivity of stream thermal regimes to variation in climatic conditions remains a critical uncertainty in understanding aquatic ecosystem responses to climate change. We used stable isotopes of hydrogen and oxygen to estimate contributions of snow and rainfall to 80 boreal streams and show that differences in snow contribution are controlled by watershed topography. Time series analysis of stream thermal regimes revealed that streams in rain-dominated, low-elevation watersheds were 5–8 times more sensitive to variation in summer air temperature compared to streams draining steeper topography whose flows were dominated by snowmelt. This effect was more pronounced across the landscape in early summer and less distinct in late summer. Thus, the impact of climate warming on freshwater thermal regimes will be spatially heterogeneous across river basins as controlled by geomorphic features. However, thermal heterogeneity may be lost with reduced snowpack and increased ratios of rain to snow in stream discharge.

1. Introduction

Snowmelt is a critical source of water that sustains summer streamflow and supplies water for people and ecosystems during drier months in many regions of the world [*Barnett et al.*, 2005]. Loss of summer water supply due to earlier onset of snowmelt, and transition to rainfall during winter months, is a potentially critical response to warming climate [*Barnett et al.*, 2005; *Stewart*, 2009; *Mantua et al.*, 2010; *Diffenbaugh et al.*, 2013; *Berghuijs et al.*, 2014]. These hydrological changes in combination with warmer air temperatures may have widespread consequences for aquatic ectothermic species that are sensitive to thermal alterations to their environment. Indirect links between snowmelt and stream temperature have been proposed, but a quantitative understanding of how snowmelt mediates stream temperature sensitivity to climatic conditions is distinctly lacking [*Mohseni and Stefan*, 1999; *Isaak et al.*, 2010, 2012; *Fellman et al.*, 2014; *Luce et al.*, 2014] and is critical for understanding the suitability of freshwater ecosystems under shifting climate regimes.

Surface water and air temperatures are both primarily heated through solar radiation; thus, air temperature is commonly used as a proxy for understanding solar heat exchange at the stream-air interface over the summer months [*Mohseni and Stefan*, 1999; *Caissie*, 2006; *Arismendi et al.*, 2014; *Garner et al.*, 2014]. In a typical energy balance model, heat exchange at the water surface is the sum of net radiation (long-wave and short-wave), sensible heat, and latent heat exchanges [*Caissie*, 2006]. Both long-wave radiation and sensible heat transfer depend directly on air temperature [*Leach and Moore*, 2010; *MacDonald et al.*, 2014] and will reflect the day-to-day and interannual variations in solar radiation (e.g., sunny days tend to be warmer than cloudy days). Latent heat transfer depends on the vapor pressure of the air, which is also a function air temperature in humid climates [*Leach and Moore*, 2010]. Relationships between air and water temperature are often linearly correlated at air temperatures between 0 and 20°C and uncorrelated at air temperatures below freezing or at warmer temperatures where streams cool by evaporative heat loss [*Mohseni and Stefan*, 1999]. Still, air temperature has been used as a reasonable proxy of solar energy exchange at the stream surface and has contributed to better understanding of physical processes that regulate stream thermal regimes.

Regional to local scale analyses reveal that the geomorphic characteristics of watersheds (e.g., elevation, slope, area, lakes, dams, and glaciers) often control water temperature relationships to air temperature [e.g., *Kelleher et al.*, 2012; *Fellman et al.*, 2014; *Luce et al.*, 2014]. However, others have shown that air to stream temperature regressions often do not adequately capture stream thermal regimes during other

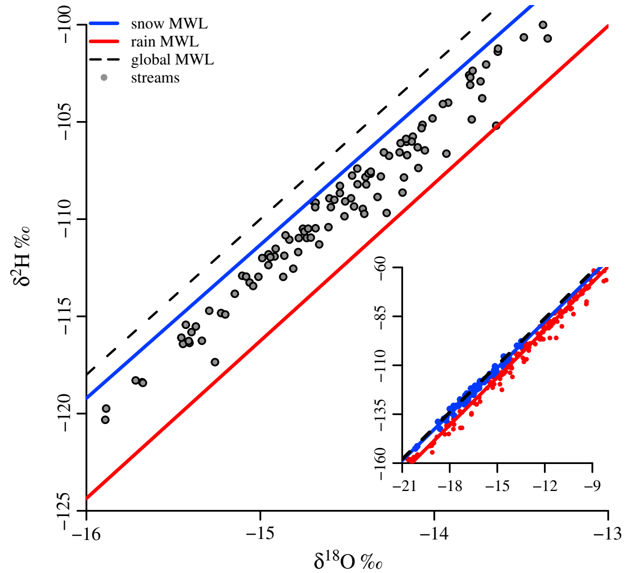

Figure 1. Water isotopes ($\delta^{18}O$ and $\delta^{2}H$ per mil) of rain, snow, and stream water. The MWL represents the local meteoric water line for snow and rain, and the dashed line indicates the global MWL $\delta^{2}H = 10 + 8 \times \delta^{18}O$ [*Craig*, 1961]. The inset plot shows water isotope values detected for snow and rain for this system.

time periods and are probably not accurate enough to predict future conditions associated with expected climate warming [*Arismendi et al.*, 2014]. This suggests that the relationships between the stream-specific temperature and air temperature are likely dynamic through time, such that associations between stream temperature and air temperature are modified by hidden processes, such a changing degree of summer snowmelt to stream discharge.

Snowmelt may mediate the sensitivity of stream thermal regimes to surface energy exchanges through lateral advection from surface runoff and subsurface flows. An increase in mountain snowmelt should provide a greater input of cold water, flowing downstream at a faster rate, and resulting in streams with greater depth and cross-sectional area, which would reduce stream temperature sensitivity to surface energy exchange [*van Vliet et al.*, 2011; *MacDonald et al.*, 2014]. For instance, *Leach and Moore* [2014] found that winter rain-on-snow events typically cool streams through lateral advective fluxes that swamp heat transfers from surface energy exchanges, bed heat conduction, and stream friction. When snow cover was absent, stormflow entered streams closer to maritime air temperatures and daily stream temperatures increased with increasing air temperatures [*Leach and Moore*, 2014]. Thus, relative inputs of snowmelt and rainfall may mediate the relationship between summer air temperature and stream temperature, but this effect has not been fully explored.

Here we evaluate the time series of daily water temperature for boreal streams in southwest Alaska to quantify relationships between air temperature and stream temperature along a gradient of snow contributions to stream discharge. This region represents one of the fastest changing climates on the globe, with large expected changes in air temperature, freezing conditions, and therefore hydrological alterations for critical fluvial habitat that support aquatic and riparian biota [*Maurer et al.*, 2007]. We quantified the extent to which watershed topography and water source (rain versus snow) mediated stream temperature sensitivity to air temperature during the ice free season, when we expect that warmer climate might impose important physiological constraints on aquatic organisms. We expected that streams in these river networks would have a common response to regional changes in air temperature but would be modified at the individual stream level by local watershed features that control the contributions of snow to streamflow.

To assess the different contribution of hydrologic sources to streams, we examined the $\delta^{18}O$ and $\delta^{2}H$ of rainfall, snowmelt, and streamflow (Figure 1) from 52 to 80 streams over three summers in three river basins in southwest Alaska (Figure 2a). We hypothesized that the $\delta^{18}O$ and $\delta^{2}H$ of streams would be more isotopically similar to snow in streams draining steeper, high-elevation basins that retain snowmelt-derived water later into the summer compared to flatter low-elevation watersheds that collect less winter snow and whose flows are dominated by summer rainfall. We used multivariate time series analysis [*Holmes et al.*, 2012] to evaluate water temperature sensitivity to changes in air temperature and examined how the sensitivity to air temperature is spatially structured across river basins to reflect a gradient of topography and water source contribution to stream discharge. We hypothesized that streams with summer flows dominated by contributions of melted snow would be decoupled from changes in air temperature compared to streams that have a higher proportion of rain in their discharge.

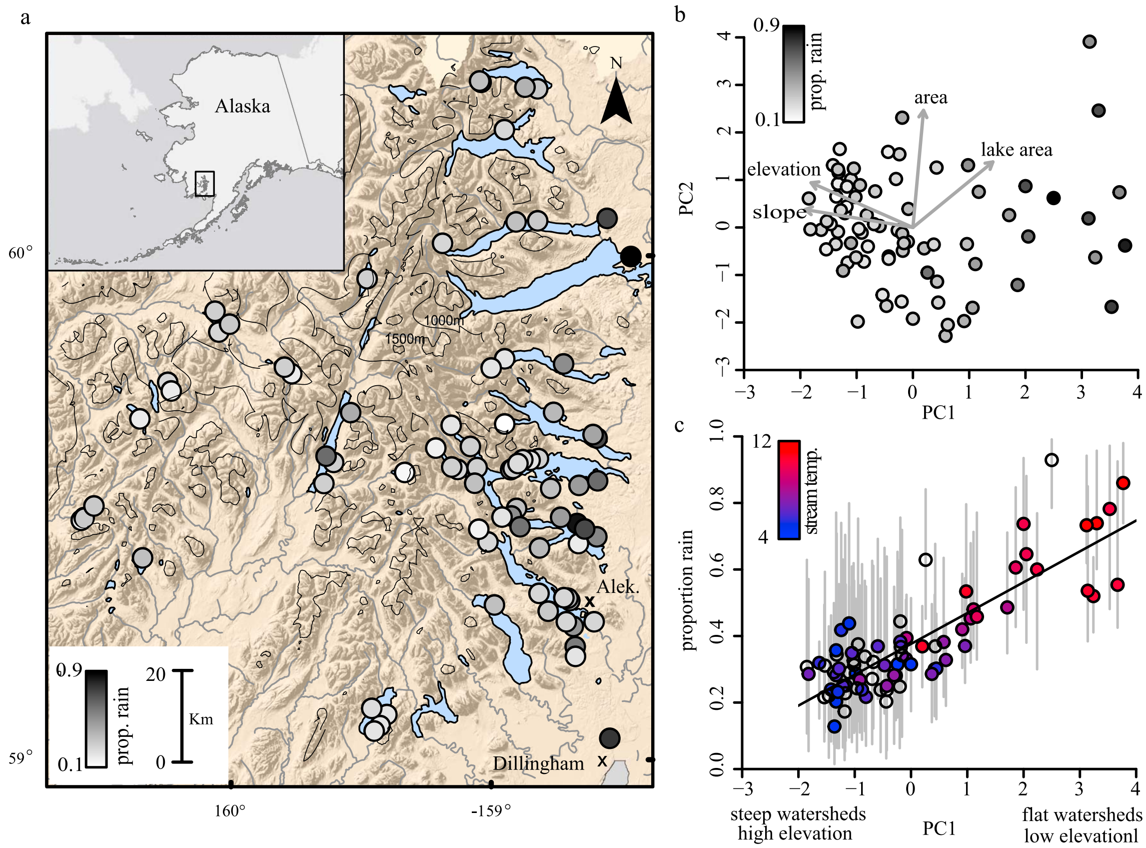

Figure 2. (a) Locations of study streams in SW Alaska and air temperature stations at Dillingham and Aleknagik (Alek.) and (b) ordination plot from principal component analysis of watershed characteristics; points are colored by the median proportion rain from mixing model. Length and direction of arrows on ordination are proportional to vector loading of watershed predictors into each principal component. (c) Median proportion and 90% credible intervals (vertical grey lines) of rain in stream discharge as a function of the first principal component. If filled, points are colored by their average stream summer temperature (°C); the open circles indicate streams lacking temperature data.

2. Methods

2.1. Stream Temperature Monitoring

This study was conducted in southwestern Alaska in the Wood, upper Nushugak, and Togiak river basins (Figure 2a). These river basins consist of several large lakes which are fed by numerous tributaries and connected by short rivers. The Ahklun mountain range stretches across this region, with typical mountain summits reaching 1000–1500 m (Figure 2a). Streams in this region are characterized by a snowmelt hydrology, with peak discharge occurring in mid-May to June. We monitored second- to fourth-order stream thermal regimes with I-Buttons® (Maxim Integrated Products, Sunnyvale, CA) and Hobo Level Loggers (Onset Computer Corp., Bourne, MA), summarized to daily averages from 1 June to 9 September. All loggers were cross calibrated and crossed checked with ambient reading in the field. Air temperature was monitored at the Dillingham, Alaska airport (Sta. PADL) and on Lake Aleknagik with a Hobo microweather station, <120 km from the most distant stream temperature logger (average distance 40 km).

2.2. Isotopic Determination of Snow and Rain Contribution to Streams

Oxygen and hydrogen stable isotopes (δ^{18}O and δ^{2}H‰) in water were used to trace the relative contributions of rain and snow to surface discharge [*Clark and Fritz*, 1997; *Brooks et al.*, 2012; *Fellman et al.*, 2014]. Streams were sampled monthly, from June to September, to characterize the spatial variation of δ^{18}O and δ^{2}H in streams during the open water season. Depth-integrated samples of the snowpack were collected in late March 2012 using snow cores across the study area. Rainfall was collected in rain gages during the summer months for daily rainfall that accumulated at least 1.3 cm or greater at lakes Nerka, Beverley, Aleknagik, and on the Togiak River. Samples were collected in duplicate using gastight 8 mL Nalgene bottles and frozen for later analysis at the University of Washington's Isolab facilities. A Picarro (L1102i and L2130i) water analyzer was used to determine the $\delta^{18}O$ and $\delta^{2}H$ of water (analytical precision σ at 0.08 and 0.6‰, respectively). Ratios of $^{18}O/^{16}O$ and $^{2}H/^{1}H$ are expressed in delta units, per mil (‰), defined in relation to Vienna standard mean ocean water.

We calculated the deuterium excess of rain and snow (i.e., the sources), and stream water isotopes, which is the orthogonal distance from global meteoric water line ($d‰ = \delta^{2}H - 8 \times \delta^{18}O$) [*Craig*, 1961; *Dansgaard*, 1964], the global average relationship between $\delta^{2}H$ and $\delta^{18}O$. Physically, $^{2}H^{1}H^{16}O$ diffuses at slightly faster rates compared to $^{1}H_2{}^{18}O$, resulting in excess deuterium in newly evaporated water [*Clark and Fritz*, 1997]. At cooler temperatures, winter precipitation is characterized by higher *d*-excess values relative to summer precipitation reflecting temperature-dependent kinetic rates of fractionation during water vapor formation [*Merlivat and Jouzel*, 1979]. We used a mixing model to estimate the contribution of snow and rain to streams with MixSIR v1.04 [*Moore and Semmens*, 2008]. This allowed us to incorporate error in our sources and generate posterior probability distributions about the median contribution of snow and rain to summer discharge.

Multivariate statistical analyses were performed to determine controls on patterns of rain and snow contribution to stream discharge. ArcGIS (v10.0, Environmental Systems Research Institute, Redlands, CA, USA) was used to estimate each stream's total watershed area, average elevation, average watershed slope (degrees) from a digital elevation model, and total area of lakes in each watershed. All habitat variables were log-transformed prior to analysis to control for differences in scale between descriptor variables. Principal component analysis [*Pearson*, 1901] on the correlation matrix was used to summarize dominant gradients of environmental variability among streams using the *vegan* [*Oksanen et al.*, 2010] and *biostats* [*McGarigal*, 2009] package in R (version 3.0.2) [*R Development Core Team*, 2011]. Stream scores on principal component axes 1 and 2 were regressed (using ordinary least squares linear regression) against rain-snow contribution to streams and compared using Akaike information criterion (AIC).

2.3. Time Series Analysis

We used dynamic factor analysis (DFA) [*Zuur et al.*, 2003] with the multivariate autoregressive state-space (MARSS) package in R [*Holmes et al.*, 2012] to evaluate common patterns in water temperature given its sensitivity to changes in air temperature. DFA is a dimension reduction technique designed specifically for time series analysis. The general idea is as follows. Given the time series of water temperature from the *N* different streams, we could simply model each stream individually but that would (1) ignore any potential dynamics shared among the streams and (2) require the estimation of many parameters per stream. DFA, on the other hand, allows us to characterize common trends among *N* time series with many fewer *M* trends. In simple terms, the DFA model is data = trends + explanatory variables + noise.

More specifically, following *Zuur et al.* [2003], we can write the DFA model as

$$\mathbf{y}_t = \mathbf{Z}x_t + \mathbf{D}\mathbf{g}_t + \mathbf{v}_t \quad (1)$$

$$x_t = x_{t-1} + w_t. \quad (2)$$

The $N \times 1$ vector of data observed at time t ($\mathbf{y}_t$) is modeled as a linear combination of the latent trend (x_t), a $P \times 1$ vector of explanatory variables ($\mathbf{g}_t$), and an $N \times 1$ vector of observation (sampling) errors ($\mathbf{v}_t$), which are distributed as a multivariate normal with mean $\mathbf{0}$ and $N \times N$ variance-covariance matrix $\mathbf{R}$. The $N \times 1$ vector $\mathbf{Z}$ and $N \times P$ matrix $\mathbf{D}$ contain the stream-specific loadings on the trend and explanatory effects, respectively.

Here the common trend is not simply a straight line but rather a random walk through time, such that the value of x at time t is simply equal to its value at time $t-1$ plus some random error w_t, which is distributed normally with mean 0 and variance q. The common trend can be thought of as an aggregate of unknown environmental drivers not captured by the explanatory variables. In order to make the model identifiable, we set $q = 1$ [*Zuur et al.*, 2003]. In addition, all data were *z*-scored to account for differences in the mean of stream temperatures among the streams.

Candidate models (i.e., models with different combinations of trends, error structures, and covariate terms) can be viewed as different hypotheses describing how thermal regimes are structured and were compared using AIC based on the maximum likelihood of the model fit. We assessed the effect of air temperature as a covariate in the model as an indication of stream sensitivity to surface heat exchange after retransforming the stream-specific effect sizes. This value indicates the °C increase in stream temperature for every °C increase in air temperature. Stream geomorphic conditions (PC1 and PC2) were then regressed (ordinary least squares regression) against factor loadings for the common trend(s) and effect sizes of air temperature.

3. Results

3.1. Topographic Association With Snow and Rain Contributions to Streams

Hydrogen and oxygen stable isotope ratios from stream water and precipitation plotted closely to the global meteoric water line (GMWL); values of δ^{18}O and δ^{2}H (‰) in streams were bound between the two relationships defined for snow and rain but were typically closer to that of the snow MWL (Figure 1). Reduced major axis regression indicated that precipitation and stream water were not measurably altered by evaporation, as the slope of the relationship between δ^{18}O and δ^{2}H was not significantly different than that expected for the GMWL; $\delta^{2}\mathrm{H} = 10 + 8.0 \times \delta^{18}\mathrm{O}$ [*Craig*, 1961] (rain 8.16 ± 0.2, snow 8.0 ± 0.2, mean ± SD; Figure 1). However, rain orthogonally separated more than snow from the GMWL, and we infer that this difference reflects temperature-dependent fractionation rates during the seasons in which the precipitation was formed [*Gat*, 1996; *Clark and Fritz*, 1997]. Oxygen and hydrogen isotopes in rain showed deuterium in excess (*d*-excess) from the GMWL (3.6 ± 2.5‰; mean ± SD), while isotope ratios in snow (8.8 ± 1.5‰) had values much closer to 10‰, that of the GMWL [*Craig*, 1961; *Dansgaard*, 1964].

Results from a two source mixing model [*Moore and Semmens*, 2008] on *d*-excess revealed wide variation in the relative contribution of rain and snow among the study streams. Water in streams draining mountainous terrain was derived mostly from snowmelt (Figure 2a). In comparison, stream water on the eastern boundary of the study region, where watersheds are flatter and at lower elevation, was composed mostly of rain. We summarized broad-scale watershed characteristics of each stream using principal component analysis and evaluated the relationship between these watershed characteristics and the rain and snowmelt contribution to summer streamflow [*Lisi et al.*, 2013]. Watershed slope and elevation had high correlation loadings (−0.94 and −0.87, respectively) on PC1 (Figure 2b). PC2 was primarily explained by strong loadings of watershed area (0.92) and lake area (0.52). However, only PC1 had strong associations with variation in water source, explaining 69% of the variation among streams (ordinary least squares, $p < 0.01$; Figure 2c). Warmer streams were also strongly associated with rain-dominated watersheds, while cooler streams were associated with snow-dominated watersheds (Figure 2c).

3.2. Time Series Analyses, Stream Sensitivity to Air Temperature.

Dynamic factor analysis (DFA) provided excellent fits to the variation of daily summer stream temperatures for 25 to 42 streams distributed across three river basins during the summers of 2011, 2012, and 2013 (model fits Figures S1–S3 in the supporting information). For each summer of data, DFA reduced the dimensionality of several stream temperature time series providing an assessment of the strength of the air temperature covariate and the common trend not explained by air temperature that together best describe the data. Further, we determined if geomorphic and water source conditions explained stream-specific associations to air temperature and the common trend. We assessed the strength of effect sizes on the air temperature time series (Figures 3a and 3c) and coefficient loadings on the common trend (Figure 3; see Figure S4 in the supporting information for coefficient loadings) for each stream and year of the study. Again, the effect sizes of air temperature indicate the Δ°C increase in stream temperature for every Δ°C increase in air temperature for each stream (Figure 3c).

Model results demonstrate that rain-dominated streams draining flatter watersheds had thermal regimes that were more sensitive to variation in air temperature compared to streams draining steeper topography dominated by snow. Stream-specific temperature sensitivities ranged by approximately fivefold to eightfold among streams, showing both cooling and warming relationships with increases in air temperatures (Figure 3c). For example, in 2013, for every 1°C increase in air temperature, the warmest stream responded by 0.65°C, while the coolest stream responded by −0.19°C—or a cooling effect of warmer air temperature. Temperature sensitivities were strongly associated with geomorphic characteristics of the watershed (PC1 loadings) during each summer ($r^2 = 0.61$, 0.68, 0.83; 2011, 2012, 2013, respectively; $P < 0.01$ all years; Figure 3c). Compared to flatter watersheds, steep watersheds that had stream thermal regimes were best

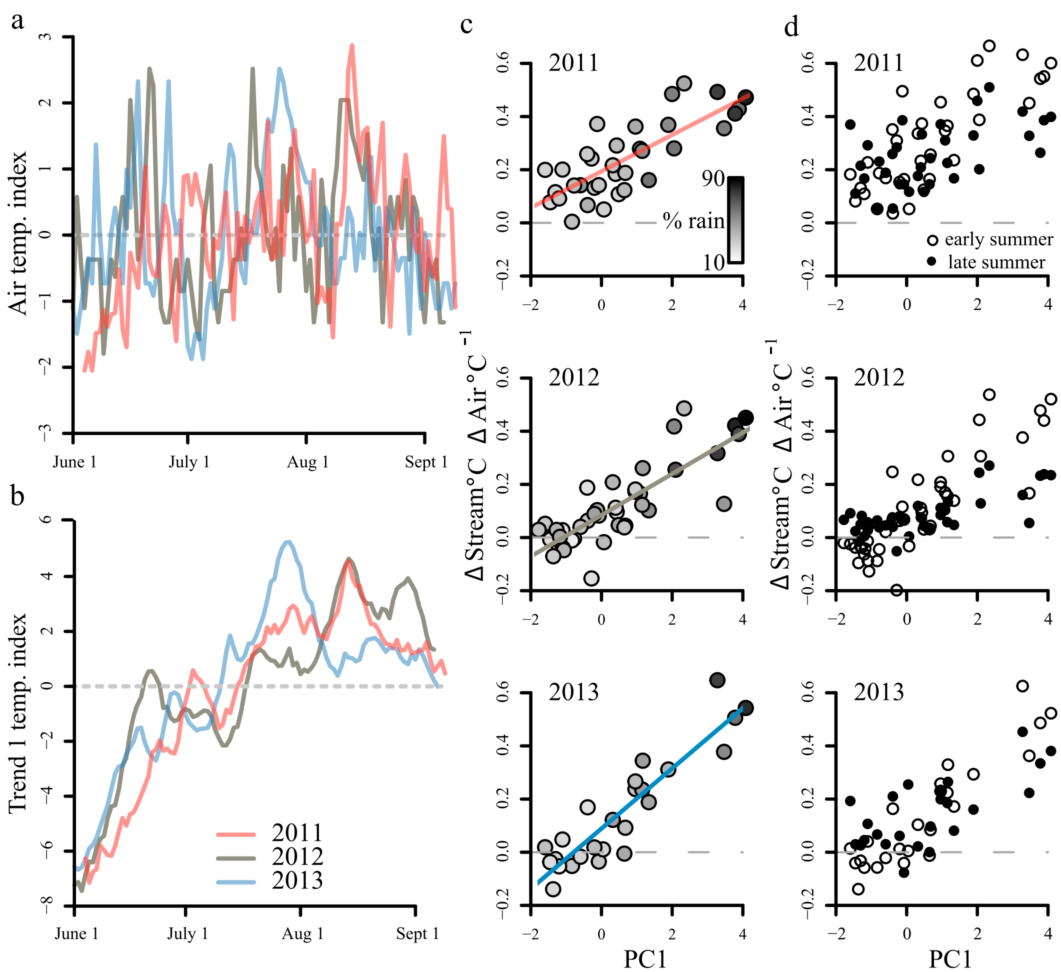

Figure 3. Results from DFA. (a) *Z*-scored air temperature time series for 3 years (2011, 2012, and 2013), (b) the most parsimonious temperature trend not explained by air temperature for the 3 years, and (c) the effect sizes of air temperature as a function of watershed topography as PC1 (column of three figures). Points are shaded by their median % rain. Year 2011 had a higher ratio of snow in all streams compared to recent years, 6% higher ratio than 2013, 4% higher than 2012, pairwise *t* test, $P < 0.01$. (d) The last column, the effect sizes of air temperature of early summer (open points) before 15 July compared to late summer after 15 July (filled points) as a function of watershed topography as PC1.

described by the common trend, reflecting their low level of thermal variation, cooler overall June and July temperatures, and snowmelt hydrology (Figure 3c). The steepest streams often showed cooling responses to increased air temperatures, a response to increased snowmelt during warm weather. Across the geomorphic gradient, we found a more homogenous response to air temperature among streams (e.g., larger intercept and lower regression slope) during the summer of 2011 (Figure 3c) that corresponded to a higher proportion of rain relative to snow in all streams during this year. With a multiple regression (analysis of covariance (ANCOVA)), we found more model support for a linear model (air temperature effect size versus PC1) that included year-specific intercepts and slopes compared to simpler models with shared intercepts or slopes ($\Delta AIC > 6$, ANCOVA). In particular, this result seemed to be driven by positive temperature sensitivities in steeper draining watersheds in 2011 (Figure 3c) compared to negative temperature sensitivities to air temperature in 2012 and 2013. This response in stream temperature regimes and lower proportion of snow in stream water was supported by local observations and snow monitoring stations (Snotel http://www.wcc.nrcs.usda.gov/snow/) that recorded below normal snow depths for the winter of 2011 compared to above normal snow depths in 2012 and 2013.

Stream-specific sensitivities to air temperature from DFA represent an average sensitivity for each stream over the entire summer. Indeed, stream-specific responses to air temperature are likely more dynamic through the summer as snow contributions to streamflow are replaced by rainfall, or as day length declines after summer solstice. To explore these relationships, we compared stream-specific summer sensitivities to air temperature before and after midsummer (15 July). This model produced an overall better fit of the data ($\Delta AIC \gg 10$) compared to the model with a single air temperature covariate for the entire summer (Table S1 in the supporting information). Stream-specific sensitivities from early summer closely matched those sensitivities produced from previous models that air temperature over the entire summer (least squares, $r^2 = 0.98$, d.f. = 98, slope = 1.15, intercept = 0). However, late summer sensitivity scores were more homogenous over the geomorphic gradient (Figure 3d). Late summer sensitivities became more sensitive in steeper, snow-dominated streams, and less sensitive in streams draining flatter, rain-dominated watersheds relative to the stream-specific sensitivity of early summer (Figure 3d).

We also found further evidence for geomorphic controls on stream thermal variation in the error structure of the variance/covariance matrix of the MARSS model, where streams with similar geomorphic characteristic (e.g., slope and elevation) features had higher levels of covariance across space (see Table S1 in the supporting information). We found substantial support ($\Delta AIC \gg 10$) for each stream having unique temperature variance and each pair of streams having a distinct covariance between their temperature responses ("unconstrained" error matrix, model 1, see Table S1 in the supporting information). This result suggests that thermal responses described by the error are not equivalent across streams but also not entirely independent. Stream pairs that shared positive covariance were often closely related in their watershed geomorphic characteristics (PC1: slope, elevation, and lake area), while negative covariance was often found between streams with larger differences in their geomorphic conditions. These relationships held even after controlling for spatial autocorrelation by geographic distance between streams (partial Mantel $r = -0.29$ (2011), -0.44 (2012), -0.53 (2013), $P < 0.01$ all years, see Figure S5 in the supporting information). These results suggest that further, undescribed variation in the modeled temperature is likely related to geomorphic conditions rather than spatial autocorrelation.

4. Discussion

Our results suggest that variation in stream thermal regimes is driven by hydrologic complexity, which is ultimately produced by watershed geomorphic features that can vary across river basins. These results indicate that for many streams in southwestern Alaska, stream thermal response to summer air temperature is controlled by the slope and elevation of watersheds, where contributions of snow buffers the thermal regimes of streams draining steeper and higher watersheds. However, our data also suggest that when snowpack is lower, streams draining steeper topography are less buffered to summer air temperatures, creating a more homogenous response to air temperature across river basins. We conceptualize that snow and snowmelt are retained in surface snowpack or within the alluvial aquifer longer into the summer in steeper basins than that of lower sloped, lower-elevation watersheds. Steeper, higher-elevation topography can also modify the heating capacity of these streams through shading, upwelling of snowmelt in groundwater, and short surface residence times in the stream channels [*Caissie*, 2006]. Lower gradient streams have longer surface water residence, higher levels of channel sinuosity, contain lake features, and thus more susceptible to changes in air temperature and solar radiation inputs during summer months [*Caissie*, 2006]. In addition to the thermal associations with air temperatures along the geomorphic gradient, we propose that the proportion of snow in stream discharge has important effects on stream thermal regimes in during the summer.

Streams draining higher-elevation watersheds may be thermally buffered by snowmelt during summer because advective fluxes of cool water overwhelm surface energy exchanges associated with increasing air temperatures [*Leach and Moore*, 2014]. Our results show that in watersheds where snowmelt contribution to streamflow was less, daily stream temperature increases with increasing air temperature. Negative thermal sensitivities from snow-dominated watersheds suggest an increasing influence of cold water from upstream snow-covered sources that melt during warm, sunny weather. In this study, surface snowpack typically melts by mid-July; thus, advective fluxes from upstream snowmelt may rapidly diminish as summer progresses. Still, snowmelt can be a key source of subsurface aquifer recharge in steeper basins that retain more snow [*MacDonald et al.*, 2014]. Throughout the summer, we found predominately snow isotope signatures in upwelling cool-water springs and seeps (2–4°C) that drained steeper watersheds and rain isotope signatures were found in warmer (8–14°C) shallow subsurface inflows draining flatter peatlands. We speculate that steeper watersheds have deeper alluvial aquifers relative to flatter basins and thus may retain lateral subsurface snowmelt inflows.

Our results also show greater contrast in sensitivities to air temperature among streams with different geomorphology in early summer and less contrast in late summer. This result supports our hypothesis that different snowmelt contributions produced by watershed geomorphic features drive thermal sensitivities to air temperature. Steeper watersheds that retain more snow had negative sensitivities to air temperature during early summer (cooling with warmer air) and positive sensitivities in late summer. We presume that these subseasonal differences are the result of a loss of snow and replacement by rainfall in streamflows in snow-dominated streams, therefore producing temperature sensitive to air temperature that was less distinct among all streams. Thus, stream-specific responses to air temperature are likely more dynamic than currently appreciated and further suggest that air to stream relationships are nonstationary over annual and subseasonal scales [*Arismendi et al.*, 2014]. Those wishing to use air temperature and stream temperature regressions to project future stream temperatures will need to consider not only how watershed geomorphology mediates the relationships over space but also how changes to hydrologic sources might influence stream to air temperature relationships through time.

The results highlight that ongoing climate change will have different effects on stream thermal regimes across river basins due to variation in the physical characteristics of stream catchments. These diverse expressions of climate are an important attribute of river systems and often an underappreciated dimension of ecosystem complexity. In this region, spatial thermal heterogeneity across a river basin is important to wild salmon spawning habitat and the terrestrial species that rely upon them. Here variation in the phenology of salmon spawn-timing, determined by variation in water temperature [*Lisi et al.*, 2013], extends the foraging season for predators and scavengers [*Ruff et al.*, 2011; *Schindler et al.*, 2013]. Ongoing climate change may affect the extent of this phenological variation because of substantial changes to snowpack as well as increased air temperatures. However, our results imply that some fine-scale thermal and hydrologic heterogeneities may be lost due to expected changes in winter snowpack, earlier timing of melt-off, and increased ratios of rainfall to snowmelt in stream discharge [*Maurer et al.*, 2007; *Scenarios Network for Alaska and Arctic Planning*, 2014].

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