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Contrasting scaling relationships of extreme precipitation and streamflow to temperature across the United States

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Abstract

Recent studies have reached inconsistent conclusions from scaling analysis about whether flood or extreme precipitation is more sensitive to warming climate. To explain the reasons behind the inconsistency, here we first used scaling analysis to illustrate how extreme daily precipitation and streamflow scale with daily air temperature across the Continental United States (CONUS). We found both similar and opposite scaling in extreme precipitation and streamflow. It indicates based on scaling analysis, the sensitivity of extreme streamflow to warming climate can be either similar, higher or lower to that of extreme precipitation. We further explored why there are contrasting scaling relationships in the CONUS. Generally, the similar scaling was found in regions where the timing of extreme precipitation and streamflow is correspondent, as well as with similar temporal evolution in extreme event timing and magnitude, e.g., the west coast and southern plains, implying extreme precipitation is the dominant driver of local floods. However, for regions with dissimilar scaling in extreme precipitation and streamflow (e.g., Rocky Mountains, southern plains), the characteristics of extreme streamflow show large difference to those of extreme precipitation, and the temporal evolution of extreme streamflow timing and magnitude are more correlated with factors/processes such as soil moisture and snowmelt. This study reflects that the contrasting scaling relationships of extreme precipitation and streamflow are oriented from the local hydro-climatological specifics. Using scaling analysis to compare the sensitivity of extreme precipitation and streamflow to warming climate is not suitable. Instead, we should focus more on local flood generating mechanisms or flood drivers when investigating floods in the changing climate.

1. Introduction

Flood is one of the most devastating and costly natural hazards. Global average annual loss (AAL) from floods reaches US $104 billion (Desai et al 2015). The loss from floods is expected to increase due to climate change and anthropogenic activities (Winsemius et al 2016, Böschl et al 2017, 2019, Ghausi and Ghosh 2020). Understanding how future floods would change with varying environments is of both great concern and importance to our societies.

In a warming climate, extreme precipitation is widely observed to have increased in both intensity and frequency (Alexander et al 2006, Westra et al 2013, O’Gorman, 2015, Berghuijs et al 2017, Hodgkins et al 2017, Mangini et al 2018, Martinez-Villalobos and Neelin 2018). The thermodynamic Clausius-Clapeyron (CC) equation describes that the water vapor holding capacity of atmosphere would increase at a rate of 6%–7%/°C with rising temperature (Trenberth et al 2003, Chan et al 2016, Ghausi and Ghosh 2020). Since atmospheric moisture tends to be precipitated out when forming extreme precipitation events, under the assumption of constant atmospheric relative humidity, it is expected that extreme precipitation would increase at a similar rate to the CC rate with global warming (Trenberth et al 2003, Chan et al 2016, Ghausi and Ghosh 2020). Some studies showed that observed extreme precipitation scales roughly according to the CC rate with rising
temperature (Westra et al 2013, Chan et al 2016, Bhattacharya et al 2017). However, more studies in fact have found the actual extreme precipitation scaling rates deviate from the CC rate (Lenderink and van Meijgaard 2008, Utsumi et al 2011, Berg et al 2013, Wasko and Sharma 2017). Generally speaking, the changing rates of extreme precipitation to temperature vary in different regions. Precipitation extremes in the high latitude areas tend to scale with temperature according to the CC relationship (Utsumi et al 2011, Panthou et al 2014, Guo et al 2020), those in the tropical areas mostly show negative scaling patterns (Utsumi et al 2011, Maeda et al 2012, Ali and Mishra 2017, Yin et al 2018), while middle-latitude areas tend to reflect peak-like scaling pattern, which means precipitation extremes increase with rising temperature until a temperature threshold, and then decrease with continual temperature rise (Utsumi et al 2011, Yin et al 2018, Gao et al 2020). On the other hand, there are also studies trying to figure out the reasons for regional inconsistency in extreme precipitation scaling or the deviations of actual scaling to the CC relation. For example, Ali et al (2018) found the regional differences in extreme precipitation scaling are associated with temperature seasonality; after removing the seasonality, 96% of the global stations exhibit positive extreme precipitation scaling. Visser et al (2020) used sub-daily temperature before extreme precipitation events for extreme precipitation scaling, and noticed this treatment can resolve the negative scaling in some Australian locations. The peak-like scaling patterns also tend to vanish if analyzing using fine temporal-resolution precipitation data, e.g., 10-min, hourly (Utsumi et al 2011, Shen and Chui 2021). However, regional difference in scaling is still a key concern in illustrating precipitation changes under warming climate.

Changes in extreme precipitation further raise concerns about how flooding will consequently vary. Flood changes under warming climate are still under debate as no coherent evidence has been observed so far (Hall et al 2014, Bloschl et al 2019). In some regions, flood trends are in line with extreme precipitation changes, e.g., both have increased historically (Hannahford and Marsh 2008, Petrov and Merz 2009, Berghuijs et al 2017, Hodgkins et al 2017, Mangini et al 2018). However, in some regions, floods are observed to have decreased despite increasing extreme precipitation (Ishak et al 2013, Do et al 2017, Sharma et al 2018, Wasko and Nathan 2019). A global flood study by Do et al (2017) shows that there are more stations worldwide reflecting decreasing flood trends than increasing trends, which is opposite to the changing trends of global precipitation.

Flood generation is a complex process governed by many factors, including climate, local watershed characteristics, stream dynamics (Mallakpour and Villarini 2015, Bloschl et al 2019). Intuitively, it is expected that floods would change at a similar rate as precipitation extremes to temperature, considering extreme precipitation is often the main trigger of floods. In recent years, studies are emerging to examine the scaling (i.e., sensitivity) of extreme streamflow (i.e., flood) to temperature directly rather than obtaining inferences from the scaling of extreme precipitation to temperature (Wasko and Sharma 2017, Yin et al 2018, Ghausi and Ghosh 2020). However, such studies are very limited. Even more unfortunately, the conclusions from current studies about global extreme streamflow scaling analysis seem to be contradictory in terms of whether extreme streamflow is more sensitive to warming climate than extreme precipitation (Wasko and Sharma 2017, Yin et al 2018, Ghausi and Ghosh 2020). Some studies (e.g., Yin et al 2018) suggested that extreme streamflow sensitivity under warming climate exceeds that of extreme precipitation globally. However, Wasko and Sharma (2017) showed in most region globally, the scaling rates of extreme streamflow are lower than those of extreme precipitation; only in most extreme cases, increases in precipitation would translate into similar changes in extreme streamflow. Wasko et al (2019) suggested that the contradictory results in Wasko and Sharma (2017) and Yin et al (2018) are due to snowmelt, which leads to increases in extreme streamflow rather than extreme precipitation. Shen and Chui (2021) conducted similar scaling comparison but divided the watersheds into different imperviousness categories. They found imperviousness has positive impacts on extreme streamflow scaling, but the impacts are limited compared to the role of climate in determining local extreme streamflow. The current limited extreme streamflow scaling studies are far from clearly explaining how floods would respond to future warming climate. More relevant studies are needed.

Therefore, this study makes further exploration about the sensitivity of extreme precipitation and streamflow to warming climate, and why contradictory conclusions were obtained in previous studies. The study was implemented over the Continental United States (CONUS). The reasons for this choice are (1) we aim at obtaining more specific conclusions from regional studies than from global studies; (2) The CONUS lies in multiple climate zones and also contains the main types of flood generating mechanism (Brunner et al 2020), thus the conclusions drawn from this study are also eligible to other regions.

2. Data and methods

2.1. Data
The daily mean temperature, precipitation and streamflow data for the watersheds across the CONUS were extracted from the Hydrometeorological Sandbox - École de technologie (HYSETS) dataset, which is a
comprehensive hydrometeorological database containing 14,425 watersheds across the North America (Arsenault et al. 2020). The watersheds whose flow peaks are impacted by anthropogenic regulations have been removed from the HYSETS by Arsenault et al (2020). The remaining watersheds are considered as reflecting the closely natural relationships between climate and streamflow (Arsenault et al. 2020). The meteorological data in the HYSETS are based on the Global Historical Climate Network Daily (GHCN-Daily) dataset (Menne et al. 2012), while daily streamflow data for the CONUS is extracted from the USGS database (USGS, 2020). Arsenault et al (2020) averaged the meteorological data from the GHCN-Daily dataset within each watershed with 1° extension using the Thiessen polygons method to obtain watershed-averaged meteorological time series. We used the watershed-averaged meteorological data in the HYSETS dataset for the analysis in this study.

We only used watersheds in the HYSETS with more than 30-year observed streamflow data in 1951–2011 and less than 15% missing data in each year. We further filtered the watersheds by watershed imperviousness below 5% (Falcone, 2017), as these watersheds are considered as minimally affected by urbanization (Lins, 2012). Since we considered climate change signal in flood timing in this study, which requires flood timing to be non-uniformly distributed (Blöschl et al. 2017, Wasko et al. 2020), the watersheds with uniformly distributed flood timing were also omitted, examined using the Rayleigh test at 0.05 significance level (Fisher, 1993, Wasko et al. 2020). Above filtering processes resulted in 2,332 watersheds preserved for this study.

Relative humidity and soil moisture data over the CONUS were used to explain the potential reasons behind the scaling patterns of extreme precipitation and streamflow to temperature. Relative humidity data were obtained from the Global Summary of the Day (GSOD) dataset from the USA National Centers for Environmental Information (NCI), which are observed data from the meteorological stations across the CONUS. The used soil moisture is simulated monthly gridded soil moisture data from the Climate Prediction Center (CPC) (van den Dool et al. 2003), with a spatial resolution of $0.5^\circ \times 0.5^\circ$.

2.2. Scaling analysis of extreme precipitation and streamflow to temperature
Similar to previous studies (e.g., Hardwick et al. 2010, Utsumi et al. 2011, Wasko and Sharma 2017, Ghausi and Ghosh, 2020), we used scaling analysis to illustrate the sensitivity of extreme precipitation and streamflow to warming climate across the CONUS. In each watershed, we first extracted precipitation and streamflow events from the observed time series based on event separation time (see supplementary materials), the maximum in each event was then paired with the coincident daily temperature. We used a binning method of 2 °C width overlapped by 1 °C to stratify the extreme precipitation-temperature and streamflow-temperature pairs for detecting scaling patterns (Utsumi et al. 2011, Wasko and Sharma 2017). We also ensured that at least 150 data were preserved in each bin so that enough samples were preserved in each bin to estimate extreme value. If there were less than 150 data in a 2 °C bin, the bin width was enlarged until reaching 150 data. In each bin, the median temperature was matched to the 99th percentiles of precipitation and streamflow. We only used bins with median temperature above 0 °C to minimize the number of snow events and snowmelt-induced floods (Mishra et al. 2012, Wasko and Sharma 2017, Ghausi and Ghosh 2020). Different scaling patterns of extreme precipitation and streamflow (the 99th percentiles) to temperature were detected using the Locally Weighted Scatterplot Smoothing (LOWESS) method (Cleveland 1979, Utsumi et al. 2011). Three examples of monotonic increase, peak-like structure and monotonic decrease scaling patterns detected using the LOWESS method are provided in figure S1 (available online at stacks.iop.org/ER/3/125008/mmedia). The actual scaling rates were estimated using exponential regression between the 99th percentiles and median temperatures (Hardwick et al. 2010). If a peak-like scaling pattern or peak point temperature was detected by the LOWESS method, exponential regression was only applied to the temperature range containing the majority of the data, with the peak point temperature as one-side boundary of the temperature range (Gao et al. 2020).

2.3. Interpretation of the scaling patterns in extreme precipitation and streamflow
To possibly explain the reasons behind the scaling patterns of extreme precipitation and streamflow to temperature, we further explored climatic and hydrological characteristics in multiple aspects, including (1) changing patterns of wet-day relative humidity with temperature (Hardwick et al. 2010, Yin et al. 2018), (2) mean annual cycle of temperature, precipitation and streamflow, (3) timing of extreme precipitation and streamflow, and (4) long-term temporal evolution of extreme streamflow timing and magnitude as well as some potential drivers (Blöschl et al. 2017, 2019). Streamflow extremes are generally caused by heavy precipitation or snowmelt events, in which antecedent soil moisture may cause some modulations (Berghuijs et al. 2016, Wasko and Nathan 2019, Wasko et al. 2020). In analyzing the temporal evolution of timing, we used the timing of annual maximum streamflow, annual maximum 7-day precipitation, first 7 days with temperature all above 0 °C (snowmelt timing) and annual maximum soil moisture. In analyzing the temporal evolution of magnitude, we used the magnitudes of annual maximum streamflow, annual maximum 7-day precipitation, averaged temperature in spring (March, April and May) and annual maximum soil moisture. Ten-year moving average
was used to filter the short-term temporal variation and highlight the long-term temporal trend (calculation methods are described in the supplementary materials).

3. Results

3.1. Scaling patterns of extreme precipitation and streamflow to temperature across the CONUS

Figure 1 shows the scaling patterns of extreme precipitation and streamflow to temperature for all watersheds across the CONUS. In this figure, the peak-like scaling patterns are represented by the peak point temperatures, and monotonic increasing and decreasing scaling patterns are marked using another two colors (black and cyan, respectively). The scaling of extreme precipitation to temperature shows clear spatial variability (figure 1(a)). Precipitation extremes in most watersheds display peak-like scaling patterns. The precipitation extremes in the midwest mostly display monotonic increasing scaling, while those in the Rocky Mountains and some in the west coast tend to decrease monotonically with rising temperature. The scaling of extreme streamflow to temperature also shows obvious spatial variation (figure 1(b)). Similar to extreme precipitation scaling, streamflow extremes in most watersheds reflect peak-like scaling patterns. In the central CONUS and Rocky Mountains, streamflow extremes in some watersheds reflect monotonic increases with rising temperature, while the majority of watersheds in the midwest and those along the west coast reflect monotonic decreases with higher temperature.

To illustrate the spatial differences in extreme precipitation and streamflow scaling, we further compared the peak point temperature (PPT) differences between extreme streamflow (q) and precipitation (p) scaling (PPT\textsubscript{q} - PPT\textsubscript{p}) (figure 1(c)). The watersheds where both the scaling of extreme precipitation and streamflow are monotonic increase or decrease are not shown as the scaling patterns are the same (i.e., PPT\textsubscript{q} - PPT\textsubscript{p} = 0). It is clearly shown in figure 1(c) that the peak point temperatures of extreme streamflow scaling are mostly higher than those of extreme precipitation in the eastern CONUS, while being lower in the central and western CONUS. The southern plains and west coast are two regions where the peak point temperatures show minor differences, indicating similar scaling patterns. The peak point temperature differences are highest and lowest in...
the Rocky Mountains and midwest, respectively, as the scaling of extreme streamflow and precipitation is opposite in the two regions, which is mostly monotonic decrease for precipitation and monotonic increase for streamflow in the Rocky mountain, and is monotonic increase for precipitation and monotonic decrease for streamflow in the midwest (figures 1(a), (b)).

Figure 1 generally reflects that both similar and opposite scaling patterns in extreme precipitation and streamflow can be detected if illustrating via scaling analysis method, thus it is hard to reach a general conclusion from scaling analysis about whether flood is more sensitive to warming climate than extreme precipitation. To illustrate why contrasting scaling patterns are occurring in different regions of the CONUS, we further focused on four regions where the four contrasting scaling patterns were detected. In addition, in order to get relatively clear climate change signal in extreme streamflow, when determining the boundaries of the four regions, we also considered that the changing trends in extreme streamflow timing and magnitude are homogeneous in the determined regions, which are shown in figure S2. The boundaries of the four regions are marked in figures 1 and S2. The watershed number and mean extreme streamflow timing and magnitude trends are provided in table S1. From the trends, it is initially reflected that the four regions are distinguished from each other.

To eliminate the influences of some other factors on the extreme precipitation or streamflow-temperature relationships, e.g., topography, land use/land cover, we aggregated the bins from all watersheds within each region and applied the LOWESS method to reflect the regional averaging scaling patterns, as shown in figure 2 (top). The scaling rates estimated using exponential regression in individual watersheds within each region are displayed in figure 2 (bottom). The conclusions reflected in the regional scaling are in line with those reflected in figure 1. In the west coast (figure 2(a)), the peak point temperatures of extreme precipitation and streamflow scaling are similar and small (~9 °C). After 9 °C, both extreme precipitation and streamflow (log-transformed) decrease linearly and at a similar rate with rising temperature. The southern plains is another region where extreme precipitation and streamflow show similar regional scaling patterns (figure 2(c)), but the peak point temperature is much higher, being around 20 °C. The Rocky Mountains and midwest are two regions where extreme precipitation and streamflow scaling is opposite. In the Rocky Mountains (figure 2(b)), extreme precipitation decreases monotonically and linearly with rising temperature, but extreme streamflow increases with temperature up to around 12 °C, and then decreases sharply with continual temperature rise. In contrast, in the midwest (figure 2(d)), regional extreme precipitation increases while extreme streamflow decreases with rising temperature in the wide temperature range (5 °C–23 °C).

The scaling rates estimated from individual watersheds (figures 2(e)–(h)) reflect similar scaling patterns as reflected by the regional averaging scaling. Specifically, the scaling rates of extreme precipitation and streamflow are relatively similar in the west coast and southern plains (figures 2(e), (g)), but are opposite in the Rocky Mountains and midwest (figures 2(f), (h)). It is worth noting that the scaling rates of extreme streamflow from individual watersheds show larger variations, while those of extreme precipitation are relatively concentrated.
This might suggest that local watershed attributes (e.g., slope, land use/land cover, soil storage) can alter the sensitivity of extreme streamflow to climate.

It is interesting that contrasting scaling patterns are found in the CONUS. The contrasting scaling patterns reflect that extreme streamflow may change both similarly and differently to extreme precipitation under warming climate. From the regional scaling analysis, some important questions remain to be answered: (1) why do the four regions display contrasting scaling patterns in extreme precipitation and streamflow? (2) why does the scaling reverse at certain temperature in some cases? Most importantly, (3) why are the temperatures of scaling inflection the same in some regions, but different in others?

3.2. Interpretation of the contrasting scaling patterns of extreme precipitation and streamflow to temperature

We tried to answer the above questions from the regional hydro-climatological perspective. We analyzed the changes of wet-day relative humidity with temperature, annual cycles of daily temperature, precipitation and streamflow, as well as timing distributions of extreme precipitation and streamflow (represented by the annual maxima) in the four regions, as shown in figure 3. We also analyzed the timing distribution of extreme precipitation and streamflow determined using peaks-over-threshold method with the 99th percentile of precipitation or streamflow time series as the threshold, the distributions (figure S3) are similar to those reflected in figures 3(c), (f), (i), (l). In analyzing wet-day relative humidity with temperature, we aggregated the relative humidity data from all stations in each region, and also applied the LOWESS method to reflect the general relationship between relative humidity and temperature. Consistent with previous studies (e.g., Hardwick et al. 2010, Yin et al. 2018, Shen and Chui 2021), the different scaling patterns of extreme precipitation to temperature tend to be explained by the changes of relative humidity with temperature (figures 3(a), (d), (g), (j)).

The temperature threshold at which precipitation scaling reverses is nearly the same temperature after which relative humidity drops with rising temperature. For example, in the southern plains, relative humidity remains stable at 80% before ~20 °C, but it decreases dramatically at higher temperature (figure 3(g)). The temperature of ~20 °C is exactly the threshold before which the extreme precipitation scales linearly and positively with temperature, but after ~20 °C, the scaling turns to be negative (figure 2(c)). A contrasting example is in the Rocky Mountains, where the relative humidity decreases monotonically with rising temperature (figure 3(d)), and extreme precipitation shows negative scaling pattern with temperature (figure 2(b)). Since it is inferred that extreme precipitation would increase exponentially with warming climate under the assumption of constant relative humidity, we do find it is the case when relative humidity is constant with temperature; but if relative humidity is not constant, extreme precipitation does not show increasing trend at higher temperatures. It should be noted that evaporation from falling precipitation also contributes to atmospheric moisture, but the majority (~85%) of atmospheric moisture is from ocean evaporation (Bigg et al. 2003). Thus, there is high confidence that the changing patterns of relative humidity determine the scaling of extreme precipitation to temperature, rather than the former determines the latter.

As for extreme streamflow scaling, we found in regions where streamflow and precipitation show similar annual cycles and the timing of extreme precipitation and streamflow is correspondent, the scaling pattern of extreme precipitation to temperature would translate into similar pattern in extreme streamflow, i.e., west coast and southern plains. For example, in the west coast, both precipitation and streamflow reach their highest values in winter (December, January, February), but decrease to the lowest in late summer (June, July, August) and early autumn (September, October, November) (figure 3(b)), extreme precipitation and streamflow are concentrated in late autumn to early spring (figure 3(c)). The similar annual cycles and extreme event distributions reflect that extreme streamflow is dominantly caused by extreme precipitation (Villarini et al. 2016, Dougherty and Rasmussen 2019, Brunner et al. 2020). So, it might suggest that in rainfall-dominated regions, the sensitivity of extreme precipitation and streamflow would stay similar, despite the nonlinear processes from climate to hydrology.

In contrast, there are also regions where (extreme) streamflow has weak connection with (extreme) precipitation, i.e., Rocky Mountains and midwest. The Rocky Mountains is a typical region where the annual cycle of streamflow and the timing distribution of extreme events are remarkably different from those of precipitation. Extreme streamflow mainly happens in late spring and early summer (figures 2(e), (f)), resulting from snowmelt (Li et al. 2019), while extreme precipitation is evenly distributed in a year. In these regions, from figure 2, we noticed that the scaling of extreme streamflow is completely different from that of extreme precipitation. Therefore, in regions where extreme streamflow is not directly caused by extreme precipitation, we cannot expect extreme streamflow would show similar changing patterns as extreme precipitation to warming climate. Instead, the changing patterns of extreme streamflow in those regions are more controlled by other processes such as snowmelt.
Since the responses of extreme streamflow to rising temperature tend to be governed by local hydroclimatological characteristics, we used the methods in Blöschl et al. (2017, 2019) to analyze the connections between extreme streamflow, extreme precipitation, soil moisture maxima and snowmelt processes in the four representative regions. Figure 3. Climatic and hydrological characteristics in the four representative regions. (a), (d), (g), (j) Changing patterns of wet-day relative humidity with daily temperature, with changing trends (red lines) estimated using the LOWESS method. The colors depict scatter intensity, with lighter color meaning denser point-scatters. (b), (e), (h), (k) Annual cycles of daily temperature, precipitation and streamflow. Grey lines represent values from individual watersheds (right Y-axis), and bold black lines are the ensemble medians in each region (left Y-axis). Please note the ranges of the two Y-axes in each sub-figure are different. (c), (f), (i), (l) Distributions of extreme precipitation and streamflow timing, which is represented by the timing of annual maximum precipitation and streamflow. The shapes of the plots depict distribution density.
regions, as shown in figure 4. The Spearman’s rank correlation coefficients between these variables are shown in table 1. We also analyzed the temporal evolution of the 95th percentile and 99th percentile of streamflow in each year (figures S4, S5, respectively), the changing patterns reflected are similar to those in figure 4. Figure 1 clearly shows that the long-term temporal evolution in extreme streamflow timing and magnitude is not solely related to temperature or precipitation, but also has strong connections with other potential drivers.

In the west coast and southern plains (figures 4(a), (c), (e), (g)), consistent with the previous conclusions, snowmelt tends to contribute little to extreme streamflow, as the first 7 days with temperature above 0 °C (snowmelt timing) always occur at the beginning of a calendar year. For example, the correlation between snowmelt timing and extreme streamflow timing is only $-0.18$ in the west coast, but it is $0.42$ between extreme precipitation and streamflow. The temporal evolutions of extreme streamflow timing and magnitude are most similar to those of extreme precipitation in these two regions, but also share some similarities with those of maximum soil moisture. For example, in the southern plains, both extreme precipitation and streamflow magnitudes increase nearly monotonically before around 1990 (figure 4(g)), but after 1990, extreme precipitation continues to increase while extreme streamflow shows a decreasing trend. The opposite changing trends might be because maximum soil moisture has decreased after around 1990.

In the Rocky Mountains, the temporal evolution of extreme streamflow timing is most similar to that of snowmelt timing, both are rather stable during the studied period, with a correlation coefficient of $0.46$.
compared to only 0.01 for extreme precipitation. In the midwest, extreme streamflow timing variation is closest to that of maximum soil moisture (with highest correlation of 0.39 compared to other variables), and is also highly similar to snowmelt timing variation before around 1986. In the Rocky Mountains, both spring and annual temperature tend to decrease in the studied period, but extreme precipitation and streamflow show slight increase. This might be because although the Rocky Mountains is a snowmelt-dominated region, increasing precipitation will cause larger snowpack, indirectly resulting in increases in extreme streamflow. In the midwest, all variable magnitudes show monotonic increasing trend during the studied period, which might reflect the midwest is a region where extreme streamflow is the result of interactions between snowmelt, extreme precipitation and soil moisture condition.

4. Discussion

This study aims at resolving why statistically analyzing the sensitivity of extreme precipitation and streamflow to temperature, different studies got contradictory conclusions. The main reason indicated in this study is that different scaling patterns of extreme precipitation and streamflow are actually the reflection of local climatic and hydrological features. Thus, instead of using scaling analysis comparing the sensitivity of extreme precipitation and streamflow to temperature, more attention should lie on the local flood generating mechanisms or flood drivers.

This study used daily data for the scaling analysis across the CONUS. Although at the regional scale, when relative humidity is constant in some regions, e.g., southern plains, midwest, extreme precipitation increase exponentially with warming climate, the increasing rate is actually lower than the CC rate. This is because daily precipitation amounts contain rainfall intervals at higher timescales (e.g., hourly). If using higher temporal-resolution precipitation data, we can expect the actual scaling rates of extreme precipitation to temperature would be close to the CC rate when relative humidity is constant (Hardwick et al 2010, Utsumi et al 2011, Shen and Chui 2021).

Extreme streamflow may not be caused by the concurrent climate events. For example, a streamflow event may be caused by precipitation event happening several days ago due to travel time (Ivancic and Shaw 2015). To account for this time lag, we assessed the correlation of extreme streamflow events and precipitation, temperature and relative humidity that are coincident or several days prior to the extreme streamflow events, as shown in figure 56. It is shown coincident precipitation is most correlated with extreme streamflow compared to precipitation events happening several days ago, e.g., in the west coast and southern plains, or changing the time lag did not significantly alter the correlation between these variables and extreme streamflow, e.g., for all variables in the Rocky Mountains and midwest, and for temperature and relative humidity in the west coast and southern plains. Therefore, for this study, the timing similarity or difference between extreme precipitation and streamflow shown in figure 3 and figure S3 truly reflects whether extreme streamflow is dominantly caused by extreme precipitation or not.

Some important factors that influence the responses of extreme precipitation and streamflow to warming climate were not considered in this study. For example, aerosol loadings can alter the links between extreme precipitation and temperature. Aerosol loadings can influence the size of droplets in the air; they can also absorb solar radiation, alter temperature changes with height, and further influence convection and cloud properties for precipitation formation (Sarangi et al 2018, Zhao et al 2019). Except for climate attributes, watershed characteristics are also non-negligible factors in extreme streamflow generation (Stein et al 2021), e.g., slope, land use/land cover, aridity. Watershed characteristics tend to be more variant among different watersheds, this may explain the wide scaling rates of extreme streamflow to temperature shown in figure 2. Even more, the role of watershed characteristics can be more dominant than climate attributes in shaping local extreme streamflow. Therefore, in extreme streamflow studies, great attention should also be given to local watershed specifics rather than solely focusing on climate attributes.

The impacts of climate change on streamflow can be indirect. This means changes in temperature and precipitation do not cause response in streamflow directly. Climate change may cause influences on some intermediate variables first, e.g., soil moisture, snowmelt, then the intermediate variables further cause influences on streamflow. This is especially the case for regions where extreme streamflow is not dominantly controlled by precipitation, while soil moisture or snowmelt process plays a key role in extreme streamflow generation, such as the Rocky Mountains and midwest in this study. So, the higher correlation between extreme streamflow and soil moisture or snowmelt process in these regions (figure 4) may also be the effects of climate change.

Last, this study extracted extreme streamflow events from historical observation, we did not investigate flood events of specific return periods directly, e.g., 20-year, 50-year or 100-year flood (Slater et al 2021, Swain et al 2020). For flood events of return periods are not very extreme, they may be within the range of historical
observation. However, for very extreme flood events, e.g., 100-year events, due to the limited length of historical record, their changes under warming climate may not be well reflected by the method framework in this study.

5. Summary and conclusions

This study investigated the scaling relationships (or sensitivity) of extreme precipitation and streamflow to temperature across the CONUS. We found both similar and opposite scaling in extreme precipitation and streamflow, including (1) similar positive scaling, (2) similar negative scaling, (3) positive precipitation scaling and negative streamflow scaling, and (4) negative precipitation scaling and positive streamflow scaling. This reflects it tends to be unreasonable to infer potential changes of flood under warming climate from the response of extreme precipitation to temperature.

The different regional scaling patterns of extreme precipitation and streamflow to temperature are attributed to local climatic and hydrological characteristics. The scaling patterns of extreme precipitation to temperature can be explained by the changes of relative humidity with temperature. Positive precipitation scaling is expected under constant relative humidity conditions. If extreme precipitation is the main trigger for extreme streamflow, the scaling of extreme precipitation to temperature tends to translate into similar scaling in extreme streamflow. However, for regions where extreme streamflow is more correlated with other drivers (e.g., soil moisture, snowmelt), the scaling of extreme streamflow has weak connection with extreme precipitation scaling.

This study reflects that the sensitivity of extreme streamflow to warming climate is determined by local hydro-climatological specifics. Using scaling analysis method to compare the sensitivity of extreme precipitation and streamflow to warming climate is not suitable. When exploring flood changes under warming climate, we should pay more attention to the local flood generating mechanisms or flood drivers (Villarini and Wasko 2021), instead of getting inferences from precipitation changes with warming climate.

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Data availability statement

All data that support the findings of this study are included within the article (and any supplementary files).

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