The dependence of precipitation and its footprint on atmospheric temperature in idealized extratropical cyclones

Samuel Phibbs1 and Ralf Toumi1

1 Space and Atmospheric Physics Group, Imperial College London, London, UK

Abstract Flood hazard is a function of the magnitude and spatial pattern of precipitation accumulation. The sensitivity of precipitation to atmospheric temperature is investigated for idealized extratropical cyclones, enabling us to examine the footprint of extreme precipitation (surface area where accumulated precipitation exceeds high thresholds) and the accumulation in different-sized catchment areas. The mean precipitation increases with temperature, with the mean increase at 5.40%/°C. The 99.9th percentile of accumulated precipitation increases at 12.7%/°C for 1 h and 9.38%/°C for 24 h, both greater than Clausius-Clapeyron scaling. The footprint of extreme precipitation grows considerably with temperature, with the relative increase generally greater for longer durations. The sensitivity of the footprint of extreme precipitation is generally super Clausius-Clapeyron. The surface area of all precipitation shrinks with increasing temperature. Greater relative changes in the number of catchment areas exceeding extreme total precipitation are found when the domain is divided into larger rather than smaller catchment areas. This indicates that fluvial flooding may increase faster than pluvial flooding from extratropical cyclones in a warming world. When the catchment areas are ranked in order of total precipitation, the 99.9th percentile is found to increase slightly above Clausius-Clapeyron expectations for all of the catchment sizes, from 9 km² to 22,500 km². This is surprising for larger catchment areas given the change in mean precipitation. We propose that this is due to spatially concentrated changes in extreme precipitation in the occluded front.

1. Introduction

Fluvial flooding is caused by high total precipitation in a given catchment area. It is therefore dependent on both heavy precipitation and spatially concentrated precipitation falling within a catchment area; in contrast, extreme precipitation in a small area is capable of causing pluvial flooding, with the risk of pluvial flooding enhanced when there is a greater area of extreme precipitation [Chen et al., 2010]. Therefore, both the surface area of precipitation and the total precipitation in catchment areas relate to flood hazard. Here, for the first time, their dependence on atmospheric temperature is investigated for extratropical cyclones.

The Clausius-Clapeyron (C-C) relation describes the water-holding capacity of the atmosphere and implies an approximate increase of 7%/°C at typical near-surface temperatures [Trenberth et al., 2003]. Global mean precipitation is not predicted to increase at this rate due to the constraint of the energy budget of the troposphere [Allen and Ingram, 2002]. However, there is growing evidence that extreme precipitation can increase at a rate greater than both the global mean and C-C scaling. This evidence comes in the form of observational and modeling studies, examples include Berg and Haerter [2013], Lenderink and Van Meijgaard [2008, 2010], Liu et al. [2009], Singleton and Toumi [2013], Westra et al. [2014], Kendon et al. [2014], Molnar et al. [2015], and Berg et al. [2013]. One suggested reason for the super C-C scaling of precipitation is that the extra latent heat release invigorates the storm [Trenberth et al., 2003]. Alternatively, it may be a result of a transition from large-scale to convective precipitation as temperature increases [Berg and Haerter, 2013]. Although, it has been shown that super C-C changes can occur in a simple entraining plume model [Loriaux et al., 2013]. There is also evidence that the extreme precipitation in extratropical cyclones will increase at a rate greater than the mean [Bengtsson et al., 2009]. A high-resolution study over the UK revealed super C-C changes for heavy summer time precipitation, with about half of the heaviest events in the summer associated with large-scale storms (embedded convection within a front, mesoscale convective systems, or squall lines) [Kendon et al., 2014]. An important advantage of modeling idealized extratropical cyclones is that any change in precipitation will be a direct result of the change in temperature, rather than a transition between the proportion of...
different types of weather systems. The sensitivity of an idealized extratropical cyclone to increased moisture has been tested previously [Booth et al., 2013]. The authors found monotonic increases in the storms’ intensification rate, central minimum pressure, extreme surface winds, and precipitation with increased moisture. In addition, they found that the horizontal scale of the storm decreased when moisture was increased beyond current levels. This study focuses on changes in precipitation with increased temperature and hence moisture, and differs from Booth et al. [2013] as the experiments are carried out at a resolution where convection is explicitly represented. This increases the detail and robustness of the results allowing several different aspects of the precipitation change with temperature to be examined: the footprint of precipitation; a comparison of how precipitation changes compare with those we would expect from the C-C relation; the spatial pattern of the extreme precipitation; and the total accumulated precipitation in catchment areas of different sizes.

2. Methodology

The Weather Research and Forecasting model (WRF v3.6.1) [Skamarock and Klemp, 2008] was used to simulate an idealized extratropical cyclone. The initial condition of all the experiments was a baroclinically unstable jet in hydrostatic balance. The domain was 1333 × 2667 with 3 km horizontal resolution, 65 \( \eta \) levels, a model top of 16 km, and the Coriolis parameter was set to be constant at 0.0001 s\(^{-1}\). The initial wind field was a zonal jet with a maximum wind speed of 67 m/s with the meridional and vertical velocity set to zero. The zonal jet and potential temperature field are shown in the supporting information. The initial relative humidity (RH) is specified as

\[
\text{RH} = 1 - 0.9 \left( \frac{Z}{8000} \right)^{1.25},
\]

where \( Z \) is the height (m) and RH is set to 0.1 above 8000 m and to have a maximum of 0.7 throughout. The lateral boundaries were periodic in the \( x \) direction and symmetric in the \( y \) direction. The surface boundary condition was set to be land at a constant temperature identical to the initial surface temperature, with a constant roughness length of 0.1 m and a moisture availability of 0.3 (where a moisture availability of 1 represents an ocean surface). To initiate the experiments a thermal perturbation (1°C) was applied. The periodic boundaries allowed the storm to cross the domain several times, meaning that some grid cells had significant precipitation because the same front had passed them twice. To prevent this from influencing the results, in postprocessing the hourly precipitation field was mapped onto a domain twice the size, so the precipitation accumulations represented a single storm passing over a large domain. The following schemes were chosen: Kessler microphysics [Kessler, 1969] and Yonsei University planetary boundary layer [Hong et al., 2006]. No shortwave or longwave physics were present. The experiments were run without a cumulus parameterization scheme. When a cumulus scheme was tested for the control run, it was responsible for a small fraction of the precipitation, showing that at 3 km resolution convection is explicitly represented; in contrast, when the control experiment was also run at 10 km and 20 km, the cumulus scheme was responsible for a significant proportion of the precipitation.

The sensitivity to atmospheric temperature was investigated by increasing the initial potential temperature uniformly vertically and horizontally in order to maintain dry stability, while including the impact of increased water vapor. The initial jet was kept the same in all experiments. The surface temperature was increased at the same rate as the atmospheric temperature. Relative humidity was held constant, as a wide variety of global climate models predict that relative humidity will stay approximately constant under global warming [Ingram, 2002]. Results from five experiments are presented for the idealized extratropical cyclone: −1°C (\( \theta - 1 \)), control (CTL), +1°C (\( \theta + 1 \)), +2°C (\( \theta + 2 \)), and +3°C (\( \theta + 3 \)).

A 5 day period was chosen in which the accumulated precipitation was examined for every 1 h, 6 h, 12 h, and 24 h period. The 99.9th percentiles of accumulated precipitation were found by ranking the grid cells with precipitation. The footprint of extreme precipitation was defined as the sum of grid cells above the 99.9th percentile for the CTL. The footprint was calculated for each accumulation period within the 5 days and then the mean was taken.

In order to compare the experiment precipitation with C-C expectations the CTL experiment precipitation \( \text{pr}_{\text{CTL}} \) was scaled by a fixed percentage for every degree of temperature change relative to the CTL \( i \), to give an expected precipitation \( X(i) \):

\[
X(i) = 1.0736 \times \text{pr}_{\text{CTL}}
\]
3. Results

In each experiment the initial baroclinically unstable jet was simulated for 10 days. The intensification and domain total precipitation are shown in Figure 1. After 72 h an extratropical cyclone began to evolve and the minimum surface pressure underwent an almost identical rapid decrease over 48 h of approximately 40 hPa in all the experiments. In the next 48 h there was a more gradual intensification of minimum surface pressure. During this period the warmer experiments generally reached a lower minimum surface pressure. From 168 h the cyclones gradually weakened. Precipitation began during the rapid intensification, with the domain total reaching its peak, in all experiments, approximately coincidental with maximum intensity. A 5 day period was chosen for the analysis period from the 72nd to the 192nd hour, as this corresponded to the development and period of maximum intensity of the extratropical cyclones but before they began to significantly dissipate. Figure 2 shows the surface pressure and precipitation of the CTL experiment at 120 h and 190 h. The cyclone is moving from west to east at approximately 15 m/s. The distribution of precipitation is similar to the conceptual Norwegian cyclone model \cite{Bjerknes1919}. In Figure 2a the most extreme precipitation is occurring southeast of the cyclone center in a narrow band associated with the surface cold front. North of this band is a wider and shorter band of intense precipitation associated with the warm front. The wider area of weaker precipitation, east of the cyclone center, is associated with the occluded front, visible in the temperature at higher levels (not shown). In Figure 2b the precipitation is largely associated with the occluded front.

Figure 3a shows that the mean precipitation increases with temperature, the average increase was 5.40%/°C. This increase is below the expected increase in specific humidity from the C-C relation 7.36%/°C. However, the evolution of the change in mean precipitation (not shown) reveals close to C-C scaling up to 150 h followed by a generally sub-C-C increase. Figure 1 also shows the domain total precipitation for the different experiments increasingly converges after 150 h. We suggest that the dynamics of the different extratropical cyclones are such that after 150 h most of the initial moisture that will be rained out has been removed, and the
extratropical cyclones then tend to the same state. To quantify how the extreme precipitation changed with temperature, the 99.9th percentiles of both the combined 1 h and 24 h accumulated precipitation over the 5 day analysis period were examined, with zero values excluded due to the size of the domain. Figure 3a shows both the 99.9th percentiles generally increasing at a rate greater than both C-C scaling and the mean. The average increases in the 99.9th percentiles were 12.7%/°C and 9.38%/°C for the combined 1 h and 24 h accumulated precipitation, respectively. In the CTL the accumulated precipitation mean (storm total) was 5.8 mm (excluding zero values), the 99.9th percentiles were 6.6 mm and 27.5 mm for the 1 h and 24 h accumulated precipitation, respectively. The values of extreme accumulated precipitation over 24 h were only approximately 3 times the 1 h values due to a combination of the speed of the extratropical cyclone and the concentration of the most extreme precipitation into a relatively narrow band behind the cold front.

The footprint of extreme precipitation was found to increase significantly with temperature. To investigate changes in the footprint of extreme precipitation the surface area above the 99.9th percentile of the CTL accumulated precipitation was determined. Figure 3b shows that this surface area grew consistently with temperature for 1 h, 6 h, 12 h, and 24 h accumulated precipitation over the 5 day analysis period. To give an idea of the scale of the absolute changes in the footprint of extreme precipitation, the surface area above the 99.9th percentile of the CTL for 1 h accumulated precipitation increased from 392,000 km² for \( \theta - 1 \) to 1,460,000 km² for \( \theta + 3 \) and the same increase for 24 h was from 36,900 km² to 270,000 km². The relative growth in the footprint of extreme precipitation with temperature was generally larger for longer accumulation periods — mean increases are 74%/°C and 282%/°C for 1 h and 24 h, respectively. Longer periods have greater absolute precipitation, so when the individual grid values are relatively increased (the C-C effect), the cumulative total (the footprint) undergoes a greater change relative to the smaller precipitation of the shorter time periods. Figure 3b also shows the increase in the footprint of extreme precipitation if the CTL precipitation increased by C-C scaling. A clear pattern of super C-C increases in the footprint of extreme precipitation emerges. Absolute increases in the footprint of extreme precipitation with temperature were progressively closer to expectations from C-C relation for accumulated precipitation over longer time periods. The fact that the 24 h footprint of extreme precipitation was closer to C-C expectations gives evidence that the longer periods have greater relative changes due to the differences in the absolute magnitude of precipitation accumulation.

Figure 2. Isobars of surface pressure (hPa) with hourly accumulated precipitation (mm) of the CTL, x and y axes are the distance from the minimum surface pressure (km) for (a) 120 h and (b) 190 h.
Figure 3. (a) Mean precipitation and the 99.9th percentiles (excluding zero values) of the combined 1 h and total 24 h accumulated precipitation, normalized to the CTL (0°C reference), with Clausius-Clapeyron scaling. (b) Surface area of 1 h, 6 h, 12 h, and 24 h accumulated precipitation above their respective CTL 99.9th percentile (6.6 mm, 16.0 mm, 23.1 mm, and 27.5 mm) over the 5 day analysis period and normalized to the CTL (0°C reference). Solid lines represent the outcome of the experiments. Dashed lines show the increase in the surface area if all the CTL precipitation is scaled with Clausius-Clapeyron expectations.

Figures 4a and 4b show the distribution of the surface area of precipitation for the 1 h and 24 h accumulated precipitation, respectively. In Figure 4b we see marked sub-C-C increases below certain values for the $\theta + 2$ and $\theta + 3$ cases. Sub-C-C increases are expected as the mean precipitation scales slightly below C-C; therefore, the super-C-C scaling of the tail of the distribution must be balanced with the sub-C-C elsewhere. The extreme 1 h accumulated precipitation is spread over a large area; therefore, the extreme 24 h precipitation is often due to persistent moderate precipitation. This combination of short duration extreme and persistent moderate precipitation that makes up the 24 h extreme precipitation, as well as the smaller sample size than the 1 h case, explains the different distribution shapes for individual temperatures. In order to examine the robustness of the results thus far, the 99th percentile was chosen and showed similar results.

The surface area of all precipitation — sum of grid cells with nonzero precipitation — was considered. In contrast to the footprint of extreme precipitation, the surface area of all precipitation was found to decrease with increasing temperature, despite the increase in precipitation with temperature (Figure 3a). Figure 5 shows that the relative decrease in surface area of all precipitation can be seen at nearly all times in the analysis period. The opposite pattern that can be seen between 72 h and 96 h occurs when the surface area of precipitation is small, as the cyclone is developing (see Figure 1b). The average decrease in the surface area of all precipitation, during the analysis period, between the experiments as temperature was increased was 1.34%. In absolute terms the decrease from CTL to $\theta + 1$ was 4,429,000 km$^2$ to 4,372,000 km$^2$.

The spatial structure of the footprint of extreme precipitation was examined. Figure 6 shows the size distribution of clusters (spatially linked grid cells) for the 1 h accumulated precipitation for $\theta - 1$, $\theta + 1$, and $\theta + 3$ as examples. Whether the precipitation clusters were linked temporally was not considered here. A clear increase in the number of larger clusters of extreme precipitation with increasing temperatures is visible. This leads to a monotonic decrease in fragmentation — number of unique clusters that make up the extreme precipitation — when normalized by the total area as shown in Figure 7. The correlation with temperature is
Figure 4. Surface area of precipitation in 1 mm intervals over the 5 day analysis period. Solid lines represent the outcome of the experiments. Dashed lines show the increase in the area if all the CTL precipitation is scaled with the Clausius-Clapeyron relation. The reference lines show the location of the 99.9th percentile: (a) 1 h accumulated precipitation; (b) 24 h accumulated precipitation.

Figure 5. Total surface area of precipitation normalized to the CTL over a 5 day analysis period.
Figure 6. Number of clusters (linked grid cells) of 1 h accumulated precipitation above the CTL 99.9th percentile, binned by size of cluster from 0 to 6500 grid cells with intervals of 500 and normalized by their respective total number of clusters. (a) $\theta - 1$; (b) $\theta + 1$; (c) $\theta + 3$.

Figure 7. Total number of independent clusters (linked grid cells) that make up the 1 h accumulated precipitation above the CTL 99.9th percentile, normalized by the total number of grid cells above the 99.9th percentile for the CTL. Solid lines represent the outcome of the experiments. Dashed lines show the increase in the surface area if all the CTL precipitation is scaled with Clausius-Clapeyron expectations.
clear, with an $R^2$ squared value of 0.75. A decrease of fragmentation is expected as a result of the C-C scaling, as previously nearby but distinct clusters of precipitation will become linked by grid cells with precipitation close to but below the 99.9th percentile of the CTL. The experiments showed a decrease in fragmentation similar to that expected from the C-C relation.

The region of the domain where the precipitation accumulated in the 5 day analysis period was identified. This entire region was then divided into equally sized squares that represented catchment areas. Different sizes of catchment areas, by varying the length of the squares, were investigated to determine the implications of the results on flood risk over different scales. Figure 8 shows how the number of catchment areas exceeding extreme accumulated precipitation totals changes with temperature for different-sized catchment areas. For each catchment area size two aspects of precipitation were considered. First, how does the number of catchment areas exceeding a fixed extreme precipitation change with temperature? Second, how does the total precipitation change in the catchment areas inundate with extreme accumulated precipitation? For the first question the fixed extreme precipitation was chosen as the 99.9th percentile of the mean precipitation in the catchment area for the CTL. The vertical lines in Figure 8 show how the number of catchment areas exceeding this metric increases with temperature, this is also shown in Figure 9a. For the warmer experiments the change is consistently above C-C relation expectations, except for the 22,500 km$^2$ area. For 9 km$^2$ the number of catchment areas exceeding 30.9 mm was 3460 for the CTL and 39,000 for $\theta + 3$; for 22,500 km$^2$ the number of catchment areas exceeding 27.4 mm was 1 for the CTL and 32 for $\theta + 3$. For the second question the mean precipitation in the 99.9th percentile catchment area for each experiment was chosen as the metric (indicated by the horizontal line in Figure 8). Figure 9b shows that this precipitation increases with temperature either above or approximately at the C-C rate for all temperature intervals. The mean precipitation accumulated in the 99.9th percentile of the 9 km$^2$ catchment area increased from 24.8 mm for $\theta + 3$ in contrast to the 30.8 mm expected from the C-C relation; the same numbers for the 22,500 km$^2$ catchment area size were 23.8 mm for the CTL and 30 for $\theta + 3$ compared to 29.3 mm expected from the C-C relation. For all the different sized catchment areas the ones with the most extreme accumulated precipitation were largely in the path of the occluded front, associated with the wider band of weaker precipitation in Figure 2a, rather than the cold or warm front. The fronts were distinguished from the vertical velocity and temperature field (not shown).
Figure 9. (a) Total number of catchment areas exceeding 99.9th percentile of mean precipitation of the CTL normalized to the CTL. (b) Mean precipitation where number of catchment areas exceeding that precipitation is equal to the number of catchment areas that exceed the 99.9th percentile of mean precipitation of the CTL.

4. Discussion

Monotonic increases in the mean and extreme precipitation with temperature, and hence moisture, are in agreement with increases in accumulated precipitation and precipitation rates found previously, for example, Booth et al. [2013]; a direct comparison with differences in the changes in large-scale and cumulus scheme precipitation is not possible, because the experiments in this study were run at a resolution where convection was explicitly represented. The mean precipitation scales close to C-C rate up to the peak precipitation, and this fits with the expectation that the heaviest rainfall events will scale with the available moisture [Ingram, 2002]. The extreme precipitation generally increased above the mean and C-C rate with increasing temperature, as seen in Figure 3a. A commonly suggested reason for this is that higher temperatures lead to higher specific humidity; in instances of heavy rainfall this will lead to additional latent heat release resulting in more uplift and hence more rainfall [Trenberth et al., 2003]. The super C-C scaling of the 99.9th percentile is similar to the above C-C scaling of precipitation in other studies [Berg and Haerter, 2013; Lenderink and Van Meijgaard, 2008, 2010; Liu et al., 2009; Singleton and Toumi, 2013; Westra et al., 2014]. Until now it has been difficult to model the large-scale conditions and resolve convection simultaneously [Westra et al., 2014], this study models both the large-scale conditions and explicitly represents convection in an idealized setup and therefore provides important further evidence for super C-C scaling of extreme precipitation. The super C-C scaling of the 99.9th percentile was stronger for the 1 h accumulated precipitation than the 24 h accumulated precipitation. This agrees with other studies where super C-C scaling was seen on shorter time scales [Lenderink and Van Meijgaard, 2008; Westra et al., 2014]. There is a greater constraint of available moisture for precipitation accumulated over longer time periods [Trenberth et al., 2003], leading to more of the extreme precipitation being made up of moderate precipitation, and this will in turn exhibit weaker super C-C increases with temperature, as seen in Figure 4.

The 99.9th percentile generally scaled above C-C for all time periods considered (not shown) and this differs from Singleton and Toumi [2013], where super C-C scaling of precipitation only existed in 1 h accumulated precipitation at high temperatures. The difference is thought to arise due to the scale of the system being
modeled. The increased propagation speed of the squall line at warmer temperatures masked the super C-C increase in extreme precipitation [Singleton and Toumi, 2013]. An extratropical cyclone has a much larger scale, and thus in this case the result is a super C-C increase in the extremes that is noticeable across a range of time periods.

A simple positive shift in a precipitation distribution will lead to large changes in the footprint of extreme precipitation relative to the original. Large increases in the footprint of extreme precipitation with temperature are found here (Figure 3b). Changes in the footprint of extreme precipitation have been compared with the change expected if the CTL precipitation is scaled with C-C. Increases in the footprint of extreme precipitation generally exceeded that expected from a C-C scaling of precipitation. This super C-C scaling of the surface area of extreme precipitation is due to the 99.9th percentile of accumulated precipitation scaling more than C-C. When the CTL precipitation is scaled by the average increase found in the 99.9th percentile, the increase in the footprint of extreme precipitation was close to the actual outcome of the experiments for all time periods considered (not shown). For 1 h accumulated precipitation the slightly greater rate of increase in the extreme precipitation with temperature compared to C-C (12.7%/°C versus 7.36%/°C), can lead to large differences in the surface area of extreme precipitation compared to those we would expect from C-C (74%/°C versus 43%/°C). The increase in the size of the footprint with temperature is closer to C-C expectations for longer accumulation periods.

The footprint of extreme precipitation has an impact on the likelihood of flooding—extreme precipitation over a larger area will increase the chance of a catchment area being inundated and precipitation exceeding the threshold for flooding occurring. The increase in the footprint of extreme precipitation is particularly large for accumulated precipitation over longer time periods. This suggests that under warming the fluvial flood risk (over larger catchments and longer duration) will increase. To fully understand how fluvial flooding might be affected we need to know whether the expansion of the footprint is spatially connected and ultimately how varyingly sized catchment areas will "see" this change in precipitation.

The decrease in the surface area of all precipitation with temperature, seen in Figure 5, is surprising given the increase in the mean precipitation. But it fits with the picture of the horizontal extent of extratropical cyclones decreasing when moisture is increased beyond today’s levels found in Booth et al. [2013].

An analysis of the clustering of the precipitation reveals how the footprint of extreme precipitation changed, as in Figure 6. An increase in the number of larger clusters was found, approximately in line with expectations from the C-C relation. This increase in the size of clusters indicates that the expansion of the footprint of extreme precipitation is likely to affect a single catchment area and hence increase the chance of a pluvial area being inundated and flooding occurring.

The experiments, with the same initial conditions, were also integrated with an ocean surface rather than a land surface. This resulted in the mean precipitation scaling closer to C-C, greater super C-C changes in the 99.9th percentile of precipitation accumulated over 1 h and 24 h, and similar substantial and generally super C-C increases in the footprint of extreme precipitation with temperature. These results are shown in Figure S2 in the supporting information. There was also both a decrease in the surface area of all precipitation and greater extremeprecipitationclusteringwithtemperature(notshown).

The impact of temperature change on precipitation accumulated on catchment areas of different sizes was investigated. It is this analysis of our experiments that gives the greatest insight into how flood risk from extratropical cyclones might change in the future. Both the number of catchment areas exceeding a fixed extreme total accumulation of precipitation and the magnitude of total accumulated precipitation in the catchment areas at the high end of the distribution were considered. The former tells us how we might expect the frequency of catchment areas inundated with extreme precipitation to change, and the latter tells us how much the most extreme events might change, both with respect to increasing temperature in extratropical cyclones. Both exhibited super C-C behavior. The relative change in the catchment areas exceeding a fixed extreme total accumulation of precipitation was generally greater for the larger catchment areas. This is consistent with the greater relative changes in the footprint of extreme precipitation for longer accumulation periods. It implies that the risk of pluvial flooding is enhanced, but smaller relative to the risk of fluvial flooding. Our finding is only relevant for floods caused by extratropical cyclones; other weather systems may respond differently to temperature.
When the sensitivity of the total precipitation across catchment areas of different sizes to temperature was compared, one might expect larger catchment areas not to exhibit super C-C behavior, as the effect of super C-C precipitation in one place and time in the catchment area could be balanced by sub-C-C behavior elsewhere. Indeed, most of the time we should expect this, as the mean precipitation scales with temperature at approximately the rate expected from the C-C relation. However, surprisingly, we find similar precipitation change across the range of different-sized catchment areas. We believe this is a result of the spatial concentration of super C-C changes in the extreme precipitation, as indicated by Figure 6. It is also interesting to note that while the most intense hourly precipitation is located in the cold front (Figure 2a), the catchment areas with the most extreme accumulated precipitation were mainly (or completely for the two larger catchment areas sizes) in the path of the occluded front where the precipitation was less intense but more sustained.

Precipitation remains one of the most difficult variables to accurately simulate in weather prediction and climate models. Global models significantly underestimate intense precipitation in the midlatitudes while compensating in light rain [Stephens et al., 2010]. A study into extratropical cyclones during warm and cold periods showed that certain increases in maximum precipitation during warm periods were only significant in the high-resolution model [Li et al., 2014]. When we carried out the same control experiment on coarser grids with a horizontal resolution of 10 km and 20 km, the extremes in hourly precipitation intensity were significantly lower than at 3 km resolution. This is most likely a result of the precipitation not being fully resolved, and the cumulus parameterization not adequately representing extreme precipitation. All of this suggests that relatively low-resolution global climate models may not fully capture increases of extremes in precipitation or the surface area of heavy precipitation with temperature in extratropical cyclones.

Precipitation in extratropical cyclones in nature is influenced by many factors that are not included in this idealized setup; examples include a wide range of initial conditions, interaction with other weather systems, and topography. These factors could lead to quite different values for footprint of extreme precipitation than those found for this idealized case. To provide further insight into the sensitivity of the extreme precipitation it would be necessary to simulate a range of past extratropical cyclones and examine their responses to temperature. It should also be noted that this study only examines the thermodynamic response of extratropical cyclones; dynamic responses to global warming, such as a shift in the storm track over Europe, may dominate changes in flooding over individual countries.

5. Conclusions

Idealized numerical extratropical cyclone experiments were used to investigate the dependence of precipitation on atmospheric temperature. There were four principal findings. First, super C-C increases in the extreme precipitation were found, for accumulations of 1 h to 24 h, for the first time in an idealized extratropical cyclone with convection explicitly represented. The high-resolution and idealized nature of the experiments allowed us to investigate the hitherto unexamined dependence of the footprint of extreme precipitation on temperature. This leads us to the second conclusion: there are substantial increases in the footprint of extreme precipitation with increases of temperature, these were generally super C-C. The increase in the footprint was accompanied by greater precipitation clustering. To assess the potential consequences of the results on flood risk caused by extratropical cyclones the domain was divided into catchment areas. Our third finding was that greater relative changes in the number of catchment areas exceeding extreme total precipitation accumulations were found for larger rather than smaller catchment areas. An implication of this finding is that extreme fluvial flooding may increase more rapidly than pluvial flooding in a warming world. Finally, super C-C behavior was found in the catchment areas inundated with some of the most extreme precipitation, this effect was present even for large catchment areas of up to 22,500 km².

References

Allen, M. R., and W. J. Ingram (2002), Constraints on future changes in climate and the hydrologic cycle, Nature, 419(6903), 224–232, doi:10.1038/nature01092.
Bengtsson, L., K. I. Hodges, and N. Keenlyside (2009), Will extratropical storms intensify in a warmer climate?, J. Clim., 22(9), 2276–2301.
Berg, P., and J. O. Haerter (2013), Unexpected increase in precipitation intensity with temperature a result of mixing of precipitation types?, Atmos. Res., 119, 56–61.
Berg, P., C. Moseley, and J. O. Haerter (2013), Strong increase in convective precipitation in response to higher temperatures, Nat. Geosci., 6(3), 181–185, doi:10.1038/ngeo1731.
Bjerknes, J. (1919), On the structure of moving cyclones, Mon. Weather Rev., 47(2), 95–99.
Booth, J. F., S. Wang, and L. Polvani (2013), Midlatitude storms in a moister world: Lessons from idealized baroclinic life cycle experiments, Clim. Dyn., 41(3–4), 787–802.

Chen, A., S. Djordjevic, J. Leandro, and D. Savic (2010), An analysis of the combined consequences of pluvial and fluvial flooding, Water Sci. Technol., 62, 1491–1498.

Hong, S.-Y., Y. Noh, and J. Dudhia (2006), A new vertical diffusion package with an explicit treatment of entrainment processes, Mon. Weather Rev., 134(9), 2318–2341.

Ingram, W. J. (2002), On the robustness of the water vapor feedback: GCM vertical resolution and formulation, J. Clim., 15(9), 917–921.

Kendon, E. J., N. M. Roberts, H. J. Fowler, M. J. Roberts, S. C. Chan, and C. A. Senior (2014), Heavier summer downpours with climate change revealed by weather forecast resolution model, Nat Clim. Change, 4(7), 570–576.

Kessler, E. (1969), On the distribution and continuity of water substance in atmospheric circulation, Meteorol. Monogr., 10(32), 84.

Lenderink, G., and E. Van Meijgaard (2010), Increase in hourly precipitation extremes beyond expectations from temperature changes, Nat. Geosci., 1(8), 511–514, doi:10.1038/ngeo262.

Lenderink, G., and E. Van Meijgaard (2010), Linking increases in hourly precipitation extremes to atmospheric temperature and moisture changes, Environ. Res. Lett., 5(2), 025208.

Li, M., T. Woollings, K. Hodges, and G. Masato (2014), Extratropical cyclones in a warmer, moister climate: A recent Atlantic analogue, Geophys. Res. Lett., 41, 8594–8601, doi:10.1002/2014GL062186.

Liu, S. C., C. Fu, C.-J. Shiu, J.-P. Chen, and F. Wu (2009), Temperature dependence of global precipitation extremes, Geophys. Res. Lett., 36, L17702, doi:10.1029/2009GL040218.

Loriaux, J. M., G. Lenderink, S. R. De Roode, and A. P. Siebesma (2013), Understanding convective extreme precipitation scaling using observations and an entraining plume model, J. Atmos. Sci., 70(11), 3641–3655.

Molnar, P., S. Fatichi, L. Gaál, J. Szolgay, and P. Burlando (2015), Storm type effects on super Clausius-Clapeyron scaling of intense rainstorm properties with air temperature, Hydrol. Earth Syst. Sci., 19(4), 1753–1766, doi:10.5194/hess-19-1753-2015.

Singleton, A., and R. Toumi (2013), Super-Clausius-Clapeyron scaling of rainfall in a model squall line, Q. J. R. Meteorol. Soc., 139(671), 334–339.

Skamarock, W. C., and J. B. Klemp (2008), A time-split nonhydrostatic atmospheric model for weather research and forecasting applications, J. Comput. Phys., 227(7), 3465–3485.

Stephens, G. L., T. Lecuyer, R. Forbes, A. Gettleman, J.-C. Golaz, A. Bodas-Salcedo, K. Suzuki, P. Gabriel, and J. Haynes (2010), Dreary state of precipitation in global models, J. Geophys. Res., 115, D24211, doi:10.1029/2010JD014532.

Trenberth, K. E., A. Dai, R. M. Rasmussen, and D. B. Parsons (2003), The changing character of precipitation, Bull. Am. Meteorol. Soc., 84(9), 1205 – 1217.

Westra, S., H. J. Fowler, J. P. Evans, L. V. Alexander, P. Berg, F. Johnson, E. J. Kendon, G. Lenderink, and N. M. Roberts (2014), Future changes to the intensity and frequency of short-duration extreme rainfall, Rev. Geophys., 52, 522 – 555, doi:10.1002/2014RG000464.