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Key Points:
- Cold streams are less sensitive to warming than warm streams
- Warm streams had greater variance in sensitivity estimates than cold streams
- Interannual scale sensitivity estimates may vary from subannual scale estimates

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Sensitivity of summer stream temperatures to climate variability in the Pacific Northwest

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Abstract Estimating the thermal response of streams to a warming climate is important for prioritizing native fish conservation efforts. While there are plentiful estimates of air temperature responses to climate change, the sensitivity of streams, particularly small headwater streams, to warming temperatures is less well understood. A substantial body of literature correlates subannual scale temperature variations in air and stream temperatures driven by annual cycles in solar angle; however, these may be a low-precision proxy for climate change driven changes in the stream energy balance. We analyzed summer stream temperature records from forested streams in the Pacific Northwest for interannual correlations to air temperature and standardized annual streamflow departures. A significant pattern emerged where cold streams always had lower sensitivities to air temperature variation, while warm streams could be insensitive or sensitive depending on geological or vegetation context. A pattern where cold streams are less sensitive to direct temperature increases is important for conservation planning, although substantial questions may yet remain for secondary effects related to flow or vegetation changes induced by climate change.

1. Introduction

Stream temperature exerts a primary influence on aquatic biota [*Caissie*, 2006; *Poole and Berman*, 2001] and is important in structuring fish species distributions, abundance, maturation rate, and life histories [*Brannon et al.*, 2004; *Dunham et al.*, 2007; *Holtby*, 1988; *Pörtner and Farrell*, 2008; *Torgersen et al.*, 1999; *Wenger et al.*, 2011a]. Global warming is expected to increase stream temperatures in the western United States by altering water and energy balances. Earlier streamflow timing [e.g., *Stewart et al.*, 2005] more closely synchronizes annual streamflow minima with the annual solar maximum [*Arismendi et al.*, 2012b], and declines in precipitation have exacerbated the decreases in summer runoff [*Leppi et al.*, 2011; *Luce and Holden*, 2009; *Luce et al.*, 2013]. At the same time, warmer air temperatures and increased atmospheric emissivity are increasing downwelling longwave radiation [*Dettinger*, 2012]. Increasing wildfire [*Holden et al.*, 2012; *Westerling et al.*, 2006] and vegetative conversions related to drought conditions [*Breshears et al.*, 2005; *van Mantgem et al.*, 2009] may further increase temperatures by decreasing shading [*Dunham et al.*, 2007; *Isaak et al.*, 2010]. Consequently, changes in stream temperatures may be one of the more significant effects of climate change on stream biota, particularly the most sensitive cold-adapted species [*Rieman et al.*, 2007; *Wenger et al.*, 2011b].

Historical trends in stream temperature around the world have shown increases in water temperature in recent decades [*Bartholow*, 2005; *Hari et al.*, 2006; *Isaak et al.*, 2012, 2010; *Kaushal et al.*, 2010; *Langan et al.*, 2001; *Morrison et al.*, 2002; *Petersen and Kitchell*, 2001]. However, because connections between climate and stream temperature are complex, not all streams are warming equally [e.g., *Arismendi et al.*, 2012a; *Hari et al.*, 2006; *van Vliet et al.*, 2010; *Webb and Nobilis*, 1997] Long-term records are exceptionally useful for interpretation of trends, but such records are rare, particularly for high elevation streams [*Arismendi et al.*, 2012a; *Isaak et al.*, 2012; *Kaushal et al.*, 2010].

Efforts have consequently turned to understanding the sensitivity of stream temperature to variations in climate using shorter records. Nonlinear regression of daily or weekly stream temperatures against air temperatures have been one of the more common approaches to assessing stream sensitivity using records of just one to a few years of length [*Kelleher et al.*, 2011; *Mohseni et al.*, 1998, 2003; *Morrill et al.*, 2005]. Some efforts have demonstrated that sensitivity varies based on the degree of groundwater contribution [e.g., *Kelleher*

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et al., 2011; *Mayer*, 2012]. Although calibrated process-based models could be applied under downscaled GCM projections, there are sufficient questions about model parameter identifiability, that some aspects of sensitivity may be more accurately assessed empirically [e.g., *Sivapalan et al*., 2003].

Inferences from empirical sensitivity assessments depend on time scales of data used in the regression. It is well understood that direct warming of streams through convective heat transfer from the air is small compared to radiative transfers [*Johnson*, 2003, 2004; *Leach and Moore*, 2010]. Thus, regressions of stream-temperature on air-temperature exploit their joint variation with incoming radiation. In the case of regressions done using intraannual temperature data, the primary source of temperature variability in both air and stream temperatures is the annual cycle of incoming shortwave radiation related to the changing solar angle. The mechanistic linkage between annual cycles of warming and the kind of warming occurring at interannual to interdecadal scales during summer months is weak, however, because solar angles between years are consistent, and changes in solar radiation from summer to summer are expected to be small relative to changes in downwelling longwave radiation [e.g., *Dettinger*, 2012]. While annual relationships based on variation in shortwave radiation provide exceptional indices of agreement [*Mohseni et al*., 2003; *Pilgrim et al*., 1998; *Stefan and Preud'homme*, 1993; *Webb et al*., 2003], standard errors of prediction can be relatively large. Consequently, direct examination of the interannual relationship between summer stream temperature and air temperature may yield a more precise measure of sensitivity that is also directly relevant to many ecological questions [e.g., *Isaak et al*., 2010; *Wenger et al*., 2011b]. As mentioned earlier, changes in streamflow may also affect sensitivity by altering a stream's assimilative capacity, and have been demonstrated to improve estimates of air temperature parameters in recent regression modeling [*Caldwell et al*., 2013; *Isaak et al*., 2010; *van Vliet et al*., 2010; *Webb et al*., 2003].

Stream-atmosphere connectivity is one way to frame the problem of summer stream sensitivity. In many mountain streams in the temperate zone, much of the water starts as snowmelt in the spring. Warming occurs when the water is exposed to solar radiation, longwave radiation from the atmosphere and, to a smaller degree, direct turbulent transfers of energy from the atmosphere. Some warming also occurs as water travels through the ground, but this is generally much less than occurs from sunlight, and such warming is not generally dependent on weather variations. Thus, streams that flow from areas with large groundwater reservoirs (with long residence times) or persistent snowpacks and glaciers, or that flow under dense forest canopies would be expected to be both generally cooler and much less sensitive to atmospheric variations. Applying this theoretical construct would suggest that cooler streams are less sensitive than warm streams, which has significant implications for aquatic species conservation strategies. It would also mean that initial approximations of sensitivity for a stream of unknown sensitivity could be derived from estimates or direct measurements of the summer stream temperature. Here we test this hypothesis using a large number of stream temperature measurements from National Forests in the Pacific Northwest United States (Figure 1).

2. Methods

2.1. Setting

The Pacific Northwest has several mountain ranges that play important roles in the region's hydroclimatology. Principal among these are the Cascade Mountains of Oregon and Washington, which run north and south just west of the center of the two states, and which contain many of the sampled streams (Figure 1). Most of the Cascades are below 2000 m elevation, but several stratovolcanoes exceeding 3000 m (and a few exceeding 4000 m) dot the range. A few small glaciers are on these volcanoes and in mountains at the north end of the range. Much of the bedrock in sampled areas is older andesite volcanics with shallow soils and bedrock with little storage. However, a few streams in the Cascades of central Oregon have more recent volcanic flows with large groundwater reservoirs that strongly influence stream temperatures [*Tague et al*., 2007].

To the west of the Cascades, elevations are generally lower, and precipitation is high, commonly greater than 2 m annually. Average summer temperatures on the coast are the coolest in the region, close to 18°C, while further inland on the western edge of the Cascades, temperatures range from 22°C to 24°C, depending on elevation and latitude. Several samples were taken in streams flowing from the Oregon Coast Range, with elevations generally below 1000 m. Winter snow cover there is brief. Bedrock is primarily a massive

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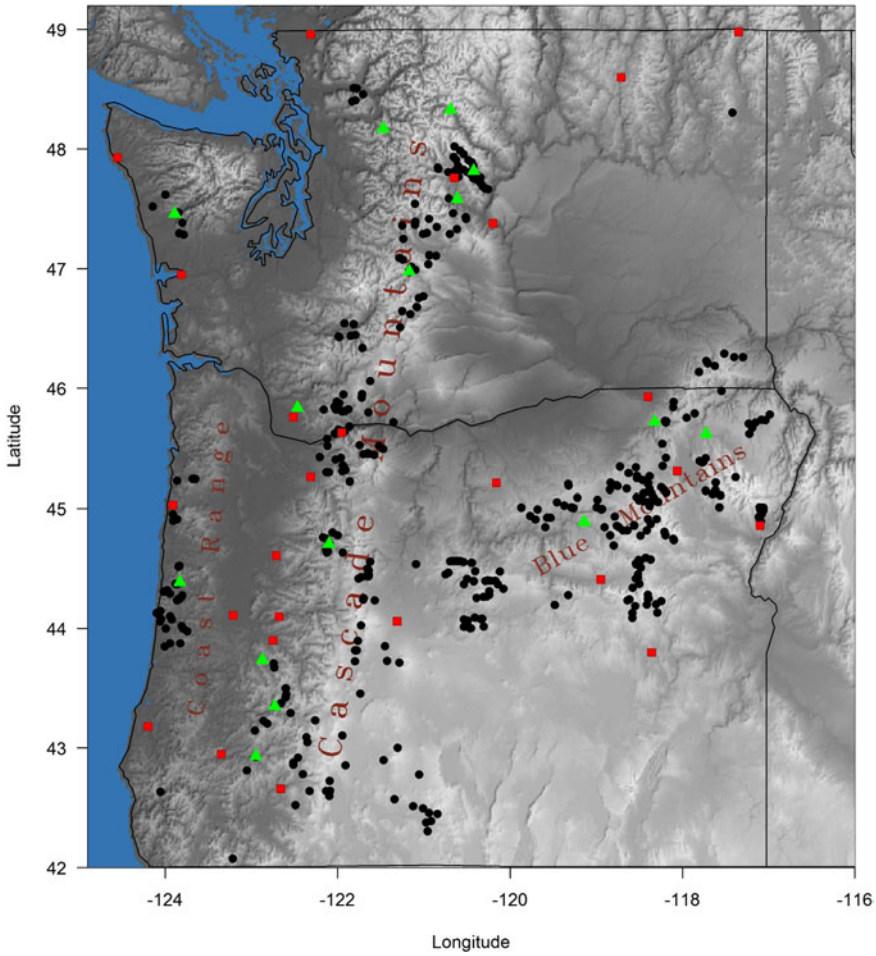

Figure 1. Map of stream temperature (black dots), air temperature (red dots), and streamflow (green triangles) stations.

sandstone with some older volcanics. Forests of the western Cascades and coast range are extremely productive, dense, and tall.

Conditions are drier and warmer to the east of the Cascades, with precipitation generally less than 1 m annually except in higher mountains, and temperatures ranging from 22°C to 28°C depending on elevation and latitude. Stream temperature stations were located in a few smaller mountain ranges, but principally draw from the Blue Mountain province. Elevations tend to be less than 2000 m, though there are a few higher peaks. Bedrock varies substantially, even over small areas, and includes flood basalts, fossil-bearing sedimentary rocks, rhyolitic tuffs, and granite batholiths. Forests in these drier areas have lower densities and tend to be shorter than the forests of the Cascades and coast range. Riparian meadows are relatively common, as are rangelands.

2.2. Stream Temperature Data

We obtained stream temperature data from a U.S. Forest Service database comprising summer stream temperatures taken from 12 National Forests in Oregon and Washington. Summer stream temperatures were collected with at least hourly resolution over the course of the summer months (June–September) and recorded on one of several brands of temperature dataloggers. Records were maintained at many sites with important fish habitat in the region to assess baseline conditions and monitor changes over time. We excluded (1) temperature records from sites in close proximity to recent tree harvest or other land management activities (e.g., intensive grazing, streamside roading) that were likely to affect temperatures, (2) sites with upstream water management activities that showed a change in practice over the period of record, (3) units near or downstream of severe natural disturbances, and (4) units known to have changed location

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within the stream in a way that could have affected temperature (e.g., if there was a change in which stream bank was monitored at a site located a short distance downstream from a tributary). Thus, we do not expect that substantial changes in land management practices affected the time series used in the analysis.

We checked each summer of data to identify erroneous readings or incomplete records. Errors included start or end dates likely to exclude the seasonal maximum and anomalous temperature readings when a sensor was exposed by low flows or removed from the stream. After excluding problematic records, we identified 256 temperature stations with at least seven summers of data between 1988 and 2010.

We calculated the maximum weekly maximum temperature (MWMT) for each summer at each site. We further checked these data for strong outliers that might indicate issues not discoverable from previous quality checks. We used the following rules to exclude time series or individual years from further analysis:

1. For stations with a temperature range greater than 5°C, we eliminated any years with temperatures greater than 1.5 times the interquartile range over the 75th percentile of the data. This may identify poor data and also prevented results from being overly influenced by a single year with large leverage. (Nine stations lost enough years to be eliminated.)

2. For stations with a temperature range less than 5°C, we eliminated those with an outlier greater than 7.5 times the interquartile range over the 75th percentile. This allowed for retaining some data on streams with small interannual variability that may technically be outliers, but were still very close to all other measured temperatures. (One station eliminated.)

After all records were examined, we retained 246 stations with record lengths averaging 10 years (range = 7–23 years; Figure 1). Some data/stations may have been eliminated in error, leading to a slight bias toward estimating lower sensitivity in warmer streams because these produce more suspicious temperatures.

2.3. Air Temperatures and Streamflows

The stream temperature stations were distributed in remote areas across the National Forests of the Pacific Northwest, mostly on streams without flow gauging and in areas without air temperature measurements. Over an area of this size, the patterns of warm and cold summers (or wet and dry) are not always synchronous, so some means of estimating the local temporal pattern in air temperature and streamflow was necessary. We applied principal components analysis to reconstruct patterns in interannual variations of the maximum weekly average temperature (MWAT) of air and annual streamflow. The exercise is similar to reconstructing a time series of maps, a common exercise in climatology and meteorology [*Lundquist and Cayan*, 2007; *Preisendorfer*, 1988], but in this case applied with the principal components as time series [e.g., as in *Holden et al*., 2011]. Because all air temperature stations measure temperature with the same scaling, we performed the principal component analysis directly to the temperature anomalies (difference from the average temperature) without rescaling. Because streamflow stations gage basins of varying size and precipitation, it was necessary to first standardize the anomalies from each station by the standard deviations to give standardized departures from the mean, amounting to an index of the relative wetness or dryness of each year. Twenty-five air temperature stations were selected from among COOP stations in the US Historical Climatology Network (USHCN) proximal to the field of stream temperature stations with records covering the 23 years available from the stream temperature sensor network (Figure 1). Similarly, 15 streamflow stations were selected from streams proximal to the stream temperature stations (Figure 1).

Principal component reconstruction is done by decomposing a matrix of observations from several stations over a series of years into a matrix of loadings, which are associated with each station, and a matrix of principal components, which are, themselves, time series. While this may sound like it is making the problem more complicated, the principal components can be thought of as archetypal patterns, and the principal components are ranked based on the relative degree of information about the original time series they contain. Furthermore, the loadings are now a spatial pattern of how strongly each station participates in each of the principal components. So if a particular station shows variations over time that look very much like the first principal component, it will have a strong loading on the first principal component. Conceptually, as stations show variations on the first principal component, additions and subtractions of linear fractions of the second and higher order principal components can be used to describe them. So if we have a

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decomposition of the observations, X_{sy}, into loadings, L_{si} and principal components P_{iy}, where y is an index of years, s is an index of stations, and i is an index of principal component rank,

$$\mathbf{X}_{sy} = \mathbf{L}_{si}\,\mathbf{P}_{iy} \tag{1}$$

or to illustrate for a specific station, a,

$$\mathbf{X}_{ay} = \mathbf{L}_{a1}\mathbf{P}_{1y} + \mathbf{L}_{a2}\mathbf{P}_{2y} + \ \ldots \ + \mathbf{L}_{an}\mathbf{P}_{ny} \tag{2}$$

where each L_{an} is the loading of station a on the nth principal component and P_{ny} is the time series of value in the nth principal component, and X_{ay} is the time series of observations at station a. Equation (2) is exact if all principal components are used, however most of the information in a matrix of weather observations with a degree of spatial correlation is typically found in the first few principal components, and a reasonable approximation of a time series can usually be made with sometimes just one or two principal components, such as found by *Holden et al*. [2011] for reconstruction of temperatures in complex terrain over a small area. We applied Parallel Analysis [*Horn*, 1965] to objectively define an optimal number of principal components for reconstruction.

Air temperature anomaly and standardized flow anomaly time series were reconstructed for the location (latitude and longitude) of each stream temperature station, and stream temperatures were regressed against the anomaly values using multiple regression. The coefficients of the regression model at each station are the sensitivities in °C/°C for air temperature and °C/SD(q) for flow. Given the relatively short records, we did not test interactions. To test our hypothesis, the sensitivities were plotted and regressed against the mean temperature, and statistical significance of individual regressions was assessed using a t test with $\alpha = 0.10$. Further examination of patterns of sensitivity provide additional insight.

3. Results

3.1. Principal Component Reconstructions

The Parallel Analysis suggested the leading two principal components (60% and 21% of the variation, respectively) from the 25 stations, however the third principal component (5% of variability) had interesting behavior with respect to the first principal component and is worth discussing with respect to how the reconstructions were ultimately done. The first PC loaded relatively equally on every station except the coastal stations, which had small loadings. For most of the region, it represents a mean behavior. The coastal stations have very little interannual variability, so load lightly on the mean. The second principal component shows a northwest-to-southeast trending pattern, and a fit of the second principal component loading versus latitude and longitude had $r^2 = 0.81$ (Figure 2). Stations along a given southwest-to-northeast contour share similar timing and temperature amplitudes. The spatial pattern may reflect a combination of latitude and coastal influence on summer temperature variability. The third principal component loads mostly on the coastal stations, the same stations with small loadings on PC1 (Figure 3); and has very little temporal variability. This relationship led to a unique approach in reconstructing temperatures for the coastal areas. We used only PCs 1 and 2 for reconstruction of air temperature patterns for coastal streams because there was no way of knowing a priori whether a particular stream or station on a stream was affected by the coastal air temperature pattern or interior, or, as is more likely, a combination across a given watershed. Consequently, interpretation of the sensitivities from measurements close to the coast should recognize that the fog and cool advected ocean air may be an implicit factor in their lower sensitivity. Conversely, streams near the coast that have high sensitivity may experience little summer fog. The PC1 loading was given as a uniform value assigned as the average of the 21 noncoastal station PC1 loadings and PC2 was regressed against latitude and longitude (Figure 2). Reconstructions of the best and worst stations using this reconstruction approach for air temperature anomalies are shown in Figure 4, excluding reconstructions for the coastal stations, which would not be expected to be well described by this particular reconstruction.

For standardized streamflow, Parallel Analysis highlighted the first two principal components, explaining 91% of the data (78% in the first PC and 13% in the second). The first PC varied in loading between 0.22

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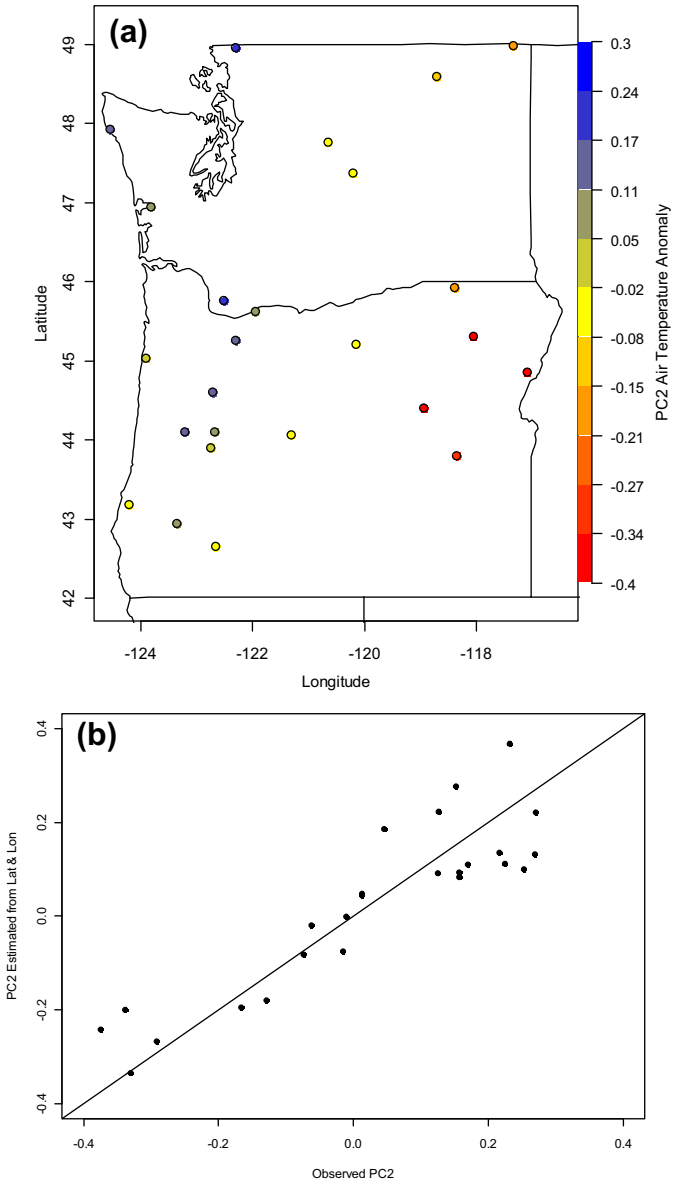

Figure 2. (a) Map of second principal component for air temperature anomalies and (b) fit of the second principal component for air temperature on latitude and longitude.

and 0.28, so represented a mean behavior in the region, while the second PC again varied from northwest to southeast with $r^2 = 0.95$ (Figure 5), a pattern broadly consistent with interannual N-S variations in storm tracks. Storms of tropical or subtropical origin tend to be the largest storms for a given season (carrying much of a year's precipitation in a single event), yielding a SW -> NE track through the region [*Neiman et al.*, 2008]. These two PCs were used to reconstruct annual flows. Figure 6 shows the stations with the best and poorest reconstruction fits using this technique.

We also attempted PC reconstruction of summer flows (defined as 15 July to 15 August), however five PCs were required to obtain similar performance (91% of the variation explained). Although it may have been possible to relate the loadings to basin characteristics, it is difficult to physically interpret and independently model more than a few PCs. There are, however, relatively strong correlations between summer and annual flows in the snowmelt dominated streams in the region [*Luce and Holden*, 2009]. Correlations for the stations used in this study (July, August, September versus annual flows) range from 0.21 to 0.86, with the majority greater than 0.73 (Figure 7). The fraction of water delivered in these months is small, ranging from 2% to 26% of the annual total, but even stations with only 2% of the flow in summer showed correlations as high as 0.7. Such behavior is related to the fact that much of the regions precipitation occurs in winter months, with summer flows sustained by release from various stores such as residual snowpack and groundwater, yielding a recession curve through summer months. This means that for many of the stations in the analysis, the annual flow serves as a reasonable index of summer flows. Lower elevation stations, particularly coastal stations, may not be as well served by this approximation.

3.2. Stream Temperature Sensitivity

The sensitivity of stream temperatures to air temperatures ranges from −0.53 to 1.95 and averaged 0.51°C/°C. This is generally consistent with the range of values for interannual variability reported by *Moore et al.* [2013] for MWAT on 22 streams in British Columbia with 6–9 years of data. Less than 2.5% of our sites have negative slope less than −0.10°C/°C, and the fit was not statistically significant at these sites. In general, colder streams exhibit lower sensitivity to air temperature variations (Figure 8), while warmer streams show

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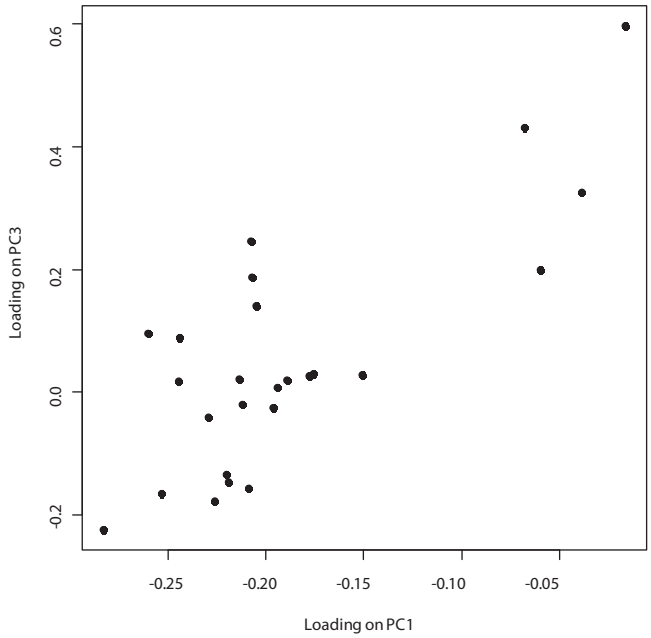

Figure 3. Station loadings for the third principal component for air temperature versus loadings for the first principal component. The four coastal weather stations show relatively stronger loading on PC3 and small loading on PC1.

greater variability in sensitivity. Overall, the pattern shows a limit wherein cold streams do not show high sensitivities, while warm streams express a range of sensitivity from high to low, with a tendency for higher sensitivity. Note that this pattern is not likely to have been influenced by the QA/QC procedures which would have been expected to produce a bias in the opposite direction, if any. The spreading pattern of sensitivity with increasing temperature is not appropriately fit using a least square error model; however, it is reasonably characterized by quantile regression. A median fit through the data has a slope of 0.028°C/°C^2, with $P = 6 \times 10^{-5}$. The upper limit of the pattern is more striking, however, and the 90th percentile sensitivity is strongly conditioned on the maximum stream temperature with a slope of 0.053°C/°C^2, with $P = 7 \times 10^{-5}$. Cold streams are much more constrained in their expressions of sensitivity, probably due to less exposure to the atmosphere and incoming radiation.

A map of stream temperature sensitivity shows a few broadscale geographic patterns (Figure 9). The mountains of eastern Oregon seem to have the majority of the most sensitive streams. Streams in the Cascades seem generally less sensitive, though with a few exceptions. Beyond that, a few of the least sensitive stations are very close to the coast, but just a short distance upstream stations can be much more sensitive. Had a coastal air temperature been used in calculating the air temperature sensitivity on these, the sensitivities would have been extreme, so it is useful to characterize the sensitivity in terms of the broader continental variations in temperature. This then frames the coastal fog belt as a source of thermal damping.

Contrasting the sensitivity map to a map of the average of stream MWMT over the period of record provides further insight into the relationships between temperature and sensitivity. Figure 10 shows broadly cooler temperatures in the higher elevations of the Cascades, the warmest temperatures in NE Oregon, and moderate temperatures along the coast. The moderate temperatures for coastal streams are related to the ocean influence and their low elevation. The low sensitivity of these same streams would explain some of the low sensitivity streams with moderate temperatures seen in Figure 8. There is also a group of streams in central Oregon with low sensitivity and warm temperatures, which may relate to geologic controls on temperature and sensitivity. There are a few small groups of cold, insensitive streams around the map. In general, the spatial structure of the sensitivity map is finer grained than the patterns in the average MWMT map, suggesting that there are specific, relatively local conditions leading to the variations in sensitivity that may not reflect the broadscale drivers of temperature, such as elevation, precipitation, vegetation cover, and streamflow seasonality.

Sensitivity to streamflow showed no clear patterns when plotted against the average MWMT (Figure 11). While most streams had the expected relationship of declining maximum temperature with increasing flow, $^1/_4$ of them had positive slopes, though generally weaker slopes than the negative slopes. Five of the eight streams with strong negative air temperature sensitivity are associated with streams that also had positive streamflow sensitivity. While the annual flow to summer flow correlation may be adequate for some stations, it seems likely that a more specific summer flow model would yield better performance. There are no discernible spatial patterns in the streamflow sensitivity.

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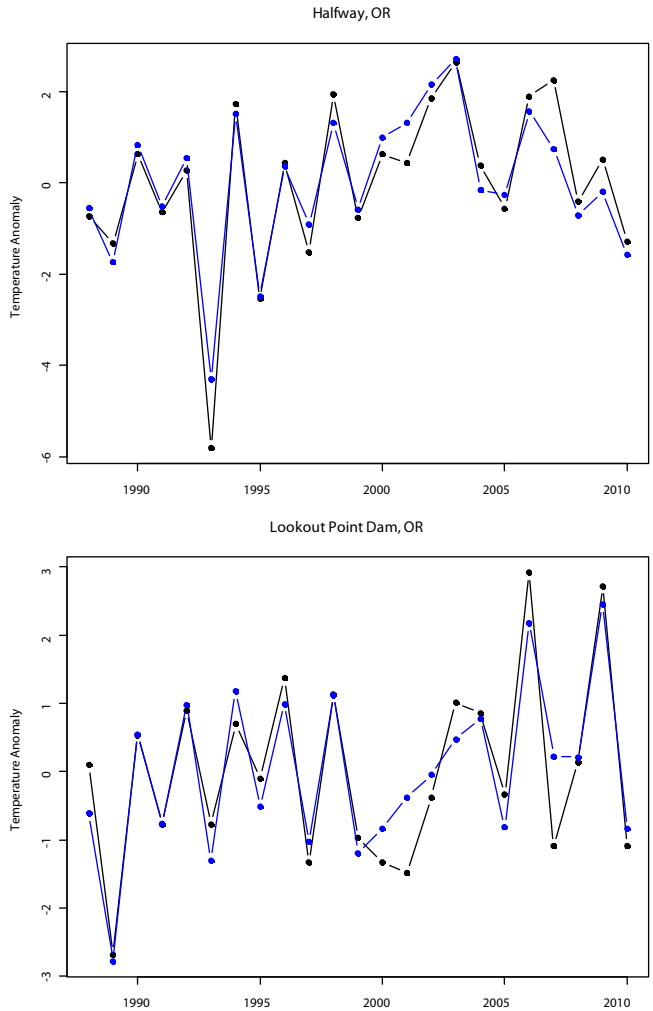

Figure 4. (top) Best and (bottom) worst PCA reconstructions of air temperature anomalies.

4. Discussion

Our results show that virtually all cold streams are insensitive, while warmer streams may or may not be sensitive, although there is a general trend of increasing sensitivity with increasing stream temperature. The finding that cold streams tend to have lower direct sensitivity to warming air temperatures has potentially important implications for conservation of aquatic species despite the fact that it appears, on the surface, to contradict other projections of the distribution of future stream warming.

Other empirical analyses have linked air temperature to stream temperature to address potential climate change effects on fishes, particularly cold water-dependent species [*Isaak et al.*, 2010; *Rieman et al.*, 2007; *Wenger et al.*, 2011b]. They have generally assumed a linear response to air temperature changes. The findings presented here suggest that some of the colder streams will be less vulnerable than predicted considering uniform sensitivity, further reinforcing the point of *Isaak and Rieman* [2013] that temperature isotherms delimiting species distributions will move more slowly in the highest elevation portions of stream networks. These remaining colder and less sensitive habitats, sometimes referred to as microrefugia [*Dobrowski*, 2011], will be important areas for conserving and protecting native organisms efforts, and, if necessary, restoration efforts to reinforce population viability. It is important to note, however, that fishes in small habitat patches separated by large warm stretches of stream may experience increased risk of extirpation by fire and other large-scale disturbances one patch at a time [*Isaak et al.*, 2010; *Rieman et al.*, 2010, 2003].

Others have attempted mechanistic modeling of stream temperatures in this region [e.g., *Cristea and Burges*, 2009; *Null et al.*, 2013; *Wu et al.*, 2012]. These studies have sometimes predicted that colder streams will be more vulnerable to warming than warmer streams. Earlier snowmelt timing, and consequent lower summer streamflow, is one mechanism noted by the mechanistic models. This could well be the case, and the weak relationships obtained with our annual flow estimates would not be adequate to test or revise that assessment (see further below in discussion on how these assessments complement the empirical analysis). However, other results show high sensitivity in cold streams even without substantial snowmelt participation, and it may well be that their equilibrium formulations [e.g., *Null et al.*, 2013; *Wu et al.*, 2012] are overly prescriptive for the sensitivity of cooler headwater streams. Projections for streams in the Sierra Mountains of California show greater sensitivity at lower elevation because projected changes in snowmelt are smaller at higher elevations [*Ficklin et al.*, 2013].

A brief review of the physics of heat transfer is useful in understanding why cold streams are likely to be less sensitive and can help explain discrepancies with mechanistic models. It is well established that

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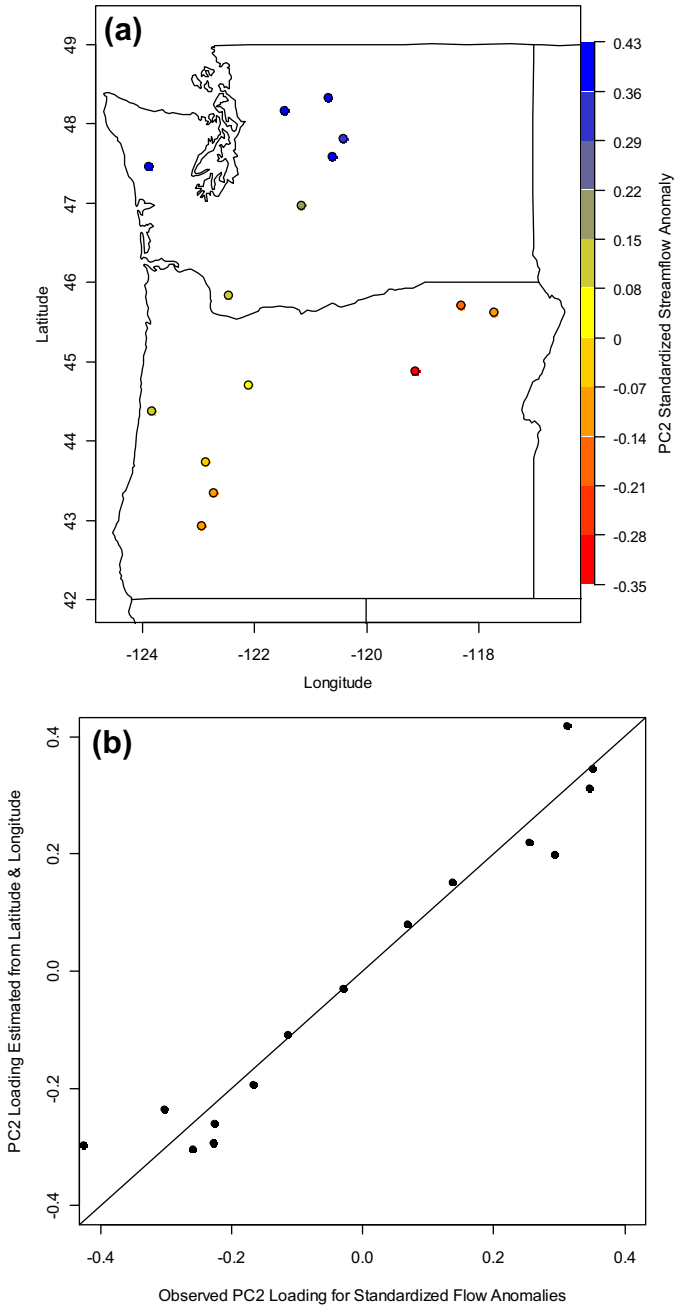

Figure 5. (a) Map of second principal component for standardized streamflow and (b) fit of the second principal component for streamflow on latitude and longitude.

changing air temperature does little to heat the stream through the direct turbulent transfer of sensible heat [*Johnson*, 2004; *Webb and Zhang*, 1997], and even then, the usual trade-off between sensible and latent heat transfer over water surfaces as the warmer air becomes drier can do much to offset any gains. One of the more direct ways for air temperature to influence energy input is through downwelling longwave radiation, which depends on the emissivity of the air column and its temperature [e.g., *Marks and Dozier*, 1979]. Increasing CO_2 and other greenhouse gases is increasing the emissivity somewhat. However, much of the expected effect on downwelling longwave radiation will result from an increase in the temperature of the air mass. *Dettinger* [2012], for example, estimated that this process would result in an increase of 30–35 W/m^2 per century under an A2 scenario over much of the western United States. Consider just a shift in the longwave radiation balance for a bowl of water that has come to equilibrium temperature with its surroundings. If we increase the downwelling longwave radiation by 35 W/m^2, a change consistent with about a 4°C temperature increase in air temperature in *Dettinger* [2012], the resulting increase in water temperature to reequilibrate would be around 6°C (depending slightly on initial temperature). This suggests a baseline "physically based" sensitivity of 1.5°C/°C, and it is readily apparent that the vast majority of observed sensitivities are substantially less than this theoretical equilibrium value. The simplest likely reason for this is that in the places the temperatures are measured, there is not the time for the water to come to equilibrium, whether it is because the river is well insulated from the atmosphere under trees or fog, or because there is a steady inflow of cool groundwater or snowmelt gradually building the stream as it flows downstream. Many of the streams in this analysis are in the headwaters, where the water is often coolest, as much of it may have seeped into the stream only a few hours before its temperature is recorded. If only a portion of the potential heating is realized under cooler air temperatures, only a portion of any additional heating will be seen under warmer temperatures.

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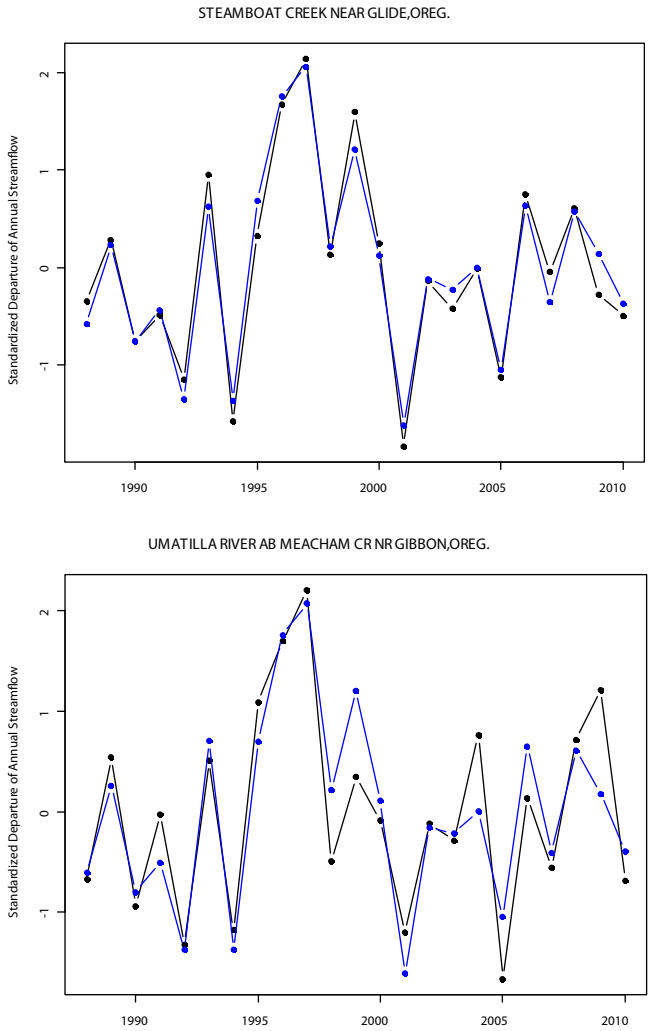

Figure 6. (top) Best and (bottom) worst PCA reconstructions of streamflow standardized anomalies.

4.1. Time Scales of Sensitivity Estimates

This review of physics also gives us a good opportunity to discuss the sensitivity estimates described in this paper in relation to sensitivities estimated from an annual cycle of temperature [*Mohseni et al.*, 1998]. At the subannual time scale, covariation in air temperatures and stream temperatures exists because both are warmed by the sun, and over the course of the year, solar loading varies dramatically. Nonlinearities in the relationship are introduced at both ends of the temperature spectrum. When the coldest air temperatures at a site are well below freezing, it is not unusual to see a substantial range of air temperatures for which the water temperature stays near 0°C. When the sun is at its highest, the solar loading changes little, and stream temperatures may change little or slowly, while air temperatures may see large excursions depending on large-scale air circulation. This will create some variations in stream temperature driven by variations in downwelling longwave radiation; however, the largest variations of air and stream temperatures on the time scale of a day to a week in the summer months are associated with clouds, which again yield larger variations in solar radiant inputs than longwave. While estimates of stream temperature sensitivity taken from one annual thermograph record or from several weeks or months of summer record would likely correlate well with the sensitivities calculated in this paper, as would a sensitivity calculated from a ratio of the diurnal temperature ranges, it would probably not match them. We believe the sensitivities calculated in this paper based on interannual variations may more accurately reflect potential changes related to long-term climatic trends, an assertion which could be examined using full-year data.

Another way to look at this same issue is to consider spring stream temperatures. According to the concept behind the subannual scale nonlinear fits, these would be expected to be the most sensitive, being on the portion of the curve with the steepest slope. However, in a specific consideration of spring temperature sensitivity in this region, most would recognize that the overwhelming amount of water might well make this season the least sensitive to increases in temperature. In an examination of stream temperature trends, *Isaak et al.* [2012] found that the magnitude of stream temperature changes in the fall and spring per degree of air temperature change were comparable to those in the summer, if not less. An approach that looks at the season and temperature of interest and then uses interannual observations to see the response to patterns has a firmer logical footing than one looking at subannual cycles confounded by variation in solar inputs.

An additional benefit to using independent samples of the warmest temperatures in each year in contrast to using data from throughout the year is that the residual errors are lower. In the models predicting the

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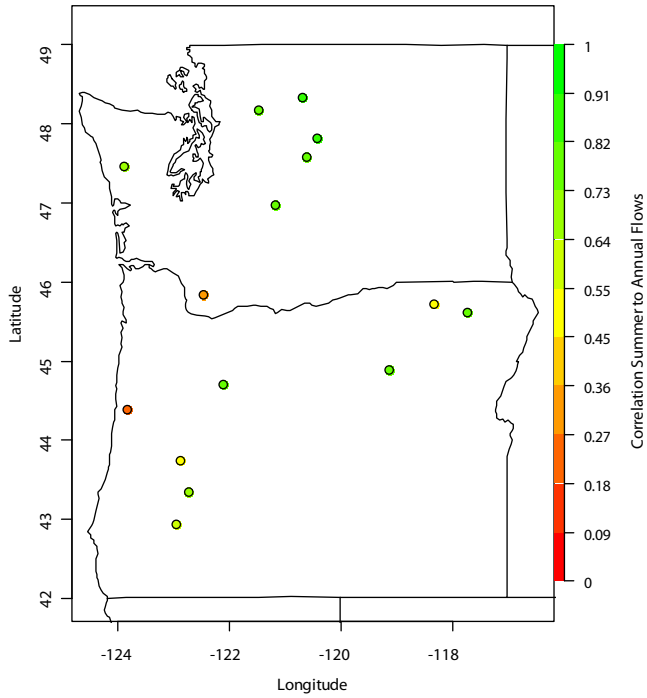

Figure 7. Map of correlation between summer (July–September) flows and annual flows.

full throw of temperatures from cold to hot, coefficients of determination or Nash-Sutcliffe scores can be quite large, yet the root mean square errors can be on the order of 1.5–2.5°C for average temperatures and 4–5°C for maximum temperatures (Table 1), which may be more biologically relevant if the concern is warming temperatures. In contrast, for the streams in this study on weekly maximums, the mean RMSE was 0.87°C with a standard deviation of 0.4°C. If the interest is in estimating increases in warm temperatures, then the approach used here more directly uses the data of interest to characterize its sensitivity.

A conceptual drawback to the approach used in this paper is that it requires multiple years of monitoring data to estimate sensitivity, but this constraint may be of less consequence as stream temperature monitoring efforts become more widespread and standardized [*Isaak and Horan*, 2011; *Isaak et al.*, 2013]. Nonetheless, additional tools are needed both to extrapolate expectations for sensitivity in other areas based on watershed and stream characteristics and to more rapidly assess sensitivity based on a few measurements of stream temperature. Although one measurement showing that a stream is cold could substantially reduce uncertainty about its sensitivity, streams with moderate to warm temperatures still have quite a range. An important question would be how measurements taken over the course of a day, a week, a season, or a year might constrain expectations for a given stream.

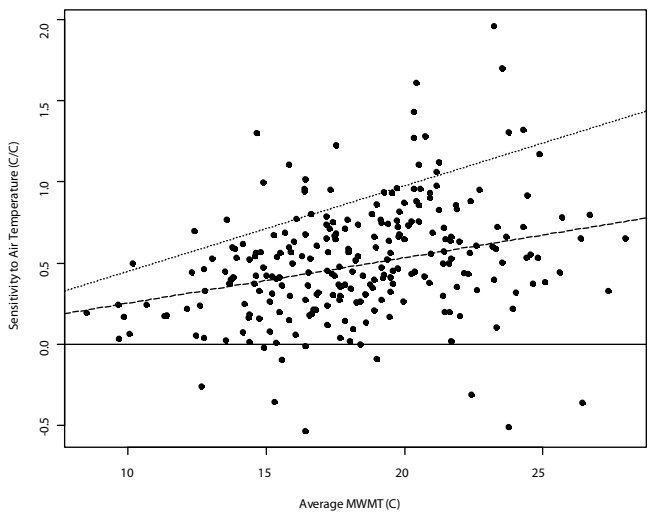

Figure 8. Sensitivity of stream temperatures to air temperature anomalies plotted against the station maximum weekly maximum temperature (MWMT). The dashed line represents the median regression with a slope of 0.028°C/°C^2, and the dotted line represents the 90th percentile regression, with a slope of 0.053°C/°C^2.

4.2. Indirect Sensitivity

"Cold" in a regional context may only be found in high elevation streams with snowmelt. The coldest streams are mainly located in the Cascade Mountains (Figure 10). In contrast, a number of different physical processes can yield relatively insensitive streams, which occur throughout the region, even in areas with warm water (Figure 9). Large groundwater reservoirs are one of the most frequently mentioned [*Kelleher et al.*, 2011; *Mayer*, 2012; *O'Driscoll and DeWalle*, 2006; *Tague et al.*, 2007]. Glaciers and permanent ice fields are other sources of plentiful cold water flow late into the summer

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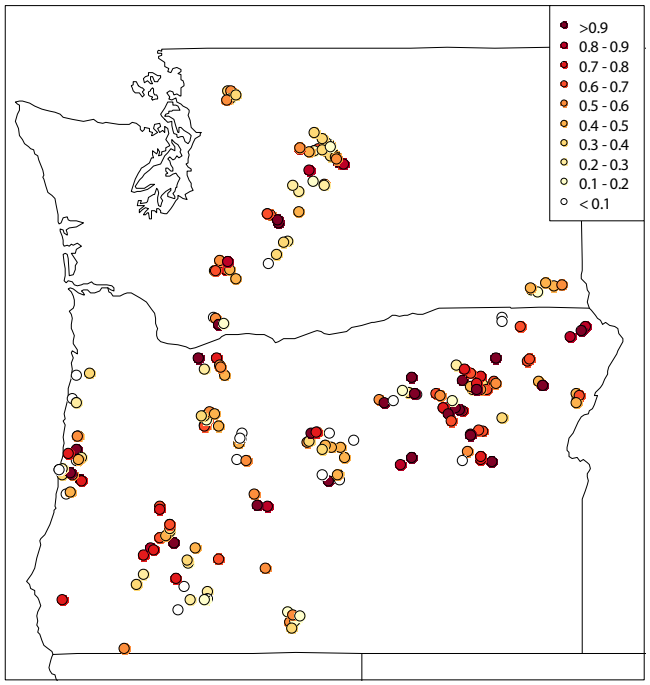

Figure 9. Map of stream temperature sensitivities (°C/°C).

[*Hari et al.*, 2006; *Webb and Nobilis*, 2007]. Snow drifts and deep snowpacks are less well recognized for their potential importance in moderation of stream temperature, though their role in sustaining streamflow later into the summer is becoming recognized [e.g., *Blöschl and Kirnbauer*, 1992; *Luce et al.*, 1998]. All of these processes involve a large, steady advected heat (cold) flux during the months of greatest heating. Other conditions (e.g., deep forest and coastal fog), lead to reduced sensitivity by moderating incoming radiant heat fluxes.

Figure 8 shows significant scatter in sensitivity at warmer temperatures. Recognition that different physical processes may control both average temperature and sensitivity in different locations, raises the question of whether some of the variation in Figure 8 can be constrained by limiting the region of analysis, and thereby the range of processes involved in temperature regulation. Figure 12 highlights the pattern of points for the set of coastal stations in Oregon, which share a common climate, geology, and forest cover. The general pattern is similar to the pattern seen across the region, but with a steeper slope (median fit slope 0.135°C/°C^2, $P = 8 \times 10^{-3}$ on 26 points). This is neither an area with the coldest streams nor the hottest. Again there is the general pattern with low sensitivity and low variability for cooler temperatures and greater variability for warmer temperatures. If the overall pattern of Figure 8 is composed of subregional patterns like this, then we can think in terms of the dominant physical processes in an area that drive general temperatures and determine which streams have low variability. The geographic distribution of the spread of thermally insensitive streams in Figure 9, augmented by knowledge about where different physical processes play out and their relationships to mean temperature, provides the insight that all of these processes can be effective at damping effects from increased energy inputs.

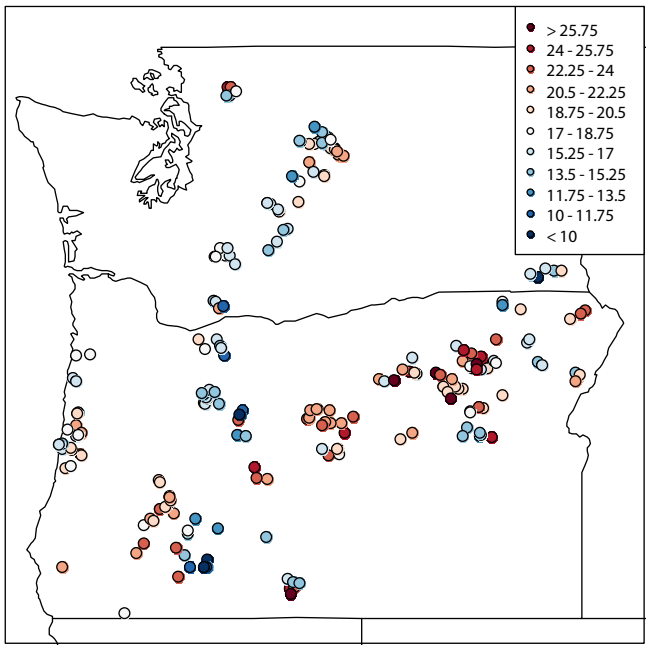

Figure 10. Map station average maximum weekly maximum stream temperatures (°C).

Although these mechanisms may buffer against increased radiant fluxes caused by or correlated to warmer air temperatures, they may be vulnerable to indirect

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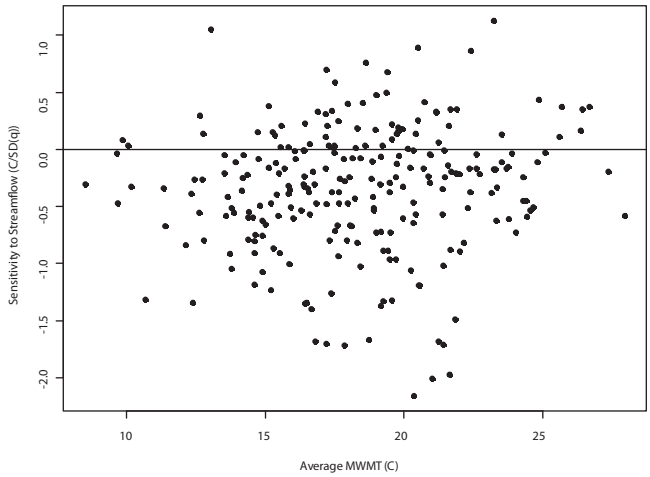

Figure 11. Sensitivity of stream temperature to standardized departures in annual streamflow.

Table 1. Root Mean Square Errors Reported for Subannual Scale Regressions of Stream Temperature on Air Temperature or Air Temperature and Flow

Study	Mean RMSE	SD of RMSE
Stefan and Preud'homme [1993]	1.49	0.67
Pilgrim et al. [1998] (Weekly Mean)	2.3	
Pilgrim et al. [1998] (Weekly Max)	3.71	
Mohseni et al. [1998] (Mean)	1.64	0.46
Mohseni et al. [1998] (Max)	4.84	
Morrill et al. [2005] (Mean)	2.2	0.47
van Vliet et al. [2010] (Daily)	2.26	0.69
Mayer [2012] (T&Q Avg)	1.12	Range: 0.06–2.16

climate change effects. The loss or widespread conversion of forest cover in regional droughts and wildfires [*Breshears et al.*, 2005; *Holden et al.*, 2012; *van Mantgem et al.*, 2009], for instance, may increase temperatures of streams in places where forest cover is an important control [*Dunham et al.*, 2007; *Isaak et al.*, 2010]. Streams protected by thick forest canopy now could become increasingly vulnerable to canopy loss, because removing the buffer against climatic forcing will expose them to more energy than removing the buffer would now.

If trends in streamflow from high elevations continue as they have over the past 60 years [*Luce and Holden*, 2009], groundwater controls on streamflow may decline in influence as well. Trends in flow of groundwater dominated rivers were among the strongest noted, because the signal was greatest in these systems and interannual variability the least. Additionally, springs are just a surface expression of deeper regional groundwater flow, so disproportionate responses can occur in spring outputs as groundwater levels decline in response to long-term precipitation declines.

Declining snowpacks in the Cascades [*Mote et al.*, 2005; *Nolin and Daly*, 2006] could potentially pose a more significant risk to stream temperature under a changing climate than would shifts in summer temperatures, particularly in smaller snowfed streams in low to midelevations [*Cristea and Burges*, 2009; *Wu et al.*, 2012]. Likewise, the loss of glaciers represents a loss of cold water [*Moore et al.*, 2009]. *Cristea and Burges* [2009] modeled the independent contributions of air temperature increase and summer flow decreases related to earlier melt from Cascade mountain snowpacks. While the air temperature sensitivities of the three streams ranged from 0.1 to 0.6°C/°C, flow related temperature changes were estimated to be on the order of 2–4 times greater than the direct air temperature related increases. While "physically based" models have

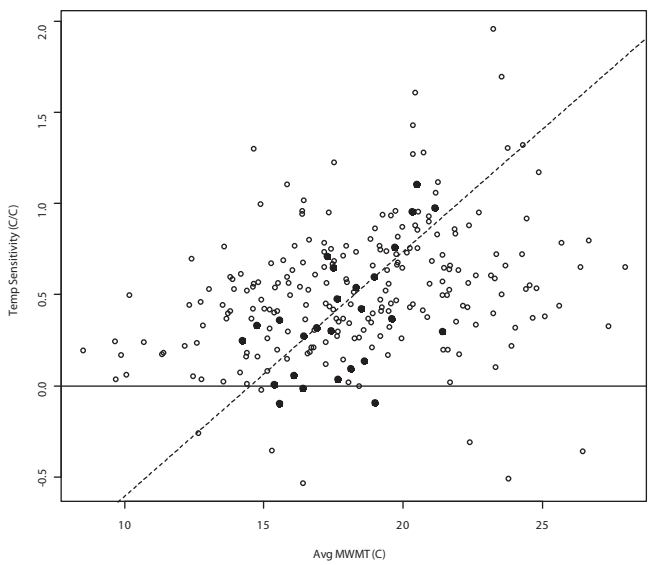

Figure 12. Same as Figure 8 but highlighting coastal streams with filled points. Dashed line is median fit of the coastal streams.

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the potential to be useful in understanding indirect sensitivity to snowpack change [e.g., *Cristea and Burges*, 2009; *Null et al*., 2013; *Wu et al*., 2012], care must be taken in interpreting the results from equilibrium model formulations in headwater areas. Cold streams imply nonequilibrium conditions with respect to external energy inputs, and empirical analyses of observed temperatures may still be one of our most reliable tools in tracking trends and climate sensitivity in such locations.

5. Conclusions

When examined at interannual time scales, colder streams are broadly less sensitive to air temperature fluctuations than warmer streams. The physical reasoning is that colder streams are generally less directly coupled to atmospheric energy exchange, which makes them less likely to respond directly to future warming. This insight may prove useful in conservation planning for cold water-dependent species.

Despite using substantially fewer data points, the interannual correlations to estimate the warmest weekly temperatures in the summer had much better standard errors of estimates than correlations between subannual scale air and stream temperature variations. Although estimating the interannual scale thermal sensitivity requires long records, on the order of a decade, it is likely that there is some correlation between subdaily, subannual, and interannual scale measurements that could be exploited to provide estimates of sensitivity with shorter records.

The strong pattern of cold streams being less sensitive may only hold for direct warming through radiative transfer. Some of the streams identified as low sensitivity may be vulnerable to secondary influences of climate change regulated by hydrology, such as riparian disturbance (e.g., fire or debris flows), earlier snowmelt (with decreased summer flows), or decreased groundwater recharge. The estimates of thermal sensitivity provided here may provide useful context for contrast with warming estimated through other processes.

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