Will extra-tropical storms intensify in a warmer climate?

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Abstract

Extra-tropical cyclones and how they may change in a warmer climate have been investigated in detail with a high-resolution version of the ECHAM5 global climate model. A spectral resolution of T213 (63 km) is used for two 32 year periods at the end of the 20th and 21st centuries, and integrated for the IPCC A1B scenario. Extremes of pressure, vorticity, wind and precipitation associated with the cyclones are investigated and compared with a lower resolution simulation. Comparison with observations of extreme wind speeds indicates that the model reproduces realistic values.

We also investigate the ability of the model to simulate extra-tropical cyclones by computing composites of intense storms and contrasting them with the same composites from ERA40. Composites of the time evolution of intense cyclones are reproduced with great fidelity, in particular the evolution of central surface pressure is almost exactly replicated, but vorticity, maximum wind speed and precipitation are higher in the model. Spatial composites also show that the distribution of pressure, winds and precipitation at different stages of the cyclone lifecycle compare well with those from ERA40, as does the vertical structure.

For the 21st century, changes in the distribution of storms are very similar to those of our previous study (Bengtsson et al., 2006). There is a small reduction in the number of cyclones but no significant changes in the extremes of wind and vorticity in both hemispheres. There are larger regional changes in agreement with previous studies.

The largest changes are in the total precipitation, where a significant increase is seen. Cumulative precipitation along the tracks of the cyclones increases by some 11% per track, or about twice the increase in global precipitation, while the extreme precipitation is close to the globally averaged increase in column water vapour (some 27%). Regionally, changes in extreme precipitation are even higher due to changes in the storm tracks.
1. Introduction

Experiments with high-resolution models indicate that tropical cyclones are likely to intensify in a warmer climate (Meehl et al., 2007), most importantly in terms of their winds and precipitation, though the number of storms will likely be fewer (Oouchi et al., 2006, Bengtsson et al., 2007). As suggested in Bengtsson et al., (2007) this dichotomy is likely related to the fact that water vapor increases more rapidly than precipitation in a warming climate. This then implies a weakening in the large-scale vertical mass flux (Held and Soden, 2006) thus providing less favorable conditions for the onset of tropical cyclones. However, when favorable conditions do occur they make use of the increased humidity. This may also happen for other convectively driven systems (Allen and Soden, 2007).

Changes in extremes of extra-tropical cyclones with climate warming are less clear (Meehl et al., 2007a). Extra-tropical cyclones form and grow via baroclinic instability, getting their main kinetic energy from the conversion of available potential energy, though some contribution may also come from latent heat release. Available potential energy is proportional to the variance of temperature in the troposphere. This is why extra-tropical cyclones are more intense during winter when the temperature variance is highest. Climate models integrated with higher concentrations of greenhouse gases generally show a reduced temperature gradient in the lower troposphere of the Northern Hemisphere (NH) at least during winter because of a stronger Arctic warming. This implies that the available potential energy will decrease. With this reasoning extra-tropical cyclones might be more intense in a colder climate. There is support for this from both observational reconstructions (Björck and Clemmensen 2004) and recent model simulations of the 17th century replicating the climate conditions during the so-called Maunder Minimum (Raible et al. 2007) at a time when the European winter climate was generally much colder than during the 20th century.

It is important to be clear what we mean by the intensity of extra-tropical storms. In many studies this is taken to be the depth of the pressure center or the magnitude of the vorticity.
However, pressure may be misleading due to changes at the larger scale whilst the parameters that matter to the public are winds and precipitation; these are the focus of this work.

To determine any observational changes in extremes of winds and precipitation, associated with cyclones, is difficult since observational records only exist for restricted regions. For winds this is mainly confined to the Northeast Atlantic and the North Sea (WASA, 1998 and references therein). A detailed analysis of these observations (Weisse et al. 2005) does not show any long-term trends but indicate considerable variability at a wide range of time scales. Even for observational reconstructions that go back to the latter part of the 19th century it is impossible to identify any persistent trend but instead periods when intense winds were more common than in other periods. (see Trenberth et al., 2007, fig 3.41). There have been only limited regional studies of changes in extreme precipitation. For example, Karl and Knight (1998) investigated the contribution of the upper 10% percentile of daily precipitation and noted a positive trend during the 20th century for an area averaged over the central part of the United States. Other studies that report observed increases in precipitation for several regions are summarized in the Climate Change Science Program (CCSP) report (CCSP, 2008, Ch. 2). Model studies by Semenov and Bengtsson (2002) show a similar increasing trends.

It has been suggested that higher Sea Surface Temperatures (SST) at middle and high latitudes will imply an intensification of extra-tropical cyclones but this is hardly likely at least in winter. The reason is that higher SST will act as a sink on available potential energy because of a more rapid warming of the cold air than the warmer air and consequently imply a reduction of the available potential energy. Rather it is more likely that it is the strengthening of the SST gradients that may favor extra-tropical cyclones (Inatsu et al., 2003). As discussed by Bengtsson et al., (2006) such changes are expected as a consequence of a future warming in the Southern Hemisphere (SH) Ocean and also over limited areas of the North Atlantic. In the North Atlantic this appears to be related to the weakening of the thermo-haline circulation. This leads to the generation of an area south of Greenland with a regionally reduced SST warming and in some coupled models a minor
cooling. The enhanced SST gradient south of the cooler region will increase the baroclinicity in the lower troposphere creating, in principle, more favorable conditions for extra-tropical cyclones.

An additional factor is tropical cyclones and meso-cyclones. Tropical cyclones can be transformed into extra-tropical storms as they migrate polewards. As tropical storms generally are intensified in a warmer climate (Oouchi et al., 2006, Bengtsson et al., 2007) this may contribute to an enhancement of storms even at higher latitudes but mainly in late summer and autumn. Meso-cyclones such as polar lows are also found in the extra-tropics. Polar lows are found at high latitudes during the winter season and develop over open water in situations of cold Arctic air masses. We might expect polar lows to be less common in a warmer climate but this may be influenced by the reduction of Arctic sea ice and they could be more common over the Arctic Ocean. Meso-cyclones, as well the convective storms that occur during the warm period months, may well be more likely to be affected by changes in latent heat release than synoptic scale systems.

There have been many studies of the impacts of a warmer climate on extra-tropical cyclones (Ulbrich et al., 2008). The most recent studies have been based on the climate scenarios (Nakicenovic et al., 2000) used for the Intergovernmental Panel on Climate Change (IPCC), 4th assessment report (AR4) (IPCC, 2007). Lambert and Fyfe (2006) assessed cyclones in a range of climate model data archived for AR4 in the Coupled Model Intercomparison Project 3 (CMIP3) archive (Meehl et al., 2007b) for a number of the scenarios. Because of CMIP3 archiving limitations their study was limited to identifying surface pressure minima in daily averaged data. Extremes were taken to be lower than 970 hPa (960 hPa) for the NH (SH). They found that the models simulated a reduction in the total number of events and an increase in the number of intense events. With their methods they did not find any changes in the location of the storm tracks and did not present any results for winds or precipitation. Leckebusch and Ulbrich (2004) investigated the relationship between cyclones and extreme windstorms over Europe under climate change using data from the HadCM3 model with a horizontal latitude-longitude resolution of 2.5° x 3.75° and the IPCC A2 and B2 scenarios (Nakicenovic et al., 2000). They compared the period 2070-2099 with
the control period 1960-1989. Extreme wind events were defined as values above the 95th percentile for daily maximum wind speeds associated with the nearest cyclone. For both scenarios, they found an enhancement of cyclone activity over the areas around the British Isles and the North Sea and a weakening in the northern part of the Norwegian Sea.

In the study of Bengtsson et al. (2006) an ensemble of three 30-year integrations of the Max Planck Institute (MPI) coupled ECHAM5/OM model (T63 spectral resolution in the atmosphere) was used to compare storm tracks for the A1B scenario for the period 2070-2099 with the control period 1960-1989. Their evaluation, based on vorticity, was for the whole globe. Their results showed a poleward shift of the extra-tropical storm tracks in both hemispheres in agreement with Yin (2005) and other Lagrangian studies, e.g. Teng et al., (2008). There was no indication of any intensification, except regional changes associated with the poleward shift. These results are similar to those obtained by Löttien et al., (2008) using the same data but a different Lagrangian analysis system. In this respect these results are just the opposite of that of Lambert and Fyfe (2006).

Evidence from such studies are summarized in the latest IPCC report (Meehl et al., 2007a, page 751) as the statement: “Model projections show fewer mid-latitude storms averaged over each hemisphere, associated with the poleward shift of the storm tracks that is particularly notable in the SH, with lower central pressures for these poleward-shifting storms. The increased wind speeds result in more extreme wave heights in those regions.” Though a bit imprecise, this statement suggests that mid-latitude cyclones will become more intense in a warmer climate.

There are several reasons why we believe it is important to reassess the way extra-tropical cyclones may be affected by a warmer climate. Firstly, as shown by WASA (1998), wind speed is not strongly correlated with lower surface pressure of a transient cyclone. Secondly, the most recent global studies of extra-tropical cyclones with AR4 models have been undertaken using data sets of coarse horizontal resolution and thirdly the data sets have suffered from a coarse time resolution making it virtually impossible to identify rapid changes in the wind speed.
Here we revisit the previous study by Bengtsson et al (2006) for extra-tropical cyclones, but use the ECHAM5 atmospheric model at the higher spectral resolution of T213. The integrations cover the period 1959-1990 and 2069-2100 using the IPCC scenario A1B (Nakicenovic et al., 2000). For a detailed description see Bengtsson et al (2007). The increased resolution makes it possible to obtain a better determination of the characteristics of intense extra-tropical storms including associated extreme winds and precipitation. We concentrate this assessment on the winter (December-February (DJF)) season of the NH but will provide some general comments on the other seasons and the SH.

The following scientific objectives will be addressed in this study:

1. What is the impact of model resolution on extra-tropical cyclones?
2. What is the typical structure and lifecycle of extra-tropical cyclones at T213 resolution in terms of characteristics such as the depth, vorticity, wind speed and precipitation? How realistically do these results agree with those from re-analyses?
3. How do intense extra-tropical cyclones respond to climate warming and what regional and seasonal differences are found?

The paper continues in section 2 with a description of the experiments, data and analysis methodologies. In section 3 a comparison between storm properties from the T63 and T213 integrations are presented. In section 4, the composite structure and evolution of the most extreme extra-tropical cyclones are compared with those from reanalyses. Section 5 describes the change in the extra-tropical storms at the end of the 21st century and is contrasted with grid point statistics of extremes. In section 6 we summarize the results of the study and in section 7 some additional comments are made.

2. The experiment, data and analysis methodologies.

The experimental setup for this study is essentially the same as that used for the study of tropical cyclones by Bengtsson et al (2007). The MPI, ECHAM5 atmosphere model (Roeckner et al. 2003) is integrated at T213 spectral resolution (63 km) with 31 levels in the vertical using the ‘time-slice’
method. This is comparable to the horizontal resolution of several limited area models. This is used to simulate the climate of two 32 year periods that are representative of the end of the 20th (1959-1990) and 21st (2069-2100) centuries, using the IPCC scenario A1B. These will hereafter be referred to as 20C and 21C respectively. To provide the boundary forcing for these integrations, SST and sea ice fraction data from one of the three T63 (208 km), ECHAM5/OM coupled model integrations for the A1B scenario produced for the IPCC AR4 (Roeckner et al., 2006) are used. The SST and sea ice fraction data are interpolated from the original T63 resolution to T213. These are repeat integrations of the ones used by Bengtsson et al (2007) to correct an error in the specification of sea ice fraction. This error has little impact on tropical cyclones but could have an impact on extra-tropical cyclones, motivating the new integrations. The results from the new integration have been compared with those of the T63 coupled integration from which the SST were taken. Results are also contrasted with the European Centre for Medium Range Weather Forecasts (ECMWF) 40-year reanalysis (ERA40) (Uppala et al., 2005) which uses a T159 resolution (125 km, linear grid) to provide verification that the model is realistic in its simulation of extra-tropical cyclones in terms of their distribution and properties, including their 3-dimensional (3D) structure and lifecycle. A limited period of data from the new ECMWF Interim reanalysis is also used for verification. This has a higher resolution than ERA40 of T255 (80 km linear grid) and uses a 4-D Var data assimilation system. Only 10 years of data are currently available (1989-1998). Reanalyses provide our best 4D view of the atmosphere though they are strongly dependent on the observations and how they are assimilated (Bengtsson et al., 2004). Inter-comparing reanalyses provides some degree of the level uncertainty. Inter-comparing cyclones between different reanalyses suggests that in the NH the reanalyses are reasonably well constrained with differences mainly associated with resolution, however in the SH there is considerably less agreement (Bromwich et al., 2007, Wang et al., 2005, Hodges et al., 2003).

The identification and tracking of the extra-tropical cyclones follows closely the study of Bengtsson et al (2006). The approach makes use of the 850hPa relative vorticity field ($\zeta_{850}$) to
identify and track the storms, the benefits of which are discussed in Hoskins and Hodges (2002). To enable reliable tracking of the storms the vorticity field is reduced in resolution to T42 and smoothed before identifying the cyclones, (see Hoskins and Hodges, 2002, for details). The T42 resolution also means that the identification is performed at a common spatial scale and thus allows a fairer comparison to be made between the T63 and T213 integrations. All seasons are considered but the emphasis of this study is on the NH winter activity.

Since the emphasis of this paper is on changes in extremes of winds and precipitation associated with extra-tropical cyclones it is not necessary to consider every identifiable storm. In the past, post-tracking filters have been applied to only retain cyclones with particular properties. For example, Bengtsson et al (2006) only considered storms that lasted for more than 2 days and traveled further than a 1000km whilst other studies have used different filters to retain or exclude more storms. We have retained the 2 day/1000km filter and tested that this has little impact on the study.

To see the full benefit of higher resolution and its impact on storm properties the full resolution properties must be extracted from the data along the storm trajectories. In the study of tropical cyclones of Bengtsson et al (2007) this was done using a B-spline interpolation and optimization approach. However, because of the more asymmetrical nature of extra-tropical cyclones we have employed additional techniques. For the determination of pressure minima, associated with the vorticity tracks, we apply the original approach where the local minimization of the Mean Sea Level Pressure (MSLP) occurs within some prescribed radius of the storm center, whilst for winds and vorticity we just do a simple search for the maximum value within the search radius (minimum value of vorticity in the SH). For variables, such as total precipitation, the area average is computed over the prescribed radius. The chosen storm centered regions are spherical caps, with areas computed accordingly. The size of the chosen regions is somewhat arbitrary and several tests have been performed with different sizes. The wind and precipitation fields of extra-tropical cyclones can extend over quite large areas, if the trailing frontal systems are included, so it might be expected
that a region with large radius would be needed to capture their full extent. After exploring different radii, a $5^\circ$ spherical arc radius was found to be adequate for capturing the extremes. Ideally the sampling region should depend on the storm size. Whilst some previous studies have attempted to estimate this, based on using MSLP or geopotential, it still requires subjective decisions as to what constitutes the storm size. It is also likely to depend on the field used, i.e. storms in vorticity are likely to look smaller than in MSLP.

We have also examined the typical lifecycle and structure of storms using composites of a number of selected extreme storms. Lifecycle composites are determined from the single value properties along the selected tracks, centered relative to the time of maximum intensity, in the T42 $\xi_{850}$, and averaged. Regional composites are produced by extracting the regions around the storm centers on a radial grid and averaging these for particular stages of the storm lifecycle. The radial grid is defined on a spherical region to prevent the kind of distortion that occurs when using projections. Additionally, the grid is aligned with the propagation direction of each individual storm by rotating the grid so that the grid preferred direction is aligned with the storm direction at each time step. Since the storm structure depends on its propagation the consistent orientation of the storms will reduce the ‘smearing’ of the composites. Full details can be found in the Appendix of Bengtsson et al (2007). Both types of composites are compared with the corresponding composites computed from extreme storms selected in the same way from the ERA40 data for the satellite period (1979-2002).

In addition to examining the properties of the storms at single levels we have also examined their full 3D vertical structure and vertical tilts. This is done by first determining the storm centers iteratively using the relative vorticity fields at a number of levels up to the tropopause (see Bengtsson et al., 2007). Each additional vorticity field is treated in exactly the same way as for the tracking. For the tilts the centers at each level are projected onto the storm direction (in spherical geometry) and the position relative to the 850hPa center determined as a geodesic angle. The composite lifecycles are then computed as before. The multi-level centers are also used to produce
the 3D composites, taking the storm vertical tilt fully into account by extending the regional composites through the depth of the troposphere.

3. Climatology and comparison with T63 and observations.

The storm climatology for the T213, 20C integration for the NH winter is shown for track density overlaid with mean intensity and genesis density in Figure 1b and d respectively. These can be contrasted with the T63 coupled integration (Figure 1a and b) for the same period from which the SST originate for the T213 integration. Figure 1 shows that there is a strong correspondence between the distribution of storms at the two resolutions as well as their genesis distributions. These can also be compared with the same distributions computed from ERA40 and shown in Figure 2a and c of Bengtsson et al (2006). This indicates that the T213 and T63 integrations show quite realistic distributions of track and genesis density. The variability of storms in the ECHAM5 model, with respect to El Niño–Southern Oscillation (ENSO) forcing, was explored by Bengtsson et al., (2006) who found the correct signal was reproduced when compared with ERA40 though somewhat to strong. We do not expect this to be any different with the T213 model as the identically same SST have been used. The biggest differences between T213 and T63 are seen in the mean intensities computed from the full resolution $\xi_{850}$ values, these show the T213 mean intensities to be roughly twice as large as those at T63. In terms of the average number of storms, during the period, the two integrations are fairly similar with 135 storms per month, excluding the tropics (north of 25°) for T63 and 138 for T213. The fact that the identified numbers of storms for the T63 and T213 integrations are so similar might at first appear strange. Previous studies using cyclone tracking have suggested that more storms are identified with increasing resolution (Blender and Schubert, 2000; Jung et al, 2006). However, if the identification is made at different resolutions then different spatial scales will be identified with smaller scale storms identified at higher resolution (Blender and Schubert, 2000).

To indicate how the resolution impacts on a range of measures of intensity, Figure 2 shows the full resolution, maximum intensity distributions for both the T63 and T213 storms, during the NH
DJF for 20C, using MSLP, $\xi_{\Delta 850}$, 925hPa wind speed and area averaged total precipitation. We have tended to use the 925hPa winds throughout this paper as a measure of near surface winds because they represent the wind field above the surface boundary layer and are specifically calculated by the model. They are a more robust quantity than the model 10m winds which are diagnostically determined. The intensity measures are constructed by finding the maximum/minimum value of the relevant parameter, determined within the $5^\circ$ degree (geodesic) region, along each track and binning the values. Figure 2 shows that the T213 maximum intensities indicate consistently more intense storms than those for T63 for all measures. Comparison with ERA40 for the satellite period (not shown), shows that the ERA40 distributions lie consistently between that of the T63 and T213 distributions for these variables.

We have also contrasted the distribution of intense cyclones, in terms of their winds, between the Pacific and the Atlantic Ocean and between ECHAM5, T213 and ERA40 and summarize the results in Table 1. As can be seen there is almost twice as many intense cyclones (>50m/s) during DJF in the Atlantic than in the Pacific for the model. If we adjust for the generally weaker winds in ERA40 (>45 m/s) the result is similar. We suggest that the contrast between Atlantic and Pacific is mainly due to enhancements of the wind speed around southern Greenland as well as exposure to cold air masses from Greenland and Northern Canada (and enhanced available potential energy) in the northern Atlantic. The interaction of synoptic scale systems with the steep orography of Greenland and the impact this has on the wind field has been studied by Moore and Renfrew (2005) using observations from the QuickSCAT satellite. Comparison of these results with those of the T213 model shows that the model at high resolution is capable of simulating these winds much more realistically than the T63 model. Tsukernik et al. (2007) in a similar tracking study have also highlighted the complex interactions that can occur between developing synoptic systems and Greenland.

Though we prefer to use winds that are computed directly by the model, we have explored the 10m wind field to determine how realistic the values are at selected sites compared with
observations. We have selected the Ekofisk oil platform in the central North Sea (56.5N, 3.2E) as being representative. This allows us to compare with the results presented by Weisse et al., 2005 (their figure 2) who used down-scaling from reanalyses with a limited area model over Europe at 50 km resolution to compare with observations. Figure 3 shows the DJF 10m wind speed for the 90, 95, 99 and 99.5 percentiles for the nearest T213 model grid point to the Ekofisk station compared with ERA40 and the observations. The observations have been sampled to the same 6 hourly frequency as the model and ERA40 data. The model results agree well, in terms of magnitudes, with both observations and Weisse et al (2005), whereas the ERA40 10m winds are in general less than those of the model and observations. This further provides us with confidence in the T213 model’s capability to produce realistic results of the properties of extra-tropical cyclones both qualitatively and quantitatively.

4. **Structure of extra-tropical storms.**

We have shown that the model is capable of producing credible spatial distributions of extra-tropical cyclones and their properties. Here we show that the model is also capable of simulating storms with a credible structure and evolution in time and space. This provides further confidence in the model when used to assess changes in extra-tropical cyclones in a warmer climate. To study this we have explored the lifecycle and structure of the extra-tropical storms in the T213 model and contrasted them with those in ERA40 for the NH winter. Since the emphasis of the study is on intense cyclones we have restricted the analysis to composites of the 100 most intense cyclones indentified from the T42, $\xi_{850}$. Any of the other intensity measures can be used as a means of selection, though they would not necessarily result in the same 100 storms. The T42, $\xi_{850}$ is used for the selection as it is the field used for the identification and tracking, is less noisy than the full resolution vorticity and has a good correspondence with the minimum pressure in the lifecycle of the storms. To ensure there is sufficient overlap in the evolution of the storms a further lifetime threshold is imposed so that storms must last for longer than 4 days. There may be more intense storms but because of their shorter lifetimes would make the composites noisy if used.
For the lifecycle compositing, each storm is centred on the time at which it reaches its maximum intensity in the T42, $\zeta_{850}$ and we use the same variables as previously discussed in section 3, for both ECHAM5, 20C and ERA40 for the satellite period. These are shown in Figure 4a. For both ECHAM5 and ERA40 the maximum pressure deepening rate can be seen to occur some 30 hours prior to when the minimum pressure occurs. We note that the maximum in the full resolution vorticity occurs prior to the minimum surface pressure by about 12 hours. This is to be expected as the scale of intense extra-tropical cyclones is such that the wind field is likely to lead the mass field in the geostrophic adjustment process (Temperton, 1973). The time scale for this adjustment is broadly proportional to the inverse of the Coriolis force (Cahn, 1945). The precipitation rapidly increases during the development of the cyclone reaching its maximum some 12 to 18 hours before the cyclone reaches its minimum pressure. At that time the precipitation is in rapid decay as the cyclone occludes and the mechanism behind the lifting of the air is collapsing. In fact the rate of diminishing precipitation is faster than the rate of increase.

It is apparent that the composite lifecycle shows that the rate of decay of the cyclone is nearly as large as the rate of intensification. This is not due to Reynolds stresses as these are much too weak, in particular over the oceans where the strongest extra-tropical cyclones occur Charney (1959, p188). Instead the rapid reduction of the cyclone’s intensity is due to the divergence of geopotential height fluxes (Orlanski and Katzfey, 1991).

The lifecycle composites are in close agreement in all respects between the model and ERA40 with the main difference that wind speed, vorticity and precipitation are less intense in ERA40. The deepening rate (as measured in surface pressure) is practically identical reaching a maximum value of 1.5 hPa/hr and clearly puts these storms into the class of “bombs” (Sanders and Gyakum, 1980). This is a robust parameter as observations of surface pressure are assimilated in ERA40. The difference in precipitation is some 20% less in ERA40. We suggest this is mainly due to spin up effects in the ERA40 as it represents precipitation during the first 6 hours of the forecast. Other possible contributing factors to the weaker intensities in ERA40 are the lower horizontal resolution
and the use of a 3D Var data assimilation system. A preliminary assessment of cyclones in the new ECMWF Interim re-analysis has also been made. Since only 10 years are available to produce lifecycle composites we have only taken the 50 most intense storms for the NH, DJF. To contrast these with ERA40 we have identified the exact same storms using a matching approach (Hodges et al., 2003) and produced the same composites. The Interim-ERA40 comparison is shown in Figure 4b. This shows similar results to those in Figure 4a except that the Interim has greater intensities for all variables than ERA40. The implication of this is that reanalyses at resolutions comparable to the T213 model, and using modern data assimilation methods, will be closer to the model. In fact reanalyses may under-predict intensities due to the data assimilation. Hence, we believe that the model is not per se over-predicting intensities (c. f. Figure 3).

The composite horizontal structure of the same 100 most intense storms in the model at different stages of their lifecycle are shown in Figure 5, with the composite storm moving to the right. Figure 5a and b show the composite structure of the cyclone in terms of the 925hPa wind field, total precipitation, MSLP and mean temperature at the time of maximum intensification, in terms T42, $\xi_{850}$. A similar result is obtained using maximum intensification in surface pressure. As the composites are of 100 cyclones, they are somewhat smoothed out. The rapidly developing composite cyclone has a broad open warm sector where the strongest winds occur (Figure 5a). The highest precipitation (Figure 5b) occurs ahead of the centre in the region of strong ascent as is typical for extra-tropical cyclones. Figure 5c and d shows the structure of the cyclone at the time of maximum precipitation. This occurs on average some 15 hours after the time of maximum intensification. The wind field has strengthened and is orientated slightly differently with the wind maximum moving in an anti-clockwise direction. Wind speeds above 25ms$^{-1}$ can now be seen practically all around the centre of the cyclone. The area of maximum precipitation has higher values and extends further around the front of the cyclone. Finally in Figure 5e and f we show the state of the cyclone at maximum intensity. The strongest winds are found behind and to the right of the cyclone in the area after the occluding cold front where maximum subsidence normally occurs.
The field of precipitation is significantly weaker and is further distorted and extended around the cyclone in an anti-clockwise direction. The same diagnostics are shown for ERA40 in Figure 6. This shows a similar progression in the wind and precipitation fields with the development of the composite storm to those for ECHAM5. However, as has already been noted the ERA40 values of wind and precipitation are generally lower, though the winds at maximum intensity are similar to those in ECHAM5. The MSLP shows an almost identical evolution and values to those for ECHAM5.

We have also explored the 3D structure of storms and their tilts, as described in section 2, using the same 100 storms. The 3D composites (not shown) are explored for the same stages of the storm lifecycle as in Figure 5 and Figure 6 and they confirm that the model is capable of reproducing similar structures for the different stages as ERA40. A more detailed discussion is left to a future publication. The tilts are shown in Figure 7a and b for ECHAM5 and ERA40 respectively, for up to 36 hours either side of the maximum intensity. The tilt gradually increases as the composite storm grows (not shown) until it reaches a maximum $\sim 4^0$ between 850 and 200hPa, roughly coincident with the maximum growth of the composite. After this it can be seen that there is a rapid change in the tilt occurring at the time when the storm has reached its maximum intensity and becomes barotropic. This rapid change to a basically barotropic state occurs over a period $\sim 12$ hours and coincides with the rapid transition of the composite storm as discussed above. In fact the tilt can actually reverse as the storm occludes as seen in both Figure 7a and b. This behavior is typical of the lifecycle of an unstable baroclinic wave (Ogura, 1957). Both the 3D composites and tilts indicate a good correspondence between ECHAM5 and ERA40.

The composite storm results discussed in this section indicate that the model is realistic at simulating the baroclinic lifecycle of extra-tropical cyclones when compared with storms identified in a modern reanalysis. This together with the results from section 3 provides us with confidence in using the model to explore extremes and their likely changes in a warmer climate. Note, that although the ECHAM and ECMWF models have a common origin they have over time almost
totally diverged in terms of dynamical core and parametrizations, so these favorable comparisons are not due to any commonality between the models.

5. Effect of climate warming

The main focus of this part of the study will be on the more serious aspects of extra-tropical cyclones, namely extreme wind and precipitation, and any changes in a warmer climate. In the previous section we concentrated on the analysis of the NH winter, here we extend the evaluation to all seasons and to both hemispheres. We first provide some general remarks on the changes in storms between 20C and 21C.

Compared to the previous study (Bengtsson et al., 2006) the changes in extra-tropical storm tracks between 20C and 21C are very similar (not shown), with a poleward migration in both hemispheres, but most marked in the SH. Some differences between the T213 and T63 results do occur, however, we judge these differences to be within the variability of the T63 ensemble and longer observational data sets. Also consistent with the previous study the number of cyclones decreases by ~3-5% between 20C and 21C. The reason for this is not clear, but could be due to a similar effect as we noted for the reduction in tropical cyclones, namely the general weakening of the vertical mass flux (Bengtsson et al., 2007). Other explanations could be the reduced temperature variance in the lower atmosphere. In general, the extreme extra-tropical cyclones have deeper central pressures in 21C than 20C (Figure 8). Our interpretation of this is that this is mainly related to the poleward movement of the storm tracks. As the large scale surrounding pressure is generally lower this will reduce the central surface pressure of the cyclones but does not enhance the wind speed (see Figure 10), which is related to surface pressure gradients. The fact is that the maximum wind speed occurs when individual cyclones reach their lowest pressure. Other measures of the storm intensity can also be considered such as the T42 and full resolution vorticities. The T42 show similar distributions to those shown in Bengtsson et al, (2006), in particular the intensification of the more extreme systems in the SH during JJA and SON as the storm track moves poleward for 21C. The full resolution vorticity results are broadly consistent with the T42 results though the
intensification in the SH is less obvious. Although differences, on the hemispheric regions so far discussed, indicate little change in the storm characteristics between 20C and 21C (apart from MSLP, associated with larger scale changes), larger differences are apparent on more regional scales as previously shown by Bengtsson et al, (2006). Larger differences are also apparent for precipitation (see later).

Contrasting the lifecycle composites of the 100 most intense extra-tropical cyclones, identified via the T42 $\xi_{850}$ intensities, in 21C with those in 20C for NH winter (Figure 9a), there is hardly any difference in vorticity with the peak wind speed slightly weaker. The minimum surface pressure on the other hand is a few hPa deeper and the peak precipitation is some 6% higher. In spite of the higher precipitation there is little effect on the maximum wind speed. The fact that the higher precipitation has limited impact on the intensity of the storms in terms of their wind speeds is emphasized if we redo the composite calculation but choose the 100 most intense precipitating storms, making sure not to include any storms of tropical origin, which behave dynamically differently. The results of this calculation, shown in Figure 9b, show a greater increase in the peak level of precipitation ($\sim$10%) in 21C compared to 20C but in terms of MSLP, vorticity and winds the composite storm is actually weaker in 21C. We suggest that the limited feedback of the increased precipitation on the intensity of the storms with respect to wind speeds etc is due to the distribution of the precipitation around the cyclone. This is not organized in such a way as to enhance the wind via a geostrophic effect. Moreover the change in the structure of the storm is very fast, in particular for precipitation, with the high precipitation decaying very rapidly as the storm occludes, the decay appears more rapid in 21C. The distribution of the composite wind patterns, precipitation and surface pressure, at different stages of the storm lifecycle for 21C (not shown) are virtually identical to what we have shown for 20C (Figure 5).

In the following sub-sections we will consider the wind and precipitation in more detail. As well as exploring the winds and precipitation associated directly with the storms we also explore these in grid point space by calculating the higher percentiles in the range 95-99.9 using the 6 hourly data.
This then gives the local change in extreme winds at every single point. Wind speed is the instantaneous value while precipitation is the accumulated value over six hours. It is possible that more extreme conditions might occur within a six-hour interval but we believe this is not important here as the objective is to compare conditions between 20C and 21C. Furthermore, the instantaneous precipitation is not a robust quantity. We have explored the extreme winds identified by both the storm centered and percentile methods and have convinced ourselves that there is a close association between the two and that the most extreme winds occur around the time when individual cyclones reach their maximum intensity. See Figure 4 and Figure 5.

5.1. Wind field

We investigate the wind at 925 hPa. As the main purpose of the study is to compare extreme winds in 20C and 21C, we think this is a more reliable approach as this is a variable predicted by the model. It is also less influenced by the boundary layer and thus a more reliable predictor. Based on a wind regression relation between 925hPa and 10m winds determined from ERA40 data we have found that 35ms$^{-1}$ at 925 hPa is approximately equivalent to 10 Beaufort for the wind at 10m and 45 ms$^{-1}$ corresponds to approximately 12 Beaufort. These are used to explore extreme winds associated with cyclones.

We summarize the results for the NH in Table 2 and
Table 3 where we have listed the number of cyclones for different regions that reach maximum wind speed of 35ms\(^{-1}\) and 45ms\(^{-1}\), respectively. For the NH as a whole there is a reduction in the number of cyclones, with wind speeds >35ms\(^{-1}\), for all seasons except JJA but the difference is hardly significant with the exception of the Atlantic. There is a general reduction over southern Europe and an increase over the Arctic. The larger number of events in the Arctic in 21C may be related to the reduced sea ice cover providing more favourable conditions for higher wind speeds. Minor changes can be found over northern Europe (>35ms\(^{-1}\)). The only significant increase is the increased number of stronger cyclones in winter for winds >45ms\(^{-1}\). The more extreme winds over northern Europe in winter can also be seen in the grid point distribution of the higher percentiles (see later). The more extreme wind speeds (>45ms\(^{-1}\)) occur predominantly over the Atlantic and Pacific oceans. The numbers for storms which have a purely extra-tropical origin are shown in brackets in Table 2 and Table 3. This has been done by excluding any systems with a tropical origin by removing any system that has its genesis in (0, 25)N. This indicates that tropical storms undergoing extra-tropical transition have a strong influence in the summer and autumn months, particularly in the Western Pacific region.

In Figure 10 we show the maximum wind speed for all storm tracks for all seasons and for the extra-tropics of both hemispheres. The extreme winds for the NH are stronger for 20C during DJF and MAM. We have explored this in detail and it seems mostly to emanate from the eastern and southern coast of Greenland. This is apparent in Figure 11 where the locations at which storms attain their maximum wind speed are displayed for maximum attained wind speeds >45ms\(^{-1}\) for the NH winter. As the Atlantic storms approach Greenland, the combination of steep orography and sharp coastline contrast leads to a secondary wind maximum at the northern and north western side of the storms. These intense winds are seen to be reduced in 21C compared with 20C as the Atlantic storm track changes its orientation away from Greenland. For 20C there is a clearer distinction between two preferred storm tracks, one northeast over the Norwegian Sea and another one
eastward over the Mediterranean. At 21C the storm tracks have a preference for an intermediate storm track over the British Isles. The cause of this change is not clear but is partly an effect of the poleward transition of the Hadley circulation and an associated weakening of the storm track over southern Europe. The enhancement of the storm track west of the British Isles we suggest is related to the strengthening SST gradient south of 50N in the central Atlantic (Bengtsson et al., 2006). For the JJA and SON periods the extreme winds are stronger for 21C. We have examined this and found that the main cause is a contribution from stronger tropical cyclones entering the extra tropics (see also Table 2 and Table 3). There is practically no difference in the extreme winds for the SH over the whole hemisphere, though regional changes are seen associated with the poleward transition of the storm tracks.

If we now consider the grid point perspective, Figure 12 shows the change in wind speed between 21C and 20C for the 99.5% percentiles (about two events every winter); the other percentiles are structurally similar. We show this for both Hemispheres and for NH winter and summer. As has been found in most previous studies (Ulbrich et al., 2008a and references therein) during the NH winter period there is a minor increase, ~1ms⁻¹, in an area stretching from the central North Atlantic towards Northern Europe covering a region between 40N and 65N. This coincides with an enhanced storm track density in the region for 21C (Bengtsson et al., 2006). In the area north and south thereof there is a reduction. Other areas of increase in winds are the eastern United States and Canada, the Arctic and eastern Asia. Large reductions occur south of 40N in both the Atlantic and Pacific and in the Mediterranean region. In our previous paper (Bengtsson et al., 2006) we noted an intensification of the storm tracks between 45° and 65° N over Europe and similar results have been found in other studies (Ulbrich et al., 2008a). The result is similar at this higher resolution, hence, the increase in extreme events in Western Europe is due to a regional enhancement in storm track density and intensity. This is likely related to increased low-level baroclinicity caused by enhanced low-level temperature gradient in the central North Atlantic related to the region of reduced SST warming south of Greenland (Bengtsson et al., 2006).
the NH summer there is a general weakening for 21C in the Atlantic sector, while the Arctic has slightly stronger winds. In the Pacific there is an apparent increase which as discussed previously we associate with changes in the intensities of tropical cyclones migrating polewards. In the SH we see a clear transition of more extreme winds towards the pole following the overall transition of the storm tracks towards higher latitudes as pointed out in Bengtsson et al., (2006). This transition occurs for all seasons.

At first sight the results from the percentile analysis is at variance with the tracking analysis shown in Figure 10 but in reality the differences are easily explained by the fact that the percentile analysis is an Eulerian analysis which is sensitive to where the storms are found and hence to the shift in the storm tracks poleward as discussed. Figure 10 indicates that there are no major changes in the intensities of the storms over a hemisphere. In fact if we compare the results in Figure 12 with the spatial statistics from the tracking, using the wind speed for intensity, then the changes in mean intensity are seen to be consistent with the changes in percentiles.

5.2. Precipitation

The global annual increase in precipitation is 6.2% in 21C compared to 20C or 1.9% per degree global warming. For column water on the other hand the increase is 26.8% or 8.3% per degree. This increase of column water follows closely the Clausius-Clapeyron relation (Pierrehumbert et al., 2007, Held and Soden, 2006).

The increase in maximum precipitation (as seen in grid point space) follows closely the increase in water vapor between 20C and 21C. The global increase is 32% (7.09 mm/hr at 21 C versus 5.36 at 20C). Although, the increase varies it is slightly higher in the tropics than in the extra-tropics (39% versus 29%) but is again generally higher along the storm tracks. For the cyclone tracks the cumulative total precipitation along the track, within a 5° geodesic radius, is higher than the total global increase in precipitation or 11.1% (28.8 mm per track versus 25.9 mm, area averaged) but not as high as for the extreme (maximum) precipitation, which amounts to almost three times this amount. We suggest that the reason for this is that the extra-tropical cyclones, because of the
dominant large-scale geostrophic processes, are much less efficient at extracting water vapor from the environment than small scale convective processes which are more capable of doing this.

In a similar way to the maximum wind speed, we have explored extreme precipitation both for the individual cyclones as well as for all individual 6 hourly grid points. Figure 13 shows the distributions of the maximum, hourly (averaged over six hours), area averaged precipitation for both hemispheres and for all seasons. There is a clear increase in the numbers of the extreme precipitating systems in 21C as is most apparent in the insets for each season in Figure 13. The relative change is broadly similar for both hemispheres and for all seasons. We have also produced distributions of the cumulative precipitation along the cyclone tracks (not shown) which show a similar behaviour to the maximum precipitation with higher extreme values for 21C than 20C and the same relative differences for both hemispheres and all seasons.

Figure 14a and b shows the geographical distribution of changes between 21C and 20C for the average precipitation in the NH for summer and winter. During winter there is an increase in precipitation north of ~45°N with the highest increase along the storm tracks and in the Arctic. At lower latitudes there is reduced precipitation. During summer the areas of reduced precipitation further increase, particularly so for Eurasia, following the weakening of the Atlantic storm track during summer. Changes in extreme precipitation for the NH winter and summer (Figure 14c and d), using the 99% percentile as a measure, have similar distributions to the mean with the largest increase along the storm tracks. It is interesting to note that the change in hourly precipitation at the 99% percentile is an order of magnitude larger than the change in the mean. Finally in Table 4 we have summarized the result in precipitation changes for different extra-tropical regions and for different percentiles and contrast them to the mean as well as the absolute maximum precipitation. There are marked differences between different areas. It is interesting to note that the increase in precipitation is larger for the NH for both summer and winter than the SH and for all the measures of extreme precipitation. For the smaller regions of the NH we note the marked reduction in mean precipitation in southern Europe which is almost halved during summer and reduced by some 20%.
in winter. However, in spite of that there is an increase in the absolute maximum showing there exist some rare occasions when the extreme precipitation will be higher irrespective of the overall reduction in total precipitation.

6. Summary of results and discussion

6.1. General

We have compared the cyclones identified in the T213 integration with those from an integration at lower horizontal resolution (T63). If we identify cyclones at a common T42 resolution the number of cyclones as a function of maximum intensity (T42) is practically identical and with no substantial change in the distribution. Differences are within the natural variability of 30-year samples (Bengtsson et al., 2006). This suggests that the T63 resolution is capable of reproducing extra-tropical cyclone distributions, but lacks only in the ability to produce more realistic extremes of wind and precipitation which is apparent when intensities at full resolution are studied.

The model results agree well with ERA40 except that the model results have stronger winds and larger precipitation. Preliminary assessment of the new Interim re-analyses shows values of these parameters are closer to the model. We have also compared the composite lifecycles and 3D structure of extreme storms in the model and ERA40 and found in general very good agreement, though with the model producing somewhat more intense storms. The weaker winds and precipitation in the ERA40 in part can be attributed to lower resolution, the impact of the data assimilation and deficiencies in the way such quantities as precipitation are determined.

The model winds have also been compared with observational data from the North Sea as described by Weisse et al. (2005). The result agrees well with the observations from Ekofisk and the other stations of the North Sea as well as with the results of Weisse et al., (2005). Incidentally, the results from the North Sea from the 20C are practically identical to those of the 21C.

The typical evolution of intense extra-tropical cyclones in 20 C and 21C is rather similar as can be seen from Figure 9a. The precipitation intensity is higher in 21C and the depth of the cyclones
slightly deeper if we choose the storms based on the T42 $\zeta_{850}$ intensities. The maximum wind speed on the other hand is somewhat weaker in 21C. We have also examined the cyclones with the strongest precipitation in 20C and 21C (Figure 9b); this shows a larger increase in precipitation between 20C and 21C but a marked reduction in intensity in terms of winds, vorticity and pressure. Consequently, *all indications are that the feedback from higher precipitation towards more intense synoptic scale extra-tropical cyclones is small.*

### 6.2. Winds

The evolution of the maximum wind of the most intense extra-tropical storms undergoes a characteristic evolution. At the time of maximum intensification the strongest winds occur in the open warm sector to the right and slightly forward of the centre (Figure 5a). At the time of maximum precipitation (Figure 5c), which occurs on average ~15 hours later the wind speed has increased and the maximum wind has moved some 45 degrees anti-clockwise. Finally, at the time of maximum intensity (Figure 5e) occurring some 15 hours later the strongest winds are found in the subsiding region behind the storm. A very similar picture occurs for 21C (not shown); maximum wind speed is practically identical and occurs at the same time and place relative to the storm center as for 20C.

In the NH the most interesting changes occur in the Atlantic region. As cyclones move towards the southern and eastern coast of Greenland the strongest winds are found around the coasts and between Greenland and Iceland. These are regions, which are known for exceptionally strong winds during the winter (Moore and Renfrew, 2005). There is a tendency to a slight weakening in this area in 21C. The reason for this is not easy to understand but it maybe related to a greater preference for the storm tracks to be orientated towards the British Isles and Scandinavia and a corresponding weakening both to the north and to the south. As discussed in Bengtsson et al., (2006) we suggest that this may be related to an enhanced SST gradient between 40-50N in the central Atlantic with associated increase in baroclinicity.
We also compared the wind speed with the model with lower resolution. While the extreme wind speed is lower in T63 this is most obvious in coastal regions where the low resolution model is unable to reproduce the marked maximum which occur in nearby coastal regions such as along the coast of Greenland and along the Norwegian coast.

6.3 Precipitation

As discussed in Bengtsson et al., (2007) the ECHAM5 model is typical of other General Circulation Models (GCM) in increasing water vapor more rapidly than precipitation. We note that precipitation will change for a number of reasons. Firstly, there is a general increase in global precipitation amounting to ~6.2%. Secondly, there is an overall change in the distribution of the precipitation with a larger percentage increase along the storm tracks and at higher latitudes in general. At the same time there is a marked reduction in the subtropics stretching further polewards during the summer season. This change follows broadly the changes in the hydrological cycle as discussed by Held and Soden (2006). Thirdly, there is a marked change in the statistical distribution of precipitation with proportionally higher precipitation at high intensity (Semenov and Bengtsson, 2002) giving rise to changes in precipitation of some 30% at the highest percentiles. The combined effect of these three effects implies considerable changes in extreme local precipitation which in some areas may amount to increases of 40 – 50 %.

We suggest that increases in extreme precipitation both in regions where there is an increase and decrease in the mean will constitute a more severe problem for society than the possible risk of higher wind speeds in some areas.

Needless to say these results are model dependent and other models are likely to provide either somewhat higher or somewhat lower global precipitation. However, the general result that the increase in annual average precipitation is much lower than the increase in column water vapor is common for all models used in the IPCC AR4 assessment (Held and Soden, 2006) as well as the response to enhance the regional extremes in precipitation. We also believe that the change in the
skewness of the distribution of precipitation over a time interval will lead to more extreme precipitation as discussed in Semenov and Bengtsson (2002).

7. Concluding remarks.

Extra-tropical cyclones have been studied in great detail using data from a T213 version of the ECHAM5 atmosphere only model forced with SST data from a coupled model at T63 resolution. We have explored the properties and structure of the cyclones and contrasted the T213 results with those from the T63 integration, modern reanalyses and other studies. Comparison with ERA40 and observations suggests that the ECHAM5 model at the T213 resolution simulates extra-tropical cyclones with great realism. This provides us with confidence in using the model to explore likely changes in extra-tropical cyclones and extremes associated with them in a warmer climate.

Needless to say, a study of this kind should be critically assessed in view of the fact that conclusions are based on only a single model. However, the ECHAM5 model has been validated extensively and has generally been found to provide realistic results (e.g. Ulbrich et al., 2008a). However, a few important questions might still be usefully considered.

1. Can we have sufficient confidence in one particular climate model?

The overall response to a warmer climate is a marked increase in column water vapor. This increase follows the Clausius-Clapeyron relation in common with other GCM. Similarly, the much slower increase in total precipitation is also common with other GCM (Held and Soden, 2006). However, also in agreement with other studies, extreme precipitation increases more rapidly than the mean precipitation (Semenov and Bengtsson, 2002). This is physically credible since convectively driven weather systems are expected to more effectively use the higher level of water vapor.

We believe we can have confidence in the change in maximum wind speed. To predict maximum wind speed in a local environment exposed to coastal and sharp orographic obstacles is not feasible with a model of the kind used here, but the model should be capable of providing reliable indicators for such predictions. We see no reason why the relation between a model
predictor and a local predictand will change in a warmer climate; consequently the local extremes are not likely to change either. Our confidence in the model is based on its ability to reproduce the typical lifecycle and structure of storms when compared with ERA40. It is also capable of reproducing the rapid change of surface pressure in the composite cyclone in virtually complete agreement with ERA40. Surface pressure is reliably determined in ERA40 as it is an observed parameter from a dense network in the NH.

2. *Is the integration sufficiently long?*

The integrations used for this study cover two 32 year periods. This is likely to be too short for a regional assessment of extreme conditions in extra-tropical cyclones (Weisse et al., 2005). However, based on the assessment of three different 30-year integrations with the same model at lower resolution (Bengtsson et al., 2006) the overall hemispheric statistics should be robust. We judge that reliable regional results will require longer integrations on a time scale of a century or the use of ensembles.

3. *Are the resolution and the physical parameterizations adequate?*

Here we are on less safe ground and it could well be the case that future ultra-high resolution using non-hydrostatic equations will give a different result. However, extra-tropical cyclones are well described even by the quasi-geostrophic equations, so we do not expect that the extra-tropical cyclones will be very different in a non-hydrostatic model. Additionally short and medium-range prediction with similar models are accurate with forecast errors more often related to errors in the initial state. In representative areas, where extreme wind statistics are available, the results agree with observations as well as with that of limited area models. Extreme precipitation on the other hand is likely to be even more pronounced in models with even higher resolution than we have investigated here and additional studies to clarify this will be needed. However our view is that an accurate resolution of the synoptic scale flow will provide more robust predictors for local extreme precipitation.
4. How confident are we that the feedback from any increase of latent heat is insignificant in generating stronger winds?

The fact that extra-tropical cyclones are most intense during the cold season and in situations with strong temperature gradients and associated strong upper air winds suggest that these conditions, that also follow from theoretical considerations, are the main drivers of extreme extra-tropical cyclones. In contrast to tropical cyclones where the release of latent heat is quasi-symmetrically organized around the cyclone the extra-tropical cyclones are different in this respect. Moreover, the evolution of an extra-tropical cyclone is characterized by a rapid transient process with a fast built up of frontal precipitation and an equally fast collapse of organized precipitation as the cyclone occludes and rapidly weakens.

In order to better explore the influence of latent heat release we identified the 100 most intense cyclones based upon precipitation intensity instead of vorticity maximum (Figure 9b). These cyclones have weaker maximum winds than those chosen using the T42 vorticity maxima but the interesting result is that there is no increase in extreme winds but rather a decrease in 21C. This further supports our view that the increase in latent heat release has only a minor influence on any likely intensification of extra-tropical cyclones.

We also calculated the composite of the available potential energy (not shown) for each of the 100 most intense cyclones and noted a slightly larger value in 21C at its maximum suggesting some effect of the higher level of latent heat release. However, this does not appear to be converted into kinetic energy as here the values at 20C and 21C were practically identical. It is our intention to pursue this further in a later study extending this work to also include the energetic interaction between cyclones and their surroundings.

Finally, there are reports of intense small-scale extra-tropical cyclones and polar lows which have features more in common with tropical cyclones, where an enhancement by latent heat cannot be excluded. We intend to investigate this in a future study.

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Captions

Table 1 Number of extreme storms in the NH, Atlantic and Pacific for the DJF period for ERA40 (1979-2001) and ECHAM5 (20C) normalized to 22 years. NH, (25, 90)N; Atlantic, (280, 360), (25, 70)N; Pacific, (120, 240), (25, 70)N.

Table 2 Number of extreme storms in the NH for all seasons (numbers are for all 32, years, 31 for DJF) with wind speeds above 35ms⁻¹ for various regions and for 20C and 21C. NH, (25, 90)N; Atlantic, (80W, 0), (25, 70)N; Atlantic/W. Europe, (90W, 40E), (25, 70)N, Pacific, (120E, 120W), (25, 70)N; Arctic, (70.0, 90.0)N; N. Europe, (10W, 40E), (47.5, 70.0)N; S. Europe, (10W, 40E), (30.0, 47.5)N. Values in brackets are excluding systems of tropical origin (those that have their genesis in (0, 25)N).

Table 3 Same as Table 2 but for wind speeds above 45ms⁻¹.

Table 4 Summary of percentage changes in precipitation between 21C and 20C for DJF and JJA for the area averages over the regions, NH (25, 90)N; SH (25, 90)S; N. America (25, 70)N, (160W, 65W); N. Atlantic (80W, 0), (25, 70)N; N. Pacific (120E, 120W), (25, 70)N; N. Europe (10W, 40E), (47.5, 70.0)N; S. Europe (10W, 40E), (30.0, 47.5)N for the time mean, maximum, 99% and 99.9% percentile.
Figure 1 Comparison of T63 and T213 tracking statistics for the DJF period for 20C: (a) T63 track density (colour), overlaid with mean intensity (line contour) computed from full resolution $\xi_{850}$, (b) T213 track density (colour), overlaid with mean intensity (line contour) computed from full resolution $\xi_{850}$, (c) T63 genesis density, (d) T213 genesis density. Densities are as number density per month per unit area with the unit area equivalent to a 5° spherical cap and intensities are in units of $10^{-5}$ s$^{-1}$ with a contour interval of 1x$10^{-5}$ s$^{-1}$ for T63 and 2.5 x$10^{-5}$ s$^{-1}$ for T213.

Figure 2 Comparison of T63 and T213 intensity distributions for different parameters for the DJF period for 20C: (a) minimum MSLP (hPa), bin width (bw) is 10hPa, (b) maximum $\xi_{850}$ (x10$^{-5}$ s$^{-1}$), bw is 5 (x10$^{-5}$)s$^{-1}$, (c) maximum 925hPa wind speed (m s$^{-1}$), bw is 5ms$^{-1}$, (d) maximum area averaged total precipitation (mm hr$^{-1}$), bw is 0.5 mm hr$^{-1}$. All parameters are determined with a 5° spherical cap region centered on the T42 storm location for all storms north of 25°. The insets show the tails of the distributions scaled to 90 months (30 DJF’s).

Figure 3 Time series of the 10m winds for the Ekofisk platform in the central North Sea (56.5N, 3.3E), for DJF centered on January, for the 90, 95, 99 and 99.5% percentiles at (a) the model grid point closest to the platform, and (b) ERA40 and the actual observations.

Figure 4 (a) Lifecycle composites of the 100 most intense storms, identified in T42 $\xi_{850}$, for the NH, DJF for T213 ECHAM5 in 20C (solid colours) and ERA40 (1979-2002) (transparent colours) and (b) lifecycle composites of the identically same 50 most intense storms in the Interim reanalysis (solid colours) and ERA40 (transparent colours). Parameters shown are MSLP (hPa) (black), $\xi_{850}$ ($10^{-5}$s$^{-1}$) (red), 925hPa winds (m s$^{-1}$) (green) and area averaged total precipitation (mm hr$^{-1}$) (blue).

Figure 5 ECHAM5 composites on a 20° radius region (geodesic) of the 100 most intense storms in T42 $\xi_{850}$ for the NH, DJF at different lifecycle stages; top is for maximum growth rate in T42 $\xi_{850}$.
middle is for maximum precipitation and bottom is for maximum T42 $\xi_{850}$ intensity. Left hand column shows the mean wind speed (colour, ms$^{-1}$) overlaid with the mean MSLP (black contours, hPa) and the mean temperature between 850 and 500hPa (dashed white contours), the right column shows the mean precipitation (colour, mm hr$^{-1}$) with the same overlays. Thick MSLP contour is at 1000hPa, c.i. is 5hPa, thick mean temperature contours are at 250K (upper line) and 270K (lower line), c.i. is 2K. Arrow indicates direction of composite storm.

Figure 6 Same as Figure 5 but for ERA40.

Figure 7 Vertical tilt lifecycle composites for the 100 most intense storms (identified in T42 $\xi_{850}$) for the NH, DJF. Time steps are shown up to 36 hours either side of the maximum intensity for (a) ECHAM5 and (b) ERA40. Tilts are in geodesic angle from the 850hPa center.

Figure 8 Distributions of minimum pressure (hPa) computed along each cyclone track for both the NH (black) and SH (grey) and for 20C (solid) and 21C (dashed). (a) DJF, (b) MAM, (c) JJA and (d) SON. Insets show the tails scaled to 90 months. Bin width is 10hPa.

Figure 9 Lifecycle composites of the 100 most intense storms for the NH, DJF for T213 ECHAM5 in 21C and 20C (transparent colour), identified from (a) their T42, $\xi_{850}$ intensities and (b) their area averaged precipitation intensities excluding any systems with tropical origins. Parameters shown are MSLP (hPa), $\xi_{850}$ ($10^{-5}$ s$^{-1}$), 925hPa winds (m s$^{-1}$) and area averaged total precipitation (mm hr$^{-1}$).

Figure 10 Same as Figure 8 except for maximum winds (ms$^{-1}$). Bin width is 5ms$^{-1}$.

Figure 11 Location of storms for the NH, DJF, when they attain their maximum wind speed at 925hPa for all storms where the maximum wind speed exceeds 45ms$^{-1}$ for (a) 20C and (b) 21C.
Figure 12 Differences between 21C and 20C of the 99.5% percentiles of the 925hPa winds for (a) the NH, DJF, (b) the NH, JJA, (c) the SH, DJF and (d) the SH, JJA.

Figure 13 Distributions of the maximum of the storm centered area averaged precipitation (mm hr$^{-1}$) for 20C and 21C and for NH and SH. (a) DJF, (b) MAM, (c) JJA, (d) SON. Bin widths are 0.5 mm hr$^{-1}$. Insets show the tails of the distributions scaled to 90 months (30 DJF’s).

Figure 14 Precipitation differences for the NH, (a) DJF time mean, (b) JJA time mean, (c) DJF 99% percentile, (d) JJA 99% percentile. Units are mm hr$^{-1}$. 
Tables.

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Table 1 Number of extreme storms in the NH, Atlantic and Pacific for the DJF period for ERA40 (1979-2001) and ECHAM5 (20C) normalized to 22 years. NH, (25, 90)N; Atlantic, (280, 360), (25, 70)N; Pacific, (120, 240), (25, 70)N.

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Table 2 Number of extreme storms in the NH for all seasons (numbers are for all 32, years, 31 for DJF) with wind speeds above 35 ms\(^{-1}\) for various regions and for 20C and 21C. NH, (25, 90)N; Atlantic, (80W, 0), (25, 70)N; Atlantic/W. Europe, (90W, 40E), (25, 70)N; Pacific, (120E, 120W), (25, 70)N; Arctic, (70.0, 90.0)N; N. Europe, (10W, 40E), (47.5, 70.0)N; S. Europe, (10W, 40E), (30.0, 47.5)N. Values in brackets are excluding systems of tropical origin (those that have their genesis in (0, 25)N).
Table 3 Same as Table 2 but for wind speeds above 45 m/s.

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<th>Pac.</th>
<th>Arctic</th>
<th>NEuro</th>
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Table 4 Summary of percentage changes in precipitation between 21C and 20C for DJF and JJA for the area averages over the regions, NH (25, 90)N; SH (25, 90)S; N. America (25, 70)N, (160W, 65W); N. Atlantic (80W, 0), (25, 70)N; N. Pacific (120E, 120W), (25, 70)N; N. Europe (10W, 40E), (47.5, 70.0)N; S. Europe (10W, 40E), (30.0, 47.5)N for the time mean, maximum, 99% and 99.9% percentile.
Figure 1 Comparison of T63 and T213 tracking statistics for the DJF period for 20C: (a) T63 track density (colour), overlaid with mean intensity (line contour) computed from full resolution $\xi_{850}$, (b) T213 track density (colour), overlaid with mean intensity (line contour) computed from full resolution $\xi_{850}$, (c) T63 genesis density, (d) T213 genesis density. Densities are as number density per month per unit area with the unit area equivalent to a $5^\circ$ spherical cap and intensities are in units of $10^{-5} \text{s}^{-1}$ with a contour interval of $1 \times 10^{-5} \text{s}^{-1}$ for T63 and $2.5 \times 10^{-5} \text{s}^{-1}$ for T213.
Figure 2 Comparison of T63 and T213 intensity distributions for different parameters for the DJF period for 20C: (a) minimum MSLP (hPa), bin width (bw) is 10hPa, (b) maximum $\xi_{850}$ ($x10^{-5}$ s$^{-1}$), bw is 5 ($x10^{-5}$s$^{-1}$), (c) maximum 925hPa wind speed (ms$^{-1}$), bw is 5ms$^{-1}$, (d) maximum area averaged total precipitation (mm hr$^{-1}$), bw is 0.5 mm hr$^{-1}$. All parameters are determined with a 5$^0$ spherical cap region centered on the T42 storm location for all storms north of 25$^0$. The insets show the tails of the distributions scaled to 90 months (30 DJF’s).
Figure 3 Time series of the 10m winds for the Ekofisk platform in the central North Sea (56.5N, 3.3E), for DJF centered on January, for the 90, 95, 99 and 99.5% percentiles at (a) the model grid point closest to the platform, and (b) ERA40 and the actual observations.
Figure 4 (a) Lifecycle composites of the 100 most intense storms, identified in T42 $\xi_{850}$, for the NH, DJF for T213 ECHAM5 in 20C (solid colours) and ERA40 (1979-2002) (transparent colours) and (b) lifecycle composites of the identically same 50 most intense storms in the Interim reanalysis (solid colours) and ERA40 (transparent colours). Parameters shown are MSLP (hPa) (black), $\xi_{850}$ ($10^{-5}$s$^{-1}$) (red), 925hPa winds (m s$^{-1}$) (green) and area averaged total precipitation (mm hr$^{-1}$) (blue).
Figure 5 ECHAM5 composites on a 20° radius region (geodesic) of the 100 most intense storms in T42 $\bar{\xi}_{850}$ for the NH, DJF at different lifecycle stages; top is for maximum growth rate in T42 $\xi_{850}$, middle is for maximum precipitation and bottom is for maximum T42 $\xi_{850}$ intensity. Left hand column shows the mean wind speed (colour, ms$^{-1}$) overlaid with the mean MSLP (black contours, hPa) and the mean temperature between 850 and 500hPa (dashed white contours), the right column shows the mean precipitation (colour, mm hr$^{-1}$) with the same overlays. Thick MSLP contour is at 1000hPa, c.i. is 5hPa, thick mean temperature contours are at 250K (upper line) and 270K (lower line), c.i. is 2K. Arrow indicates direction of composite storm.
Figure 6 Same as Figure 5 but for ERA40.
Figure 7 Vertical tilt lifecycle composites for the 100 most intense storms (identified in T42 $\zeta_{850}$) for the NH, DJF. Time steps are shown up to 36 hours either side of the maximum intensity for (a) ECHAM5 and (b) ERA40. Tilts are in geodesic angle from the 850hPa center.
Figure 8 Distributions of minimum pressure (hPa) computed along each cyclone track for both the NH (black) and SH (grey) and for 20C (solid) and 21C (dashed). (a) DJF, (b) MAM, (c) JJA and (d) SON. Insets show the tails scaled to 90 months. Bin width is 10hPa.
Figure 9 Lifecycle composites of the 100 most intense storms for the NH, DJF for T213 ECHAM5 in 21C and 20C (transparent colour), identified from (a) their T42, $\zeta_{850}$ intensities and (b) their area averaged precipitation intensities excluding any systems with tropical origins. Parameters shown are MSLP (hPa), $\zeta_{850}$ ($10^{-5}$ s$^{-1}$), 925hPa winds (m s$^{-1}$) and area averaged total precipitation (mm hr$^{-1}$).
Figure 10 Same as Figure 8 except for maximum winds (m s$^{-1}$). Bin width is 5 m s$^{-1}$. 
Figure 11 Location of storms for the NH, DJF, when they attain their maximum wind speed at 925hPa for all storms where the maximum wind speed exceeds $45\text{ms}^{-1}$ for (a) 20C and (b) 21C.
Figure 12 Differences between 21C and 20C of the 99.5% percentiles of the 925hPa winds for (a) the NH, DJF, (b) the NH, JJA, (c) the SH, DJF and (d) the SH, JJA.
Figure 13 Distributions of the maximum of the storm centered area averaged precipitation (mm hr⁻¹) for 20C and 21C and for NH and SH. (a) DJF, (b) MAM, (c) JJA, (d) SON. Bin widths are 0.5 mm hr⁻¹. Insets show the tails of the distributions scaled to 90 months (30 DJF’s).
Figure 14 Precipitation differences for the NH, (a) DJF time mean, (b) JJA time mean, (c) DJF 99% percentile, (d) JJA 99% percentile. Units are mm hr$^{-1}$. 
