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Downturn in scaling of UK extreme rainfall with temperature for future hottest days

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Extreme daily precipitation is thought to increase with temperature at a rate of 6.5% per K according to the Clausius-Clapeyron relationship between temperature and saturation vapour pressure [1]. A wide range of scaling relationships has been observed globally for extreme daily and hourly precipitation, with evidence of scaling above 6.5% per K for sub-daily extreme precipitation in some regions [2, 3, 4]. Only high-resolution climate models can simulate this scaling relationship [5]. Here we examine the scaling of hourly extreme precipitation intensities in a future climate, using experiments with a model for the southern UK with kilometre-scale resolution [6]. Our model simulates the present-day scaling relationship at 6.5% per K, in agreement with observations. The simulated overall future increase in extreme precipitation follows the same relationship. However, UK extreme precipitation intensities decline at temperatures above about 22℃ – a temperature range that is not well sampled in the present-day integration – as a result of a more frequent occurrence of anticyclonic weather systems. Anticyclones produce more days with strong daytime

heating, but are not favourable to the development of deep intense convective storms. We conclude that future extreme hourly precipitation intensities cannot simply be extrapolated from present-day temperature scaling, and demonstrate the pitfalls of using regional surface temperature as a scaling variable.

The Clausius-Clapeyron (C-C; see Supplementary Information) relation describes the rate of change of saturated vapour pressure with temperature. For water vapour, this is $\approx$ 6–7% per K near the Earth surface. Assuming constant relative humidity (RH), this sets a scale for change in extreme daily precipitation intensities [1]. The observed relationship globally between annual maximum daily precipitation and global mean temperature is about 5.9–7.7% per K, consistent with the C-C relation [7]. As a consequence, temperature changes become a potential predictor for extreme daily precipitation change. Here we assume a C-C scaling of $\gamma \approx 6.5\%$ per K. Changes in weather patterns, moisture availability [8] and thermodynamic stability [9, 10, 11] may lead to scalings that depart substantially from γ.

Most climate models require convective parameterisations (CP) to represent convection, but CP is not designed to represent individual storms and their precipitation [12]. Even moderately high-resolution model versions are unable to replicate observed scaling at temperatures above 20℃ [2]. Since warmer temperatures are typically associated with a greater proportion of convective compared to stratiform precipitation [13], the inadequacies of the model CP become increasingly important as temperature rises; such models are known to show inconsistent temperature and humidity changes [10, 14]. However, a "convective-permitting" (no-CP) high-resolution regional climate model (RCM) is able to replicate observed super-C-C scaling for hourly precipitation over the Alps for a present-day simulation [5].

The C-C hypothesis has been tested in many observational and modelling studies. General circulation model (GCM) projections produce sub-C-C scaling for daily mean precipitation; only daily extreme precipitation scalings are consistent with the C-C hypothesis [10, 11, 15, 16]. At regional and local scales, observed scalings of extreme hourly precipitation can exceed γ ("super-scaling") [2, 3]. Some studies have found super-scaling only for sub-hourly totals [4]. A decline in extreme hourly precipitation intensities at temperature exceeding $\approx$ 23℃ is found in some subtropical and tropical regions [4, 17]. These previous studies have examined surface temperature and humidity as scaling variables because they are well observed.

However, surface conditions may be strongly modulated by solar heating, and do not necessarily provide useful information about the thermodynamic and dynamic properties of the storms. Studies looking at tropospheric vertical velocity and humidity have shown a positive scaling with extreme daily precipitation [10, 11]. For Europe, projections of summer daily and hourly extreme precipitation projections show both increases and decreases for different parts of Europe [18, 19].

UK C-C scaling of extreme precipitation intensities has been examined in observations showing super-scaling for anti-cyclonic conditions in summer [20], but the same has not been examined for model simulations. The UK Met Office (UKMO) has recently completed continuous high-resolution (convective-permitting 1.5-km southern UK and CP-enabled 12-km European) present-day (G-P) and future-climate (G-F) RCM simulations [6, 21] with large-scale conditions provided by HadGEM3 GCM integrations [22]. Despite larger mean precipitation biases, the 1.5-km model has significantly improved physical realism in simulating extreme precipitation intensities compared to coarser-resolution model simulations [21], and shows future intensification of both summer and winter extremes [6]. Here, we examine its ability to represent observed C-C scaling relationships for the southern UK from radar observations [23], and the extent to which these scalings may apply with future climate change.

Relationships between hourly extreme precipitation intensities and daily temperatures for summer (Jun-Jul-Aug; JJA) and winter (Dec-Jan-Feb; DJF) over southern UK land points are shown in Figure 1 for radar, present-day, and future-climate integrations. For summer, the 99^{th} percentile precipitation increases in the radar observations at γ — the same rate as for UK gauge observations [20], and this scaling is only captured by the 1.5-km model (Figure 1b). The 12-km model is unable to capture the observed scaling, producing relatively constant intensities with temperature increases. However, both the 12-km and 1.5-km models are able to capture the observed scaling in winter (Figure 1d and e), which is less than γ. Neither observed nor present-day climate integrations show a clear relationship between temperature and extreme daily precipitation (see Supplementary Figure 1).

Unlike other analyses of observed data [2] and model simulations [5], there is little evidence of "super-scaling" in either radar observations or model simulations for southern UK as an average, but local scaling variations do exist (see Supplementary Figure 2). We do not account for scaling variations under different weather regimes and conditions [13, 20]. However, the fact that

the 1.5-km model can capture the observed scaling gives us more confidence in its use to estimate future scalings.

Scaling relationships for future simulations are shown in the right panels of Figure 1. A ≈ 5℃ warming and an average 30% increase for the 99$^{\text{th}}$ percentile of summer precipitation intensities for the 1.5-km model across the full temperature range is consistent with the C-C hypothesis. The Alps-region analysis [24] — examining the change in 99.9th percentile precipitation extreme between control and future simulations — provides a smaller $3-6$% change. However, unlike our "wet"-value-only analysis, they use both "wet" and "dry" values to calculate percentiles; therefore, the changes reflect a combination of changes in both frequency and intensity. For the future climate, we find both the 1.5-km and 12-km models simulate declines in hourly extreme precipitation intensities at temperatures in excess of ≈ 22℃ over the southern UK. Such declines have been observed globally for daily intensities, whereas hourly precipitation declines are mostly found in the subtropics and tropics [4, 17]. The UK declines occur at the higher end of the future temperature range, which is not sampled in the present-day climate simulations. Over the overlapping temperature range, the scaling of the 1.5-km future-climate simulation is comparable with the present-day simulation.

A warmer future climate may indeed be more favourable to extreme precipitation, but our simulations indicate that this does not extend to the warmest summer days as seen in observational and modelling studies [4, 14]. Two explanations have been put forward for the declines in the observed scaling relation at high temperatures: i) the decrease in extreme precipitation event duration [4]; ii) the suppression of RH [17]. For the former, sub-hourly extreme precipitation intensities continue to intensify with temperature increases [4]. For the latter, a decrease in RH has been found over land since the late 1990s [25], and climate models project reductions in extra-tropical troposphere RH [26]. Here, we examine the role of RH.

Precipitation totals are ultimately constrained by moisture availability, which may be much less than the temperature-dependent maximum. Here we turn to the near-surface dew point T_d [14], and wet-bulb potential temperature θ_w as humidity measures (see Methods). In Figure 2, we plot the variation of the 99$^{\text{th}}$ percentile of 1-hour summer precipitation ($P_{\text{1-hr}}$) with near-surface T_d and 850-hPa θ_w for both the 1.5-km and 12-km models. Unlike Figure 1, the results here are obtained using a "non-local" method — whereby the highest near-surface T_d and 850-hPa θ_w is found near the pre-

cipitating grid point (see Methods).

Both models sample a similar range of T_d and 850-hPa θ_w. All but the future 12-km simulation show increases of extreme hourly precipitation intensities with T_d and θ_w in accordance with the γ scaling. For near-surface T_d, the scaling here is less than previous work [14]; a full discussion is in the Supplementary Information, including justification for the "non-local" methodology and the usage of 850-hPa θ_w. The 1.5-km future simulation shows a general extension of the γ scaling for the present-day climate into a higher T_d and θ_w range. This extension is similar for the 12-km model, however at the highest T_d ($18 + ℃$), extreme hourly precipitation intensities decline with T_d.

The scaling decline for the 12-km RCM is not seen with 850-hPa θ_w, but is evident with 850-hPa temperatures and local surface dew points (Supplementary Figure 3 and 4). The differences are likely related to the use of CP by the 12-km model [10], but a detailed examination is beyond the scope of this paper. Nevertheless, results indicate model-simulated scalings with upper-air humidities and temperatures are generally more consistent than the scalings with surface humidities and temperatures. For the 1.5-km RCM, the hot day scaling declines in the future simulations are associated with reduced-RH days and thus a higher saturation barrier. Model-simulated relationships between precipitation, θ_w and surface air temperature are explored further in Supplementary Figure 5.

Since the large-scale flow influences local weather conditions, we use mean sea-level pressure (MSLP) as a diagnostic for examining that influence. In Figure 3, we plot the differences in model-simulated MSLP between wet days with high surface air temperature and wet days with high θ_w. Compared to wet-high θ_w days, wet-hot days over the southern UK are associated with higher MSLP anomalies. The positive MSLP anomalies are indicative of less fronts [27], more anticyclonic conditions with drier and hotter weather, reducing RH, and inhibiting deep convection. The MSLP anomalies also suggest weaker low-level westerlies over much of the UK, and imply reduced moisture from the warm Atlantic (Supplementary Figure 6). Changes in MSLP between the future- and present-day integrations reveal a shift to more anticyclonic conditions over the UK in the future (Supplementary Figure 7). This change comes from the driving GCM, and appears to be the driver of the downturn in scaling at high temperatures in the 1.5-km RCM through its effect on reducing absolute and relative humidity. Such anticyclonic shifts are found in other GCM projections for Europe, suggesting this downturn

may be a general feature of projections for this region and potentially other regions with similar shifts (see Supplementary Information).

In conclusion, we have shown that the 1.5-km convective-permitting RCM can simulate UK observed C-C scaling of hourly extreme precipitation intensities. This is true in both summer and winter, unlike the lower-resolution 12-km RCM, which is only able to simulate the winter scaling. Despite an average intensification of extreme precipitation at γ, present-day hourly-precipitation scaling does not extend into higher temperatures ($22 + ℃$) sampled only by the future climate integration. In particular, for surface temperatures exceeding $\approx 22℃$, hourly extreme precipitation declines with increasing temperature. This breakdown in scaling has been observed around the globe for sub-daily and daily extreme precipitation. It, therefore, has implications for precipitation intensities globally as well as the UK.

We have demonstrated that 850-hPa θ_w is an improved predictor for changes in precipitation extremes as it can account for relative humidity changes, giving a more consistent scaling between future- and present-day climate integrations compared to surface temperatures and dew points. The C-C hypothesis assumes that moisture availability is temperature limited, but our simulations indicate that, at the regional scale, weather patterns provide important moderators of moisture availability.

In the case of the UK under global warming, the driving GCM simulates a shift to a more anticyclonic MSLP regime causing more very hot days and less convection. Similar changes may also be a factor for observed declines in scaling at lower latitudes [4, 17]. We conclude that although changes to extreme hourly precipitation are dominated by thermodynamic changes, dynamical changes can have important regional effects. Therefore, regional surface temperatures cannot be used alone to extrapolate changes in extreme precipitation intensity. Realistic large-scale circulation patterns and variability from GCMs will be crucial for good regional predictions.

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Author Contributions SCC co-wrote the manuscript, conducted data analysis and data visualisation. EJK co-wrote the manuscript, supplied the data, performed the necessary model simulations, and advised on data analysis methodologies and result interpretation. NMR co-wrote the manuscript, and advised on analysis methodologies and result interpretation. HJF advised on the analysis methodologies, discussed the results, co-wrote the manuscript, and commented on the manuscript. SB advised on observational aspects of the present study, discussed the results, and commented on the manuscript.

Competing Interests None

Methods

The UKMO high-resolution 1.5- and 12-km RCM simulations used in the present analysis have been fully described in other papers [6, 21]. The 1.5-km model simulations are centred over the southern United Kingdom [21] with an inner domain that roughly follows Supplementary Figure 2. Unlike the 12-km RCM, the 1.5-km model does not use convective parameterisation (CP), and is able to explicitly represent individual showers or storms on the model grid. The 1.5-km simulations are driven by 12-km European-domain simulations (also analysed here), which themselves are driven by 60-km HadGEM3 GCM simulations for the present- and future-climate. The future HadGEM3 simulation follows the IPCC Representative Concentration Pathway 8.5W/m^2 "business as usual" climate forcing scenario for the end of 21st century [28]. There is no ocean model coupling; future sea surface temperatures (SST) are superposition of another coupled model's SST projections and observed

SSTs between 1997 and 2009, in which the latter serves as the SST baseline that drives the present-climate GCM simulation [22]. The analysis domain is the same as in previous studies [21]. Only land points are examined. Results here focus on hourly precipitation; model-simulated sub-hourly precipitation was unavailable to test duration changes [4], and may be examined in future work.

For observed near-surface air temperature, we have used the National Climate Information Centre 5-km gridded daily temperature observations [UK5; 29]. Daily maximum (T_{max}) and minimum (T_{min}) surface air temperatures are available from 1960, but we only use the data that overlap with the radar period. The daily-averaged surface air temperatures (T_{avg}) are estimated by:

$$T_{\mathrm{avg}} = \frac{T_{\mathrm{max}} + T_{\mathrm{min}}}{2} \tag{1}$$

The UK5 daily values are taken between 09Z - 09Z. The RCM-simulated daily averaged near-surface air temperature is computed with the same formulation. However, UKMO models use a 00Z - 24Z period to determine daily minimums and maximums.

For gridded hourly precipitation, we have used the UKMO gridded quality - controlled gauge - calibrated Radarnet composite radar observations, which are available from 2003 [23]. Hence, our observational temperature-precipitation analysis only covers the 2003-2012 period. The same radar data have been previously used to examine UK precipitation extremes, and a discussion of the pros and cons for using radar for climate research can be found there and in other related papers [30, 31]. On average, radar tends to underestimate intensities of intense rainfall due to attenuation of the beam [23]. As in previous studies [2, 14, 17], we use the daily maximum of 1-hour total precipitation $P_{\mathrm{max,1\text{-}hr}}$ with a "wet" threshold of 0.1mm/hr. A minimum "wet" threshold is adopted as we wish to separate intensity and frequency changes [19]. It is the intensity which has been hypothesised to scale with saturation vapour pressure, whilst frequency is less clearly related [1]. To be consistent with the air temperature data, the daily window for $P_{\mathrm{max,1\text{-}hr}}$ is as for T_{max} and T_{min}. All data are first interpolated to a common 12-km grid prior to any analysis including the selection of $P_{\mathrm{max,1\text{-}hr}}$. Spatial averaging is used to interpolate precipitation.

For the C-C scaling analysis, the methodology used for Figure 1 is based

on the equal-sample-size binning method that is used in an Australian observational study [17], but modifications of the original methodology are applied for gridded data. Here $P_{\text{max,1-hr}}$ values at each grid point and its eight neighbouring grid points are binned according to their "local" (central grid point) T_{avg}. Each bin has a distribution of $P_{\text{max,1-hr}}$ values from which its percentiles are estimated. The temperature dependency of the percentiles can be then estimated. We require each T_{avg} bin to have approximately an equal number of data pairs.

The key differences with the Australian work are:

- We have used shorter datasets that are gridded. We conduct 3-by-3 adjacent-grid-point pooling to increase our sample size, and the binning is conducted with the pooled samples
- We use 20 bins instead of the 12 bins used in the Australian study

The overall scaling relationship for the southern UK is estimated by superposing all individual 3-by-3 grid-box relationships using hexagon scatter density plots, and fitting a LOcally-wEighted Scatterplot Smoothing (LOESS) non-parametric regression line [4, 32]. We stratify our data by seasons (summer and winter) to gain some separation between summer convective precipitation and winter large-scale precipitation [17, 33]. The above method diagnoses scaling changes with temperature variability within a stationary (present or future) climate regime, and differs with the Alps-region future-climate projection analysis [24] which focuses on average percentile changes between the future and present climate.

Surface dew point temperature T_d is a measure of surface specific humidity translated to temperature with the C-C relationship [14]. Wet-bulb potential temperature θ_w is a combined temperature-humidity measure that is unaffected by adiabatic motion or phase changes, and is commonly used in weather forecasting as an air mass indicator (see Supplementary Information). Model-simulated θ_w and surface T_d are available every 6 hours and day respectively. To estimate the scaling relationship with θ_w and surface dew point (T_d) in Figure 2, we adopt the following approach. For each 6-hourly period, we first select the maximum over-land 1-hr precipitation within our region of interest ($P_{\text{1-hr}}$; same 0.1mm/hr "wet" threshold applies). We then

determine the largest 850-hPa θ_w or T_d in a 7-by-7 box centred over the selected grid point. The surface air temperature under the precipitating grid box is recorded for Figure 3. The key difference with the method we have used for temperature is that there is only one precipitation-(θ_w-or-T_d) pair per 6-hour period as opposed to as many pairs as precipitating grid boxes per day. The goal is to associate a precipitation intensity with a θ_w or T_d that is representative of the air mass supplying the cloud. After all precipitation-(θ_w-or-T_d) pairs are collected, the pairs are binned according to their θ_w or T_d. Bins with fewer than 200 samples are eliminated. The 95% confidence intervals (CI) of the percentile are estimated by "bootstrapping" [34]:

1. Randomly draw bin samples with replacement to create a new bin with the same number of samples; this is completed 5000 times
2. 5000 new percentiles are estimated
3. The 95% CI is taken to be the 2.5% and 97.5% percentiles from the new 5000 estimates

The above method departs from the temperature-precipitation analysis to reduce double counting: each precipitating grid point — possibly from the same precipitating system — would find a common non-local θ_w maximum. Double counting would cause erroneous uncertainty underestimation and over-confidence of the presented results. However, the sample number for one value per 6 hours is substantially less than using all wet values per wet day if one assumes there are at least 4 wet land grid points per wet day. Sensitivity tests have shown that results do not change substantially at lower isobaric level (925-hPa), and θ_w box sizes are chosen by gradual increase of box size until results are stabilised. The non-local method is adopted as the existence of precipitation and clouds are likely to modify "local" θ_w and T_d. The non-local method aims to sample more representative values for θ_w and T_d that are being fed into the cloud.

Code availability. The R and IDL source code used in the analysis can be requested from the lead author, but installations of additional software packages may be required. R is available as free-and-open-source software, and IDL is a propriety product available from Exelis Visual Information Solutions [35, 36]. The model and observational data used are under Crown

copyright of the UK Met Office, and access may be requested from the Met Office.

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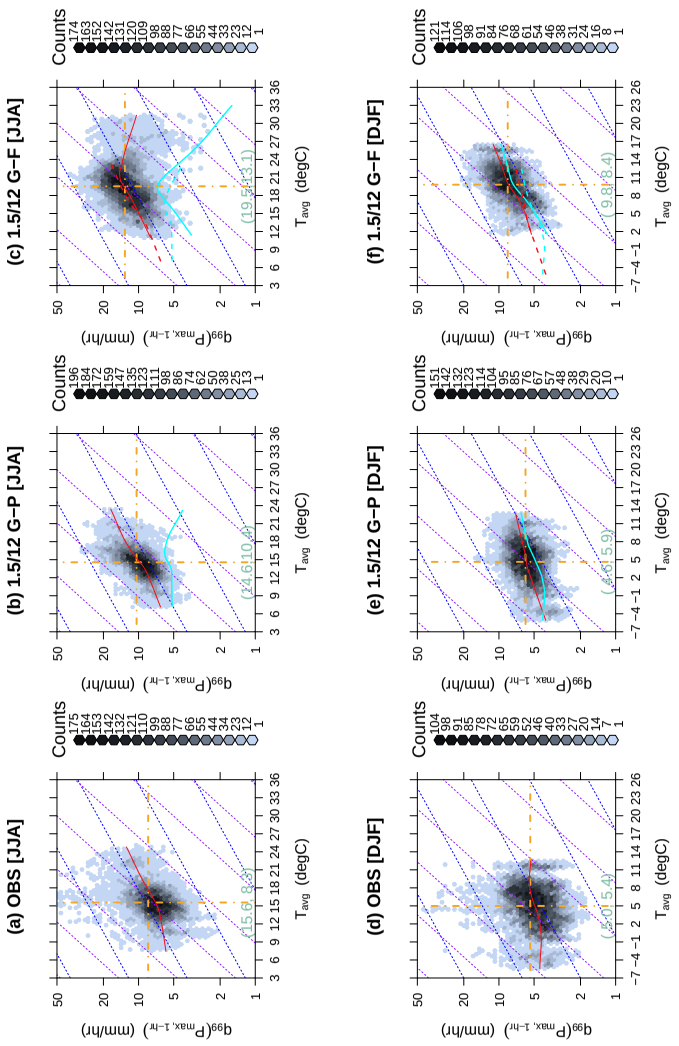

Figure 1: The relationship between surface air temperature and extreme 1-hr precipitation intensities over land. The observed (left), model-simulated present- (centre) and future-climate (right) JJA (top) and DJF (bottom) relationships between mean surface air temperature (T_{avg}) and the 99$^{\text{th}}$ percentile of daily maximum 1-hr precipitation ($P_{\text{max,1-hr}}$). Red lines indicate the relationship between the two variables for the observations and 1.5-km RCM, and the orange lines indicate the mean. The 12-km RCM relationships are indicated by cyan lines. Only "wet" values (daily maximum 1-hr intensity exceeding 0.1mm/hr; see Methods) are examined. The relationship for the present-climate simulation is shown in the right (future-climate) panel as a dotted line. Blue and purple dashes indicate 1× and 2× C-C scaling respectively.

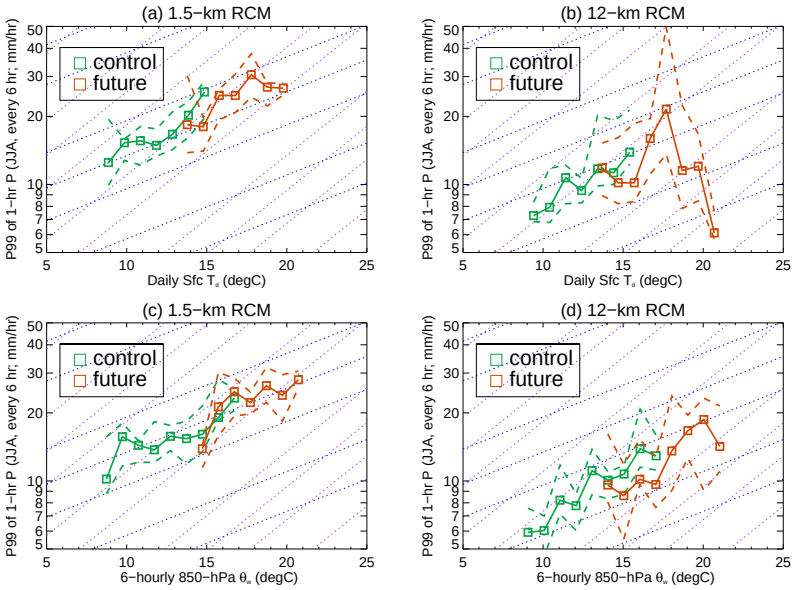

Figure 2: The summer near-surface dew point / wet-bulb potential temperature relationships with extreme 1-hr precipitation intensities. Near-surface dew point temperature T_d (upper panels) is the daily-averaged value, and wet-bulb potential temperature θ_w (lower panels) is the instantaneous value at 850 hPa measured every 6 hours. For both T_d and θ_w, the highest value within a 7x7 box centred around the precipitating grid point is selected. Coloured dashes indicate the 95% confidence interval of the relationship. The left and right panels are for 1.5- and 12-km RCM respectively. The black and purple dashes indicate 1× and 2× C-C scaling.

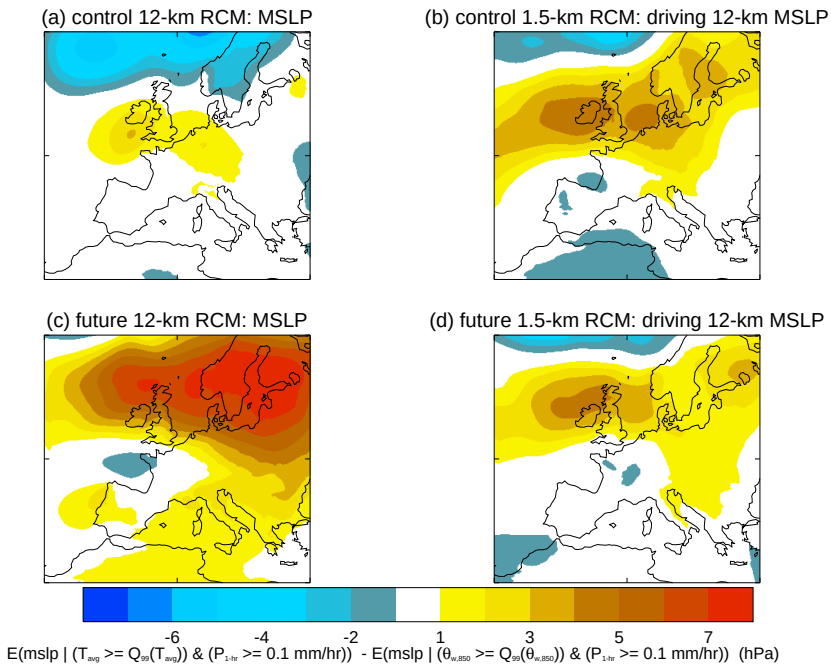

Figure 3: The mean sea-level pressure differences between wet hot days and wet hot humid days with high 850-hPa θ_w over the southern UK. These are defined as days from any month with precipitation $\geq$ 0.1mm/dy, and T_{avg} or 850-hPa θ_w exceeding the 99$^{\text{th}}$ percentile, respectively. Only days with at least one southern UK land grid point exceeding 0.1mm/dy precipitation are selected. For 850-hPa θ_w, we use the values from the non-local scaling analysis. However, for T_{avg}, we use the values at the precipitating grid box. For the 1.5-km RCM case, we use the MSLP that is prescribed by the driving 12-km RCM. Units are in hecto-pascals (hPa).