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Linking increases in hourly precipitation extremes to atmospheric temperature and moisture changes

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Abstract

Relations between hourly precipitation extremes and atmospheric temperature and moisture derived for the present-day climate are studied with the aim of understanding the behavior (and the uncertainty in predictions) of hourly precipitation extremes in a changing climate.

A dependency of hourly precipitation extremes on the daily mean 2 m temperature of approximately two times the Clausius–Clapeyron (CC) relation is found for temperatures above 10 °C. This is a robust relation obtained in four observational records across western Europe. A dependency following the CC relation can be explained by the observed increase in atmospheric (absolute) humidity with temperature, whereas the enhanced dependency (compared to the CC relation) appears to be caused by dynamical feedbacks owing to excess latent heat release in extreme showers.

Integrations with the KNMI regional climate model RACMO2 at 25 km grid spacing show that changes in hourly precipitation extremes may indeed considerably exceed the prediction from the CC relation. The results suggests that increases of +70% or even more are possible by the end of this century. However, a different regional model (CLM operated at ETHZ) predicts much smaller increases; this is probably caused by a too strong sensitivity of this model to a decrease in relative humidity.

Keywords: extreme precipitation, climate change, water vapor, Clausius–Clapeyron relation

 Online supplementary data available from stacks.iop.org/ERL/5/025208/mmedia

1. Introduction

Events of extreme precipitation have a huge influence on society. They are associated with flooding, erosion and water damage, and may have impacts on transport and safety. It is commonly expected that precipitation extremes will increase as the climate warms (Trenberth *et al* 2003, Pall *et al* 2007, Groisman *et al* 2005, Emori and Brown 2005).

The primary reason why precipitation extremes are expected to increase follows from the fact that a warmer atmosphere can ‘hold’ more moisture. The increase in the moisture-holding capacity of the atmosphere with temperature occurs at a rate given by the Clausius–Clapeyron relation: approximately 7% per degree temperature rise. If the relative

humidity in the future climate remains approximately the same as in the present-day climate—which is generally expected based on model results and physical arguments (Held and Soden 2006)—the amount of water vapor in the atmosphere will increase at the same rate. Now, the commonly used argument is that in extreme precipitation events all water vapor in the air (or a constant fraction thereof) is converted to rain. Hence, extreme precipitation will scale with the Clausius–Clapeyron relation.

There is model evidence from global climate models (GCMs) that precipitation extremes indeed increase at the rate predicted by the Clausius–Clapeyron (hereafter CC) relation (Allen and Ingram 2002, Pall *et al* 2007). However, a CC scaling is not generally obtained in models. Also, despite

the conceptual understanding as outlined above, there is no reason why extremes should follow the CC relation exactly. For instance, changes in the dynamics of the atmosphere and the convective cloud, the size of the convective cloud, and the moist adiabatic temperature profile could cause deviations from the CC scaling (Trenberth *et al* 2003, O’Gorman and Schneider 2009).

Lenderink and van Meijgaard (2008) found a dependency of approximately 7% per degree for temperatures below and 14% per degree for temperatures above 10°C for extreme hourly precipitation in De Bilt (The Netherlands). Thus for higher temperatures the dependency of hourly precipitation extremes on temperature is about twice as large as predicted by the CC relation.

The cause of this super CC scaling in the observations is under debate (Haerter and Berg 2009, Lenderink and van Meijgaard 2009). In our opinion, the explanation can be found in the physics of precipitating convective clouds. The strength of the upward motions in these clouds is (largely) determined by latent heat release in the cloud. More rainfall formation implies more condensation and therefore more latent heat release, which leads to stronger cloud updrafts feeding back again onto the rainfall formation.

Alternatively, changes in the frequency of different precipitation types with temperature may affect the scaling as well. In mid-latitude regions, precipitation at low temperatures mostly occurs in the winter season with the prevalence of large-scale cyclones. In these synoptic systems rainfall intensities are generally low and durations relatively long (typically 4–6 h). Conversely, precipitation at high temperatures occurs mainly in summer with the frequent occurrence of convective rainfall characterized by high intensities and short durations (order 1 h). Therefore, the change from low temperatures to higher temperatures involves both a change from the winter to the summer season as well as a change from predominantly large-scale to convective precipitation. This could statistically induce an enhanced (compared to the CC relation) temperature dependency of hourly precipitation (Haerter and Berg 2009).

Another point of debate is whether precipitation extremes relate to atmospheric moisture (Berg *et al* 2009), or to the maximum moisture-holding capacity of the atmosphere, and therefore to temperature. A dependency on temperature could also result from changes in atmospheric stability, with higher surface temperatures leading to larger vertical instability and, therefore, more vigorous convection.

It is important to establish the cause of the super CC scaling because this has direct consequences as to whether (and how) the relation between hourly precipitation extremes and temperature derived from the present-day climate will manifest itself in the climate change signal. If the super CC scaling is caused by the temperature dependency of the relative frequency of large-scale versus convective precipitation, as proposed by Haerter and Berg (2009), then the scaling might not be a valid predictor of climate change as this relative frequency is not likely to change much as climate changes. However, if the super CC scaling is due to how convective precipitation relates to increased atmospheric moisture, then it could be useful as a predictor of climate change.

Indeed, in a climate scenario simulation with the regional climate model RACMO2, an increase of extreme hourly precipitation close to two times the CC relation has been found (Lenderink and van Meijgaard 2008). However, the spread in model predicted changes of precipitation extremes, in particular for the summer season, is generally considerable (Frei *et al* 2006, Fowler *et al* 2007), and it is not clear how robust such a response is.

In this paper we focus on western Europe, which we define as the land area north of the Alps (>47°N), south of Norway (<60°N) and west of western Poland (<20°E). This area is characterized by moderately high temperatures in summer (as opposed to the lower temperatures in northern Europe) and is not strongly affected by widespread shortage of soil water in summer (as opposed to the dry conditions in southern Europe and the Mediterranean). For this area we aim to address the following questions:

- (i) How robust is the super CC scaling of observed hourly precipitation extremes?
- (ii) What is the role of atmospheric moisture in the observed scaling?
- (iii) To what extent is the observed scaling influenced by statistical effects due to the temperature dependency of the frequency of occurrence of the different precipitation types?
- (iv) Can model predictions of future changes in precipitation extremes (including model spread) be understood from relations obtained from (simulations of) the present-day climate?

Finally, we want to emphasize that in this study we take a macro-physics point of view, and that the role of cloud micro-physics is not considered here.

2. Observations

2.1. Dependency on temperature

We repeated the analysis in Lenderink and van Meijgaard (2008) using, besides the data of De Bilt, three other data sources: (i) data from 1950 to 2005 measured at Ukkel (Belgium), (ii) data from 1981 to 2008 measured at Bern, Basel and Zurich (Switzerland), and (iii) data from 27 stations within The Netherlands from 1995 to August 2009 (see the supplementary data available at stacks.iop.org/ERL/5/025208/mmedia). For the latter two data sources, the time series of the different stations are pooled to obtain one data set.

The analysis consists of two steps. First, the hourly precipitation data is divided into bins of 2°C width based on the daily mean temperature (see the supplementary data available at stacks.iop.org/ERL/5/025208/mmedia). From the binned data the 90th, 99th and 99.9th percentiles of the distribution of wet events are computed. In addition to the percentiles computed from the raw data, the 99th and 99.9th percentile are also computed from a generalized Pareto distribution fitted to the upper 4% of the data. Uncertainty bands are computed using the bootstrap. In contrast to Lenderink and van Meijgaard (2008) we used

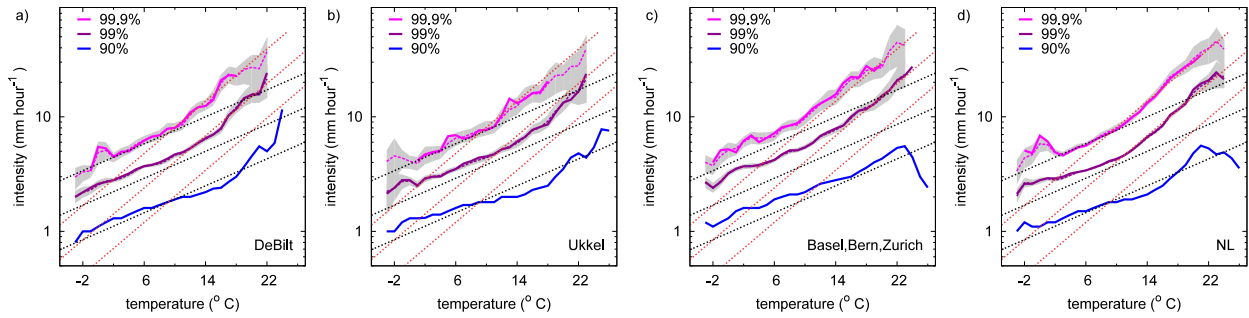


Figure 1. Dependencies of different extreme percentiles (90th–99.9th) of the distribution of observed hourly precipitation on temperature in four different data sets (De Bilt: source KNMI; Ukkel: source RMI; Bern, Basel and Zurich: source MeteoSwiss, NL: source KNMI). The gray shading plotted for the 99.9th and 99th percentiles denotes the 90% confidence intervals estimated by the bootstrap. Solid lines are percentiles computed from the raw data, whereas dotted lines are computed from the GPD fit. Exponential relations given by a 7% and a 14% increase per degree are given by the black and red stippled lines, respectively. Note the logarithmic y-axis.

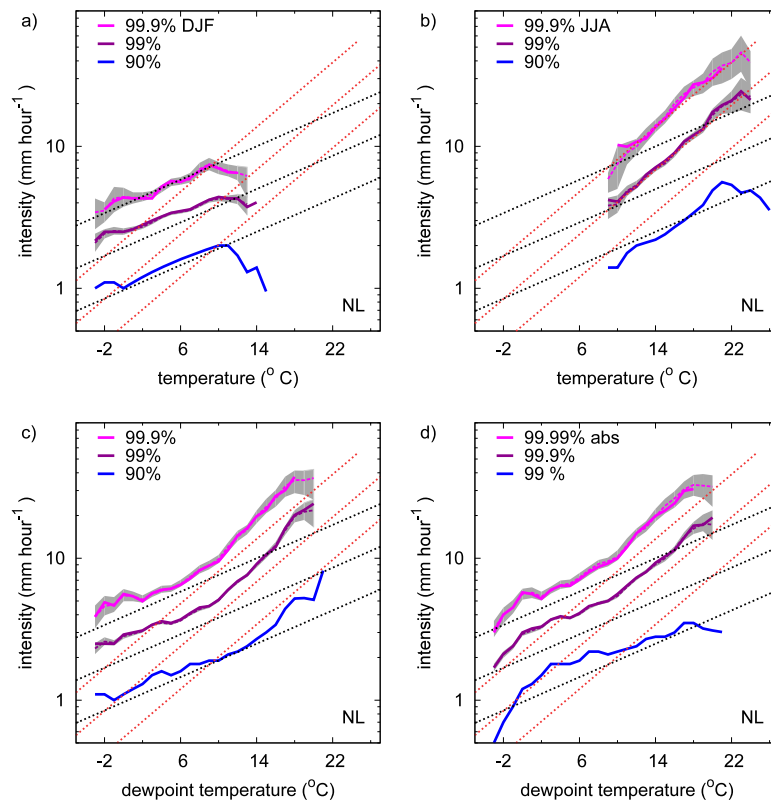


Figure 2. Dependency of hourly precipitation extremes on temperature computed from data of (a) the winter season DJF and (b) the summer season JJA. Dependency of hourly precipitation extremes on the dew point temperature, with the percentiles derived for only wet events (c) and for all events (d). Lines and shading are similar to figure 1.

overlapping bins, using steps of 1° . This is done because the arbitrary choice of putting the bin boundaries at even or at uneven numbers has a small, yet non-negligible, influence on the plots.

The 99th and 99.9th percentiles computed from all different data sets display a similar temperature dependency to the observations at De Bilt (figure 1). Above 10°C a temperature dependency of hourly precipitation extremes close to or slightly below two times the CC relation is found.

The data from 27 stations in The Netherlands over the last 15 years, in total approximately 400 years of hourly precipitation data, displays the most well defined behavior,

with a temperature dependency very close to two times the CC relation (figure 1(d)). We will therefore continue with this data set (in following labeled NL).

Results for the winter period (DJF) show a dependency of the hourly precipitation extremes, mostly along the CC line, except for temperatures above 10°C (figure 2(a)). The summer period (JJA) is characterized by a double CC scaling for the whole temperature range between 10 and 22°C (figure 2(b)).

2.2. Dependency on moisture

As a measure of atmospheric moisture we use the 2 m dew point temperature T_d , which is defined as the temperature to

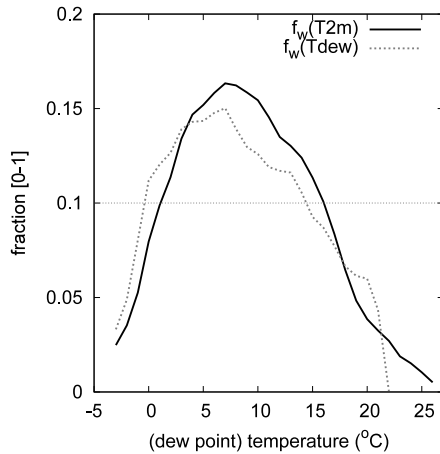


Figure 3. Fraction of wet events (number of wet hours divided by total hours in each temperature bin) as a function of temperature and dew point temperature.

which an air parcel must be cooled (at constant pressure) to reach saturation. A 1° increase in dew point temperature is equivalent to an increase of $\sim 7\%$ in moisture content. A constant relative humidity is approximately equivalent to a constant dew point depression, which is defined by $T - T_d$.

Using the dew point temperature instead of the temperature, similar results are obtained (figure 2(c)). The dependency of hourly precipitation extremes on the dew point temperature even slightly exceeds the double CC relation (approximately 17% per degree). Computing the percentiles from all hours in each bin, instead of only the wet ones, a similar scaling is obtained (figure 2(d)). Note the difference in the plotted percentiles to account for the average fraction of wet hours of $\sim 10\%$ (see also below). The insensitivity of the dew-point-temperature-based scaling to whether the percentiles are computed from all hours or only from wet hours is interesting for the following reasons.

First, the robust scaling obtained with the dew point temperature contrasts with the scaling obtained using the temperature. In the latter case we have to select only the wet hours to obtain a robust scaling (see the supplementary data available at stacks.iop.org/ERL/5/025208/mmedia). In the case of the dew point temperature this selection has become almost irrelevant. This supports the hypothesis that it is actually the increase in atmospheric moisture that determines the strong increase of the intensity of precipitation extremes with temperature.

Second, it rules out the possibility that the super CC scaling is due to a change in the *relative* frequency of convective versus large-scale events with temperature, as proposed by Haerter and Berg (2009). A decrease in the number of (less intense) large-scale precipitation events with temperature, while the number and intensity of the convective events remain unchanged, leads to an increase in the extreme percentiles of the distribution of wet events—the same extreme event becomes less rare in a smaller population of rainfall events. Such a statistical effect cannot occur when the percentiles are computed from all hours (or days). The possibility remains that an increase in the

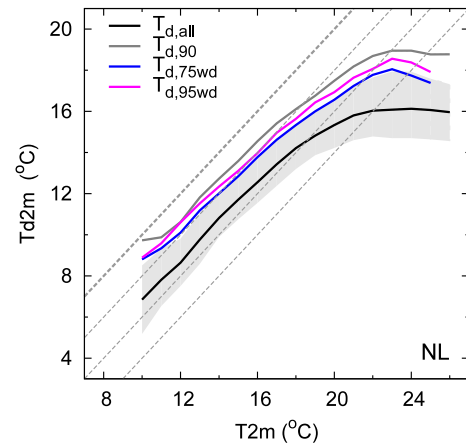


Figure 4. Relation between temperature and dew point temperature. Shown are the mean dew point temperature (with in shading the 25–75% interquartile range), the 90th percentile $T_{d,90}$ and the conditional mean dew point temperature for events with precipitation amounts greater than the 75th ($T_{d,75wd}$, blue) and 95th percentile ($T_{d,95wd}$, magenta) of the distribution of precipitation. Also shown are the saturation curve, where T_d equals T (thick stippled gray line), and lines with a constant dew point depression, $T - T_d$ (thin stippled gray lines).

frequency of *convective* events with temperature causes the enhanced temperature dependency, but this is not very likely since the fraction of wet events decreases with temperature (figure 3).

At approximately 15°C the fraction of wet events f_w is 10% (figure 3). At that temperature the 99.9th percentile of the distribution of the wet events ($P_{99.9w}$) equals the 99.99th percentile of the distribution of all events ($P_{99.99}$). Differences between the scaling of $P_{99.9w}$ and $P_{99.99}$ (and other similar pairs) relate to the deviations of f_w from 0.1. These deviations are shown to be important for the scaling of the lower percentiles, but barely affect the scaling for the highest ones (figures 2(c) and (d)). We note that this low sensitivity to frequency changes can be understood from the approximately exponential decay of the tail of the distribution of precipitation, for which the probability of exceedance is given by $\text{Pr}(X > p) = \exp(-\lambda p)$.

Figure 4 further explores the relation between temperature and dew point temperature, and their connection with precipitation extremes in summer. From all events within each temperature bin, we computed the mean dew point temperature ($T_{d,all}$), the 90th percentile of the dew point temperature ($T_{d,90}$), and the conditional mean dew point temperature belonging to the days with daily precipitation greater than the 75th percentile and 95th percentile (of wet days) of precipitation in that bin ($T_{d,75wd}$ and $T_{d,95wd}$, respectively). Here we used daily data in order to be able to compare to the model results in section 3 (for which we do not have hourly data), but we note that hourly data give almost identical results.

Up to 18°C the dependency of the mean dew point temperature $T_{d,all}$ on temperature is approximately 0.9. Thus, a 1° temperature rise implies a 6–7% increase in atmospheric moisture. The slope of $T_{d,all}$ starts to decrease above 18°C , and $T_{d,all}$ levels off for temperatures above 22°C , which is

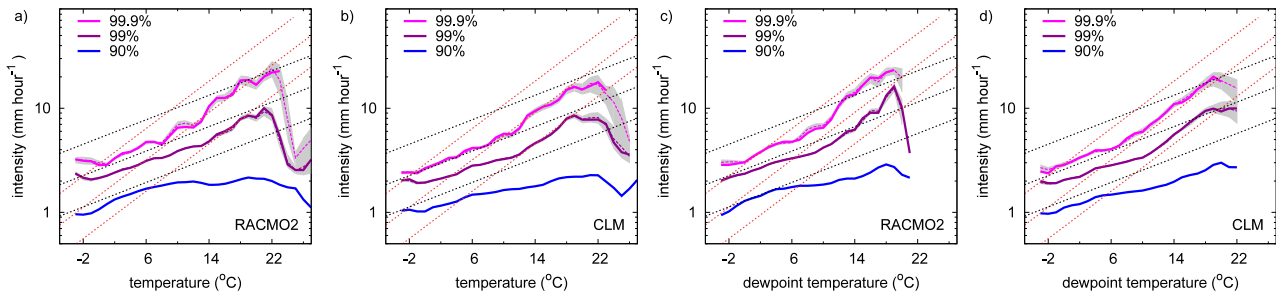


Figure 5. Scaling of the maximum hourly precipitation intensity with temperature and dew point temperature obtained from the output for NL in RACMO2 and CLM (lines and shading similar as figure 1).

likely caused by drying out of the soil and/or the atmospheric circulation changes. The three (conditional) samples $T_{d,90}$, $T_{d,75wd}$ and $T_{d,95wd}$, follow curves that are approximately parallel to $T_{d,all}$, but with positive offsets of 2° – 4° .

In the temperature range from 10 to 18°C the slope of dew point temperature on days with heavy precipitation, $T_{d,75wd}$ and $T_{d,95wd}$, is between 0.8 and 0.9. Thus each degree temperature rise implies a 0.8° – 0.9° rise in dew point temperature on heavy rain days. This is consistent with the 20% higher dependency of the hourly precipitation extremes on the dew point temperature compared to temperature (figures 1(d) and 2(c)). The other data sources also show larger dependencies on the dew point temperature.

We note that each degree rise in dew point temperature implies a 1° rise in the temperature on days with heavy rain. This is not shown, but is obtained by repeating the analysis of figure 4 with T and T_d interchanged; that is, binning on T_d . Therefore, if the increase in precipitation intensity was caused by temperature, then one would expect the same dependency on temperature as on the dew point temperature.

3. Models

3.1. Analysis of model characteristics for the present-day climate

Within the EU funded project ENSEMBLES (Hewitt and Griggs 2004), a large number of regional climate simulations for the present day and future climate have been performed at 25 km grid spacing. Details of the experimental setup and the models can be found in RT3 (2009). An analysis of the results of this ensemble is in progress. Here we show results of two regional models, RACMO2 run at KNMI and a version of CLM run at ETHZ. These models are chosen because they are representative of the upper and lower bounds in the change of precipitation extremes (hourly as well as daily) within the model ensemble.

In the ENSEMBLES data base only the daily maximum of the hourly precipitation intensity is stored. In the observations, the scaling of the maximum hourly intensity and of the hourly intensity display similar behavior, however with less noise for the latter (see also Lenderink and van Meijgaard 2008). The results in section 2 have therefore been based on the hourly intensity, whereas here we use the daily maximum of the hourly intensity. We used output of 64 grid points

within the Netherlands (NL) from the period 1961 to 2000 in simulations of CLM and RACMO2 driven by the ERA40 re-analysis (Uppala *et al* 2005).

For the 99.9th percentile both models reproduce the double CC scaling of hourly precipitation extremes above 10°C reasonably well (figure 5). There are subtle differences between the two models for temperatures near 20°C . RACMO2 appears to be able to retain a stronger temperature dependency, whereas the CLM appears to level off at a somewhat lower temperature. Analysis of the dependency of the hourly precipitation extremes on the dew point temperature does not reveal large differences between the two models either, and both models capture the observed relation between moisture and precipitation intensity.

The strong reduction in modeled precipitation intensity for temperatures above 22°C (figure 5) is likely due to model errors. But we note that for temperatures above $\sim 22^{\circ}\text{C}$ rather anomalous atmospheric conditions (e.g. with severe soil drying and/or strong high pressure systems) could also suppress the occurrence and intensity of precipitation extremes. Moreover, in the observations there are hints of a fall off in intensity (increase) in the highest temperature range.

Model differences, and discrepancies between the models and the observations, are more apparent for the relations between temperature, dew point temperature and extreme precipitation shown in figure 6. Results are shown for The Netherlands (NL) and Switzerland (CH) (the latter uses observations of Bern, Basel and Zurich and the model results from eight grid points north of the Alps).

For CLM the average dew point temperature on days with heavy precipitation are too high for temperatures above 20°C . In fact, $T_{d,75wd}$ and $T_{d,95wd}$ follow almost linear relationships, with a slope close to one and close to the saturation curve. Thus, it appears that CLM is only able to produce extreme events when the atmosphere at 2 m is close to saturation. Conversely, the slope of the curve of the mean dew point temperature $T_{d,all}$ is too small, and the divergence between $T_{d,all}$ and $T_{d,95wd}$ (or $T_{d,75wd}$) potentially reduces the probability of heavy rain at high temperatures. However, the upper 10% of the distribution of dew point temperature remains relatively close to saturation ($T_{d,90}$ in figure 6) and extremes therefore remain possible at high temperatures. Finally, we note that RACMO2 shows similar deficiencies, yet generally to a (much) smaller degree.

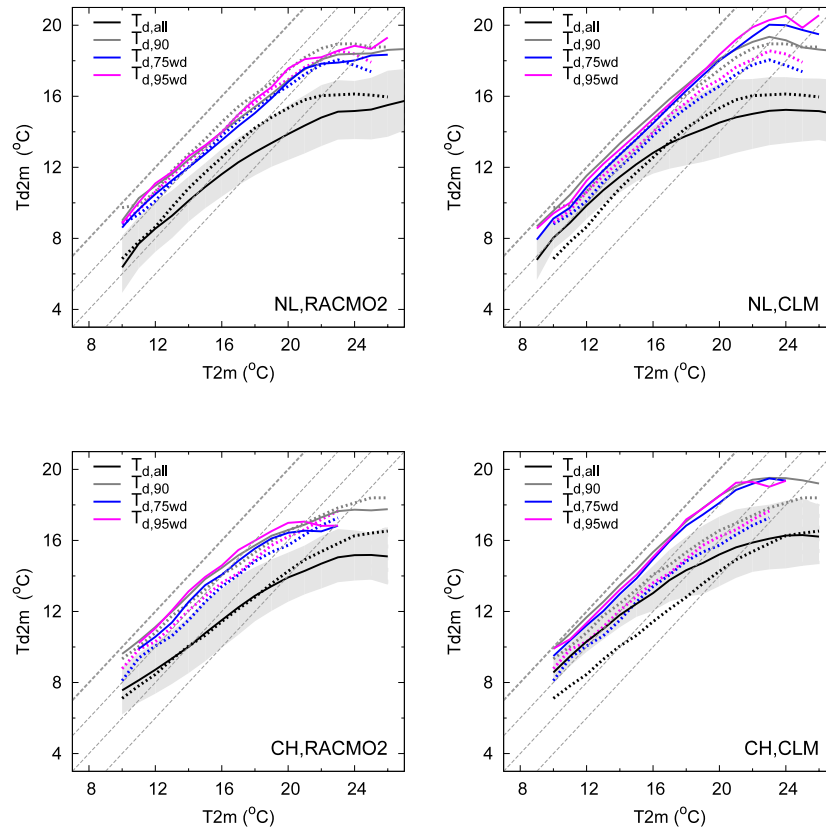


Figure 6. As figure 4, but now for the model results (solid lines) in comparison with the observations (stippled lines) for The Netherlands (NL) and the northern part of Switzerland (CH).

The relation between the relative humidity (measured by the dew point depression) and precipitation extremes is further explored by computing the cumulative frequency distribution (CFD) of the dew point depression $T - T_d$ for all days in summer (D_a) together with the conditional distributions for days with heavy precipitation exceeding the 90th and 99th (absolute) percentile (D_{p90} and D_{p99} , respectively). With this measure we do not focus as before on the distributions conditioned on temperature, but on the distributions of all events within the summer season.

The difference between the distribution of all days, D_a , and the distributions for days with heavy precipitation, D_{p90} and D_{p99} , provides a measure of the dependency of precipitation extremes on relative humidity. When the distributions of D_a , D_{p90} and D_{p99} coincide *relative* humidity is irrelevant. On the other hand, if D_{p90} and D_{p99} are close to zero, extreme events are restricted to days with high relative humidity.

Figure 7 confirms that in CLM extreme precipitation events in summer primarily occur on days with very high relative humidity (low $T - T_d$). However, the number of days with high relative humidity is also overestimated considerably, which compensates for the apparent high sensitivity of the model to (decreases in) relative humidity. In particular for Switzerland (CH) the difference between CLM and the observations is very large. Results from RACMO2 are generally (much) closer to the observations. However, we note

that for the dryer conditions in CH, RACMO2 also appears too sensitive to relative humidity.

3.2. Climate change signal

Figure 8 shows the change in the 99.9th percentile of maximum hourly precipitation in summer between 2071–2100 and 1971–2000. Three different simulations with the above regional models (using identical model configurations as the ERA40 integrations) are shown: RACMO2 driven by boundary conditions from the global climate models ECHAM5 (hereafter RACMO2-E) and MIROC3.2 (K-1 model developers 2004) (hereafter RACMO2-M), and CLM driven by boundary conditions from HadCM3 (hereafter CLM-H). The global models simulations have been performed as contributions to the 4th IPCC Assessment Report, and use the A1b emission scenario.

Despite that RACMO2-E and CLM-H are forced by different global models, they share a rather similar climate change signal for temperature (figure 8) and atmospheric circulation (with a change towards a dryer circulation in western Europe due to an increase in pressure over the British Isles and a decrease over the Mediterranean). The simulations, however, display rather large differences in the climate change signal of hourly precipitation extremes. Whereas the climate change signal in RACMO2-E is approximately 30–40% for western Europe, in CLM-H it is only half of this. With the weaker circulation change (van Ulden and van Oldenborgh

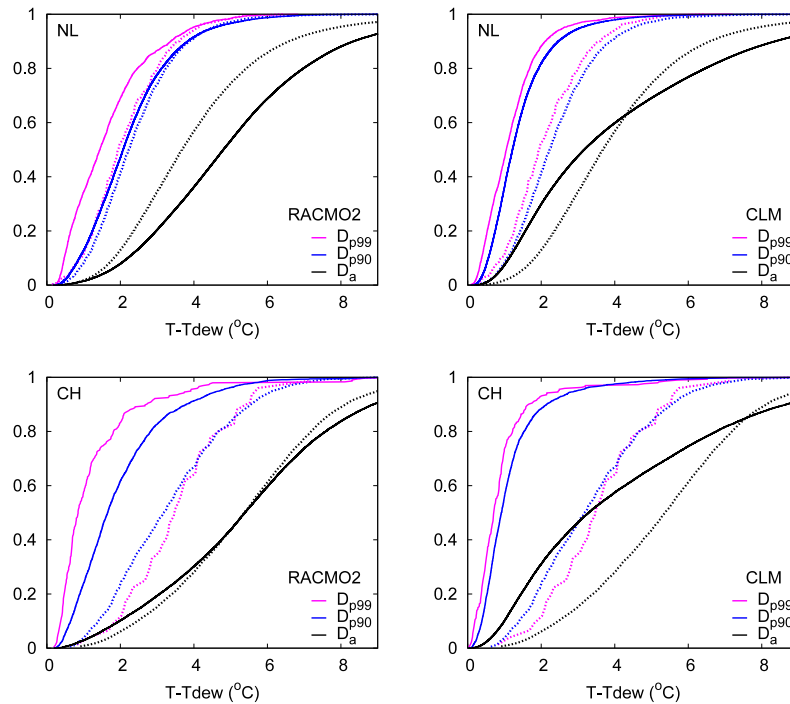


Figure 7. Cumulative distribution of the dew point depression, $T - T_d$, for all days and days with precipitation amounts exceeding the 90 and 99th percentile of the distribution of precipitation events. Shown are results for summer for The Netherlands (NL) and the northern part of Switzerland (CH) for the observations (stippled lines) and the models (solid lines).

2006) and the larger temperature increase of 3.5–5.0 °C in the MIROC forced integration, the predicted increase in hourly precipitation extremes in RACMO2 even amounts up to +60–80% in large parts of Europe. Both RACMO2 simulations display an increase of approximately +11–14% per degree temperature rise for western Europe, exceeding the CC relation by a factor 1.5–2.

The difference in climate change signal in hourly precipitation extremes between CLM and RACMO2 cannot be explained by differences in the models' response to increases in moisture, since both models display a similar dependency of precipitation extremes on moisture (see figures 5(c) and (d)). Instead, the difference appears primarily related to the changes in relative humidity. The distribution of the dew point depression (relative humidity) changes significantly from the control to the future period. For CLM (RACMO2) the decrease in the fraction of days with $T - T_d < 2$ is –15% (–5%) in NL and –45% (–40%) in CH. However, the cumulative frequency distributions of the dew point depression conditioned on days with heavy rain, D_{p90} and D_{p99} , hardly changes from the control to the future time period. In fact, these are very close to those derived from the ERA40 driven simulation (figure 7). Therefore, the aforementioned strong sensitivity of CLM to decreases in relative humidity could well explain the low increase in hourly precipitation extremes.

4. Conclusions and discussion

We investigated the relations between hourly precipitation extremes, temperature and moisture in observations and regional climate model results, thus extending on the work by

Lenderink and van Meijgaard (2008). We found that for the observations:

- (i) The temperature dependency of hourly precipitation extremes of approximately two times the CC relation appears robust. It is found in four independent data sources in western Europe.
- (ii) This scaling appears to be primarily linked to the increase in atmospheric moisture (or equivalently, the dew point temperature) with temperature. Two findings support this. First, the dependency of the hourly precipitation extremes on the dew point temperature reveals a more robust scaling behavior for the different percentiles and, second, this dependency is (slightly) stronger (compared to the temperature dependency) in all data sets.
- (iii) The statistical influence on the scaling of frequency changes of different precipitation types with temperature is likely small. The primary reason for the super CC scaling appears to relate to the physics of the convective events themselves.

Two climate scenario integrations with the regional model RACMO2 show an increase of hourly precipitation extremes that considerably exceeds the prediction from the CC relation. Results from a different regional model, CLM, show a much lower response, which appears to be caused by the strong sensitivity of this model to a reduction in relative humidity.

It is interesting that the version of CLM discussed here employs a closure of the convection scheme based on moisture convergence. Hohenegger *et al* (2009) showed that this closure led to a strong positive feedback with soil drying leading to

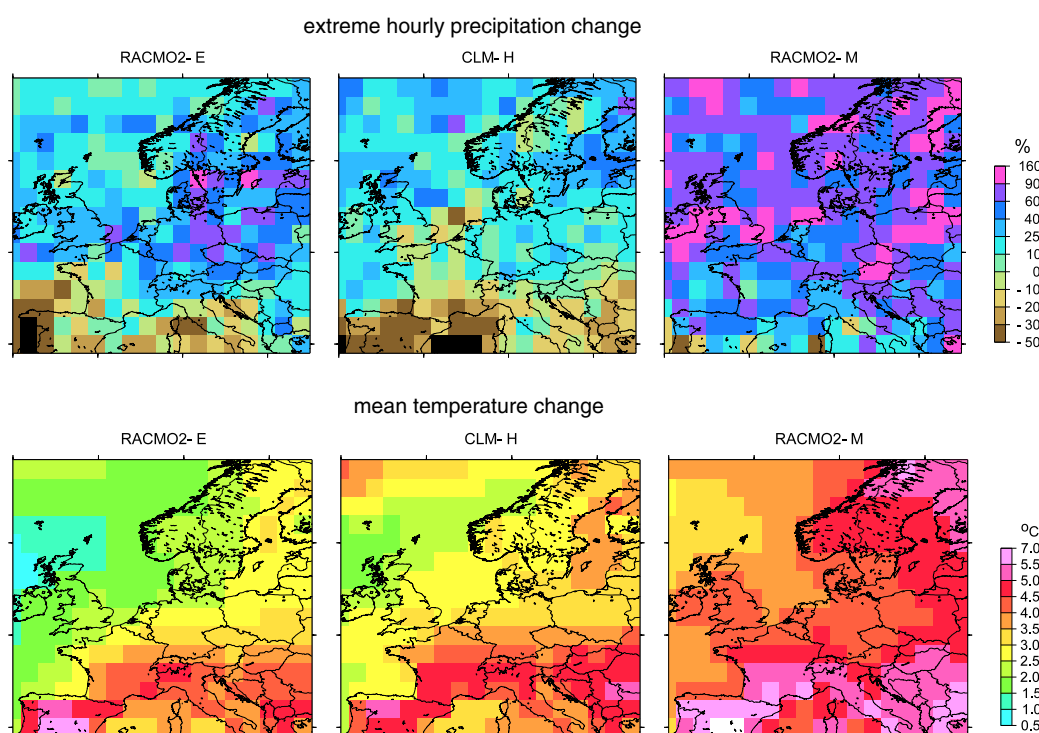


Figure 8. Climate response in summer (2071–2100 with respect to 1971–2000) in three different regional climate integrations. Shown are the relative change in the 99.9th percentile of maximum hourly precipitation (upper panels) and the mean temperature response (lower panels). The data is pooled in boxes of $2^\circ \times 2^\circ$ longitude–latitude in order to improve the statistics for precipitation extremes.

a reduction in precipitation, opposing the results of a non-hydrostatic version of the model that gave rise to a negative soil moisture feedback.

From the results of the two regional climate models the following conceptual picture emerges. Two main factors play an important role in determining changes of hourly precipitation extremes in the climate scenario integrations: first, the response of the model to absolute humidity (dew point temperature), and second the response to relative humidity (dew point depression). Scaling relations derived from simulations of the present-day climate show that the two models considered here capture the response to absolute humidity reasonably well, although they tend to underestimate the dependency for the lower percentiles and in the high (dew point) temperature range (figure 5). The response to relative humidity is not so well represented, in particular in CLM (figures 6 and 7).

The characteristics of the two models considered here do not necessarily apply in general. However, we think that by employing the conceptual framework and the scaling relations presented in this paper a considerable part of the uncertainty in predictions of future (hourly) precipitation extremes can be understood.

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