Clim Dyn (2007) 28:351–363
DOI 10.1007/s00382-006-0180-2

Testing the Clausius–Clapeyron constraint on changes in extreme precipitation under CO_2 warming

P. Pall · M. R. Allen · D. A. Stone

Received: 25 May 2005 / Accepted: 21 July 2006 / Published online: 30 August 2006

Abstract Increases in extreme precipitation greater than in the mean under increased greenhouse gases have been reported in many climate models both on global and regional scales. It has been proposed in a previous study that whereas global-mean precipitation change is primarily constrained by the global energy budget, the heaviest events can be expected when effectively all the moisture in a volume of air is precipitated out, suggesting the intensity of these events increases with availability of moisture, and significantly faster than the global mean. Thus under conditions of constant relative humidity one might expect the Clausius–Clapeyron relation to give a constraint on changes in the uppermost quantiles of precipitation distributions. This study examines if the phenomenon manifests on regional and seasonal scales also. Zonal analysis of daily precipitation in the HadCM3 model under a transient CO_2 forcing scenario shows increased extreme precipitation in the tropics accompanied by increased drying at lower percentiles. At mid- to high-latitudes there is increased precipitation over all percentiles. The greatest agreement with Clausius–Clapeyron predicted change occurs at mid-latitudes. This pattern is consistent with other climate model projections, and suggests that regions in which the nature of the ambient flows change little give the greatest agreement with Clausius–Clapeyron prediction. This is borne out by repeating the analyses at gridbox level and over season. Furthermore, it is found that Clausius–Clapeyron predicted change in extreme precipitation is a better predictor than directly using the change in mean precipitation, particularly between 60°N and 60°S. This could explain why extreme precipitation changes may be more detectable then mean changes.

P. Pall (✉) · M. R. Allen · D. A. Stone
Atmospheric, Oceanic and Planetary Physics,
Department of Physics, University of Oxford,
Clarendon Laboratory, Parks Road,
OX1 3PU Oxford, UK
e-mail: pall@atm.ox.ac.uk

1 Introduction

Increased precipitation intensity, albeit with certain regional variations, in a future climate with increased greenhouse gases was one of the earliest climate model results regarding precipitation extremes, and remains a consistent result with improved and more detailed models (Meehl et al. 2000b; Cubasch and Meehl 2001, and references therein). In fact it has been found in many climate models that increases in extreme precipitation are greater than the that of the mean over near-global domains (e.g Hennessy et al. 1997; Kharin and Zwiers 2000; Semenov and Bengtsson 2002; Watterson and Dix 2003; Hegerl et al. 2004; Wehner 2004; Emori et al. 2005). In regional studies also, models have exhibited larger increases in heavy or extreme precipitation amounts versus moderate or mean amounts (e.g. Wilby and Wigley 2002; Räisänen et al. 2004).

The observational record also suggests that disproportionate increases in extremes of precipitation may have already occurred in parts of the world, such as for the United States (Karl and Knight 1998), United Kingdom (Osborn et al. 2000) and Australia (Suppiah and Hennessy 1998); and see Easterling et al. (2000) and Folland and Karl (2001) for a review. Complementary to this, statistical modelling studies for precipitation akin to those of Groisman et al. (1999) highlight the large impact that even modest increases in the mean of precipitation distributions can have on the return periods of the extremes.

A recent Intergovernmental Panel on Climate Change (IPCC) workshop on extremes (IPCC 2002), recognized precipitation as one of the key climate variables of impact and highlighted that understanding the distribution of precipitation is key to understanding its extremes. It concurrently, however, identified weakness in the the IPCC Third Assessment Report (TAR) relating to extreme precipitation which were due to availability, quality and definition of data. The records suffer from spatio-temporal inhomogeneity, changes in instrumentation and practices, lack of digitization and barriers to international access to data (Groisman and Legates 1994; Easterling et al. 2000); and the period since the late 1970s whence satellites provide more extensive coverage (e.g. Xie and Arkin 1997; Adler et al. 2003) is arguably too short to draw conclusions regarding decadal-scale and above variations (Folland and Karl 2001). The comparatively low resolution of climate models is also an inhibiting factor in determining agreement with available observations (McAvaney 2001; Osborn and Hulme 1997; Kiktev et al. 2003). A further problem is that the timing of the diurnal cycle in model precipitation is generally between 2 and 4 h before that in nature, which is related to design of convection schemes and simulation of cloud cover (Dai et al. 1999; Dai and Trenberth 2004; Betts and Jakob 2002; and see Trenberth and Dai 2003).

Allen and Ingram (2002) argue that the only objective probabilistic forecast of global temperature change is provided by the constraint of consistency with actual climate observations, and demonstrate how the spread of models in the CMIP-2 experiment (Meehl et al. 2000a) may be substantially underestimating the true spread on future temperature forecasts. This conclusion is confirmed by the findings of Collins et al. (2006) whose perturbed physics ensemble, despite not representing uncertainty in ocean properties, fails to overlap the CMIP-2 range. Moreover, Allen and Ingram (2002) state that the problem could be expected to be worse for variables like precipitation which are not so well constrained by the available data, and note the problem of observed changes being dominated by sulphate aerosol forcing (natural and anthropogenic) where it occurs.

With regard to all this, Cubasch and Meehl (2001) conclude that ‘‘there is more internal variability and model differences and less common signal indicating lower reliability in the changes of precipitation compared to temperature’’. McAvaney (2001) further state that the lack of consistent methodologies used in analyses of extreme events prevents a ready intercomparison of results between models and that future IPCC assessments would be greatly assisted if common approaches were adopted.

In searching, then, for a common approach to determining the change in extreme precipitation and a consistent methodology one may tackle the problem by searching for constraints on the climate system, and particularly on the hydrological cycle, that could simplify the problem.

In quantifying the expected increase in extreme precipitation Allen and Ingram (2002) follow Trenberth (1999b) in arguing that under global warming, and under the constraint of constant relative humidity (e.g. Ingram 2002; Mitchell et al. 1987; and see Semenov and Bengtsson 2002) the Clausius–Clapeyron relation implies that specific humidity and hence atmospheric moisture would increase roughly exponentially with temperature (approximately $6.5\%\mathrm{K}^{-1}$, Boer 1993). Then if the heaviest rainfall events are likely to occur when effectively all the moisture in a volume of air is precipitated out, and the intensity of these events increases with the availability of moisture, we might expect the uppermost quantiles of the rainfall distribution to be constrained to increase also as Clausius–Clapeyron. This would apply if changes in the ambient flows have little impact on moisture convergence, which is most likely at higher latitudes. In the tropics, the flows leading to precipitation are themselves largely driven by the latent heat released by precipitation, and still larger *super Clausius–Clapeyron* increases might occur.

The overall intensity, however, of the hydrological cycle and hence global-mean precipitation is primarily constrained by the availability of energy not moisture. Specifically, it is determined by the ability of the troposphere to radiate away latent heat released from precipitation. The relevant energy balance is that of: solar heating of the surface (partially offset by radiative cooling) and radiative cooling of the troposphere being balanced by an upward latent heat flux, such that evaporation cools the surface and precipitation heats the troposphere. It is shown in Allen and Ingram (2002) that, for a range of model simulations under equilibrium CO_2 doubling, this global energy constraint dictates an approximately $3.4\%\ \mathrm{K}^{-1}$ increase in global-mean precipitation (In fact that regression between precipitation and temperature does not pass through the origin, having a negative offset. Hence for a given model the absolute mean precipitation change per degree warming will be less than if this). This is indeed a projected increase slower than the increase in the extreme (~6.5% K^{-1}).

It was then demonstrated that these concepts may be emergent globally at least in one coupled climate model, namely the Third Hadley Centre Coupled Model, HadCM3 (Pope et al. 2000; Gordon et al. 2000. Figure 1 (after their Fig. 4) shows their return level plot illustrating, for the upper 50 percentiles of both a transient CO_2 and control HadCM3 simulation, the change in precipitation at those percentiles.

We re-iterate the main features of this figure. Firstly one sees that above approximately the 90th percentile (1-in-10 day events and above) precipitation events in a transient climate are becoming more intense at a given percentile, and a given intensity of event is becoming more probable, compared to the corresponding control climate. In the context of a constraint, there appears to be a convergence of the ratio of transient to control precipitation to about a 26% increase at the highest end of the distribution which is indeed in the region expected from the Clausius–Clapeyron relationship. Furthermore, one could account for the slightly higher than predicted precipitation change value as being due to the dominance of the tropics (generally the major source of extreme daily precipitation) on the signal at the top percentiles and super Clausius–Clapeyron conditions here.

Secondly, we note that such increases at the extreme are more than double the increase in the global-mean precipitation (i.e. the change summed over all percentiles). Indeed the increase at the heaviest rain events is large enough that the energy constraint on the total implies that on only 1 day in 10 does precipitation increase (transient and control distributions cross around the 90th percentile as noted above), with accompanying decreases lower down in the distribution. The implication is that the change in mean precipitation would be a poorer indicator of the change in extreme precipitation, than would using the Clausius–Clapeyron prediction. This analysis, however, was only at the global annual level and is unsatisfying in the sense that behaviour at the extremes is dominated by tropical precipitation, with 70 and 72% of the precipitation above the 90th percentile occurring between 30°N and 30°S in the transient and control climates respectively (rising to 85% for precipitation above the 99.9th percentile in both climates). So relatively little can be deduced about other regions here.

The purpose of this paper is to extend the analysis of Allen and Ingram (2002) (henceforth referred to as 'AI02') to investigate, on regional and seasonal scales, the degree to which a 'Clausius–Clapeyron constraint' is a better predictor of changes in extreme precipitation, than is using the change in mean precipitation. If the Clausius–Clapeyron constraint is indeed an

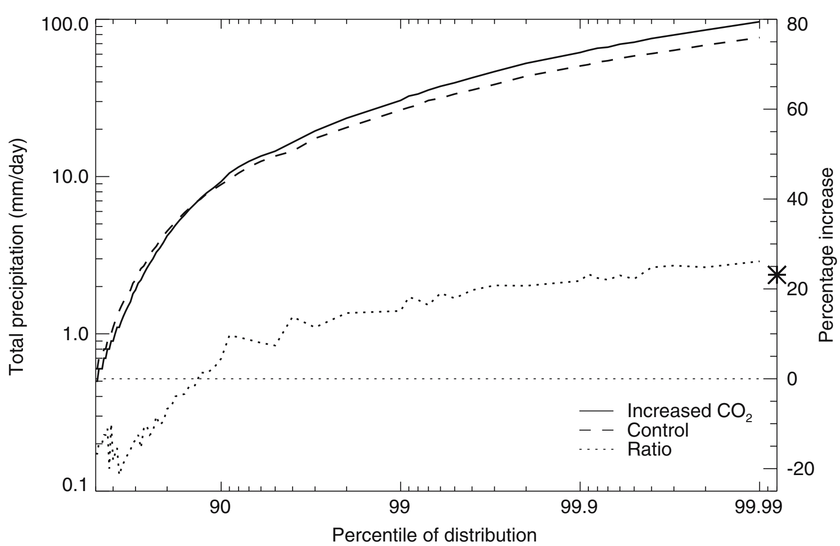

Fig. 1 After Allen and Ingram (2002), Fig. 4. Log–log plot of the change in distribution of global (all gridpoints and seasons) daily precipitation at years 2070–2100 of a transient HadCM3 climate change simulation. The *solid curve* shows the distribution in the transient simulation, where CO_2 levels have increased by a factor of approximately 2.7. The *dashed curve* shows the corresponding control simulation. Their ratio is shown by the *dotted curve* and right-hand (linear) axis. The global-mean warming is 3.6 K and the tropical-mean is 3.3 K, giving a Clausius–Clapeyron predicted limit on this ratio of about 23% as shown by the starred point. Note that Allen and Ingram (2002) *incorrectly*: stated this period as being around the time of CO_2 doubling; found a 22% Clausius–Clapeyron predicted limit; labelled the left-hand axis as 'Intensity')

uncontentious physical principle applicable to individual parcels of air then one might expect, within the timescale of local moisture transport, its emergence on regional scales also. We reproduce the analysis of Fig. 1 by model latitude, gridbox and season, using the the same data as was used in AI02.

The structure of the remainder of this paper is then as follows. Section 2 gives details of the climate model, forcing scenario and data used. Section 3 presents results of the precipitation changes found and their agreement with Clausius–Clapeyron predicted change. Section 4 provides a discussion of these results and the conclusions.

2 Experiment data and design

We use the same data used by AI02 in producing their Fig. 4, generated with the HadCM3 model. The atmospheric component of the model has 19 levels with a horizontal resolution of 2.5° latitude by 3.75° longitude, producing a global grid of 73×96 gridboxes. The large-scale precipitation and cloud scheme is formulated in terms of an explicit cloud water variable following Smith (1990). Johns et al. (2003) state that in general the model is realistic in capturing patterns of mean seasonal precipitation, particularly for DJF and JJA when judged against the CMAP climatology of Xie and Arkin (1997). The oceanic component has a higher horizontal resolution which makes it possible to represent important details in the oceanic current structures relevant to precipitation.

A first set of data was generated by performing a transient climate change integration under IPCC forcing scenario gIS92a (Leggett et al. 1992), often referred to as a ''business-as-usual'' scenario representing a plausible course for global energy under a public policy that gives no consideration to climate change concerns; the 'g' prefix indicating that contributions due to greenhouse gases only (and not due additionally to the direct effect of sulphate aerosols) were taken. A 'time-slice' of 30-years of the transient data was taken between years 2070 and 2100, at which point CO_2 levels had increased by a factor of approximately 2.7. A corresponding control integration without the above forcing scenario provided the second set of data. The time slice taken from this control was for 240-years. Although the two time slices do not correspond, the control is generally thought to be relatively stable before and during such a time period and the lengthier slice simply amounts to greater population of the stationary control density function.

The daily precipitation data over these transient and control slices was then provided in a binned format (J.M. Gregory, personal communication). Specifically, as histograms of precipitation for each model gridpoint, with bins of non-uniform width giving emphasis on resolving the smallest precipitation amounts (where most of the mass of the histogram was expected to lie). Table 1 shows the distribution of these bin widths. It is important to note that, whilst this binning allows us to more easily compute changes in percentiles of precipitation distribution, there is only spatial and no temporal information on variability of precipitation within these slices. This gridbox-resolution data does, however, allow us to perform our regional analyses. Further, the precipitation data was composed by season and in later analysis we dis-aggregate accordingly to perform seasonal analyses.

Mean temperature over the time-slices, used to compute the Clausius–Clapeyron predicted change in saturation vapour pressure (SVP) and hence moisture availability, was also provided on the same spatial grid. As a minor note, seasonal control mean-temperature was only available for a 100-year period as opposed to the 240-year transient data but this was not deemed problematic due to the stationarity of the control integration mentioned above.

With this data we repeat the analysis of AI02 by model latitude and gridbox to generate the analogous return level plots of change in extreme precipitation, as well as determining the change in mean. Note the abscissa can also be readily interpreted as a representing return times of the daily precipitation events, with an event at the $p \times 100$th percentile having a return time of, $T = 1/(1 - p)$ days. We also examine the seasonal change in the most extreme events.

3 Results

In this section we present results of how well change in extreme precipitation, under CO_2 warming, is being predicted by both the Clausius–Clapeyron relationship

Table 1 Distribution of bins for capturing daily precipitation at each model gridpoint, showing the non-uniformity in their width

Precipitation range (mm)	-0.05 to $9.\dot{9}$	10.0 to $208.\dot{9}$	209.0 to $508.\dot{9}$
Mid-bin value of first bin (mm)	0.0	10.5	359.0
Bin width (mm)	0.1	1.0	300.0

The last bin has a comparatively large range in an attempt to capture any extremely rare precipitation events that had not been captured by the already extreme bin values preceding it and by the change in mean precipitation at zonal and gridbox scales. Seasonal changes in the uppermost extremes are also presented.

3.1 Zonal analysis

3.1.1 Selected latitudes: tropical- versus high-latitudes

The global data is dis-aggregated by latitude so that now, at each latitude, we aggregate only the data over each gridbox at that latitude and then investigate its distribution. Return-level plots analogous to Fig. 1 are then generated for selected tropical- and high-latitudes (at 15°N and 65°N, respectively). This was done as a means of highlighting any differences in precipitation distribution change between regions well known to differ in their climatology. The results are shown in Fig. 2.

It is seen that at the tropical latitude behaviour similar to the global case is being exhibited—i.e. we have a cross-over of the curves at approximately the 90th percentile and apparent convergence of their ratio to some limit. At the high-latitude, however, there is a more or less uniform increase over all upper 50 percentiles and any convergence of the ratio is less clear.

3.1.2 All latitudes

Extending the analysis, one can generate the ‘Ratio’ curves of Fig. 2 at every model latitude (as well as now also over all percentiles). The result is encapsulated in Fig. 3.

There are three main features of interest in this plot. Firstly, one sees that between 60°N and 60°S there are distinct regions of crossover from precipitation decrease to increase with increases only at the highest

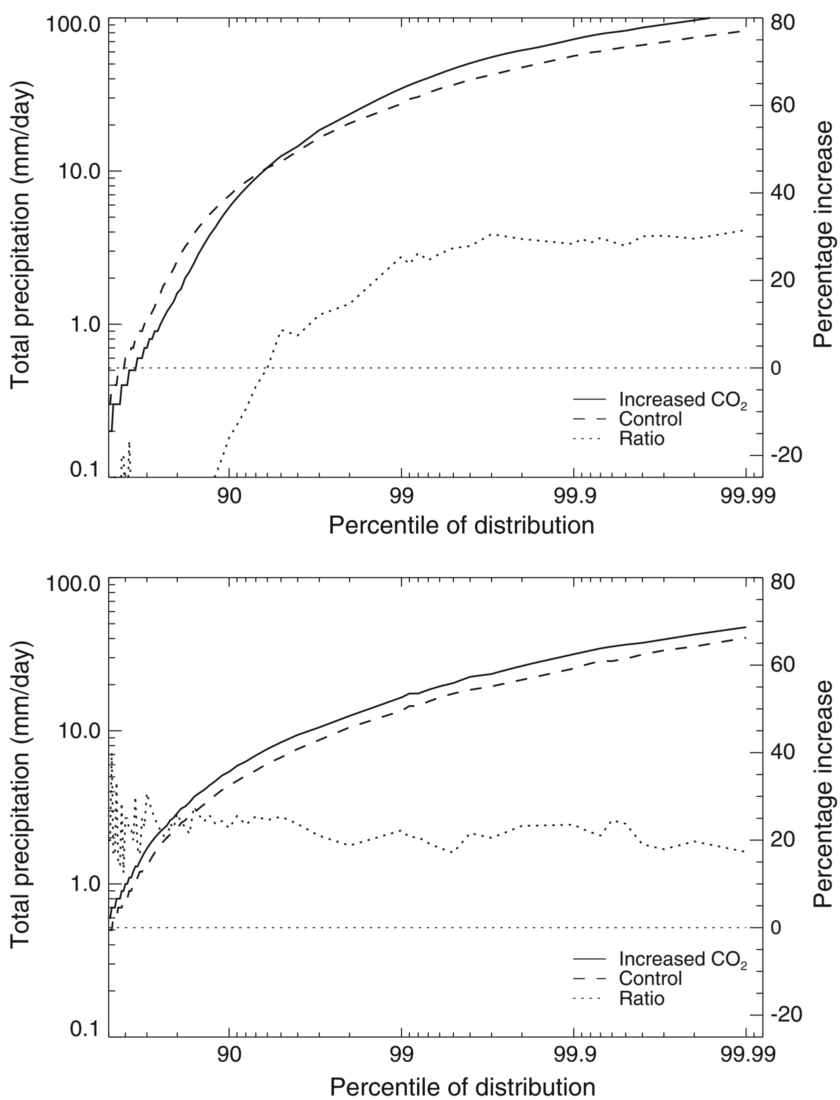

Fig. 2 As for Fig. 1 but now for the change in distribution at model latitudes of: (*top*) 15°N—a tropical case, (*bottom*) 65°N—a high-latitude case

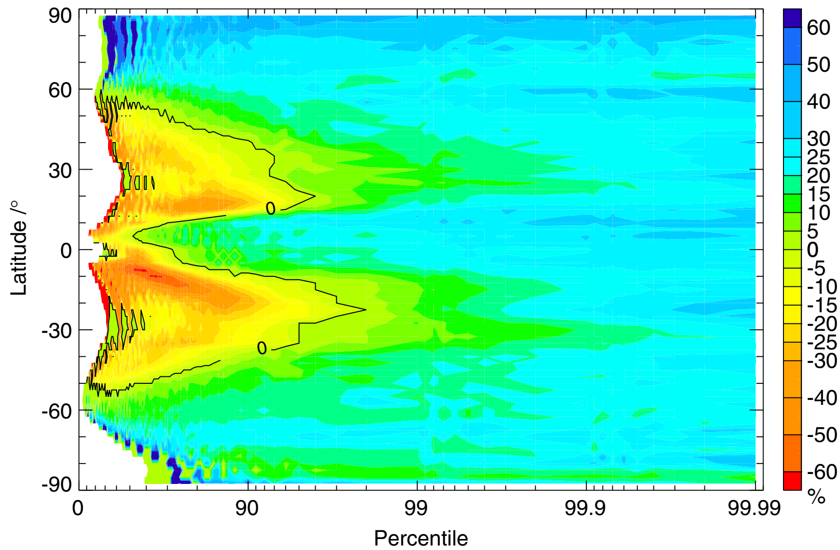

Fig. 3 Log-linear plot of the change in zonally aggregated precipitation, for all percentiles of the underlying distributions and over all latitude. *Green–blue* regions correspond to precipitation increases, and *yellow–red* regions to precipitation decreases. The zero precipitation change regions are contoured. The 'Ratio' (*dotted*) curves of Fig. 2 would correspond to vertical cross-sections through this contour plot at **a** 15°N and **b** 65°N respectively. The digitization at lower percentiles is due to discrete binning of the data. The white regions are due to zero (0–0.05 mm/day bin) control precipitation in the model at these percentiles and so an undefined ratio

percentiles, akin to the global behaviour. Secondly, at higher latitudes there is no crossover and precipitation increases over all percentiles. Thirdly, it appears that at around the 95th percentile is where we see the latest crossover, meaning at these latitudes (~30°N and 30°S), precipitation increases on only 1 day in 20. This pattern occurs in both hemispheres.

Before comparing the changes in extreme precipitation with mean changes and the Clausius–Clapeyron prediction, we proceed with further dis-aggregation of the data.

3.2 Gridbox analysis

The global data is now dis-aggregated by gridbox and for each gridbox the change in extreme and mean precipitation determined. The results are shown in Fig. 4. Note that the 99.9th percentile is chosen to represent the extreme to give reasonable sampling. Specifically, given that for daily precipitation over the 30-year time slice we have 10,800 counts per gridbox (model years are 360 days long), and at the 99.9th percentile we are looking for the signal of a 1-in-1,000 day event, the sampling can expected to be of order 10. Though this produces a relatively noisy spatial pattern of change (which has been smoothed in the figure) it is generally robust to patterns found at both lower and higher extreme percentiles (e.g. at 99th and 99.99th percentiles; not shown) which are simply clearer or unclearer due to higher or lower sampling respectively. This robustness might have been expected given the relatively low variability of precipitation change in Fig. 3 above the 99th percentile.

Both patterns of change in extreme and mean precipitation show pronounced regions of increases and decreases, suggesting changes in strength of convergence and subsidence or the atmospheric circulation (e.g Held and Soden 2006; Meehl et al. 2005), and the pattern of the latter generally agrees well with the mean-precipitation change pattern observed in in the CMIP-2 (AI02, Fig. 6). The greatest agreement between the two is in the intertropical convergence zone and subtropical high pressure belts over the tropical oceans. More interestingly, however, there are much larger regions of disagreement particularly at mid- to high-latitudes where the change in mean would underestimate change in extreme by of order 20–30%, and in some regions even estimate the wrong sign.

The general pattern of change in the gridbox-resolution precipitation seems coherent with the changes exhibited by the latitudinally aggregated data (Fig. 3). At mid- to high-latitudes there are increases in precipitation, and at low latitudes there are pronounced regions of drying indicative of the precipitation decreases at lower percentiles of the change distribution.

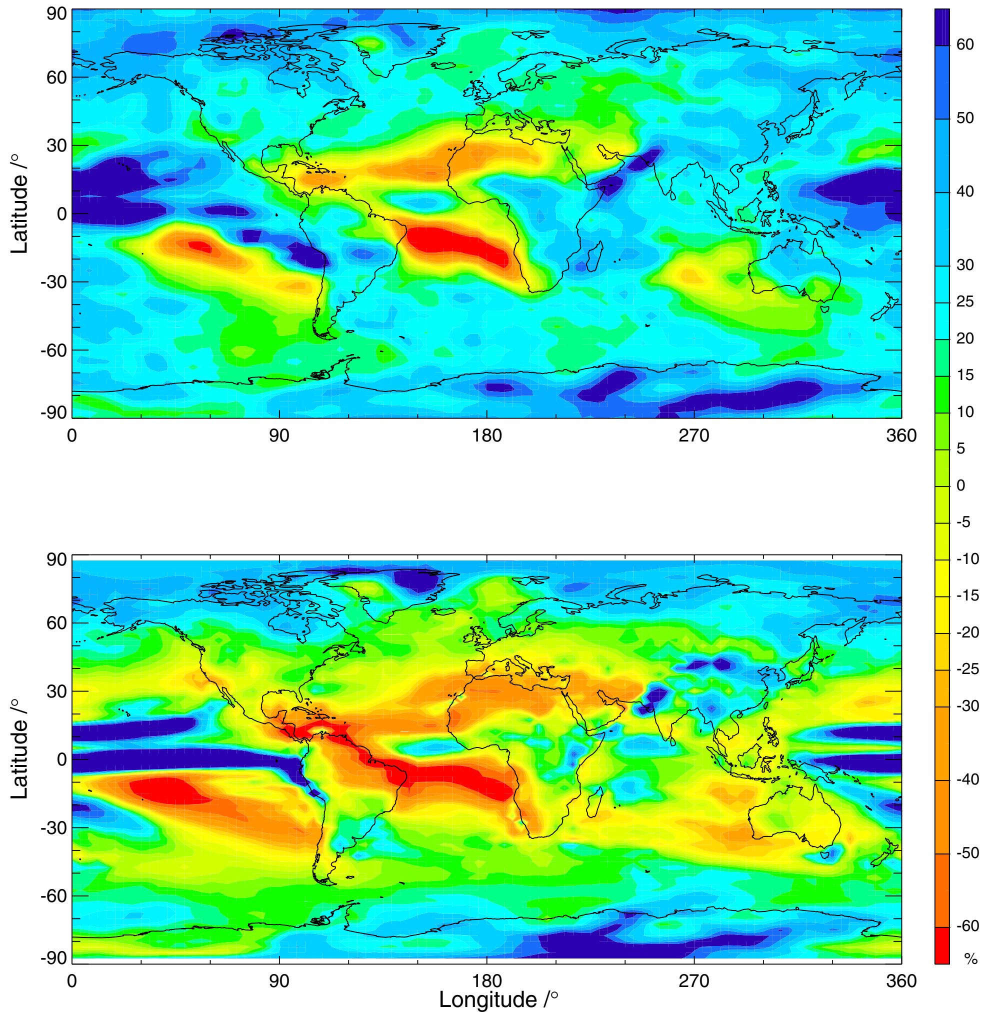

Fig. 4 Maps at model gridbox resolution of: (*top*) change in precipitation at the 99.9th percentile (1-in-1,000-day event) of the underlying distribution; (*bottom*) change in mean precipitation. *Green–blue* regions correspond to increases, and *yellow–red* regions to decreases. The top figure has been smoothed with a 7.5° × 11.25° lat-lon boxcar filter

3.3 Zonal comparison: extreme and mean change along with the Clausius–Clapeyron prediction

By means of further comparison with the changes found in latitudinally-aggregated precipitation, we now take the zonal-mean of the gridbox-resolution precipitation change maps shown in Fig. 4. This is compared to the latitudinally-aggregated precipitation change at the 99.9th percentile from Fig. 3. I.e. we calculate the average percentage change in precipitation, at the 99.9th percentile, *over each gridbox within a latitude band*, and compare this to computing changes at that same percentile found from having *first aggregated the data in that band*. Also now shown is the zonal-mean of the Clausius–Clapeyron predicted change in extreme precipitation computed from gridbox-resolution surface temperature. The result is encapsulated in Fig. 5.

Both the gridbox-resolution and latitudinal-resolution changes in extreme precipitation match with the Clausius–Clapeyron prediction to varying extents, and generally at mid-latitudes. The gridbox-resolution curve also exhibits minima in the tropical latitudes. We can reason this as being due to the effect of subsidence regions at mid- to low-latitudes that is now being picked up in the gridbox-resolution data. In these subsidence regions there are large decreases in precipitation at the high percentiles, and these provide a negative contribution when one takes the zonal-mean for that latitude, thus lowering the zonal-mean curve at that latitude. For the latitudinal-resolution curve, however, all gridboxes within a latitude circle contribute to the aggregate distribution at that latitude, so the changes at high percentiles are most easily influenced by gridboxes which already have relatively heavy events. This behaviour might most easily be explained if heavy precipitation increases more in regions which already have more heavy precipitation events and less in in regions which have fewer heavy precip events.

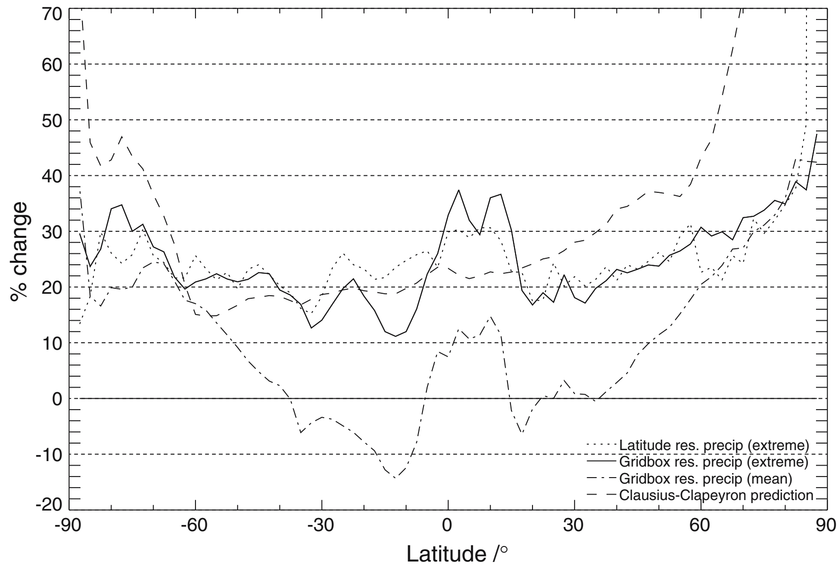

Fig. 5 Gridbox- versus latitudinal-resolution change in extreme precipitation, along with change in mean and Clausius–Clapeyron prediction. The comparison is made between change in 99.9th percentile precipitation of *latitudinally-aggregated* data (*dotted curve*), and the *zonal-mean* of the change in 99.9th percentile precipitation of gridbox-resolution data (*solid curve*). Also shown is the zonal-mean of gridbox-resolution change in mean precipitation (*dashed-dotted curve*), and the Clausius–Clapeyron predicted change (dashed curve)

Importantly, we see that the changes in extreme precipitation are generally better matched by the Clausius–Clapeyron prediction than the change in mean precipitation over all latitudes 60°N–60°S, with the mean precipitation being either close to zero or the wrong sign 15–35°N and 5–40°S, and the best prediction is at southern mid-latitudes. At high-latitudes, however, this is not so clear and there are dramatic increases in the prediction. It is interesting to note from the non-linearity of the Clausius–Clapeyron relation that for a *given* change in temperature, the smallest percentage change in SVP would occur in warmer regions and the largest in cooler regions. Specifically, the functional form of the relationship between SVP and Temperature, T, is: SVP $\propto exp(-L_v/(R_w T))$, where L_v is the specific latent heat of vapourization and R_w is the specific gas constant for water vapour. Hence the change dSVP/SVP is proportional to $1/T^2 dT$ and so for a given change in temperature dT, the percentage change in SVP will be small or large when T is large (warm) or small (cold) respectively [If one also accounts beforehand for the temperature dependence of L_v when deriving the functional form of the Clausius–Clapeyron relation, as has been done in this study, the above result still holds as a $1/T^2$ term again appears which dominates at small T]. This concept is summarised in Fig. 6 and partially explains why we see dramatic increases in the Clausisus–Clapeyron prediction curve at higher latitudes. The important accompanying explanation is there are actually larger *absolute* changes in temperature at higher latitudes (particularly in the north-polar regions) than at lower- and tropical latitudes.

3.4 Seasonal analysis

The analysis has thus far been limited to daily precipitation changes over the entire annual cycle. Given that intra-annual variability of extreme precipitation is well know to occur in many regions of the world, a seasonal analysis of the same data may help in further understanding what affects the validity of any constraint. Also note that since we now investigate by season our histogram population over the 30y time slice is reduced by factor four, so that for the transient simulation we now have 2,700 counts per gridbox compared to the previous 10,800. Again we restrict ourselves to look for precipitation changes up to only the 99.9th percentile (1-in-1,000 day events) to have adequate, if somewhat low, sampling.

We again examine zonal precipitation changes, following the procedure of Sect. 3.3. Figure 7 shows, by season, the change in extreme precipitation (again from both latitudinal and gridbox resolution data), mean precipitation, and the corresponding Clausius–Clapeyron prediction.

Again, changes in extreme precipitation generally have better agreement with the Clausius–Clapeyron prediction, than do changes in mean. Greatest regions of agreement with prediction are again at mid-latitudes, particularly in the southern hemisphere. There is also a notable decrease in predicted values for the northern hemisphere summer, reflecting the relatively small increase in temperature here typical for this model. One may also wonder how closely the mean of these seasonally dis-aggregated curves reproduces the annual curve of Fig. 5, as one might expect impacts due

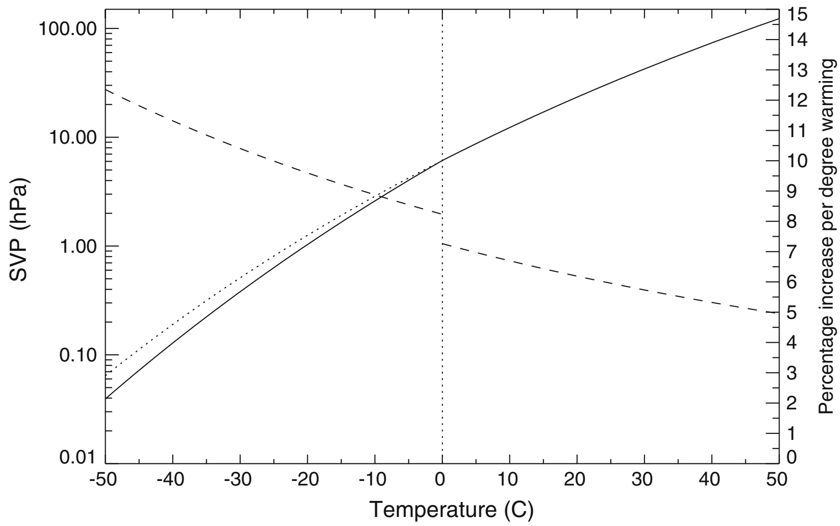

Fig. 6 Relationship between SVP, percentage SVP change, and temperature. SVP (*solid curve*) is expressed on a log scale (*left hand axis*). The *vertical dotted line* separates the vapour–liquid and vapour–ice transition regimes (above and below the triple point of water respectively), with the *dotted curve* showing the deviation of the SVP for the vapour–ice transition from the SVP of a hypothetically continuing vapour–liquid phase transition. The percentage increase in SVP per degree warming (*dashed curve*) is given on a linear scale (*right hand axis*) and has a discontinuity at the triple point, due to the change in regime. For warmer regions, the increase is smaller than for cooler regions

to dis-aggregation by time. It appears (not shown), however, to make little difference.

4 Discussion and conclusions

In the introduction we highlighted that many climate models show increases in extreme precipitation greater than that for the mean—the implication being that change in the latter may be a poor predictor of change in the former. We also acknowledged difficulties in using current climate models and the observational record for predicting changes in extreme precipitation under future climate change, and suggested that one should also search for physical constraints on changes in extreme precipitation. We then put forward a physical argument, following AI02, for why one might expect the change at the uppermost quantiles of precipitation distribution to be constrained by moisture availability, that is, by the Clausius–Clapeyron relationship, rather than energy budget considerations as is the case for the change in mean. Moreover we argued that this is an ‘uncontentious physical principle’ which one might expect to emerge in models of future climate.

To find evidence for this emergence the analysis of AI02 was extended. In AI02 encouraging results for the convergence of change in extreme daily precipitation to a limit predicted by the Clausius–Clapeyron relation had been observed in HadCM3. Though this change was indeed greater than the change in mean it was, however, observed on a global scale and of wider applicability would be the emergence of the Clausius–Clapeyron constraint on regional scales. Thus the extension of the analysis was to regional and seasonal scales also. We will now discuss the resulting morphology of change and predicted change presented in Sect. 3. It would seem that in searching for any regional robustness, one has to consider both the thermodynamics and dynamics.

At latitudinal-resolution we saw patterns of precipitation increase and decrease (Figs. 2, 3): at low- to mid-latitudes there is similar behaviour to that at global resolution in that increases in precipitation are observed only at higher percentiles of the change distribution, with decreases at lower percentiles. This is not unreasonable given that the tropics, generally the major source of extreme daily precipitation, dominate the global signal at the extremes as was stated in the introduction (85% of the global precipitation at the 99.9th percentile originating from between 30°N and 30°S in both transient and control climates. At high-latitudes there are increases over all percentiles and this behaviour is plausible because though globally the precipitation change integral is bounded by energy constraints this is not a requirement locally. At any particular local latitude increases at the extremes greater than the mean increase do not have to be accompanied by decreases at lower percentiles.

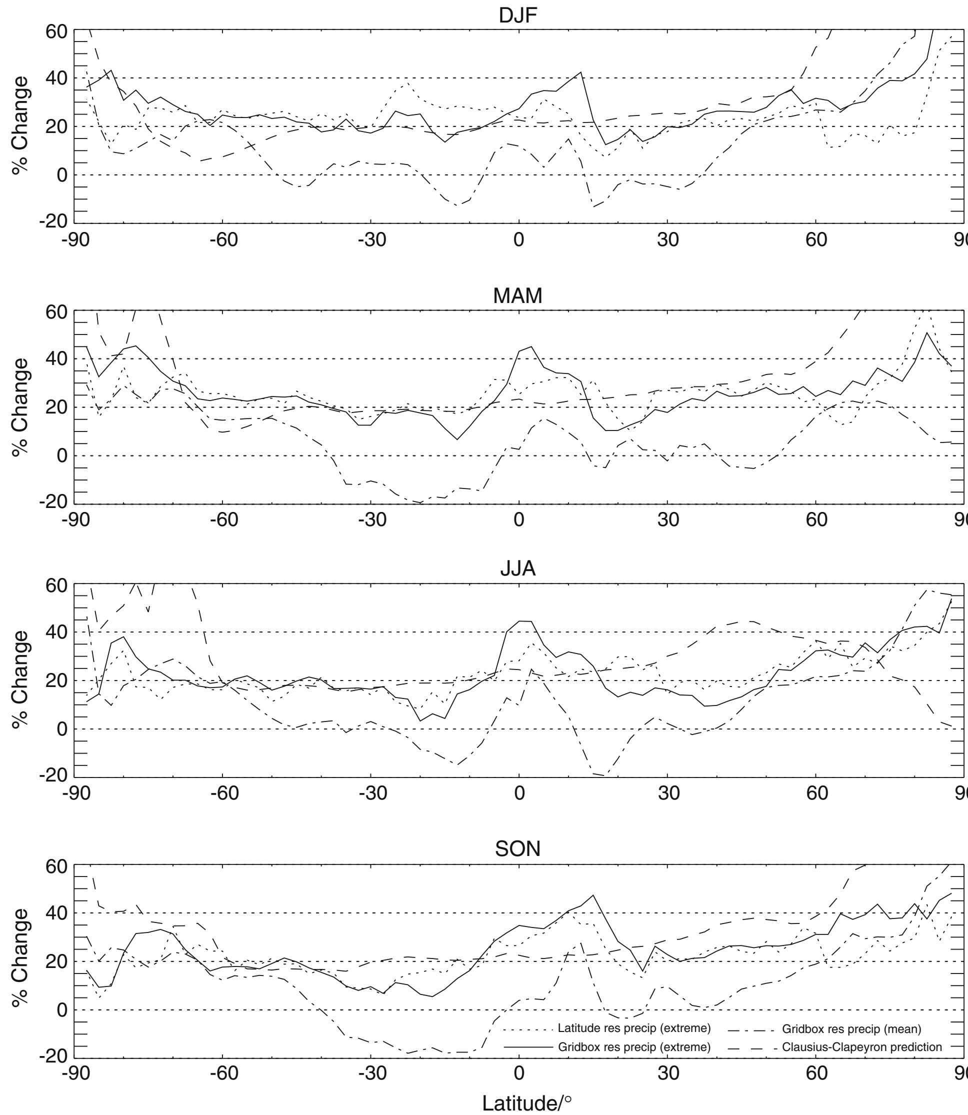

Fig. 7 Seasonality of extreme and mean precipitation change, along with Clausius–Clapeyron prediction. Curves for each season have been generated as for Fig. 5

This general pattern is consistent with a shift at low- to mid-latitudes toward more intense convective events resulting in fewer light non-convective events and fewer wet days; and at high-latitudes to simply an increase in the intensity of non-convective events. A physical explanation can attributed, following Hennessy et al. (1997), to the increase, under conditions of (near) constant relative humidity at the surface, of moisture convergence and hence precipitation in regions of persistent low-level convergence [for example, low-latitudes and in the inter-tropical convergence zone (ITCZ)]. Furthermore the subsidence associated with convection generally dries out the boundary layer and troposphere hence reducing the frequency of super-saturation and the occurrence, and perhaps amount, of large-scale precipitation. At higher latitudes latent heat is a less important part of the energy budget and changes can be closer to Clausius–Clapeyron predicted over all percentiles.

Watterson and Dix (2003) also found the shift in distribution of wet day amounts toward heavier events at nearly all latitudes, even those where the dry day frequency increases. Even with less sophisticated models, McGuffie et al. (1999) find increases in moderately extreme daily precipitation in mid-latitudes, with an increase in frequency of very extreme precipitation events in the Intertropical Convergence Zone.

At gridbox-resolution, we again saw a general increase in precipitation at mid- to high- latitudes and decreases at low-latitudes (Fig. 4). In short, the general tendency was for precipitation to increase in regions where it is already high and decrease where evaporation is already high, which is just what one might expect from the increased specific humidities leading to increased atmospheric fluxes of humidity from its sources to its sinks (AI02). Over the tropical oceans there is also a northward shift in the ITCZ associated with a displacement of the thermal equator in that direction, which is symptomatic of this model (Johns et al. 2003).

As alluded to in the introduction we would expect our thermodynamic constraint theory to hold in regions where the nature of the ambient flows have changed little under under CO_2 warming, the argument assuming that local temperature changes give rise to local changes in atmospheric moisture capacity and subsequent precipitation. The fact that, when comparing zonal profiles in Fig. 5, we see departures between the thermodynamic prediction and actual extreme precipitation, particularly at higher latitudes, implies that the dynamics are changing or that other factors must be considered.

Indeed Meehl et al. (2005) diagnose in multi-model transient simulations that at mid- to higher-latitudes, in addition to increases in water vapour from local sources, advective effects associated with changes in atmospheric circulation produce increases in mean and extreme precipitation—some of which are caused by a polward shift in the storm tracks (Yin 2005). We also speculate that, at the poles, we may also be effectively seeing changes to the boundary conditions for this problem due to changes in sea-ice distribution which would affect moisture locally available via evaporation from the underlying ocean and/or possibly affect the atmospheric circulation there.

In the tropics, we see departures supposedly due to the aforementioned northward shift in thermal equator so that there are relatively large increases in precipitation at 0–20°N with concurrent decreases at 0–20°S. Under the implied convergence in this increase region, we may be seeing the super Clausius–Clapeyron feedback on precipitation where the flows leading to precipitation are themselves driven by the latent heat released from precipitation and thus larger than predicted values, though there is no simple a priori way to estimate this. A change in the nature of the transient flow in models, particularly in the tropics, is also suggested by Held and Soden (2006) as due to the inability of the radiatively constrained hydrological cycle to respond to Clausisus–Clapeyron order increases in lower troposheric water vapour. The result is a weaking of the circulation due to decreases in convective mass flux to the free troposphere and subsequent horizontal redistribution of moisture transport. Evidence for this type of dynamical weakening is found in a modified atmosphere-only version of HadCM3 by Emori and Brown (2005).

Other factors for discrepency between predicted and modelled extreme precipitation may be due to not all the available moisture being used up in the heavy precipitation events, even at the extreme percentiles considered, or possible multiple cycling of evaporation and precipitation within a day (albeit less likely in the extra-tropics).

Conversely, where there is good agreement between predictied and extreme precipitation at southern mid- to high-latitudes this could be reasoned by arguing that not only is moisture more readily available in these regions from the oceans relative to the northern hemisphere, but also the associated lack of orography (which otherwise redirects flow) means that flows are generally more zonal here so that moisture may remain more local over a latitude circle. Consequently one may argue that, particularly for the northern hemisphere, repeating this analysis exclusively over the oceans may yield better results.

Importantly, we see that if the spatial pattern of change in mean precipitation were used to directly predict the change in extreme then there would be many regions, particularly at low- to mid-latitudes and even high-latitudes, of under-prediction by of order 20–30% and even predictions of the wrong sign. Parallel analysis by season (Fig. 7) produced results which, whilst in some cases reinforcing the above findings, did not add anything new to them.

The main conclusion, then, is that we have identified the *thermodynamic* Clausius–Clapeyron constraint as providing good prediction of future changes in extreme precipitation only over mid-latitude regions. Importantly, though, in these regions as well as in the tropics the constraint is a better predictictor for future changes in extreme precipitation than is directly using the change in mean precipitation. In regions where there are departures from the prediction, we suggest that *dynamic* factors, that is, changes in the nature of the atmospheric circulation and hence moisture are expected to dominate, with other multi-model studies providing supporting evidence for such changes in circulation.

Indeed, Trenberth (1999a) demonstrates using observations that, for 1,000-km scales, a global annual average of less than 20% of precipitation is 'recycled' from the local evaporation within each domain, with highest recycling rates occurring over the sub-tropics and convergence zones where advective moisture fluxes are small, and low values occurring over southern oceans, the North Pacific, and the eastern equatorial Pacific, where moisture fluxes are at maximum. Trenberth and Dai (2003) further explain that about 6 times the locally available moisture can fall in one day, and from scales 3 to 5 times the radius of the precipitating area. It is also noted that surface characteristics and vegetation greatly affect the moisture available over land for evaporation as a source for precipitation there, reinforcing the failure of our theory in very dry regions (e.g. Koster et al. 2004).

Having said all this, it has recently been found across some climate models under CO_2 warming that both increases in mean and extreme precipitation are dominated by thermodynamic rather than dynamic factors, again with increases in global-mean extreme precipitation larger than the mean (Emori and Brown 2005). Clausius–Clapeyron order (~6.5% K^{-1}) increases have also been projected in nested regional models over Europe (Frei et al. 1998: Increases in precipitation frequency consistent with a 15% increase in precipitation intensity, for an imposed uniform 2 K warming under constant relative humidity) and Australia (Hennessy et al. 1998: 15% increase in 1-in-20 year daily rainfall events for a 0.5–2.7 K warming by 2050), and in a high resolution nested hurricane model (Knutson et al. 2001, 31% and 19% for maximum local and areal-averaged precipitation respectively, given a temperature rise of 2.2–2.7 K; in fact these changes are larger and again accountable by super Clausius–Clapeyron considerations). In the hurricane case, changes were also attributed to a higher vapour mixing ratio under CO_2 warming and subsequent convergence.

Coupled with AI02, the findings of this study suggest that, at least in mid-latitudes, the Clausius–Clapeyron relation provides a useful constraint on changes at least in mid-latitude extreme precipitation, and importantly, more usefully so than the change in mean precipitation. An important implication is that in these regions one could then take the generally better defined changes in simulated temperature under CO_2 warming, scaled according to the ratio of simulated twentieth century temperatures to those observed, and apply the Clausius–Clapeyron constraint to infer the changes in extreme precipitation. Given the weaknesses identified in the last Intergovernmental Panel on Climate Change Assessment Report relating to direct model simulation of extreme precipitation events and in observing changes in extremes directly, this could be a worthwhile result.

Nevertheless, more systematic studies need to be conducted in order to establish whether this result is robust across models and hence the extent to which it may be expected to apply to the real world. For example, a more sophisticated experiment design would test the constraint across all major forcings (greenhouse gas or otherwise) for the twentieth century where the characteristics of model precipitation are known (Lambert et al. 2004, 2005) to vary differently in their response to each separate forcing.

Acknowledgments The authors especially wish to thank Jonathan Gregory for provision of data and assistance with analysis. We also thank him, along with William Ingram, for insightful discussion. Furthermore, we are grateful to the reviewers for helpful suggestions and Steve Jewson and Robert Muir-Wood for advice on interpretation of results. P.P. was supported by a CASE Award studentship from the UK Natural Environment Research Council with Risk Management Solutions Ltd. M.R.A. received support from the US NOAA/DoE International Detection and Attribution Group. D.A.S. was supported by the UK Department for Environment, Food, and Rural Affairs and a Wellcome Trust Showcase Award.

References

Adler R, Huffman G, Chang A, Ferraro R, Xie P, Janowiak J, Rudolf B, Schneider U, Curtis S, Bolvin D, Gruber A, Susskind J, Arkin P, Nelkin E (2003) The version-2 global precipitation climatology project (GPCP) monthly precipitation analysis (1979-present). J Hydrometeorol 4:1147–1167

Allen M, Ingram W (2002) Constraints on future changes in climate and the hydrologic cycle. Nature 419:224–232

Betts A, Jakob C (2002) Evaluation of the diurnal cycle of precipitation, surface thermodynamics, and surface fluxes in the ECMWF model using LBA data. J Geophys Res 107:8045

Boer G (1993) Climate change and the regulation of the surface moisture and energy budgets. Clim Dyn 8:225–239

Collins M, Booth B, Harris G, Murphy J, Sexton D, Webb M (2006) Towards quantifying uncertainty in transient climate change. Clim Dyn (in press)

Cubasch U, Meehl GA (2001) Projections of future climate change. In: Houghton JT et al (eds) Climate change 2001, The scientific basis, chap. 9. Cambridge University Press, Cambridge, pp 527–582

Dai A, Trenberth K (2004) The diurnal cycle and its depiction in the community climate system model. J Clim 17:930–951

Dai A, Giorgi F, Trenberth K (1999) Observed and model-simulated diurnal cycles of precipitation over the contiguous United States. J Geophys Res 104:6377–6402

Easterling D, Evans J, Groisman P, Karl TR, Kunkel K, Ambenje P (2000) Observed variability and trends in extreme climatic events: a brief review. Bull Am Meteor Soc 81:417–426

Emori S, Brown S (2005) Dynamic and thermodynamic changes in mean and extreme precipitation under changed climate. Geophys Res Lett 32:L17 706

Emori S, Hasegawa A, Suzuki T, Dairaku K (2005) Validation, parameterization dependence, and future projection of daily precipitation simulated with a high-resolution atmospheric gcm. Geophys Res Lett 32:L06708

Folland CK, Karl TR (2001) Observed climate variability and change. In: Houghton JT et al (eds) Climate change 2001, The scientific basis, chap. 2, Cambridge University Press, Cambridge, pp 99–181

Frei C, Schär C, Lüthi D, Davies H (1998) Heavy precipitation processes in a warmer climate. Geophysical Res Lett25:1431–1434

Gordon C, Cooper C, Senior C, Banks H, Gregory J, Johns T, Mitchell J, Wood R (2000) The simulation of SST, sea ice extents and ocean heat transports in a version of the Hadley Centre coupled model without flux adjustments. Clim Dyn 16:147–168

Groisman P, Legates D (1994) The accuracy of United States precipitation data. Bull Am Meteor Soc 75:215–228
Groisman P, Karl T, Easterling D, Knight RW, Jamason P, Hennessy K, Suppiah R, Page C, Wibig J, Fortuniak K, Razuvaev V, Dougal A, Førland E, Zhai P (1999) Changes in the probability of heavy precipitation: important indicators of climatic change. Clim Change 42:243–283
Hegerl G, Zwiers F, Stott P, Kharin V (2004) Detectability of anthropogenic changes in annual temperature and precipitation extremes. J Clim 17:3683–3700
Held I, Soden B (2006) Robust responses of the hydrological cycle to global warming. J Clim (accepted)
Hennessy K, Gregory J, Mitchell J (1997) Changes in daily precipitation under enhanced greenhouse conditions. Clim Dyn 13:667–680
Hennessy K, Whetton P, Katzfety J, McGregor J, Jones R, Page C, Nguygen K (1998) Fine resolution climate change scenarios for New South Wales, annual report 1997–1998: research undertaken for the New South Wales environment protection authority. Available from CSIRO Private Bag 1, Mordialloc, Victoria 3195, Australia
Ingram W (2002) On the robustness of the water vapor feedback: Gcm vertical resolution and formulation. J Clim 15:1917–1921
IPCC (2002) Report on: Workshop on changes in extreme weather and climate events. Beijing, China, 11–13 June 2002. IPCC supporting material Available from http://www.ipcc.ch/pub/support.htm
Johns T, Gregory J, Ingram W, Johnson C, Jones A, Lowe J, Mitchell J, Roberts D, Sexton D, Stevenson D, Tett S, Woodage M (2003) Anthropogenic climate change for 1860 to 2100 simulated with the HadCM3 model under updated emissions scenarios. Clim Dyn 20:583–612
Karl T, Knight R (1998) Secular trends of precipitation amount, frequency and intensity in the USA. Bull Am Meteor Soc 79:231–242
Kharin V, Zwiers F (2000) Changes in the extremes in an ensemble of transient climate simulations with a coupled atmosphere-ocean GCM. J Clim 13:3760–3788
Kiktev D, Sexton D, Alexander L, Folland C (2003) Comparison of modeled and observed trends in indicies of daily climate extremes. J Clim 16:3560–3571
Knutson T, Tuleya R, Shen W, Ginis I (2001) Impact of CO_2 induced warming on hurricane intensities as simulated in a hurricane model with ocean coupling. J Clim 14:2458–2468
Koster R, Dirmeyer P, Guo Z, Bonan G, Chan E, Cox P, Gordon C, Kanae S, Kowalczyk E, Lawrence D, Liu P, Lu CH, Malyshev S, McAvaney B, Mitchell K, Mocko D, Ok T, Oleson K, Pitman A, Sud Y, Taylor C, Verseghy D, Vasic R, Xue Y, Yamada T (2004) Regions of strong coupling between soil moisture and precipitation. Science 305:1138–1140
Lambert F, Stott P, Allen M, Palmer M (2004) Detection and attribution of changes in 20th century land precipitation. Geophys Res Lett 31:L10 203
Lambert F, Gillett N, Stone D, Huntingford C (2005) Attribution studies of observed land precipitation changes with nine coupled models. Geophys Res Lett 32:L18 704
Leggett J, Pepper WJ, Swart R (1992) Emissions scenarios for the IPCC: an update. In: Houghton JT et al (eds) Climate change 1992, the supplementary report to the IPCC scientific assessment, chap. 3, Cambridge University Press, Cambridge, pp 69–75
McAvaney B (2001) Model evaluation. In: Houghton JT et al (eds) Climate change 2001, the scientific basis, chap. 8. Cambridge University Press, Cambridge, pp 471–526
McGuffie K, Henderson-Sellers A, Holbrook N, Kothavala Z, Balachova O, Hoekstra J (1999) Assessing simulations of daily temperature and precipitation variability with global climate models for present and enhanced greenhouse climates. Int J Climatol 19:1–26
Meehl G, Boer G, Covey C, Latif M, Stouffer R (2000a) The coupled model intercomparison project (CMIP). Bull Am Meteor Soc 81:313–318
Meehl G, Zwiers F, Evans J, Knutson T, Mearns L, Whetton P (2000b) Trends in extreme weather and climate events: issues related to modelling extremes in projections of future climate change. Bull Am Meteor Soc 81:427–436
Meehl G, Arblaster J, Tebaldi C (2005) Understanding future patterns of increased precipitation intensity in climate model simulations. Geophys Res Lett 32:L18 719
Mitchell J, Wilson C, Cunnington W (1987) On CO_2 climate sensitivity and model dependence of results. Q J R Mereorol Soc 113:293–322
Osborn T, Hulme M (1997) Development of a relationship between station and grid-box rainday frequencies for climate model evaluation. J Clim 10:1885–1908
Osborn T, Hulme M, Jones P, Basnett T (2000) Observed trends in the daily intensity of United Kingdom precipitation. Int J Climatol 20:347–364
Pope V, Gallani ML, Rowntree P, Stratton R (2000) The impact of new physical parameterizations in the Hadley Centre climate model: HadAM3. Clim Dyn 16:123–146
Räisänen J, Hansson U, Ullerstig A, Döscher R, Graham L, Jones C, Meier H, Samuelsson P, Willén U (2004) European climate in the late twenty-first century: regional simulations with two driving global models and two forcing secnarios. J Clim 22:13–31
Semenov V, Bengtsson L (2002) Secular trends in daily precipitation characteristics: greenhouse gas simulation with a coupled AOGCM. Clim Dyn 19(2):123–140
Smith R (1990) A scheme for predicting layer clouds and their water content in a general circulation model. Q J R Mereorol Soc 116:435–460
Suppiah R, Hennessy K (1998) Trends in total rainfall, heavy rain events and number of dry days in australia, 1910–1990. Int J Climatol 18:1141–1164
Trenberth K (1999a) Atmospheric moisture recycling: role of advection and local evaporation. J Clim 12:1368–1381
Trenberth K (1999b) Conceptual framework for changes of extremes of the hydrological cycle with climate change. Clim Change 42:327–339
Trenberth K, Dai A (2003) The changing character of precipitation. Bull Am Meteor Soc 84:1205–1217
Watterson I, Dix M (2003) Simulated changes due to global warming in daily precipitation means and extremes and their interpretation using the gamma distribution. J Geophys Res 108:4379
Wehner M (2004) Predicted 21st century changes in seasonal extreme precipitation events in the Parallel Climate Model. J Clim 17:4281–4290
Wilby R, Wigley T (2002) Future changes in the distribution of daily precipitation totals across the United States. Geophys Res Lett 29:391–394
Xie P, Arkin P (1997) Global precipitation: a 17-year monthly analysis based on gauge observations, satellite estimates and numerical model outputs. Bull Am Meteor Soc 78:2539–2558
Yin J (2005) A consistent poleward shift of the storm tracks in simulations of 21st century climate. Geophys Res Lett 32:L18 701