

The intensity of precipitation during extratropical cyclones in global warming simulations: a link to cyclone intensity?

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ABSTRACT

Simulations of global warming over the coming century from two CSIRO GCMs are analysed to assess changes in the intensity of extratropical cyclones, and the potential role of increased latent heating associated with precipitation during cyclones. A simple surface cyclone detection scheme is applied to a four-member ensemble of simulations from the Mark 2 GCM, under rising greenhouse gas concentrations. The seasonal distribution of cyclones appears broadly realistic during 1961–1990. By 2071–2100, with 3 K global warming, numbers over 20°N to 70°N decrease by 6% in winter and 2% annually, with similar results for the south. The average intensity of cyclones, from relative central pressure and other measures, is largely unchanged however. 30-yr extremes of dynamic intensity also show little clear change, including values averaged over continents.

Mean rain rates at cyclone centres are typically at least double rates from all days. Rates during cyclones increase by an average 14% in the northern winter under global warming. Rates over adjacent grid squares and during the previous day increase similarly, as do extreme rates. Results from simulations of the higher-resolution (1.8° grid) Mark 3 GCM are similar, with widespread increases in rain rates but not in cyclone intensity. The analyses suggest that latent heating during storms increases, as anticipated due to the increased moisture capacity of the warmer atmosphere. However, any role for enhanced heating in storm development in the GCMs is apparently masked by other factors. An exception is a 5% increase in extreme intensity around 55°S in Mark 3, despite decreased numbers of lows, a factor assessed using extreme value theory. Further studies with yet higher-resolution models may be needed to examine the potential realism of these results, particularly with regard to extremes at smaller scale.

1. Introduction

A number of studies (e.g. Zhang and Wang, 1997; Sinclair and Watterson, 1999) of simulations of future climate change by GCMs have indicated that the overall frequency of cyclones (low-pressure events or ‘storms’), in the middle to high latitudes may decrease with global warming over the coming century (typically 3 K; Cubasch et al., 2001) due to increased greenhouse gases. Such decreases can result from a decrease in the pole-to-equator temperature gradient, and hence baroclinicity in the mid-latitudes, although the relationship of storm activity to other indices remains a field of active research (Paciorek et al., 2002).

Studies of GCMs of moderate resolution (2°–4° grid spacing) by Hall et al. (1994) and Carnell and Senior (1998) have also found an increase in storm intensity in certain ‘storm track’ regions. This was consistent with a potential increase in storm enhancement by the latent energy release associated with pre-

cipitation, which could increase as a result of the larger moisture content of the warmer atmosphere (Stocker et al., 2001, p. 432). Such enhancement, relative to dry or no latent heat conditions, has been demonstrated in simulations of storms with regional models (e.g. Kuo et al., 1995; Revell and Ridley, 1995; Ahmadi-givi et al., 2004). More recently, Kaas et al. (2001) reported a shift in the wintertime Atlantic storm track simulated by a high-resolution (~1.1°) global model, associated with decreased Arctic mean surface pressure. However, rather limited change in extreme storminess and wind speeds was evident. Even when individual storms were resimulated at higher resolution, the additional moisture (some 25%) made available by increasing temperatures by 3 K had only a small effect, in contrast to the difference compared to runs with no latent heat. Such a non-linear dependence on moisture appears comparable to the effects of moisture on large-scale baroclinic life cycles found by Gutowski et al. (1992).

In any case, many studies (e.g. Hennessy et al., 1997; Kharin and Zwiers, 2000) have found that the frequency distribution of daily precipitation from GCMs does change. While there are

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fewer days with rain at many places, there are widespread increases in the intensity of rainfall. Watterson and Dix (2003) have shown this to be the case in an ensemble of global warming simulations of the CSIRO Mark 2 (or Mk2) coupled atmosphere–ocean GCM. Using multiple representations of the warmer period 2071–2100 allowed even 30-yr extremes to be shown to increase, with some statistical significance. Similar relative changes in extreme rainfall occur in the new, higher-resolution CSIRO Mark 3 (Mk3) model. Such increases in extreme rainfall or winds, particularly over inhabited land, are of importance to the assessment of the impacts of climate change.

There appears to be no global modelling study that has considered the rainfall associated with cyclones, however. Such an analysis may shed light on the potential for intense rainfall to be generated coincidentally with intense cyclones, and hence the potential for a latent heat enhancement. This paper describes analyses of cyclones and rainfall from both the Mk2 and Mk3 simulations, using the data sets described in the following section. The data do not include latent heating, and a study of cyclone intensification is not attempted. A simple surface cyclone event criterion is used, and various measures of both dynamical and rainfall intensity are assessed. Following Watterson and Dix (2003), a statistical theory for extreme events, based on the gamma distribution and outlined in Appendix A, is also considered. The focus is on relative changes of intensity, particularly in the winter hemisphere, and on extremes over land. Mean baroclinicity and extreme surface wind speeds are also considered. Methods and selected results for Mk2 are described in Section 3. Results for Mk3 are in Section 4. A discussion follows, and conclusions are given in Section 6.

2. Models and data sets

2.1. Mk2

Originally presented by Gordon and O'Farrell (1997), Mk2 is a now-standard coupled atmosphere–ocean GCM simulating atmospheric dynamical processes on nine sigma levels, with a modest R21 spectral horizontal representation. The ensemble of simulations of the period 1871–2100 described by Watterson and Dix (2003) is considered here, with runs differing only through perturbations of the initial state. The model climate changes from radiative forcing (Watterson and Dix, 2005) by rising greenhouse gas (GHG) concentrations, as well as changes in ozone and sulphate aerosol (through surface albedo adjustments), following the 'A2' scenario of Cubasch et al. (2001) beyond 1990, with steady growth in GHG. The cyclone analysis has been performed for four simulations of the period 1961–1990, and four of the 2071–2100 period. The global mean increase of surface air temperature between these periods is 3.49 K.

The simulated data are on the model's Gaussian grid of 64 longitudes and 56 latitudes, approximately 5.6° by 3.2° , in large monthly files. The dynamical data are available 12-hourly, at

1200 and 2400 UTC, being the end of each model day. The pressure-level data used were determined by interpolation from sigma level values. Where 1000 hPa is beneath the surface, the lowest sigma level temperature, from around 170 m altitude, is used in the height calculation, while the level winds are used unchanged. The precipitation data set is 24-h accumulated 'rain', including simulated frozen precipitation, at the end of each day. In addition, each value effectively represents an average over the grid square surrounding each grid point.

The 1961–1990 climate is very similar to that of the 'control' simulation from the slab-ocean version of Mk2, which Sinclair and Watterson (1999) found to contain quite realistic cyclone activity at the length-scale resolved. This included the position of storm tracks in both hemispheres and the location and frequency of cyclogenesis.

2.2. Mk3

The Mk3 coupled GCM is a substantial upgrade to Mk2, with the resolution of the atmosphere increased to 18 sigma levels (the lowest around 45 m) and T63 or 1.8° by 1.8° . Other relevant differences are the adoption of the stratiform cloud and precipitation scheme of Rotsteyn (1997) and the convection scheme of Gregory and Rowntree (1990). The GCM and its control (unforced) simulation are described fully by Gordon et al. (2002).

A single simulation of change under the A2 scenario over the full 1871–2100 period is available. As described by Watterson and Dix (2005), the resulting radiative forcing is similar to that of the Mk2 case, except for a smaller cooling effect from aerosol changes. Nevertheless, the smaller sensitivity of Mk3 results in a reduced warming, especially at higher latitudes. The global warming from averages over 2071–2100 relative to 1961–1990 is 2.57 K, but this underestimates the forced response because of a slight continuing drift in the control run. Allowing for the corresponding periods of that run, the inferred response is 2.77 K.

In order to be able to compare cyclone analyses from multiple 30-yr periods, we can consider a series of nine periods of the control run to represent the climate of the recent past. The average mean temperature of these is 0.4 K lower than the 1961–1990 value. Only two additional periods of similar mean temperature to the A2 2071–2100 case are available. One is an extension of the run beyond 2100 with constant forcing, which is 0.7 K warmer (averaged over 2101–2130). The other is the 2071–2100 period from the preliminary A2 simulation described by Gordon et al. (2002), which is 0.5 K cooler because the forcing started only at 1961. Large-scale averages of the analysis results from these three periods suggest that any significant differences between the cases are linear in mean temperatures, so we use the three as an ensemble whose average approximately represents the 2071–2100 period. The mean warming between these two ensembles is then 3.07 K, 11% more than the 1961–1990 to 2071–2100 forced change, and 12% less than that from Mk2.

Under linearity, the change results from Mk3 that follow could be reduced by 10% to give a more accurate result for the two periods.

For all these 30-yr cases, the atmospheric and rainfall data are of the same form as for Mk2. The data are available only on a grid reduced to 96 longitudes and 48 latitudes, by taking values from every second point in each direction of the Gaussian grid. Given that simulated diffusive processes must smooth the smallest scales represented in a GCM, use of this 'reduced grid' should not be too detrimental. A further full-grid data set for just 2400 UTC sea level pressure and daily accumulated rain is also considered. The Mk3 control and A2 data sets will also be assessed by the Intergovernmental Panel on Climate Change (IPCC; see http://www-pcmdi.llnl.gov/ipcc/about_ipcc.php).

2.3. ERA40 Reanalyses

While the focus of the paper is on relative changes in storminess simulated by CSIRO global models, some comparison with observational results for 1961–1990 is warranted. For this purpose daily (1200 UTC) 1000-hPa data from the European Centre for Medium-Range Weather Forecasts (ECMWF) ERA40 reanalysis project (Kållberg et al., 2004) are used. These are on a 2.5° grid, representing fields spectrally truncated to T63, so nominally matching the Mk3 resolution.

3. Analysis of Mk2 data

3.1. Cyclone analysis

Following Zhang and Wang (1997) and Paciorek et al. (2002) we consider an 'index' of surface cyclone occurrence based on local lows in the 12-hourly 1000-hPa height (Z) fields. Events with the height at the cyclone centre lower than those of the surrounding eight points on the data grid, and satisfying the local 1000-hPa relative vorticity criterion $(f/|f|)\xi > 1 \times 10^{-5} \text{ s}^{-1}$, are counted. The vorticity is calculated using simple differencing of 1000-hPa winds from the four grid points adjacent to the centre. The monthly index of cyclone activity is half the number over all times, so representing the time in days within each month when a cyclone centre is present in a grid square. Each seasonal index is the sum of the index from the relevant three months. For convenience, we will use this as a measure of cyclone numbers or frequency in the unit 'cyclone-days per season'.

The vorticity criterion is the same as that used by Sinclair and Watterson (1999) although their storm centres were determined from ξ local extrema. No persistence or movement criteria are used here, however, partly for simplicity but also because they are of less relevance to impacts of lows over land. Counts of events with no ξ criterion have also been made. Many such lows occur near the equator, over certain coasts and land points with high topography (and large extrapolation to 1000 hPa) or subject to 'heat lows' in summer, and which may be unrelated to baroclinic

cyclogenesis. The vorticity criterion removes most such weak lows, particularly in the tropics, but it removes few in the oceanic storm track regions. Some odd concentrations of lows remain, in particular in central Asia, which Paciorek et al. (2002) excluded from consideration. The detection scheme is not ideal at very high latitudes, given the narrowing of grid squares, and the focus here, as for Paciorek et al. (2002), is on events within latitudes 20° and 70° in each hemisphere. Nevertheless, the scheme passes surface lows at all latitudes, except at the equator and the poles, and nearby results are not excluded from plots. Naturally, neither GCM can simulate the extremes associated with smaller-scale tropical cyclones and tornadoes.

The analysis has been performed for all simulated months during the 30-yr periods under consideration. During the cyclone detection processing, local and neighbouring values of various quantities during the standard cyclone events were accumulated or summed in each month, for further analysis. Monthly extremes of such quantities, as well as 'surface' wind speed from the lowest sigma level, were also extracted.

At each grid square, averages over the four simulations were determined for each season. The resulting maps of the occurrence index for local winter in 1961–1990 show a concentration of events in the expected storm tracks, similar to that depicted by Sinclair and Watterson (1999). The Northern Hemisphere (NH) case for December–February (DJF) is shown in Fig. 1. Following Paciorek et al. (2002), these are shown simply as numbers per grid square (without normalization by square area). There is some statistical uncertainty of values, for example at squares in the central North Atlantic storm track, specifically the region bounded by 40–20°W and 40–55°N, there are typically some 130 half-day counts in 30 seasons, with a range among the four runs of 125–145. Variations among adjacent grid squares are at least partly a result of the uncertainty remaining in the average results. Grid square shadings, rather than contours, are used in the plots for clarity. Statistical significance of changes in ensemble means can be assessed through a t -test approach, but we will mostly consider regional or zonal means for which the statistical uncertainty is much reduced.

The sum over grid squares around each latitude circle, then divided by the grid square latitudinal span, gives the number of cyclone days per degree latitude. These are shown in Fig. 2 for latitudes poleward of 20° in winter, with June–July (JJA) for the Southern Hemisphere (SH). For comparison, corresponding values for the ERA40 data set are also shown (with those for Mk3 discussed later). Following Paciorek et al. (2002), in the calculation of vorticity used in the criterion for ERA40, winds from points 5° away are used. The distributions are similar to those from Mk2, with lows particularly frequent in the sub-Antarctic trough. Numbers for ERA40 are generally larger, consistent with the higher resolution of the data. In northern latitudes, numbers are larger over ocean, but smaller over land. In fact, the numbers over land are also larger if winds from the nearest grid points are used, as this leads to higher local vorticities. Experiments

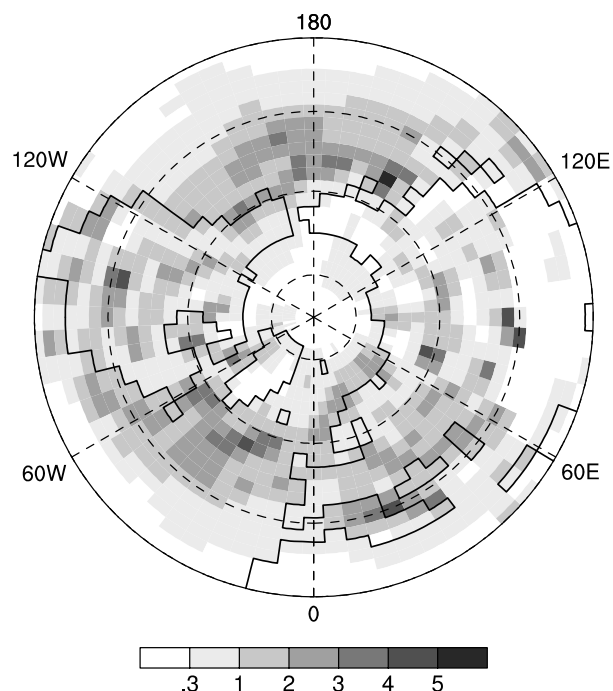


Fig. 1. Frequency of cyclone events in each grid square from Mk2 for the NH during the winter season, in days per season. Simple polar plots, bounded by 20°, are shown, including the model coastal outline.

indicate that the extremes considered shortly are not sensitive to the detection criteria, in any case.

Zonal frequencies of cyclones determined by the scheme for Mk2 for both 1961–1990 and 2071–2100, including annual totals at all latitudes, are shown in Fig. 3. Numbers are slightly smaller at most latitudes in the warmer climate, for both annual and winter cases. There is little change overall in either index at high latitudes. Total numbers, summed from grid squares over 20°N to 70°N decrease 6% in DJF, and 2% annually, while over 70°S to 20°S the decreases are 6% (JJA) and 5% (annual). Larger decreases of typically 10% occur in the mid-latitude storm track regions in winter. Some regions of increase occur at very high latitudes in both hemispheres, evidently associated with widespread decreases of climatological mean sea level pressure, which reach 4 hPa or more in winter. As the patterns of change are quite similar to those in Mk3, further discussion on the explanation of the changes is deferred.

3.2. Cyclone intensity

We turn to measures of dynamical intensity of the counted cyclone events, which will be paralleled by the analysis of rainfall intensity later. Mean values of such measures for each season and climate can be readily computed from the monthly accumulations mentioned above. Again, these are then averaged over the ensemble. A familiar measure is the central (sea level) pres-

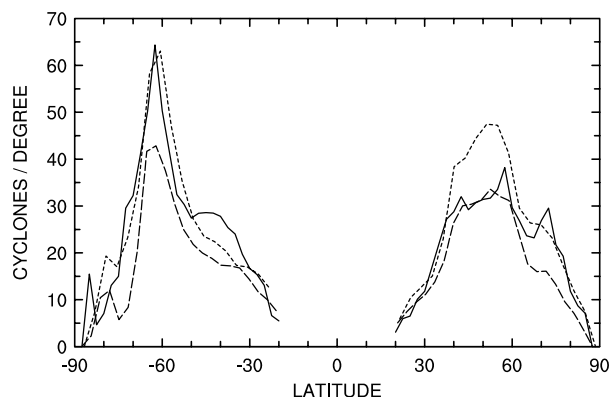


Fig. 2. Total numbers of cyclones in days per (local) winter season, per degree of latitude for 1961–1990, from Mk2 (long-dashed line), Mk3 (short-dashed line) and ERA40 (solid line).

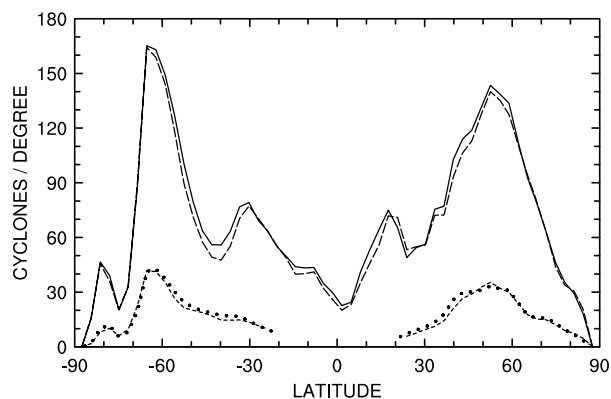


Fig. 3. Total cyclone numbers from Mk2 in days per year, per degree of latitude for 1961–1990 (solid line) and 2071–2100 (long-dashed line). The winter season values are also shown for 1961–1990 (dotted line) and 2071–2100 (short-dashed line).

sure of the low. As anticipated, the mean central pressures are deepest in mid to high latitudes in winter. Plots of changes in the central pressures appear quite similar to those of the climatological mean pressure. With respect to the intensity of storm-related winds, the central pressure relative to the climatological mean (usually a negative quantity) is likely more important, and we focus on this. The average values of relative central pressure from the model are typically -8 to -18 hPa, being larger in winter, over the storm tracks and at high latitudes, as seen in Fig. 4a. The values confirm that the surface cyclones over land, as over ocean, are of substantial depth as lows. There are certain points where numbers or depth appear affected by topography, nevertheless. The change in the average relative pressure with the warmer climate for each NH square in DJF is shown in Fig. 4b. There are few regions of increase in this measure of local intensity. Zonal average values indicate typically reduced intensity (by 0.5 hPa) in the NH in all seasons. However, weakly increased intensity occurs over the Southern Ocean (not shown). Similar comments

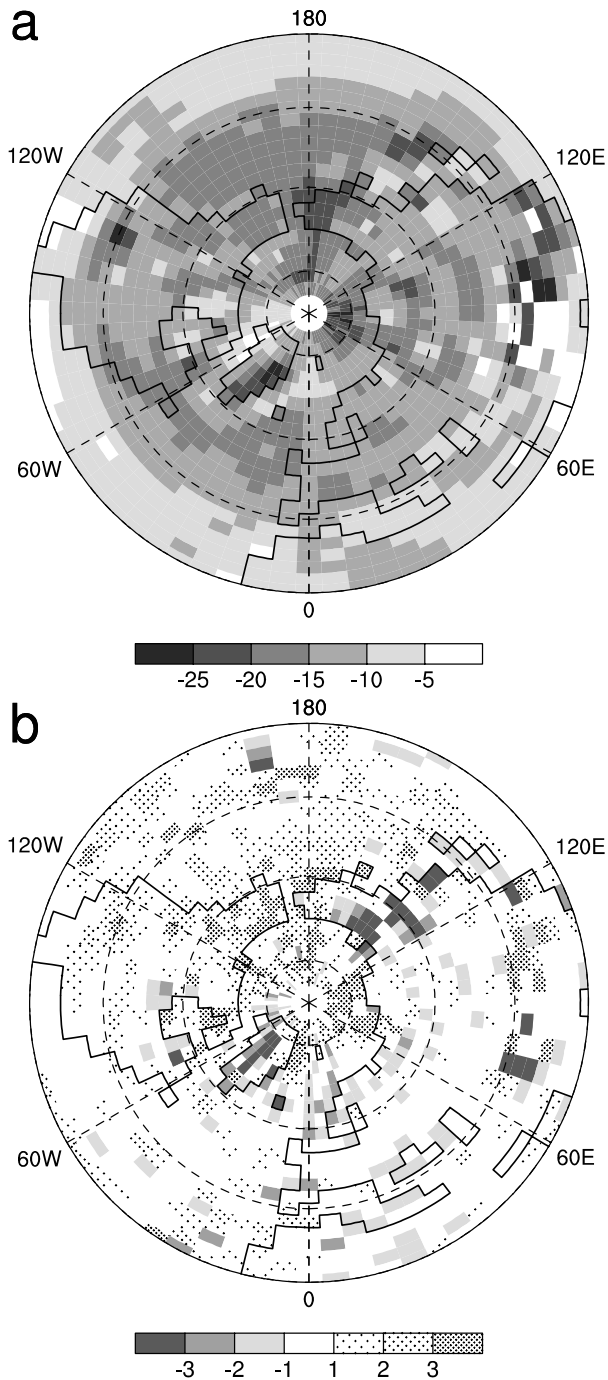


Fig. 4. Relative central pressure averaged over NH cyclones from Mk2 during DJF (in hPa): (a) 1961–1990 and (b) change to 2071–2100, with a negative value meaning deeper lows.

pertain to the measure used in the intensity criterion, the local 1000-hPa vorticity, considered further for Mk3.

A further measure of dynamical intensity is the depression of the central 1000-hPa Z relative to the minimum of the eight surrounding grid point values – the quantity tested for cyclone

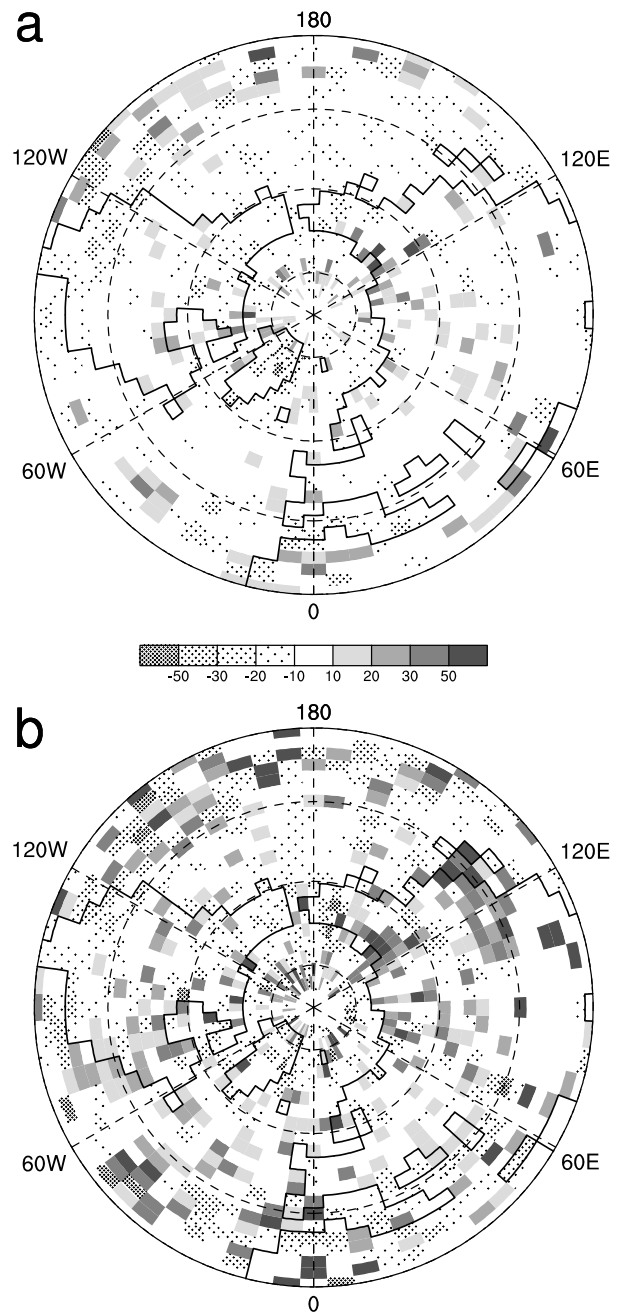


Fig. 5. Per cent changes in statistics of Z-depression for cyclones from Mk2 during DJF: (a) mean and (b) scale parameter. Grid squares with fewer than 0.1 cyclone days per season in 1961–1990 are shown unshaded, as are squares (b) where the scale is undetermined. The scale bar applies to both panels, as in subsequent figures.

occurrence. This ‘Z-depression’ is indicative, to some extent, of height gradients and hence local geostrophic wind anomalies, and of relative vorticity for deep lows. The averages are typically 8 m in mid-latitudes, and down to 3 m elsewhere. The percentage changes, shown in Fig. 5a for the NH in DJF, are small practically everywhere, consistent with previous measures. Accumulations

of the square of Z-depression have also been determined, and the variance over all cyclone days assessed. This variance also changes little.

As Z-depression is strictly positive, these two ‘moment’ statistics are worth interpreting in terms of the gamma distribution theory of Watterson and Dix (2003), briefly described in Appendix A. The ratio of the variance to the mean gives the scale parameter of the distribution, a dimensional quantity that relates to the size of values and even to the extremes in the case of a true gamma distribution (Katz, 1999). This has been determined for both climates, for each month of the year, except at grid points where there are 10 or fewer half-day events in the ensemble. Values of the scale parameter are typically a little smaller than the means. Changes in the winter month averages, shown in Fig. 5b, are mostly small on a regional basis. There is evidently a little more noisiness in this statistic than in the means. Annual, zonal averages (not shown) of the scale parameter peak in the mid-latitudes at around 7 m, and there are again only small changes, consistent with those of the previous measures. The non-dimensional shape parameter (see Appendix A) varies little with latitude or climate, around 1.2.

With respect to impacts of storms and severe winds, the extremes of the various measures over the 30-yr periods are perhaps more important than the means. These were determined simply by taking maxima from the relevant individual monthly extremes at each grid point. Having only a sample of four such extremes for each climate inevitably leads to considerable statistical uncertainty at single points, but this may be effectively reduced by spatial averaging. The zonal means of the 30-yr extreme relative central pressure and Z-depression values at each grid point are thus shown in Fig. 6. The largest mean values occur in the mid to high latitudes, except that Z-depression is constrained by the decreasing grid square width towards the poles. Mid-latitude values determined by restricting the data to the winter months are less extreme, naturally, although barely so for pressure. In general, the dynamical extremes are weaker in summer than the other seasons. Changes in both winter and all-season values are relatively small. Maps of changes in the extremes (not shown) are noisy, but also suggest little clear change in intensity. However, increases are more common than decreases in the SH. A small increase in intensity is evident around 55°S in the zonal mean results in Fig. 6. To focus on cyclones over inhabited land, percentage changes of six continental means of the 30-yr extreme fields are given in Table 1 for each of the above three measures. All changes are small. This is true as well for the extreme surface wind speed from all times (not just the counted cyclones) also given in Table 1.

Watterson and Dix (2003) found that for daily rainfall, extremes estimated from extreme value theory (see Appendix A) have less statistical uncertainty than simulated extremes. The theoretical expected values for the 30-yr extremes for Z-depression have thus been determined, as in the previous study using the gamma distribution parameters and storm numbers. The results

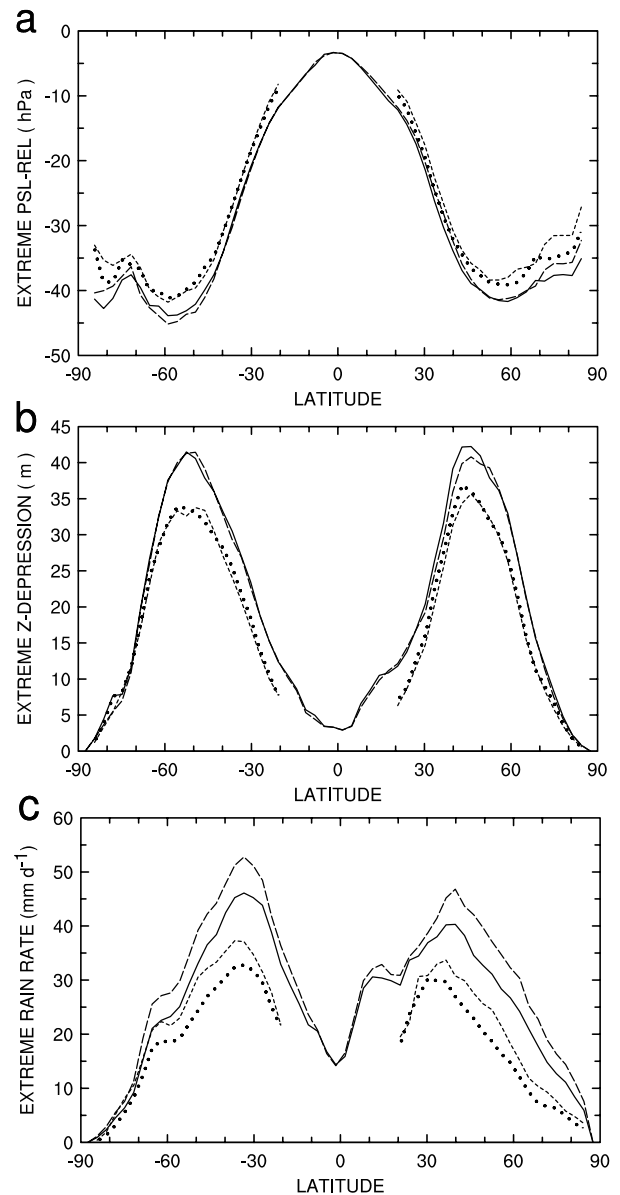


Fig. 6. Zonal means for cyclones from Mk2 of 30-yr extreme (from any season): (a) relative central pressure, (b) Z-depression, and (c) local rain rate (past 24 h) for 1961–1990 (solid line) and 2071–2100 (long-dashed line). The winter season values are also shown for 1961–1990 (dotted line) and 2071–2100 (short-dashed line).

are very similar overall to the simulated extremes for both climates (as shown for the Mk3 case, shortly). It is worthwhile then considering the percentage changes in the theoretical values at each grid point. The European region is illustrated in Fig. 7a, with the field exhibiting less noise than that from the simulated extremes. There are only a few increases in extreme Z-depression of 10% in the North Atlantic, while some small decreases occur in the Mediterranean region.

Table 1. Percentage changes in regional mean values of 30-yr extremes (from any season) from Mk2. Regions are representative of Europe, Asia, North America, Southern Africa, Australia and South America: consisting of the relevant land grid squares centred within 70°N and 20°N (30°N for Europe, which is separated from Asia by 60°E) or 60°S and 20°S. Quantities are relative central pressure (psl-rel), 1000-hPa vorticity (vort), 1000-hPa Z-depression (Z-dep), surface wind speed (wind) (all from 12-hourly data) and local rain rate from the past 24 h (rain)

	Eur.	Asia	N Am	S Af	Aust	S Am
psl-rel	2	-3	-5	-1	0	-1
vort	1	1	-1	0	1	-2
Z-dep	-1	-1	-6	-1	6	2
wind	1	1	-2	1	2	0
rain	14	13	14	-14	3	2

Summarizing from the various dynamical measures, there is no clear change to the intensity of Mk2 cyclone events, either on average or in the extremes. However, there are some changes in the pattern of occurrence, with small decreases in mid-latitude numbers being the clearest signal.

3.3. Local rainfall intensity

As a basic analysis, estimates of the current ‘daily rain rate’ at the cyclone centre, in mm per day, have been made by simply taking the coincident whole-day rain, in the case of 1200 UTC cyclones, and the average of the past and following days rain, in the 2400 UTC case. Seasonal averages over the ensemble for this quantity are evaluated as previously.

We first note that the rainfall is non-zero at nearly every cyclone centre over much of the globe. The exceptions are mostly for surface lows over subtropical land, where low relative humidity in the lower troposphere often prevents condensation at the centres. Maps of the seasonal average rain rate during cyclones (not shown) have similar patterns to those of mean rainfall from all days, but values are typically doubled.

Turning to the warmer climate of 2071–2100 under the A2 scenario, most cyclones still have non-zero rain, but there is a slight decrease in frequency of rain overall, as there is for all days (Watterson and Dix, 2003). Significant decreases in the subtropics and the NH mid-latitude land during summer coincide with decreases in low-level relative humidity. The mean rain rate during cyclones increases substantially at higher latitudes, particularly in winter as shown in Fig. 8a. The percentage changes during cyclones are similar to those for all days. As discussed by Watterson and Dix (2003), global increases in absolute humidity evidently allow heavier rain in cyclones, unless a local reduction in relative humidity counters this effect. To match the Z-depression analysis, a gamma distribution fit of the rain amounts from cyclone events with non-zero rain has also

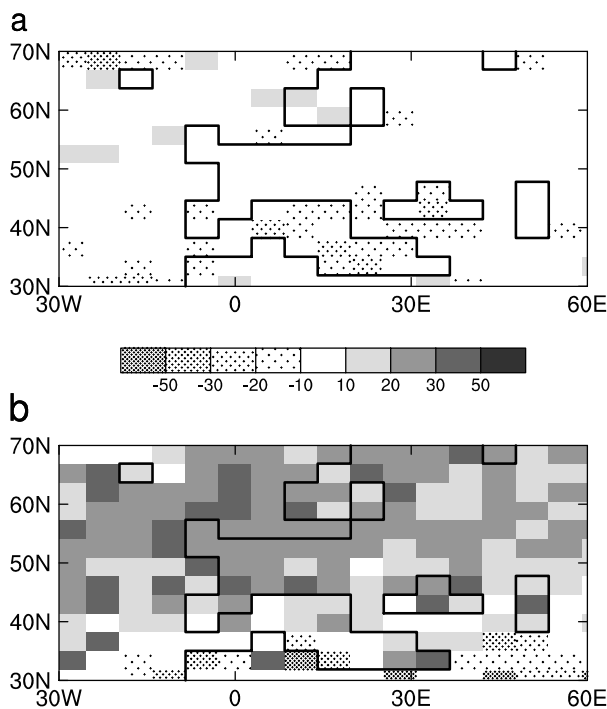


Fig. 7. Per cent change in theoretical 30-yr extremes (from any season) over Europe for cyclones from Mk2 during any season: (a) Z-depression and (b) local rain rate. Grid squares with values less than 0.1 hPa (a) or 0.1 mm d⁻¹ (b) in 1961–1990 are shown unshaded.

been undertaken. The changes in the scale parameter for the NH in DJF are again shown, in Fig. 8b. There is a fairly uniform increase in the larger rain amounts, as quantified by this parameter, much as in the all-day case. There is some reduction in the land–ocean contrast compared to the mean change. Subtropical values often increase, where mean rain per low decreases. Both mean and scale parameter changes contrast greatly with the small Z-depression changes in Fig. 5.

Consistent with the widespread increases in the scale parameter in all months, both theoretical and simulated extremes of 30-yr rain during cyclones mostly increase. Changes in theoretical values for the European region shown in Fig. 7b are mostly increases of 20%. The exception is in the subtropics where decreases can occur. We will consider the simulated extremes after looking at rainfall statistics in more detail.

3.4. Other rainfall statistics

We have seen that daily rain rates at cyclone centres mostly increase in the warmer climate, consistent with the increase in absolute humidity. The dynamical intensity of cyclones changes little, however, which suggests that this intensity is not strongly influenced by the changes in latent heating associated with rainfall in the model. It is evident, though, that the daily rain rate centred on the cyclone both in space and time may not be

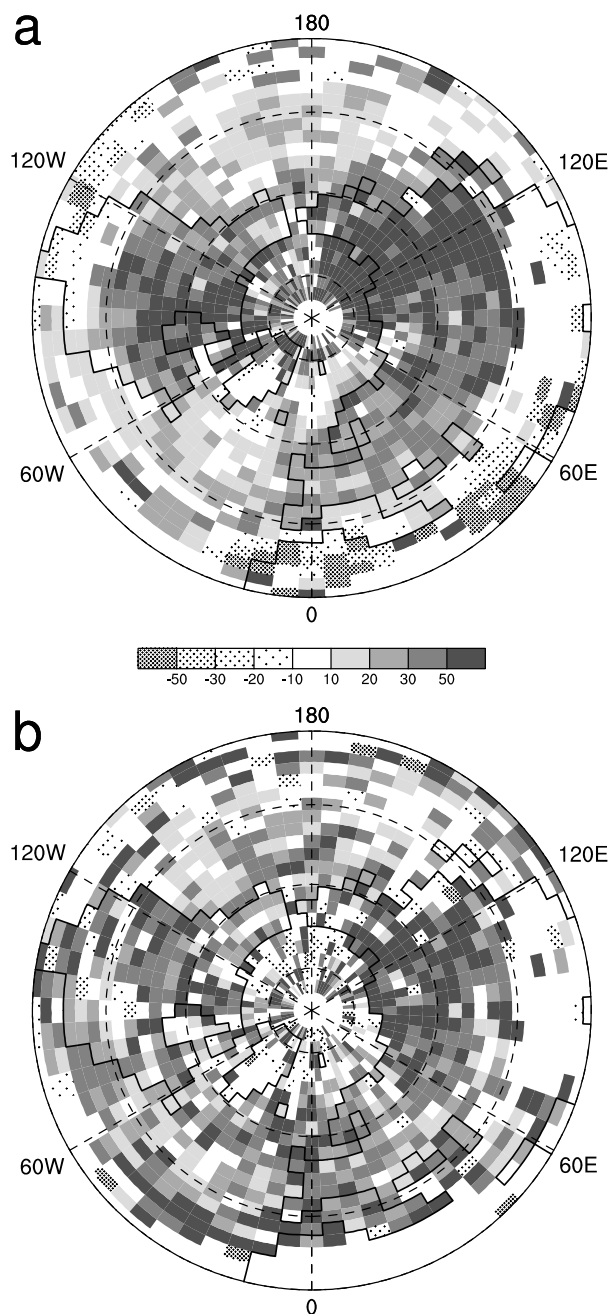


Fig. 8. Per cent changes in statistics of local rain rate for NH cyclones from Mk2 in DJF: (a) mean and (b) scale parameter. Grid squares with fewer than 0.1 cyclone days per season in 1961–1990 are shown unshaded, as are squares (b) where the scale is undetermined.

representative of latent heating during the intensification of the storms. The specification of the relevant heating is likely to depend on the region, so it is difficult to address in this global study. However, a brief consideration of case studies in the literature suggests that this heating would often be encompassed by rainfall during the previous 24 h and within Mk2 grid squares

Table 2. Rain rates and changes for grid squares local and adjacent to cyclone centres from Mk2 during DJF averaged over 20°N to 70°N. The rows refer to relative latitude number, and the columns to relative longitude number, for each statistic

	Rain (mm d^{-1})			Change (%)		
	–1	0	+1	–1	0	+1
+1	3.0	5.9	7.2	12	13	17
0	3.3	6.0	6.6	10	14	17
–1	3.1	4.3	4.3	9	14	16

adjacent to the cyclone centre. Given the limitations of the data set available, we present only some related means and extremes.

Consider first rain rates at the grid squares adjacent to the cyclone centre, averaged over the 1200 and 2400 UTC events as previously. Over much of the globe, rainfall tends to be larger in the squares eastward of the centre. Away from the tropics and particularly in winter, it also tends to be larger in the squares poleward of the centre. This is true of the NH winter averages, given in Table 2. These variations are physically reasonable, given the likelihood of the poleward moving air to the east of a centre being warmer and more humid than the air to the west. This is also consistent with the poleward flux of moisture associated with transient motions in general. Turning to the change in rain rates from the 1961–1990 period to the 2071–2100 period, we find that on average rainfall in all eight squares surrounding the centres increases (Table 2). The percentage increases in the rainfall in the squares to the east tends to be greater than that of the centre, and those to the west less.

Given the typical movement of lows in the mid-latitudes, eastwards at 10 m s^{-1} or about one grid square width in 12 h (Sinclair and Watterson, 1999), prior rainfall at the centre is likely to be included in the value to the west. The smaller percentage increase there may indicate more modest increases in latent heating during cyclone development than those implied by the central values (e.g. Fig. 8a). However, the average increase is still significant. Further, similar increases are seen for values in the centre of the North Atlantic storm track where cyclone intensification may be expected.

Another indication of variation in rain rates over the cyclone life cycle can be obtained by considering the 24-h rain for the 1200 and 2400 UTC storms separately. The 2400 UTC case represents entirely rain up to this time (the following day's rain is not used here), while the 1200 UTC case includes rainfall after the time of cyclone detection. The former is thus likely to be more representative of rainfall during intensification (although we have not been able to exclude rain from decaying storms). The 2400 UTC values would also contain more rain to the east of an eastward moving centre. Consistent with this, the winter NH average rates for 2400 UTC are a little larger than those for

Table 3. Rain rates (in mm d^{-1}) and changes (%) at cyclones during DJF averaged over 20°N to 70°N , from both Mk2 and Mk3. Mean refers to averages over all cyclones, Extreme to extremes from 30 seasons of cyclones. C refers to the 24-h rainfall centred on cyclones at 1200 UTC, and P to rain during the previous 24 h for cyclones at 2400 UTC

	Mk2		Mk3	
	Rain	Change	Rain	Change
Mean-C	6.5	14	7.0	11
Mean-P	8.3	16	7.7	13
Extreme-C	18.2	14	18.0	13
Extreme-P	22.3	14	19.9	15

1200 UTC, as seen in Table 3. On going to the warmer climate, however, the percentage increases in the two cases are very similar. These comparisons are true of the 30-winter extreme daily rain in cyclones, also in Table 3.

With the 2400 UTC extremes being larger we now focus on these. The zonal means of all-season and also winter extreme values from the 2400 UTC data in both climates are shown in Fig. 6c. Increases of around 15% occur at all latitudes, except in the subtropics. Again, these contrast with the very small dynamical intensity changes, a weak exception being those of the Southern Ocean (Figs. 6a and b). Finally, the changes in continental averages of all-season extreme rain are given in Table 1. For the northern regions, the changes are again large. The southern cases are more subtropical and have only small increases, and even decreases for southern Africa, as with the theoretical values for far North Africa shown in Fig. 7b. We will return to this issue in considering the higher-resolution GCM.

4. Analysis of Mk3 data

4.1. Cyclone analysis

The full analysis has been applied to the Mk3 reduced-grid data set, and we consider averages over the nine control and three warm climate periods. When a surface low is detected at a reduced-grid point, the actual minimum may have occurred in an adjacent point of the full grid. We will nevertheless assume that the count represents lows over an augmented grid square based on the reduced grid. The distribution of cyclone events in the control winter season of each hemisphere is shown in Fig. 9, with zonal totals included in Fig. 2. The simulated locations of the wintertime oceanic storm tracks appear broadly realistic. Consistent with the higher resolution of the model, numbers exceed those in Mk2. They are similar to those in ERA40 over the ocean, although a little low in the South Atlantic and in the Australian region. The 3.6° spacing of winds used in the vorticity calculation partly explains a larger number over land. The

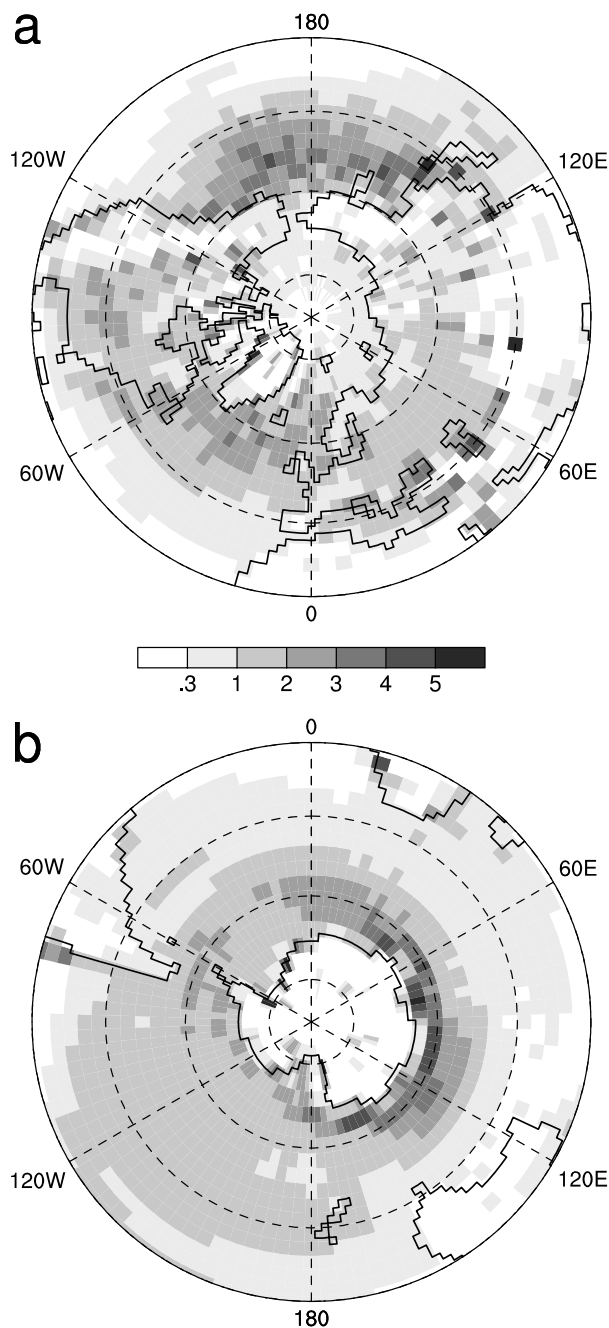


Fig. 9. Frequency of cyclone events in each square (of the reduced grid) from Mk3 during the local winter season (in days per season): (a) NH in DJF; (b) SH in JJA.

winds are also on average weaker in ERA40 over land in these 1000-hPa data sets. Presumably, model vertical resolution and interpolation methods play a role.

Percentage differences in occurrence on going to the warmer climate are shown in Fig. 10. As with Mk2, there are broad decreases in the mid-latitudes in both hemispheres. Total numbers over 20°N to 70°N decrease 6% in DJF, and 4% annually, and

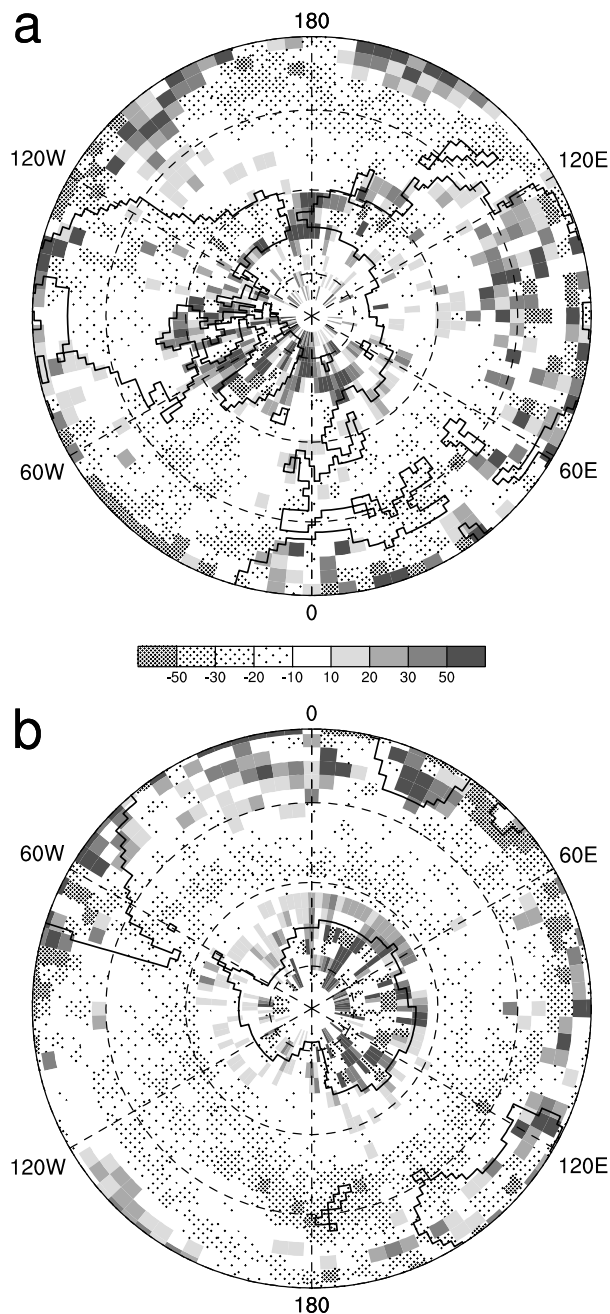


Fig. 10. Per cent change in number of cyclone days from Mk3 during the winter season: (a) NH in DJF; (b) SH in JJA.

over 70°S to 20°S the decreases are 8% (JJA) and 6% (annual). Numbers in the centre of the North Atlantic storm track decrease 20% in DJF and 9% annually. These changes appear largely consistent with broad decreases during all seasons in mid-latitude baroclinicity, as quantified by the mean Eady growth rate evaluated from the 12-hourly data at 500 hPa following Paciorek et al. (2002). Annual means of the growth rate (all in unit per day) drop from 0.68 to 0.63 at 45°N, and 0.74 to 0.67 at 45°S.

More specifically to the NH winter storm tracks, the Mk3 DJF mean field is very similar to that for observational data shown by Paciorek et al. (2002), with peaks in growth rate on the upstream side of the Pacific and Atlantic storm tracks. The Atlantic peak drops some 10%, although the Pacific change is smaller. In the SH winter, a band of increase in growth rate matches the increased numbers in the South Atlantic at 30°S. At high latitudes, regions of increase in numbers of lows such as at Hudson Bay (Fig. 10a) seem explained by lowered climatological pressure, specifically in the Arctic and sub-Antarctic where pressure often decreases 3 hPa or more. Likewise, pressure increases some 2 hPa in all seasons around 45°S, where there are fewer surface lows. Both growth rate increases and mean pressure decreases in winter over north-eastern Asia around 40°N, so the increased numbers there may not be merely topographic effects.

4.2. Means of dynamical and rainfall intensity

The same suite of quantities averaged over cyclone events for Mk2 is available for the Mk3 data set. To a large extent the results are similar, despite the greater resolution of the GCM. The presentation highlights the contrasting dynamical and rainfall intensities. With vorticity used in the storm intensity criterion, its average is only a little larger in Mk3 than in Mk2; however, other quantities can differ more. As a difference between heights two Gaussian grid points away in the Mk3 analysis, rather than one, Z-depression is apparently less constrained and control mean values are rather larger than in Mk2. However, changes remain small overall (not shown).

The average relative central pressures for the NH in winter are a little deeper for Mk3 than those for Mk2 in Fig. 4a. The percentage changes, shown in Fig. 11a, are again generally small. There is some agreement with Mk2 in increased intensity over parts of Europe (the shading here has comparable meaning with that in Fig. 4b). Changes in mean local daily rainfall rate, shown in Fig. 11b, are positive outside the subtropics. As for Mk2, mid-latitude increases are somewhat larger over land than over ocean. This matches the mean warming, which in Mk3 is typically 4 K or more over land and less than 2 K over ocean. A small region of cooling south of Greenland corresponds to no change in mean rainfall intensity. Values in the central Atlantic storm track increase typically 15%, nevertheless. The percentage changes for the SH in winter are shown in Fig. 12. Again, there is a substantial contrast between pressure and rainfall changes south of 40°S. Average rain rates during lows decrease over most land at lower latitudes, however.

The analysis of mean rainfall rates in grid squares adjacent to lows yielded rather similar results to those for Mk2. There is a subtle difference in that the rain data are for a smaller Gaussian grid area than in Mk2, but cyclone centres are likely to move through several squares within the 24-h accumulation period. Mean central rates are broadly similar to Mk2 – the overall precipitation rate is a little lower in Mk3 than Mk2. For local rainfall,

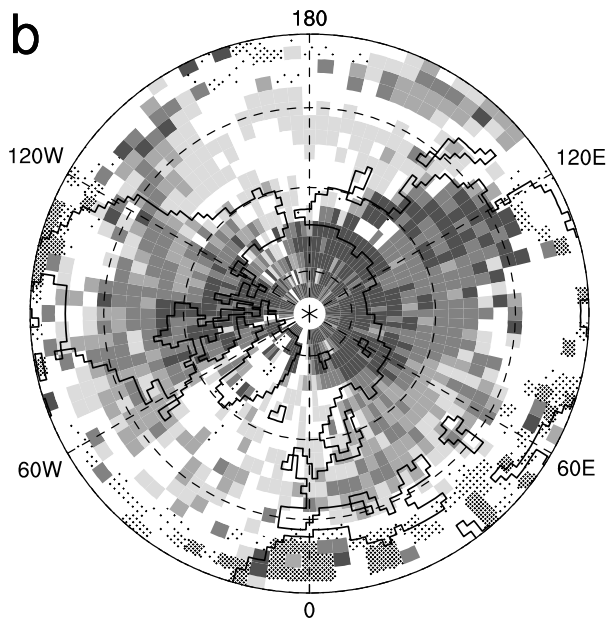
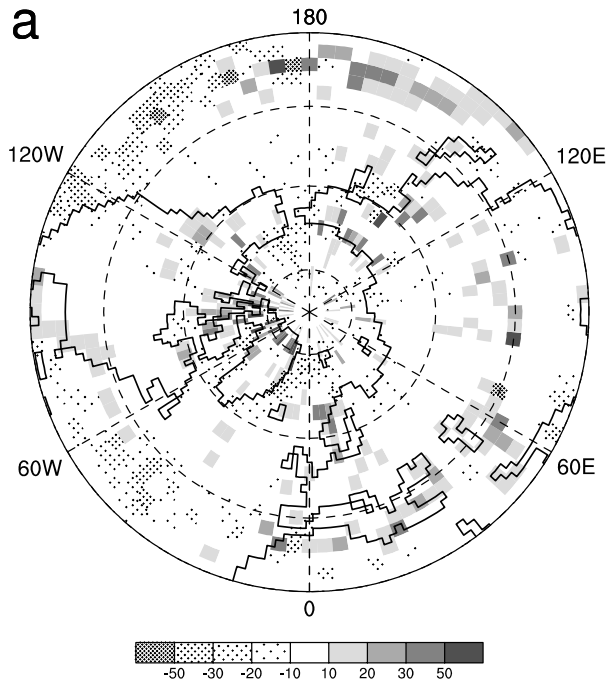


Fig. 11. Per cent changes for cyclones from Mk3 in NH during DJF: (a) average relative central pressure; (b) average local rain rate. While the values are from data on the reduced Gaussian grid, shading is extended to cover the reduced-grid squares. Squares with fewer than 0.1 cyclone days per season in the control are shown unshaded.

at the centres of the surface low, the partition of daily values centred or prior to the low was again made. A similar contrast was obtained, and the NH winter means are included in Table 3 for comparison with Mk2. The percentage increases are similar to Mk2, particularly if scaled by the mean warming (Section 2).

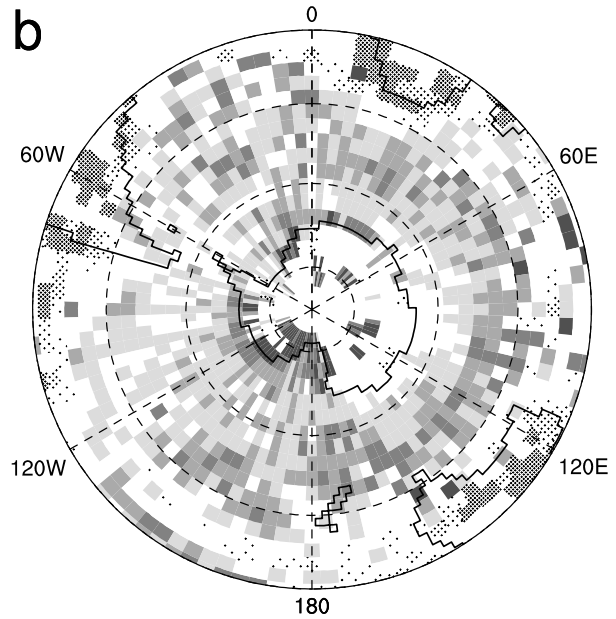
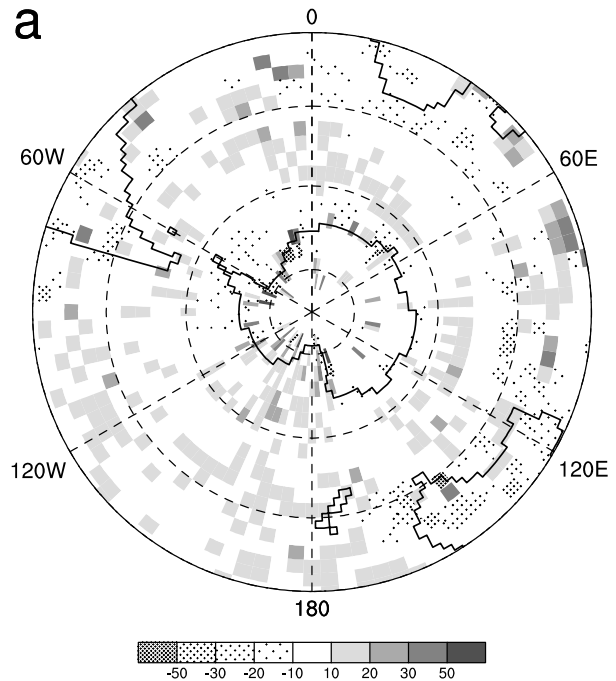


Fig. 12. As Fig. 11, but for SH in JJA.

Before turning to extremes, we note that the Mk3 gamma distribution parameters and their changes also have much in common with Mk2. There is some indication of heavier intense rain events in warmer conditions, with the scale parameter for local rain being larger in Mk3 in the tropics and in summer. The central Atlantic scale parameter for DJF increases 20%, again consistent with less spatial contrast in the change in the scale compared to the mean.

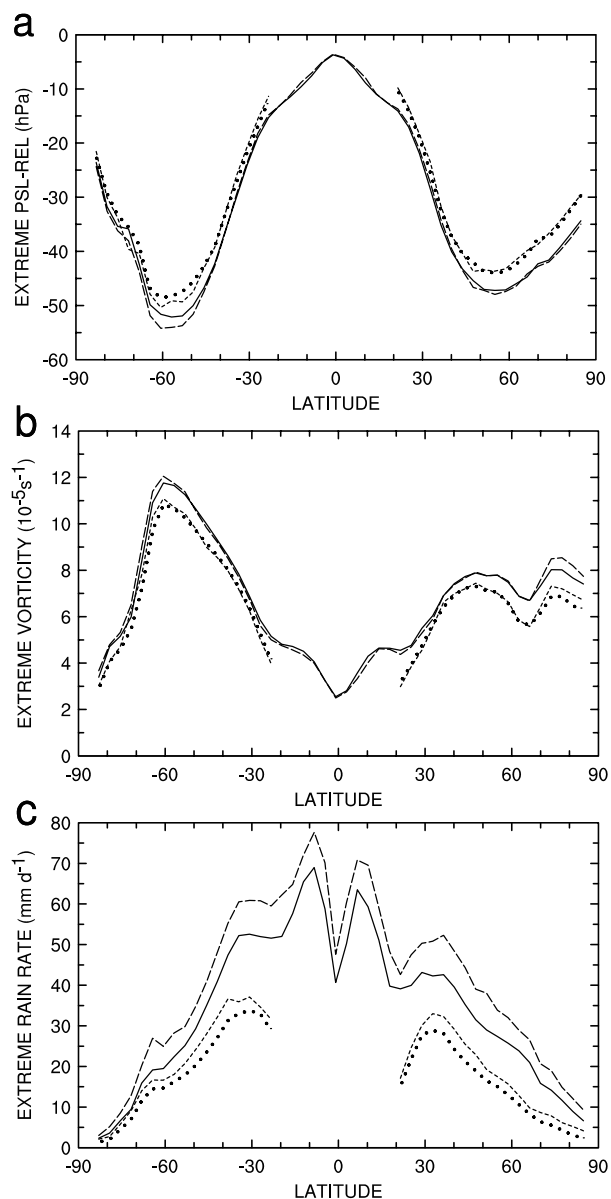


Fig. 13. Zonal means for cyclones from Mk3 of 30-yr extremes (in any season) of (a) relative central pressure, (b) 1000-hPa vorticity and (c) local rain rate for control (solid line) and 2071–2100 (long-dashed line). The winter season values are also shown for control (dotted line) and 2071–2100 (short-dashed line).

4.3. Extremes associated with cyclones

We consider first zonal means of extreme central pressure relative to climatology, shown in Fig. 13a. These reach a greater intensity than in Mk2, as expected from the improved resolution. (As will be illustrated shortly, the extreme at an individual point is influenced by the number of events occurring there, but the actual numbers of lows per point on the data grid are mostly similar in the two GCMs.) Changes on going to the warmer climate periods

are on average small. There is some enhancement of the modest deepening over the Southern Ocean seen in Mk2. The change at 55°S for JJA is 5%.

With vorticity being a field that can be expected to be more sensitive to changes at smaller scales than pressure, is it worth focusing on this field in the higher-resolution model. The 30-season extremes for DJF, averaged over the nine-member control ensemble, are shown for the NH in Fig. 14, together with the noisier single-period ERA40 field. The ERA40 values here were calculated using 1000-hPa winds from the nearest grid points. The largest values occur in the oceanic storm tracks, with the Mk3 result being of similar magnitude to ERA40 over the Atlantic. The tendency for the ERA40 1000-hPa winds to be smaller over land is evident in the comparison. Overall, the model result is encouragingly realistic, for the resolved length-scale. The changes from global warming (Fig. 14c) are generally small with no mid-latitude regional trend, except for north-eastern Asia. Central Atlantic values increase only 2% on average. Applying a *t*-test indicates that there are few points where the 80% confidence level for a non-zero change is reached, the average magnitude of such a change being 30%. However, these appear to occur merely at the rate expected by chance. The lack of change holds for the mid-latitude zonal means in both winter and annual cases shown in Fig. 13b. Small increases in intensity are seen at very high latitudes.

Average extremes of local rainfall, centred and prior to the lows, are given for the NH winter in Table 3. The percentage increases given are similar to Mk2. The prior 24-h rain values from 2400 UTC lows are again used for the zonal mean results shown in Fig. 13c. The mid- and high-latitude values are similar to Mk2 for both climates. Tropical values are substantially larger in Mk3. In part this is due to a greater number of lows counted, but also to the greater intensity in Mk3 of the parametrized convective rain that dominates there.

Continental averages of extremes associated with cyclones have been determined, including those of Z-depression and surface winds from all times. The percentage increases are given in Table 4. As for Mk2, there is a sharp contrast between a lack of change in dynamical extremes and increases in the rainfall extremes. This extends to the southern continents, which being more subtropical have extremes largely determined by warmer season values and convective rain.

Results from an analysis of the full-grid data set of pressure and daily rain are also included in Table 4. For this analysis surface lows were determined from the pressure field, with no vorticity criterion. Central pressure, 'pressure depression' (determined as for Z-depression), and rain per low were evaluated. With deep lows occurring around a quarter as often for these grid points as for those in the reduced-grid analysis, for comparable extremes the squares were first combined into two by two sets. Zonal means of the resulting extremes are very similar to those in Fig. 13. The continental means change by similar percentages. The pressure depression result, which tests gradients at the

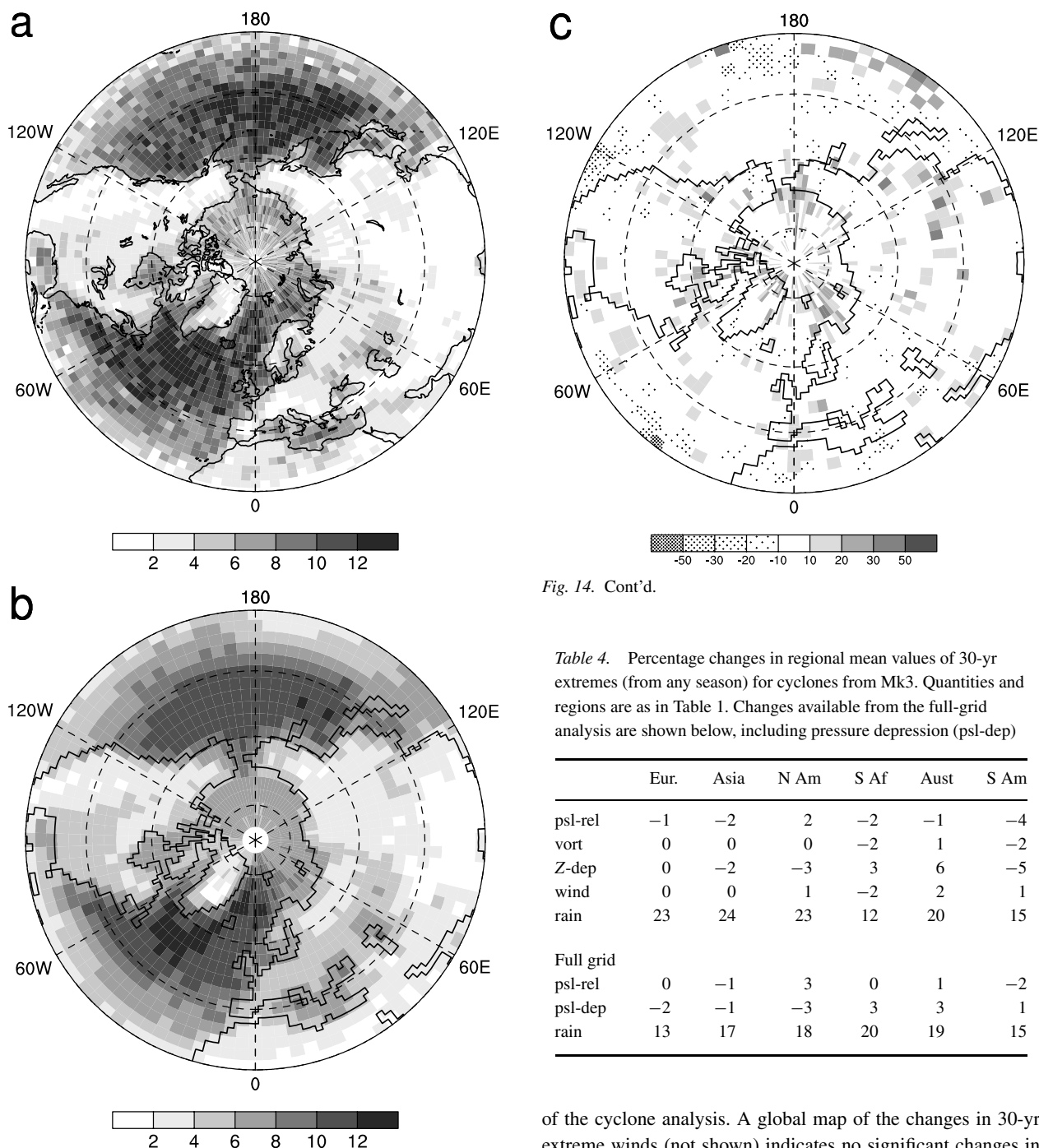


Fig. 14. Extreme 1000-hPa vorticity at cyclones in NH from 30 winters, unit 10^{-5} s^{-1} : (a) from ERA40 reanalyses for 1961–1990, (b) simulated by Mk3 (control), and (c) per cent change from Mk3, with grid squares with fewer than 0.1 cyclone days per season in the control shown unshaded.

smallest scale from the Mk3, suggests that there is no signal of change missing from the original reduced-grid analysis. In any case, the extreme surface winds considered in Table 4 are simulated values from the Gaussian grid points sampled, independent

Fig. 14. Cont'd.

Table 4. Percentage changes in regional mean values of 30-yr extremes (from any season) for cyclones from Mk3. Quantities and regions are as in Table 1. Changes available from the full-grid analysis are shown below, including pressure depression (psl-dep)

	Eur.	Asia	N Am	S Af	Aust	S Am
psl-rel	−1	−2	2	−2	−1	−4
vort	0	0	0	−2	1	−2
Z-dep	0	−2	−3	3	6	−5
wind	0	0	1	−2	2	1
rain	23	24	23	12	20	15
Full grid						
psl-rel	0	−1	3	0	1	−2
psl-dep	−2	−1	−3	3	3	1
rain	13	17	18	20	19	15

of the cyclone analysis. A global map of the changes in 30-yr extreme winds (not shown) indicates no significant changes in wind speed over land, except an increase of 10% in tropical north-western Australia. Global mean speed increases 0.8% but this results largely from increases over the Southern and Arctic Oceans, reaching 6% for zonal averages at 55°S and 8% at 75°N .

Finally, we consider extreme value theory applied to the Z-depression values from Mk3. The zonal mean of the simulated 30-yr extremes, shown in Fig. 15, are larger than in Mk2 (Fig. 6b). However, there is little change in the warmer climate (not shown), even at 55°S . Zonal means of the theoretical 30-yr extremes

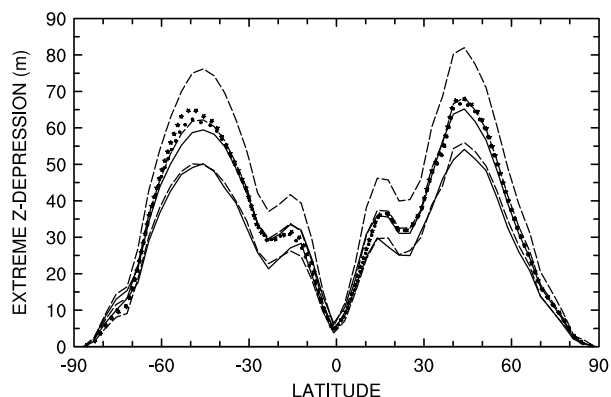


Fig. 15. Zonal means for cyclones from Mk3 of extremes (in any season) of Z-depression. Solid lines are simulated extremes from 10- and 30-yr periods of the control climate (naturally, increasing with period length). Dashed lines are theoretical extremes for the control climate for 10-, 30- and 100-yr periods. Dots show theoretical values for the warmer climate, while stars are those when the gamma parameters are changed, but the event frequency remains that of the control climate.

are also shown. These are very close to the simulated values, being slightly larger in mid-latitudes. Warmer climate values are again very similar. The theory can then be used to test the effect of the decreased numbers of cyclones. Recalculating the values for the warmer climate but with storm numbers fixed at the control values (Fig. 15) does confirm an effect of numbers on extremes. The largest change is 5% at around 55°S, where the annual numbers decrease 25%. A decrease of 3% in the central North Atlantic extremes can be inferred to result from the decrease in numbers alone. It is likely that extremes of other storm-related quantities are reduced similarly. The effect of an even larger decrease in numbers on extremes can be seen from the simulation, if the extremes are taken from only decadal periods (see Fig. 15). The theory provides a good match for this period. It indicates the potential for 100-yr extremes, also shown, to significantly exceed those from 30-yr.

5. Discussion

In both global models considered here, there are widespread increases from 1961–1990 to 2071–2100 in both mean and extreme rainfall intensities associated with extratropical surface lows, particularly over NH land. This provides further support for the expectation of greater magnitudes of floods under global warming. However, there is little change overall in the dynamical intensities of the simulated cyclones. This tends to discount a direct link between such intensities in the GCMs. There is, of course, a clear relationship between cyclones and rainfall, as evidenced by the average daily rain at or near cyclone centres being often three times the mean rain from all days (e.g. the Mk2 win-

ter averages in Table 3 compared with the climatological value of 2.5 mm d^{-1}).

A relationship between rainfall and dynamical intensities can also be sought from regression analyses. Considering spatial variation between the cyclone mean fields from Mk2, we have for example a correlation $r = 0.36$ between rain rate and Z-depression for 20°–70°N in winter from 1961–1990. However, for fields of percentage changes between these quantities (see Figs. 5a and 8a), we have only $r = 0.14$. Such values are typical of other rainfall and dynamical mean and extreme measures, other seasons and the SH. Whatever leads to the present climate relations is evidently not sufficiently strong to maintain an important link in the changes.

Temporal relationships between daily rain rate and Z-depression at each grid point can be seen from an extension of the previous moment analyses using additional covariance statistics. In the winter case, the correlations determined from these statistics are moderate, with $r > 0.3$ over most regions with frequent rain, in particular the oceanic storm tracks. While this is true of the warmer climate also, the corresponding linear regression coefficient is different. For example, with Z-depression fitted by $a + bP$ (P being rain), the average b over the region with $r > 0.3$ decreases by 18%. Thus, the relationship changes in such a way that P can increase on average but the Z-depression does not change. Again, comparable effects occur for other cases.

One cannot necessarily infer cause and effect from such analyses, especially given the modest correlations that result, and lags in space and time are not considered. Nevertheless, in both spatial and temporal analyses, the relationship is largely consistent with rain intensities responding to the dynamical intensity of storms. This response is larger when there is more atmospheric moisture, including the effect of global warming. This can explain the contrast between the two intensities in the overall changes.

While our analysis cannot specifically target the latent heating during cyclone intensification, it does suggest that latent heating associated with precipitation during cyclones is enhanced in the warmer climate. However, any widespread enhancement of extreme dynamical intensity due to it is small enough to be countered by other effects. It is quite possible that the process is simply not sufficiently resolved by even the higher-resolution GCM. Effects that are resolved include a reduction in baroclinicity, rises in mid-latitude pressure, and the statistical effect of a smaller number of lows, illustrated by the extreme value theory. An apparent exception is in the Southern Ocean around 55°S, where extreme relative central pressure magnitudes increase by an average 5% in Mk3, and a little less in Mk2, while extreme cyclone rain increases some 20%. These percentage increases appear comparable to those for the simulations of high-latitude storms reported by Kaas et al. (2001). In the North Atlantic, storm track dynamical extremes are barely changed, with decreased numbers likely to have countered any effects of a more modest rainfall increase.

Of course, neither GCM studied here is able to simulate the most intense winds and rain rates associated with cyclones, even extratropical ones. Downscaling from such resolved grid scale extremes to those applicable locally is a major topic, which is beyond the present study. To the extent that Mk3 produces similar contrasts in rain and wind changes to Mk2, this modest dynamical downscaling suggests little increased intensity of cyclones. Closer analyses of life cycles of many individual storms would be needed to confirm whether latent heating associated with intensification really does increase at the rate suggested by the present study. Simulations with a very high-resolution model may be needed to explore its effect on extreme intensity at local scales.

6. Conclusions

Extratropical cyclones simulated by two CSIRO GCMs, and the rain rates associated with them, are analysed to assess changes in storm intensities under global warming, and the potential role of increased latent heating. The simple surface cyclone detection scheme applied here produces broadly realistic distributions of events in the ensemble of simulations of the 1961–1990 climate by Mark 2. By 2071–2100, with warming of 3.5 K under the A2 forcing scenario, numbers decrease by a few per cent in the mid-latitudes. Some high-latitude locations show a small increase in numbers, mostly where mean pressure decreases. There is little change overall in various measures of dynamical storm intensity, including relative central pressure and height depression (relative to surrounding values), either in the mean or in 30-yr extremes.

Mean rain rates at cyclone centres are usually several times the climatological rain rate, with rates at points to the east and poleward usually larger again. The local rain rate at centres increases by typically 15% at most latitudes, and over 30% for northern high-latitude land in winter. There are some decreases in the dry subtropics, associated with decreasing relative humidity. Extreme rain rates also increase broadly. Theoretical extremes based on the gamma distribution for daily rain show similar increases, contrasting a lack of change in the height depression extremes. Rain rates adjacent to the cyclones and those in the prior 24 h also increase.

The new Mark 3 GCM, with higher resolution, produces broadly similar changes for a global warming of 3.1 K (relative to a control run climate). Rain rates associated with subtropical surface lows increase more strongly, but there is again little evidence of changed dynamical intensity of storms. A weak exception to this is around 55°S, where increases of 5% in extreme relative pressure are seen.

In both models, rain rates clearly respond to the presence of storms and to their dynamical intensity. This response is stronger when there is greater atmospheric moisture, including the effect of global warming. This is consistent with the contrast between the changes in the rain rates and dynamical quantities. The enhanced latent heating associated with the anomalous rain rates

may have some influence on the simulated storm intensities, but this does not seem strong enough in either GCM to have a clear effect on storm intensity overall, except perhaps in offsetting the statistical effect of smaller numbers of events on 30-yr extremes.

However, these rain and pressure intensities, even in the higher-resolution model, are much smaller than intensities observed at smaller scales local to real storms. Much higher resolution models may be needed to assess convincingly whether the anticipated moisture increases of around 25% can lead to significant latent heat enhancement of extreme cyclone intensities.

7. Acknowledgments

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8. Appendix A: Extreme Value Theory

The gamma distribution has often been used to approximate distributions of daily rain amounts z on rain days (e. g. Katz, 1999; Semenov and Bengtsson, 2002). Its density function is given by

$$f(z) = [(z/\beta)^{\alpha-1} e^{-z/\beta}] / [\beta \Gamma(\alpha)],$$

where α is the shape parameter, β is the scale parameter (in the units of z), and Γ denotes the gamma function. We denote the (cumulative) distribution function by F .

Watterson and Dix (2003) found that with regard to more extreme amounts, fitting the parameters by moments was effective, with the mean intensity equated to $\alpha \beta$ and the variance to $\alpha \beta^2$. The approach is applied to both rain rate and Z -depression during cyclone events in this paper, for each grid point and with a separate distribution for each of the 12 months of the year and each ensemble. For rain, cyclones with non-zero local rain are counted as events.

Katz (1999) showed that the distribution of the extreme value from a large number of samples N taken from a gamma distribution is the Gumbel distribution with distribution function

$$G(z) = \exp\{-e^{[-(z-u)/\lambda]}\}.$$

The parameters u and λ are solutions of

$$1 - F(u) = 1/N \quad \text{and} \quad 1/\lambda = Nf(u).$$

In applying this to the various monthly cases here, N can be written as the total number of possible independent event times in a period, multiplied by the frequency of event occurrence, averaged over each ensemble. For Mk2, the results are found to be

typically a little closer to the simulated extremes assuming daily events are independent. For Mk3, assuming 12-hourly events gives a better match for Z-depression, possibly because storms are less likely to persist on the smaller grid points. In any case, with regard to percentage changes the sampling frequency makes negligible difference.

Following Watterson and Dix (2003), the extreme from any month, the annual case, has a distribution given by the product of the 12 monthly G (ignoring months with no events or undetermined parameters). The expected or mean value of this distribution is evaluated numerically.

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