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To cite this article: Andrew Bui et al 2019 Environ. Res. Lett. 14 074025

View the article online for updates and enhancements.
Environmental Research Letters

LETTER

The relationship of atmospheric air temperature and dew point temperature to extreme rainfall

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Keywords: precipitation, atmospheric temperature, atmospheric dew point, scaling, Clausius–Clapeyron, climate change

Abstract

To understand the expected changes of extreme rainfalls due to climate change, the sensitivity of rainfall to surface temperature is often calculated. However, as surface temperatures may not be a good indicator of atmospheric moisture, an alternative is to use atmospheric temperatures, but the use of atmospheric temperatures lacks precedent. Using radiosonde atmospheric temperature data at a range of geopotential heights from 34 weather stations across Australia and its territories, we examine whether atmospheric temperature can improve our understanding of rainfall-temperature sensitivities. There is considerable variability in the calculated sensitivity when using atmospheric air temperature, while atmospheric dew point temperature showed robust positive sensitivities, similar to when surface dew point temperature measurements were used. We conclude atmospheric dew point temperature may be a promising candidate for future investigations of empirically calculated sensitivities of rainfall to temperature but does not appear superior to the use of surface dew point temperature measurements.

1. Introduction

Climate change represents one of the most pressing issues facing society due to its effect on meteorological and hydrological events. Rising temperatures can directly impact rainfall patterns by increasing the atmospheric moisture holding capacity leading to increased rainfall extremes (Trenberth et al. 2003, O’Gorman 2015). Hence, understanding the relationship between temperature and extreme rainfall is a key step towards understanding the effects of climate change on rainfall. The association between rainfall and temperature, termed ‘scaling’, is linked to the Clausius–Clapeyron (C–C) relationship, which is the theoretical exponential increase in the saturation vapour pressure of approximately 7%/°C. If the saturation vapour pressure increases at this rate it is plausible to suggest the maximum rainfall intensity should increase at a similar rate (Trenberth 2011, Westra et al. 2014). This scaling relationship is based on two core assumptions; that relative humidity will remain approximately constant in the future (Soden and Held 2006), and that extreme rainfall events precipitate all available moisture (Lenderink and van Meijgaard 2010). However, the possible violation of these assumptions, alongside artefacts introduced due to the use of surface temperatures in the calculations results in deviations from the C–C relationship being observed.

Scaling of rainfall extremes much higher than C–C (super C–C) has been found across a variety of climates and regions (Lenderink and van Meijgaard 2008, Liu et al. 2009, Mishra et al. 2012), particularly for short rainfall durations (Hardwick Jones et al. 2010, Lenderink et al. 2011, Panthou et al. 2014, Busuioc et al. 2016, Wasko et al. 2018, Wibig and Piotrowski 2018). Super C–C scaling has generally been attributed to the convective nature of short duration storms (Berg and Haerter 2013, Berg et al. 2013, Park and Min 2017). During a storm the natural release of latent heat could cause additional moisture convergence in the system, invigorating the storm and increasing the rainfall intensity beyond the theoretical C–C value (Trenberth et al. 2003, Westra et al. 2014). Using
lightning (Molnar et al 2015) or cloud type (Berg and Haerter 2013) as a proxy for identifying convective events, super C–C scaling rates have been found for convective storms. When convective storms are removed from the storm sample, the scaling reduces to close or less than C–C. Hence changes in atmospheric circulations which are associated with convection also affect the scaling calculated (Blenkinsop et al 2015, Chan et al 2016).

Negative scaling has also been identified in many regions across the world (Utsumi et al 2011, Wasko et al 2016) and is particularly associated with the higher temperatures experienced in tropical regions (Hardwick Jones et al 2010, Maeda et al 2012). Negative scaling contradicts observed historical increases in tropical rainfall extremes (Donat et al 2016, Guerreiro et al 2018). This has led to the explanation that, at higher temperatures, the relative humidity decreases, as there is no more moisture to be sourced (Hardwick Jones et al 2010) and evaporation limitations (Priestley 1965, Roderick et al 2019) result in a reversal from positive to negative scaling (Drobinski et al 2016, 2018, Gao et al 2018). Indeed the most negative scaling is found in regions with the largest humidity limitations (Wasko et al 2015). However, other explanations are also possible. Warmer surface temperatures generally occur on less cloudy days with less rainfall, resulting in negative scaling (Trenberth and Shea 2005), which would account for the observation that at higher temperatures the proportion of the day experiencing rainfall is also smaller (Utsumi et al 2011). It has also been shown that storms of shorter duration generally occur when temperatures are higher. As these storms also have lower rainfall intensities due to their shorter duration there is a resulting negative bias in the scaling relationship calculated (Wasko et al 2015).

Negative scaling has also been explained by considering the timing of measured temperatures in relation to the storm arrival. Temperatures can decrease during a rainfall event for many reasons, including evaporative cooling, or movement of cold air associated with the rainfall event (Bao et al 2017). As a result the temperature coincident with the rainfall may not truly reflect the temperature that occurred when the rainfall was actually generated (Ali and Mishra 2017, Bao et al 2017). Using the temperature three days prior to rainfall events has returned positive scaling in some locations where previously negatively scaling was calculated (Ali and Mishra 2017). However, as it has been refuted that local cooling is the reason for negative scaling (Barbero et al 2017), there continues to be an assertion that moisture limitations are the most likely physical explanation for negative scaling (Lenderink et al 2018).

To overcome moisture limitations at higher temperatures, two alternatives in the literature have been proposed for calculating rainfall–temperature sensitivities; surface dew point temperature and atmospheric air temperature. It has been strongly advocated that dew point temperature (e.g Lenderink and Attema 2015, Barbero et al 2017) should be used as, by definition, the dew point is the temperature a parcel of air needs to be cooled at constant pressure for saturation (100% relative humidity) to occur. Hence, a 1 °C increase in dew point temperature can be interpreted as equivalent to a 7% increase in atmospheric moisture content (Lenderink and van Meijgaard 2010). A multitude of studies have investigated the sensitivity of extreme rainfall with dew point temperature (Lenderink and van Meijgaard 2010, Lenderink et al 2011, Panthou et al 2014, Ali and Mishra 2017, Park and Min 2017, Ali et al 2018, Wasko et al 2018). Most recently a global study (Ali et al 2018) showed that, for the majority of tropical areas, positive scaling closer to the C–C relationship is obtained when surface dew point temperature is used (instead of surface air temperature). However, one of the main arguments, for not using surface temperatures (air or dew point) is that they are impacted by the rainfall from the storm event.

An alternative proposed is to use an upper troposphere temperature at a height sufficient enough to avoid fluctuations due to the storm event (Mishra et al 2012) and prevent the dominance of solar surface heating (Chan et al 2016). Using atmospheric temperature is physically more consistent with rainfall causing processes. Rainfall totals are constrained by the amount of moisture in the atmosphere and hence exhibit a very strong correlation to the integrated water vapour (Roderick et al 2019). The integrated water vapour is a function of the mean temperature in the water vapour column (Hagemann et al 2003), and rainfall resulting from convection is dependent on the atmospheric temperature (Neelin et al 2009). Hence a change in atmospheric air temperature (as opposed to surface temperature) should better capture a change in the saturation vapour pressure that is linked to extreme rainfall.

The use of reanalysis atmospheric air temperature at 850 hPA above India largely resulted in positive scaling, in line with expectations, including instances of super C–C scaling slightly above 7%/°C (Ali and Mishra 2017). To that extent climate modelled 850 hPA air temperatures have also resulted in positive scaling across the United Kingdom where surface temperatures did not (Chan et al 2016). Atmospheric air temperature has also been used for developing non-stationary IDF curves (Ali and Mishra 2017, Golroudbary et al 2019).

The notion that surface temperature may (a) not be physically well linked to atmospheric moisture, and (b) subject to many (statistical) artefacts, gives credence to the use of atmospheric temperatures. However, there is currently a lack of precedent for using atmospheric temperatures for rainfall–temperature scaling calculations. Our aim is to investigate the relationship of rainfall with atmospheric air temperature...
and atmospheric dew point temperature at various geopotential heights to see whether there is merit in using atmospheric temperatures over surface temperatures for calculating the scaling of rainfall with temperature.

2. Data and methods

We use data sourced from the Australian Bureau of Meteorology at 34 locations around Australia and its surrounding territories. Atmospheric air temperature and atmospheric dew point temperature is measured using balloon-borne radiosonde instruments up to an altitude of 25 km. The frequency of measurements varies but generally occurs every 6 h. The locations of these sites are summarised in Table S1 and are available online at stacks.iop.org/ERL/14/074025/mmedia (and later presented in figure 2).

Daily rainfall is reported at each of the radiosonde sites at the ground at 9 am local time as a daily accumulation and is measured through either a manually read rain gauge or automatic weather station. Daily surface air and dew point temperatures are also sourced from the Australian Bureau of Meteorology at each of the radiosonde measurement locations.

The stations are spread evenly across the continent of Australia covering a range of atmospheric and climatic conditions (Linacre and Geerts 1997, Peel et al. 2007). This includes tropical regions in the north where rainfall is summer dominant, and temperate regions in the south-east and south-west where rainfall is winter dominant. In central Australia the climate is arid desert. The territorial locations are largely tropical (Cocos Island, Norfolk Island, and Lord Howe Island), however Macquarie Island is Antarctic tundra. Hence, this data set represents a diverse range of climates.

For each station, the atmospheric air temperatures and atmospheric dew point temperatures were averaged for the 24 h concurrent with the period of any non-zero rainfall observation. These pairs were then binned on geopotential height. The tropopause marks the end of the troposphere at approximately 11 km above sea level, and also the upper limit of most weather (Ahrrens 2003), with the stratosphere above this exhibiting a temperature inversion at approximately 20 km above sea level. Although the role of cloud processes in the stratosphere is limited (Wallace and Hobbs 2006) these elevations are included for completeness. The data was divided into 8 geopotential intervals of 2.5 km, from the surface 0 km, up to 20 km, that is 0–2.5, 2.5–5, 5–7.5, 7.5–10, 10–12.5, 12.5–15, 15–17.5, and 17.5–20 km. Within each of these geopotential intervals the rainfall-temperature pairs were then binned on temperature. Robust scaling was calculated by removing values that deviated more than 30 °C from the mean temperature for that geopotential height. The rainfall-temperature data pairs were binned into up to 12 equal temperature intervals, with intervals containing less than 20 events excluded from the analysis. Consistent with previous similar studies, the 95th percentile of the rainfalls in each bin was calculated, and a linear regression was fitted using the resulting data points (e.g. Ali et al. 2018) (equation (1)).

$$\ln(R) = \beta_0 + \beta_1 T,$$

where R and T are the vectors (of length 12) of the rainfall percentile and temperature pairs calculated for each temperature bin for each geopotential height interval. The temperature adopted in this case was the mid-point of each interval. The scaling for the 95th percentile was then calculated for each geopotential interval by:

$$\text{Scaling} [\%/{\degree}C] = 100 \times (e^{\beta_1} - 1).$$

This resulted in 8 estimates of scaling for both air and dew point temperature at each site, one for each of the varying levels of geopotential height. For simplicity, when we refer to scaling, we refer to that calculated for the 95th percentile. Scaling for surface temperature and rainfall pairs is similarly calculated using equations (1) and (2). The results using surface temperatures repeat those previously published (Hardwick Jones et al. 2010, Herath et al. 2018, Wasko et al. 2018) and are presented for comparison to the scaling calculated using atmospheric temperatures. They are thus discussed after presentation of the scaling using atmospheric temperatures.

Finally, although the above method provided robust results, given the possibility of bias due to arbitrary bin size choice, quantile regression was also used to compare and verify the scaling results (Wasko and Sharma 2014, Ali et al. 2018). Quantile regression directly fits the relationship in equation (1) using all the rainfall-temperature pairs in a geopotential interval negating the need for binning. Results were broadly similar so only the binning results are discussed with results using quantile regression presented in the supplementary information.

3. Results

3.1. Scaling using atmospheric temperatures for geopotential interval 0–2.5 km

We begin by presenting results for the scaling of rainfall with atmospheric air temperature for Sydney Airport (−33.946, 151.1731) located in the temperate south-east of Australia and Darwin Airport (−12.4239, 130.8925) in the tropical north for the lowest geopotential interval (0–2.5 km). Sydney is presented here because it has a long record and has often been reported in previous similar studies; Darwin is presented because it is located in the tropical north of Australia and is well known to
have negative scaling (Hardwick Jones et al 2010, Barbero et al 2017, Wasko et al 2018). Here, even when atmospheric air temperature is used (figure 1(a)), the scaling in Darwin remains strongly negative (−8.1%/°C). Likewise, for Sydney (figure 1(b)), a temperate region where the rainfall scaling may be expected to be positive it is also weakly negative when atmospheric air temperature is used (−1.0%/°C). There is a suggestion of the reversal from positive to negative scaling (Drobinski et al 2016) for Sydney (figure 1(b)) but in general sites around Australia do not straddle the positive to negative scaling temperature change point (Wasko et al 2015). The values calculated correspond closely to the results presented using surface air temperatures in similar studies (Wasko et al 2018) suggesting atmospheric air temperature does not necessarily provide more robust estimates of the relationship of rainfall than surface temperature. However, when atmospheric dew point temperature is used the scaling direction reverses from negative to positive. The scaling is now 6.9%/°C and 7.1%/°C for Darwin (figure 1(c)) and Sydney (figure 1(d)) respectively. These values are similar to the Clausius–Clapeyron relationship.

The scaling using both atmospheric air and dew point temperature for all the stations is presented in figure 2. The sensitivity of atmospheric air temperature at the lowest interval (0–2.5 km), is mixed with both positive and negative scaling rates observed and no discernible spatial patterns (figure 2(a)). Only about half the sites have scaling rates that are significantly different from zero. There is a slight concentration of positive scaling along the eastern regions of Australia and some negative sensitivities on the western seaboard, similar to previous studies using surface temperatures and daily rainfall (Wasko et al 2016, Herath et al 2018). In the tropics there is a mix of positive and negative scaling, which may suggest improved results for locations with higher atmospheric moisture when atmospheric temperature is used, as previously using surface temperatures resulted in exclusively negative scaling (Hardwick Jones et al 2010, Wasko et al 2018).

In contrast, when the scaling is calculated using atmospheric dew point temperature (figure 2(b)) all sites have positive scaling, with the scaling at the majority of sites statistically significant. Most sites now
have a scaling above 6%/°C. This includes stations on the Australian mainland which could arguably have greater moisture limitations, and be more likely to have negative scaling, compared to those stations located on islands in the Pacific, Indian and Antarctic Oceans, which are less likely to be affected by moisture limitations (and have positive scaling with atmospheric air temperature). The results suggest that atmospheric dew point temperature gives robust estimates of the rainfall scaling relationship with temperature and is likely a better measure of the atmospheric moisture in the atmosphere than air temperature.

3.2. Scaling using atmospheric temperatures for geopotential height above 2.5 km
Results for all geopotential intervals are presented in figure 3. Figure 3(a) presents the scaling with atmospheric air temperature, and figure 3(b) with atmospheric dew point temperature. As discussed in section 3.1, at the lowest geopotential height (0–2.5 km) there is significant variability in the scaling (shown in figure 2). As the geopotential height increases more stations show positive rainfall scaling with atmospheric air temperature, particularly up to a height of 10–12.5 km. Stations in the west that had negative scaling at lower atmospheric heights exhibit positive scaling at 10–12.5 km. But some stations on the east coast of Australia continue to remain negative at this geopotential height. There is a change in the direction of the scaling starting approximately at 12.5 km with a shift to more negative scaling. This is most evident for eastern seaboard and island stations. As the tropopause is located at around 15–20 km this may explain the sudden change in gradient sign. The tropopause is characterised by stagnating or even gradually increasing temperature. As the temperature is monotonically decreasing from surface level to the tropopause, a shift in sign above the tropopause for the temperature gradient could be responsible for the change in scaling sign also.

The scaling of rainfall with atmospheric dew point temperature (figure 3(b)) is generally positive for all geopotential levels. Below 5 km, almost all stations have positive, statistically significant scaling. There is consistent positive scaling, generally above 6%/°C at a height below 5 km with limited variability between sites (compared to the atmospheric air temperature results presented above). At a geopotential height above 5 km some sites in the west of Australia have negative scaling. However, these stations are in the minority and the negative scaling is generally not statistically significant. The number of sites with statistically significant relationships reduces markedly above 15 km, again suggesting that above the tropopause the relationship between rainfall and atmospheric dew point temperature may not be as robust.

Figure 2. Rainfall–temperature scaling for Australia and surrounding territories using temperature for geopotential interval 0–2.5 km (a) atmospheric air temperature (b) atmospheric dew point temperature. Black outlines represent sites with statistically significant results (p-value < 0.05) while grey outlines represent sites that are not statistically significant (p-value ≥ 0.05).
To understand the variability and median scaling the results for all the stations are summarised as box plots in figure 4. The median scaling with atmospheric air temperature below 12.5 km is reasonably constant, between 2% and 4%/°C but there is large variability in the scaling with many sites exhibiting negative scaling. A clear decline in the scaling above 10 km is evident. In contrast, the atmospheric dew point scaling is almost always positive. The median scaling is highest, approximately 7%/°C, in the 0–2.5 km geopotential interval. It then decreases to approximately a constant median value between 2% and 4%/°C. Although the median values for the scaling calculated using the air and dew point atmospheric temperatures for the heights between 2.5 and 12.5 km are similar, the scaling calculated with air temperature exhibits much greater variability. The smallest interquartile ranges are observed in the geopotential intervals of 5–12.5 km. This suggest that the scaling calculated at these heights may be the most robust, but at these heights the absolute value of the scaling is lower than the scaling calculated closer to the surface. Above the tropopause the scaling with atmospheric dew point temperature remains largely consistent with the results at lower geopotential heights, but with greater variability.

3.3. Comparison to scaling calculated using surface temperatures

The scaling calculated using surface air and dew point temperatures was compared in figures 3 and 4 to the atmospheric temperatures results. The scaling calculated using surface temperatures is presented in detail in Wasko et al (2018) and hence the surface temperature scaling is presented here only as a reference. The scaling calculated using surface air temperature varies little to that calculated using atmospheric air temperature. There is some evidence that the atmospheric air temperature produces more sites with positive scaling, but the evidence of this is limited as the box plot interquartile ranges presented in figure 4 broadly overlap. Scaling calculated using surface dew point temperature has a median of almost 10%/°C, which is greater than the approximately 7%/°C median calculated using atmospheric dew point temperature. The greater than C–C scaling calculated at the surface suggests that storm invigoration is possible that results in rainfall increasing at a faster rate than the saturation vapour pressure increase with higher temperatures (Lenderink and van Meijgaard 2008, Wasko and Sharma 2015, Lochbihler et al 2017).
4. Discussion

Debate exists about the methods employed to calculate rainfall-temperature scaling (Wasko and Sharma 2014, Ali et al. 2018). There is evidence that quantile regression is resilient to many forms of bias as it considers the data as a whole, in contrast to the discretization of binning. Figure S1 replicates the results presented in figure 4 using quantile regression. The overall results using quantile regression were very similar to equal-width binning, with positive atmospheric dew point temperature scaling observed through all geopotential height intervals. As seen with equal-width binning, for the atmospheric air temperature a shift of scaling direction from positive to negative is displayed at 12.5 km and above. The spread of interquartile ranges observed using quantile regression is similar. The only difference is slightly more positive scaling observed throughout, but overall, this suggests that the results are robust regardless of the method employed to calculate the scaling.

Many studies make use of differing percentiles, such as the 90th and 99th percentiles. Given the similarity of the results presented here using the 95th percentile to studies which used the 99th percentile (e.g. Herath et al. 2018, Wasko et al. 2018) we do not expect the results or conclusions to change if a different percentile was adopted. The shift in scaling direction for atmospheric air temperature at the tropopause and increasing variability above the tropopause for atmospheric dew point temperature suggests there are additional factors that may need to be investigated before we fully understand how this atmospheric region affects empirical calculation of the rainfall-temperature scaling, however, we note that less 'weather' occurs at these heights.

Atmospheric air temperature is strongly linked to the amount of water in the atmospheric column (Hagemann et al. 2003, Neelin et al. 2009) and hence it could be expected that atmospheric air temperature would be strongly related to rainfall. The lack of a strong relationship between atmospheric air temperature and rainfall, but a strong relationship between atmospheric dew point temperature and rainfall suggests that rainfall is dependent on the relative humidity (Hardwick Jones et al. 2010, Wasko et al. 2015) and dew point temperature better captures this dependence (Lenderink and van Meijgaard 2010, Lenderink et al. 2011). The high variability in the scaling of rainfall with atmospheric air temperature may be a result of the scaling being affected by atmospheric circulations (Blenkinsop et al. 2013) and the type of weather event, in particular convective versus stratiform events (Berg et al. 2013). The fact that atmospheric dew point temperature changes are more robust suggests dew point may be a sufficient explanatory variable without considering the type of weather event.

5. Conclusions

The topic of rainfall scaling has been a very active area of research for the past decade. The fact that scaling calculated using surface air temperatures has increased variability due to surface conditions (e.g. Bao et al. 2017, Barbero et al. 2017) is highlighted by the negative scaling observed in tropical areas in Australia (e.g. Hardwick Jones et al. 2010, Wasko et al. 2015). The use of atmospheric air temperatures produced mixed scaling results with many stations also showing negative scaling. Historical increases in extreme rainfall with climatic change (Westra and Sisson 2011,
Guerreiro et al. (2018) naturally imply that we should expect to see positive extreme rainfall temperature sensitivities. The calculation of negative scaling using air temperature demonstrates that sensitivities of rainfall to both surface and atmospheric air temperature are not good indicators of the changing nature of rainfall with higher temperatures.

Conversely atmospheric dew point temperature performed robustly, yielding a more uniform positive scaling with less variation than atmospheric air temperature. The scaling was generally in the order of 2%–5%/°C regardless of the geopotential height employed, with a median of 7%/°C for a geopotential height interval of 0–2.5 km. A scaling of 7%/°C is consistent with C–C, however the scaling was still less than that found using surface dew point temperature. The super C–C scaling calculated using dew point temperature suggests the possibility of intensifying rainfall patterns with higher temperatures (Lenderink and van Meijgaard 2008, Wasko and Sharma 2015).

Although there is little consensus on what value of rainfall-temperature scaling we should expect (Zhang et al. 2017), the robust results using atmospheric dew point temperature support the assertion that a 1 °C increase in dew point temperature may be interpreted as an increase atmospheric moisture content and translates to an increase in the extreme rainfall (Lenderink and van Meijgaard 2010) at daily scales. As atmospheric temperatures are measured at fewer locations than surface temperatures, the use of atmospheric temperature will also result in less spatial definition for the rainfall-temperature scaling. The results here suggest that atmospheric dew point temperature is a more robust indicator of change in atmospheric moisture and rainfall than atmospheric air temperature but is unlikely to add additional value over the use of surface dew point measurements for scaling calculations.

Acknowledgments

Daily rainfall and surface air temperature data can be downloaded from the Australian Bureau of Meteorology website https://bom.gov.au/climate/data/stations/. Upper air data and dew point temperature data can be requested at cost from the Australian Bureau of Meteorology from the above site. Conrad Wasko acknowledges funding from the University of Melbourne McKenzie Postdoctoral Fellowships Program. Fiona Johnson is supported by a UNSW Scientia Fellowship and Australian Research Council Discovery Project DP150100411.

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