Northern Hemisphere Extratropical Cyclones in a Warming Climate in the HiGEM High-Resolution Climate Model

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ABSTRACT

Changes to the Northern Hemisphere winter (December–February) extratropical storm tracks and cyclones in a warming climate are investigated. Two idealized climate change experiments with the High Resolution Global Environmental Model version 1.1 (HiGEM1.1), a doubled CO2 and a quadrupled CO2 experiment, are compared against a present-day control run. An objective feature tracking method is used and a focus is given to regional changes. The climatology of extratropical storm tracks from the control run is shown to be in good agreement with the 40-yr European Centre for Medium-Range Weather Forecasts Re-Analysis (ERA-40), while the frequency distribution of cyclone intensity also compares well.

In both simulations the mean climate changes are generally consistent with the simulations of the Intergovernmental Panel on Climate Change Fourth Assessment Report (AR4) models, with strongly enhanced surface warming at the winter pole and reduced lower-tropospheric warming over the North Atlantic Ocean associated with the slowdown of the meridional overturning circulation. The circulation changes in the North Atlantic are different between the two idealized simulations with different CO2 forcings. In the North Atlantic the storm tracks are influenced by the slowdown of the MOC, the enhanced surface polar warming, and the enhanced upper tropical-troposphere warming, giving a northeastward shift of the storm tracks in the $2 \times CO2$ experiment but no shift in the $4 \times CO2$ experiment.

Over the Pacific, in the $2 \times CO2$ experiment, changes in the mean climate are associated with local temperature changes, while in the $4 \times CO2$ experiment the changes in the Pacific are impacted by the weakened tropical circulation. The storm-track changes are consistent with the shifts in the zonal wind.

Total cyclone numbers are found to decrease over the Northern Hemisphere with increasing CO2 forcing. Changes in cyclone intensity are found using 850-hPa vorticity, mean sea level pressure, and 850-hPa winds. The intensity of the Northern Hemisphere cyclones is found to decrease relative to the control.

1. Introduction

The day-to-day variability of weather in the midlatitudes is strongly affected by extratropical cyclones. Intense extratropical cyclones can have large socioeconomic impacts associated with their strong winds and heavy rain. For example, storm Kyrill, which affected Europe in January 2007 (Fink et al. 2009), caused a loss of more than €1 billion due to its strong winds (Vitolo et al. 2009). In December 1999 three intense extratropical cyclones (Anatol, Lothar, and Martin) affected Europe, causing €18.5 billion of economic damage (Munich Re 2002; Mailier et al. 2006). As the climate changes in response to increased CO2 forcing, it is possible that the socioeconomic impacts associated with extratropical cyclones will also change.
There have been a number of previous studies investigating Northern Hemisphere storm-track changes in a warming climate (e.g., Bengtsson and Ulbrich 2004; Bengtsson et al. 2006; Pinto et al. 2007; Bengtsson et al. 2009). A consistent result from the studies is that there is a general decrease in the total number of cyclones occurring in a warmer climate. This is thought to be due to the enhanced surface warming at higher latitudes and weaker surface warming at low latitudes, leading to a decreased meridional temperature gradient and associated decreased low-level baroclinicity. Due to enhanced moisture content of the atmosphere, the storms would also be more efficient at transporting heat (Held 1993), so less storms could perform the same heat transport.

It might also be expected that, in a warming climate, the availability of extra moisture would lead to increased latent heating in extratropical cyclones and, therefore, more intense extratropical cyclones (in terms of pressure and winds). Although the study of Lambert and Fyfe (2006) (which used counts of low pressure centers) found an increase in the number of intense cyclones across models from the Coupled Model Intercomparison Project phase 3 (CMIP3), recent studies have found either no change or even a decrease in the number of the intense cyclones (e.g., Bengtsson et al. 2009; Watterson 2006; Bengtsson et al. 2009). The HiGEM1.1 control simulation was run in Roberts et al. (2009) and Shaffrey et al. (2009). The HiGEM1.1 control simulation was run in Roberts et al. (2009) and Shaffrey et al. (2009).

At present it is unclear why some climate models predict an increase in extratropical cyclones, while others predict a decrease. One question is whether climate models have sufficient horizontal resolution to be capable of capturing the synoptic-scale features of intense extratropical cyclones. To address this, the response of the Northern Hemisphere extratropical cyclones to idealized climate forcing will be studied in a high-resolution coupled climate model [High Resolution Global Environmental Model (HiGEM), Shaffrey et al. (2009)] that is capable of capturing the observed synoptic-scale structure of intense extratropical cyclones (Catto et al. 2010).

A secondary issue to be addressed in this study is the relative influence of regional-scale versus global-scale processes in driving change in extratropical storm tracks. The Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4) “Summary for Policymakers” (Bernstein et al. 2007) suggests that in the future there will be a “poleward shift of extra-tropical storm tracks with consequent changes in wind, precipitation and temperature patterns.” This poleward shift was found by Yin (2005) to be a feature of the multimodel mean from the AR4 models, when using zonally averaged eddy kinetic energy as a measure of storm-track activity.

There are, however, some large differences between the regional responses of the CMIP3 models. Bengtsson et al. (2006) and Ulbrich et al. (2008) found longitudinal differences in the storm-track changes for the A1B scenario in ECHAM5 and 23 of the IPCC models. The study of Lainé et al. (2009) (which used counts of low pressure centers) found an increase in the number of intense cyclones across models from the Coupled Model Intercomparison Project phase 3 (CMIP3), recent studies have found either no change or even a decrease in the number of the intense cyclones (e.g., Bengtsson et al. 2009; Lainé et al. 2009; Woollings 2010). This study will also address how the relative importance of regional processes, and hence the response of the extratropical storm tracks, may change with time.

The paper is structured as follows. In section 2 the model simulations, observationally constrained data, and objective feature tracking methodology are described. The present day climatology of storm tracks from HiGEM1.1 is evaluated in section 3. The mean circulation changes in the two simulations and the changes to the storm tracks are described in sections 4 and 5, respectively, with a focus on the North Atlantic in section 6. Section 7 will focus on changes in storm intensity, and the conclusions are discussed in section 8.

2. Data and methodology

a. Model simulations

This study uses HiGEM1.1, a high-resolution coupled climate model based on the Met Office Hadley Centre Global Environmental Model version 1 (HadGEM1) (Johns et al. 2006; Ringer et al. 2006), run on the Earth Simulator in Japan as part of the UK–Japan Climate Collaboration (UJCC). The horizontal resolution of the atmosphere component of HiGEM1.1 is 0.83° latitude × 1.25° longitude, which is higher than the typical resolution of the climate models used in the IPCC AR4 (Randall et al. 2007), and the atmosphere has 36 levels. The ocean component also has a higher resolution than many of the IPCC AR4 models at 1/3° × 1/3° with 40 vertical levels. More details of HiGEM1.1 can be found in Roberts et al. (2009) and Shaffrey et al. (2009). The HiGEM1.1 control simulation was run...
CO₂ levels were stabilized after 35 years at a computational expense of such a high-resolution model. The transient run was performed at 2% per year. The transient run was performed with CO₂ levels increasing by 2% per year to reduce the simulation length, owing to the computational expense of such a high-resolution model. The CO₂ levels were stabilized after 35 years at 2 × CO₂ levels and run for a further 30 years. The transient run was continued and the CO₂ stabilized at 4 × CO₂ levels after 70 years. Again, this was run for a further 30 years after CO₂ stabilization. The two runs with the stabilized CO₂ levels will be referred to as the 2 × CO₂ and 4 × CO₂ experiments. It is important to remember that, although the CO₂ levels are constant in these experiments, the climate of the model will not be in equilibrium, much like the transient scenario runs used for the IPCC AR4.

An assessment of the total cyclone numbers in the control, 2 × CO₂, and 4 × CO₂ simulations for the entire NH suggests that there is large interannual variability in the number of cyclones. The standard deviations of yearly December–February (DJF) cyclone numbers in the NH are 18.0, 18.5, and 15.1 cyclones for the control, 2 × CO₂, and 4 × CO₂ simulations, respectively. The corresponding linear trends for the same period are 0.3, −0.2, and 0.1 cyclones per year. This indicates that, compared to the interannual variability, the trends in cyclone numbers are small, and the trends in the climate change simulations are of similar magnitude to the control simulation. This gives some confidence that the climate of the simulations is not substantially drifting.

**FIG. 1.** Schematic showing the idealized climate simulations.

using present-day radiative forcings. The concentrations of greenhouse gases are held constant at 345 ppmv for CO₂, 1656 ppbv for CH₄, and 307 ppbv for NO₂. Aerosol and ozone concentrations are also held constant for present day values. After the first 20 years the net top-of-atmosphere radiation and the upper ocean are considered to be spun up, so years 21–70 of the simulation have been used here [consistent with the previous study of Catto et al. (2010)].

The idealized climate change simulations performed with HiGEM1.1 are shown schematically in Fig. 1. From the control simulation, a transient climate change run was performed with CO₂ levels increasing by 2% per year. The transient run was performed at 2% per year to reduce the simulation length, owing to the computational expense of such a high-resolution model. The CO₂ levels were stabilized after 35 years at 2 × CO₂ levels and run for a further 30 years. The transient run was continued and the CO₂ stabilized at 4 × CO₂ levels after 70 years. Again, this was run for a further 30 years after CO₂ stabilization. The two runs with the stabilized CO₂ levels will be referred to as the 2 × CO₂ and 4 × CO₂ experiments. It is important to remember that, although the CO₂ levels are constant in these experiments, the climate of the model will not be in equilibrium, much like the transient scenario runs used for the IPCC AR4.

The period 1979–2002 from ERA-40 is used as this period is more strongly constrained by observations that include upper-air satellite data (Bengtsson et al. 2004). To investigate the impact of resolution of the global reanalysis, ERA-Interim (which also has improved data assimilation and other improvements over ERA-40) was used to evaluate the distribution of storm characteristics in HiGEM.

**b. Reanalysis data**

The two sources of observationally constrained data used in this study against which the present-day control simulation from HiGEM1.1 is evaluated are the 40-yr European Centre for Medium-Range Weather Forecasts Re-Analysis (ERA-40) (Uppala et al. 2005), which has a resolution of T159 (~125 km), and the ECMWF Interim reanalysis (ERA-Interim) (Simmons et al. 2007; Uppala et al. 2008) with a resolution of T255 (~80 km). The period 1979–2002 from ERA-40 is used as this period is more strongly constrained by observations that include upper-air satellite data (Bengtsson et al. 2004). To investigate the impact of resolution of the global reanalysis, ERA-Interim (which also has improved data assimilation and other improvements over ERA-40) was used to evaluate the distribution of storm characteristics in HiGEM.

**c. Cyclone identification and feature tracking**

Objective feature tracking is used to identify and track individual cyclones and has been shown to be a useful method for investigating the changes in extratropical cyclone activity in a changing climate (e.g., Leckebusch and Ulbrich 2004; Bengtsson et al. 2006; Pinto et al. 2007; Bengtsson et al. 2009). The tracking algorithm used here is that of Hodges (1994, 1995, 1999) and Hoskins and Hodges (2002), identifying cyclones as maxima of 850-hPa relative vorticity in the 6-hourly data from the Northern Hemisphere. Before the tracking is performed, the relative vorticity field is spectrally truncated at T42. This removes the noise inherent in such a small-scale field and allows features at the synoptic scale to be identified. To remove the large-scale planetary wave signal, a background field with wavenumbers less than or equal to 5 is removed (e.g., Hoskins and Hodges 2002; Catto et al. 2010). A first-guess track of 6-hourly vorticity features is found using a nearest-neighbor approach. Then the smoothest set of tracks is achieved by minimizing a cost function by iteratively swapping points between tracks to provide the largest gain in track smoothness. Cyclones with lifetimes of at least 2 days and that travel farther than 1000 km during their lifetime are used in the calculation of the statistics. Similar to the studies of Bengtsson et al. (2009) and Catto et al. (2010) and other studies, the focus is on the NH winter season (DJF). Estimates of significance of differences between the different simulations are made using a nonparametric Monte Carlo approach (Hodges 2008). This method resamples the full tracks 2000 times to produce a probability distribution of possible differences from which the p values can be found.
Frequency distributions of certain variables attributed to the cyclone tracks (identified by the tracking algorithm) have been produced, allowing the effect of climate change on the extrema to be examined. These frequency distributions were produced having referenced the tracks identified at the T42 resolution back to the full resolution fields. The use of the full resolution fields has been done so that extreme values, which are the most important for regional impacts, can be seen. In this study, the full resolution minimum mean sea level pressure (MSLP), maximum 850-hPa vorticity, and maximum 850-hPa wind speeds along the tracks are used to evaluate extratropical cyclone intensity. Each of these measures characterizes cyclone intensity in a different way. The minimum MSLP has been used in previous studies (e.g., Leckebusch and Ulbrich 2004 and references therein). The extraction of the full resolution vorticity and wind fields was performed using a simple search for the maximum within a specified radial region (on the sphere) from the T42 feature point. This was chosen as $5^\circ$, consistent with Bengtsson et al. (2009). The MSLP minima associated with the tracks were found using a minimization technique (Bengtsson et al. 2009).

3. Storm tracks in the present climate

Before the results of the climate change experiments are presented, it is essential to evaluate the present-day simulation against observations. The synoptic-scale structure of intense NH extratropical cyclones in HiGEM was previously evaluated in Catto et al. (2010). It was found that HiGEM had sufficient horizontal resolution to capture the dynamical structure of intense extratropical cyclone found in the ERA-40 reanalyses, although there were some differences in the vertical structure of the moisture in the cyclones. A brief evaluation of the extratropical storm tracks in the control run is presented in this section. This allows the results of the idealized climate simulations to be interpreted with the biases of the model in mind. The diagnostics used for this purpose are the genesis density (Figs. 2a,b), which gives a measure of the number of cyclone tracks starting in a particular region, and the track density (Figs. 2c–e), which is similar to a traditional count of the number of systems passing through a grid box per unit area. See Hoskins and Hodges (2002) for further description of the tracking statistics.

The main genesis regions identified in ERA-40 (Fig. 2a) over the Gulf Stream and Kuroshio, and to a lesser extent in the mid-Atlantic and mid-Pacific, are well represented in HiGEM1.1 (Fig. 2b). As seen in Figs. 2c,d, the Atlantic and Pacific track density features are also spatially well represented in HiGEM1.1 compared to ERA-40. There are many small-scale features that are well represented in the track density, particularly over western Asia and Siberia (associated with local features such as the Ural Mountains and the Caspian Sea), in the lee of the Rocky Mountains, and in the Mediterranean.

Despite many of these small-scale features being well represented in the model, there are some differences. The track density is slightly higher in HiGEM1.1 than ERA-40 in the region between Iceland and Greenland and over the Gulf Stream, as seen by the differences for HiGEM1.1 minus ERA-40 shown in Fig. 2e. Over Mongolia and northern China both genesis density and track density in HiGEM1.1 are lower than in ERA-40. Farther south, over southern China, the genesis and track densities are higher in the model. These two signals are related to the southward displacement of the jet over South Asia seen in HiGEM (Shaffrey et al. 2009). The track density in HiGEM1.1 is too high over the central and western Pacific, related to the larger values of genesis just to the east of Japan and in the central Pacific.

The representation of the intensities of cyclones can be evaluated using frequency distributions (e.g., Bengtsson et al. 2009). The frequency distributions are shown in Figs. 3a–c for MSLP and 850-hPa vorticity and winds, respectively. The frequency distributions of minimum MSLP (Fig. 3a) show that there is very close agreement between HiGEM1.1, ERA-Interim, and ERA-40. MSLP is a spatially smooth variable, so it is not strongly affected by the differences in resolution [the impact of dataset resolution on storm-track statistics has been previously examined by Blender and Schubert (2000)]. The tails of the frequency distributions scaled to 10 yr show that the higher-resolution datasets have slightly more cyclones with very low pressure.

The frequency distributions of maximum 850-hPa vorticity show larger differences between the three datasets. All three distributions peak around 10 to 20 ($\times 10^{-5}$ s$^{-1}$). However, HiGEM1.1 and ERA-Interim have more intense cyclones with maximum vorticities $>30 \times 10^{-5}$ s$^{-1}$ reached along the tracks. This is similar to the findings of Bengtsson et al. (2009), who compared the distributions of maximum vorticity along the tracks in two different resolution versions of the ECHAM5 model. The upper tail (scaled to 10 yr) of the maximum vorticity
FIG. 2. Northern Hemisphere DJF genesis density for (a) ERA-40 and (b) HiGEM1.1; NH track density for (c) ERA-40, (d) HiGEM1.1 control, and (e) HiGEM1.1 control minus ERA-40. Units are cyclones per month per 5° spherical cap. Color bars show contour intervals. Period used for ERA-40 is the satellite period (1979–2002).
Fig. 3. Northern hemisphere distributions (number of storms per month) and tails of the distributions scaled to 30 months (10 winters) of (a) minimum MSLP (hPa, bin size 10 hPa), (b) maximum 850-hPa relative vorticity \((10^{-2} \text{ s}^{-1})\), (bin size \(10 \times 10^{-5} \text{ s}^{-1}\)), and (c) maximum 850-hPa wind speed (m s\(^{-1}\), bin size 5 m s\(^{-1}\)) occurring along the tracks found using 850-hPa vorticity in the HiGEM control (solid black line), ERA-Interim (1989–2008; dashed black line), and ERA-40 (1979–2002; dashed gray line).
distribution in Fig. 3b shows that HiGEM1.1 is capable of capturing some of the extreme values seen in ERA-Interim.

The maximum wind speed at 850 hPa, the distributions of which are shown in Fig. 3c, shows some effect of resolution. The distribution of maximum values from ERA-40 has slightly lower values compared to HiGEM1.1 and ERA-Interim. The distribution from HiGEM does lie in between the two reanalysis distributions, indicating that HiGEM1.1 is capable of capturing the wind speeds associated with the extratropical cyclones within the range of the observationally constrained data. In general, HiGEM1.1 is good at capturing both the spatial distribution and the intensity of the extratropical cyclones compared to the reanalysis datasets.

4. Changes to the mean climate

There is a strong relationship between the mean flow of the atmosphere and extratropical storm tracks; therefore, it is important to characterize changes in the mean climate. The differences between the $2 \times CO_2$ and $4 \times CO_2$ experiments and the present-day control run for global surface temperature, zonal-mean tropospheric temperatures, and the upper-level zonal winds are discussed in this section.

a. Surface temperature change

There is general surface warming in DJF in both $2 \times CO_2$ and $4 \times CO_2$ experiments relative to the present day control, as seen in Figs. 4a and 4b. As expected, there is larger warming over land than over ocean because of the different heat capacities and feedbacks related to surface moisture availability (e.g., Joshi et al. 2008; Dong et al. 2009). In the NH there is a very large amount of warming over the very high latitudes. This high-latitude warming is as large as 18 K in the $4 \times CO_2$ experiment (Fig. 4c) due to the reduction and eventual disappearance of sea ice (the sea ice albedo feedback). These features are broadly consistent with the results from the IPCC AR4 model projections (Meehl et al. 2007).

There is a region of reduced warming in the North Atlantic in DJF in both $2 \times CO_2$ and $4 \times CO_2$ experiments relative to the present day control, associated with the slowdown of the Atlantic meridional overturning circulation (MOC) (e.g., Hu et al. 2009). This reduced warming signal in the North Atlantic is consistent with that seen in the IPCC AR4 multimodel mean (Meehl et al. 2007). The reduced warming seen in HiGEM1.1 leads to an increased meridional temperature gradient over a small region of the North Atlantic. The impact of this localized increase in the meridional temperature gradient will be addressed in later sections.

In the $4 \times CO_2$ experiment there is increased warming over the eastern and central tropical Pacific relative to that seen in the subtropics. This is associated with a decrease in the strength of the tropical Walker circulation (Vecchi and Soden 2007). The changes in the meridional temperature gradients associated with the warming in the tropical Pacific influence the low-level baroclinicity in the North Pacific, and subsequently the Pacific extratropical storm tracks.

b. Circulation change

Figure 5a shows the DJF zonal mean temperature and zonal wind from the control simulation. The changes in the $2 \times CO_2$ and $4 \times CO_2$ experiments relative to the present day control can be seen in Figs. 5b and 5c. The magnitude of the temperature response is approximately double in the $4 \times CO_2$ run compared to the $2 \times CO_2$ run (note the change of scale in Fig. 5). Since the tropical atmospheric lapse rate remains close to a moist adiabat, the zonal mean temperature response shows enhanced warming in the tropical upper troposphere in both $2 \times CO_2$ and $4 \times CO_2$ runs. There is also enhanced warming at lower levels over the Arctic, associated with sea-ice
albedo feedback, in both of the climate change experiments.

Figure 5 also shows the change in the zonal-mean zonal wind in the $2 \times CO_2$ and $4 \times CO_2$ experiments. In the NH, there is an upward shift in the jet in both experiments (Figs. 5b,c). This is consistent with the results found by Lorenz and DeWeaver (2007) and Yin (2005) for the multimodel mean from the IPCC AR4 models.

However, neither of the simulations shows a strong poleward shift. The response in the NH is baroclinic. At some latitudes the wind response changes sign with height and is generally a very small signal at lower levels. The baroclinic wind response in the NH in DJF is associated with the change in meridional temperature gradient response with height (Woollings 2008); that is, at low levels the temperature gradient is reduced, while at upper levels it is increased.

The present-day control spatial pattern of upper-level (250 hPa) zonal wind for DJF can be seen in Fig. 6a, and the changes in the $2 \times CO_2$ and $4 \times CO_2$ experiments in Figs. 6b and 6c, respectively. The pattern of change in the NH is not zonally symmetric and in DJF is different between $2 \times CO_2$ and $4 \times CO_2$. In the Pacific Fig. 6b shows a strengthening of the jet over Japan and the west
Pacific in the $2 \times CO_2$ experiment relative to the present-day control. During DJF in the North Pacific, in the $4 \times CO_2$ experiment, there is a dipole indicating a strengthening of the jet between $20^\circ$ and $40^\circ$N and a weakening farther north. This pattern of upper-level zonal wind change is consistent with a weakening of the tropical Walker circulation, as seen in Vecchi and Soden (2007), and gives a pattern similar to that seen during the mature phase of El Niño events (e.g., Wang 2002).

In the Atlantic, in the $2 \times CO_2$ experiment, there is an enhancement of the upper-level (250 hPa) jet over the North Atlantic between $40^\circ$ and $60^\circ$N and a weakening farther north. In the $4 \times CO_2$ experiment, the pattern in the North Atlantic is quite different from the $2 \times CO_2$ experiment, with $4 \times CO_2$ showing an enhancement of the jet at low latitudes to the east of Florida. Differences between the $2 \times CO_2$ and $4 \times CO_2$ experiments seen in the North Atlantic will be discussed in more detail in section 6.

5. Extratropical storm-track changes

In this section changes to the extratropical storm tracks are discussed, and these are related to the mean circulation changes described in the previous section.

a. The $2 \times CO_2$ experiment

The results for the NH winter track density and cyclogenesis are shown in Fig. 7. Changes to the position of the genesis regions will have an impact on the downstream track density. In the $2 \times CO_2$ simulation it can be seen in the West Pacific that there has been a northward shift of the track density relative to the present day control (Fig. 7d). There are significant decreases (at the 95% level) of up to 1.4 cyclones per month per $5^\circ$ spherical cap to the south of Japan and increases over Japan and to the east of Japan. This is related to the northward shift of the upper-level jet in this area (Fig. 6c) and is consistent with the increase in genesis to the east of Japan. Figure 8a shows the difference in track density between the $2 \times CO_2$ and the control experiments for cyclones with genesis in a $10^\circ$ circular region over the Kuroshio (indicated by the thick dashed ring in Fig. 8), and confirms that the genesis to the east of Japan does indeed contribute to the increased track density in the North Pacific. There is also a large decrease in track density of up to three cyclones per month per $5^\circ$ spherical cap in the mid-Pacific (Fig. 7d, significant at the 95% level), consistent with a slight decrease in genesis density in that area (Fig. 7c).

The Mediterranean storm track shows a weakening in the $2 \times CO_2$ simulation with up to 1.4 cyclones per month per $5^\circ$ spherical cap less than in the present-day control simulation. This has a downstream impact on the Middle East with larger decreases in track density in this region. The weakened storm track in the Middle East and into South Asia is associated with the weakening of the subtropical jet in this region, as seen in Fig. 6b. This feature is also seen in many of the CMIP3 models (Christensen et al. 2007) and in other studies (e.g., Bengtsson et al. 2006; Pinto et al. 2007; Raible et al. 2010).

In the $2 \times CO_2$ experiment, over the North Atlantic there is a slight northeastward shift in the track density relative to the present day control. This shift is associated with the change in the upper-level jet, which is strengthened across a latitude of $50^\circ$N from Newfoundland to the United Kingdom (Fig. 6d). Previous studies have shown the link between changes in the upper-level jet and storm tracks, for example, Pinto et al. (2009).

b. The $4 \times CO_2$ experiment

Over the North Pacific the northward shift in track density is more pronounced than in the $2 \times CO_2$ simulation and the decrease in track density along the $40^\circ$ latitude band is significant at the 95% level. There is also an extension of the storm track into the east Pacific with a large increase in track density in this region. This feature is clearly associated with the upper-level zonal wind changes in this region, which are consistent with the pattern of changes associated with a weaker tropical Walker circulation (Vecchi and Soden 2007).

There is a small increase in genesis density (Fig. 7e) in the east Pacific in the $4 \times CO_2$ experiment relative to the present day control, which would contribute to the increased track density. However, there is also a decrease in the lysis density between $60^\circ$ and $70^\circ$E (not shown), suggesting that perhaps the cyclones generated in the west of the basin are tracking farther across the Pacific because of the stronger jet. This can be confirmed by looking at the difference in track density associated with only the cyclones with cyclogenesis in a $10^\circ$ circular region over the Kuroshio (indicated by the black ring in Fig. 8b) between the $4 \times CO_2$ experiment and the present day control. In the $4 \times CO_2$ experiment, there is much higher track density in the east of the Pacific basin associated with these cyclones (Fig. 8) compared to the present day control.

The decrease in track density in the Mediterranean is very large in the $4 \times CO_2$ experiment, being more than double the change in the $2 \times CO_2$ run (Fig. 7f). The maximum decrease of more than three cyclones per month per $5^\circ$ spherical cap is significant. As in the $2 \times CO_2$ experiment, there is a weakening of the subtropical jet in this region (Fig. 6c).
 FIG. 7. Tracking statistics for the NH in DJF: genesis density for (a) the control, (c) $2 \times CO_2$ – control, and (e) $4 \times CO_2$ – control and track density for (b) control, (d) $2 \times CO_2$ – control, and (f) $4 \times CO_2$ – control, contour intervals given by the color bars. Note the change in contour interval between the $2 \times CO_2$ and $4 \times CO_2$ experiments for track density. Shading shows regions where the changes are significant at the 95% level, as calculated using the nonparametric method of Hodges (2008).
Despite the stronger CO\textsubscript{2} forcing, the response of the North Atlantic storm track in HiGEM is much weaker in the 4 \times \text{CO}_2 than in the 2 \times \text{CO}_2 experiment. In the 2 \times \text{CO}_2 experiment a northeastward shift of the North Atlantic storm track was clearly seen (Figs. 7c and 7d). In the 4 \times \text{CO}_2 run, only small decreases in track density and genesis density can be seen over the Gulf Stream (Figs. 7e and 7f). These small decreases are associated with reductions in the land–sea temperature contrast and subsequent decreased baroclinicity. This interesting disparity is discussed in more detail in the next section.

6. Changes over the North Atlantic

This section focuses on the mechanisms that lead to the response of the North Atlantic storm track in the 4 \times \text{CO}_2 experiment being weaker than the response seen in the 2 \times \text{CO}_2 experiment. Figures 9a and 9c show the zonally averaged warming over the North Atlantic sector. There is enhanced warming in the tropical upper troposphere and in the polar lower troposphere in both 2 \times \text{CO}_2 and 4 \times \text{CO}_2 experiments relative to the present day control. These areas of enhanced warming act to increase (decrease) the meridional temperature gradient at upper (lower) levels.

Both 2 \times \text{CO}_2 and 4 \times \text{CO}_2 experiments show a reduced warming in the midlatitude lower troposphere. This is associated with reduced surface warming of SST in the North Atlantic (Fig. 4). This reduced warming is broadly consistent with the results from the AR4 models and is associated with the slowdown of the MOC in response to anthropogenic forcing. In the HiGEM1.1 experiments, the MOC undergoes a slowdown from an average transport of 21 Sv (Sv = 10\textsuperscript{6} m\textsuperscript{3} s\textsuperscript{-1}) in the present day control to 17 Sv in the 2 \times \text{CO}_2 experiment and to 15 Sv in 4 \times \text{CO}_2. The study of Brayshaw et al. (2009) also found that a weakening of the MOC (through freshwater hosing) led to reduced temperatures in the North Atlantic and a change to the storm tracks, consistent with the surface temperature gradients in the region.

In the 2 \times \text{CO}_2 experiment (Fig. 9a), the reduced warming is centered at \~55°N. In the subtropics this produces an increased lower-troposphere temperature gradient at the same latitude as the enhanced upper-level gradient. To the north there is a strongly weakened meridional temperature gradient. This leads to a barotropic northward shift in the zonal wind, which can be seen in Fig. 9b. Consistent with this there is a northeastward shift in the track density (Fig. 7d) at lower levels, which is approximately collocated with the shift in the upper-level jet (Fig. 6b).

In the 4 \times \text{CO}_2 experiment (Fig. 9c), the region of reduced warming occurs over a greater latitude range centered on 45°N, thereby moving the position of the lower-troposphere temperature gradient increase farther south. The relative strength of the temperature gradient is also less than in the 2 \times \text{CO}_2 experiment, giving a much smaller signal at low levels. The displacement of the upper- and lower-level increased meridional temperature gradient is consistent with a baroclinic response in the zonal wind. The lower-level zonal wind over the North Atlantic, therefore, is not shifted northeastward as in the 2 \times \text{CO}_2 experiment.

The patterns of temperature change relative to the present day control are the same in the two experiments at upper levels, associated with the strongly enhanced warming in the tropical upper troposphere. However, the patterns of temperature change are different near the surface between the two experiments, associated
with different zonally averaged responses in the zonal wind over the North Atlantic. The northeastward shift of the storm track in the $2 \times CO_2$ run, but no evident shift of the storm track in the $4 \times CO_2$ run, is consistent with changes to the temperature gradients and zonal winds. These results indicate that, owing to competing influences, regional changes to the extratropical storm tracks seen in the $2 \times CO_2$ experiment are not necessarily amplified in the $4 \times CO_2$ experiment. This is somewhat different from previous studies such as that of Pinto et al. (2007) who found that the same patterns of change increased in magnitude with increasing CO2 forcing, although experiments with such large forcing were not examined. This suggests that, as CO2 forcing becomes very large, the response is nonlinear.

7. Changes in extratropical cyclone intensity

In this section the changes in extratropical cyclone intensity and number that occur under increased CO2 concentrations in HiGEM are investigated. Initially changes to the total number of cyclones are considered. In the NH for DJF, the mean number of cyclones in the present-day control simulation is 127.9 per month, in the $2 \times CO_2$ run 126.0 per month, and in the $4 \times CO_2$ run 123.9 per month. The small decrease in the total number of cyclones from the present day control to the $2 \times CO_2$ run is not significant at the 95% level (according to a two-sample Student’s $t$ test). The decrease of four cyclones per month (3%) from the control to the $4 \times CO_2$ run is significant. The reduction in extratropical cyclone numbers is consistent with other climate model studies (e.g., Bengtsson et al. 2009) and is associated with the reduced meridional temperature gradient and low-level baroclinicity.

As mentioned in the introduction, previous climate model studies have reported conflicting results concerning how the intensity of extratropical cyclones are predicted to change in response to anthropogenic forcing. Lambert and Fyfe (2006) found an increase in cyclone intensity with a warming climate. However, other studies have found decreases in extratropical cyclone intensity
over the NH (e.g., Leckebusch and Ulbrich 2004; Pinto et al. 2007; Bengtsson et al. 2006; Bengtsson et al. 2009). The results from the HiGEM idealized warming scenario are of interest since HiGEM has a higher horizontal resolution than typical climate models. It has already been shown, for example, that HiGEM is capable of capturing the observed synoptic-scale structure of intense extratropical cyclones (Catto et al. 2010).

Frequency distributions of extratropical cyclone intensity are presented in Fig. 10. Figure 10a shows histograms of the minimum MSLP along each of the cyclone tracks in the Northern Hemisphere for the present day control and the $2 \times \text{CO}_2$ and the $4 \times \text{CO}_2$ experiments. It can be seen that there is very little difference in the MSLP distributions between the control and the idealized experiments. The tails of the distributions reveal a very slight increase in the number of cyclones with extreme minimum MSLP with increasing CO$_2$ levels; however, the sampling in the tails is small. The impact of removing a background field, similar to Bengtsson et al. (2009) and Donohoe and Battisti (2009), has been investigated and found to have no effect on the conclusions reached here. It has also been found that normalization of the distributions by the total cyclone numbers does not change the tails of the distributions.

The distributions of maximum 850-hPa vorticity are shown in Fig. 10b. Again, the overall shape of the distributions is very similar between the three models runs, with a peak between 10 and 20 ($\times 10^{-5}$ s$^{-1}$). The tails of the frequency distributions (scaled to 10 yr) indicate a small systematic decrease in intensity with increasing CO$_2$ levels.

Although MSLP and vorticity are often used as dynamical measures of intensity, they are not as relevant as wind speeds for climate impacts. Distributions of the wind speeds at 850 hPa are shown in Fig. 10c. Ideally, a level closer to the surface would be used, as in Bengtsson et al. (2009); however, 850 hPa is the lowest level on which wind speeds were archived at 6-hourly intervals for the HiGEM1.1 simulations. The control and $2 \times \text{CO}_2$ experiment have very similar distributions of wind speed, even in the tail, with the reduction in total numbers the prominent feature. The wind speed distribution from the $4 \times \text{CO}_2$ experiment shows more of a shift to lower wind speeds, a result also seen in the tail of the distribution.

The results found here, indicating very little difference in the intensity in terms of MSLP and a decrease in terms of vorticity and wind speeds with increased CO$_2$ forcing, are consistent with the findings of Bengtsson et al. (2009) and Watterson (2006). Della-Marta and Pinto (2009) and Sienz et al. (2010), using extreme value analysis, also did not find any significant changes for extreme cyclones over the North Atlantic, based on several measures of cyclone intensity. However, over smaller regions, particularly close to Europe, increased intensity of cyclones was seen by Della-Marta and Pinto (2009), in agreement with earlier results from Pinto et al. (2007). These changes are consistent with an extension of the jet stream toward Europe. The use of extreme value statistics, such as generalized extreme value distributions, and return values, such as those used in Della-Marta and Pinto (2009) and Sienz et al. (2010), have proved useful in quantifying the changes to the most intense cyclones in a warming climate, but this analysis is beyond the scope of the present study. Lambert and Fyfe (2006) found an increase in the number of intense cyclones in the IPCC AR4 models (with intensity defined by MSLP) in contrast to the studies previously mentioned. This difference may be due to the exact definition of intense cyclones in the various studies (Ulbrich et al. 2009).

8. Discussion and conclusions

Two idealized climate warming experiments were performed with the HiGEM fully coupled climate model with CO$_2$ levels of two and four times the present-day control levels. These experiments have been used to investigate the response of the Northern Hemisphere extratropical storm tracks to increased CO$_2$ concentrations. There have been other studies on this aspect of climate change; however, the model used here is of higher resolution than typical of the IPCC AR4 models. The use of two different experiments allows the nonlinearity of the system to be compared for increasing CO$_2$ levels.

There are two main components to the changes in extratropical cyclones that are of interest. The first is the spatial distribution of storm tracks. This is an important consideration, as there may be shifts in the location of extreme rainfall, flooding, strong winds, or droughts associated with increased or decreased cyclone numbers. This spatial distribution is strongly influenced by changes in the mean climate, such as surface and tropospheric temperatures. The second important component is the change to the intensity of winds associated with extratropical cyclones. This would also have socioeconomic consequences in terms of potential damage due to the cyclones.

The main conclusions of this study are as follows.

- The zonal mean circulation changes in the $2 \times \text{CO}_2$ and $4 \times \text{CO}_2$ experiments are consistent with the upward shift of the upper-level jet, seen in the multimodel mean of the IPCC AR4 models (Yin 2005; Lorenz and DeWeaver 2007; Woollings 2008); however, there is little evidence of a zonal-mean poleward shift.
There is a reduction in track density over the Mediterranean and the Middle East, as has been seen in other studies (IPCC, Christensen et al. 2007).

In the North Pacific the storm track shows a northward shift in both $2 \times CO_2$ and $4 \times CO_2$ experiments. There is a large increase in track density on the eastern side.
of the Pacific in the $4 \times CO_2$ experiment, which is most likely due to the zonally extended upper-level jet, associated with the weakened tropical Walker circulation (Vecchi and Soden 2007).

- Over the North Atlantic there is a northward shift of the storm track in the $2 \times CO_2$ experiment that gives an increased track density over the United Kingdom, similar to the results of Leckebusch and Ulbrich (2004), Bengtsson et al. (2006), and Pinto et al. (2007). This is consistent with the barotropic response of the zonal winds to temperature changes; there is enhanced warming in the tropical upper troposphere and at the surface over the North Pole and a narrowly confined region of reduced lower-troposphere warming in the midlatitude North Atlantic due to the slowdown of the MOC.

- In the $4 \times CO_2$ experiment, the reduced lower-troposphere warming is more widespread, leading to a cancellation of the upper- and lower-level meridional temperature gradients. This is consistent with the baroclinic structure in the response of the zonal wind with very little signal near the surface. This is also consistent with no shift in the low-level track density over the North Atlantic in the $4 \times CO_2$ experiment and the track density near the western region of the United Kingdom staying close to the present-day control values.

- In agreement with many other studies (e.g., Geng and Sugi 2003; Leckebusch and Ulbrich 2004; Lambert and Fyfe 2006; Bengtsson et al. 2009), there is a decrease in cyclone numbers in both of the climate warming experiments due to a decrease in baroclinicity associated with enhanced polar warming.

- The intensity (in terms of vorticity, wind speed, and mean sea level pressure) of the most extreme cyclones in HiGEM1.1 decreases slightly in the NH with increasing CO$_2$ levels, similar to Bengtsson et al. (2009).

That the number of intense extratropical cyclones slightly decreases with increasing CO$_2$ concentrations is consistent with the results of Bengtsson et al. (2009), Watterson (2006), and Geng and Sugi (2003), but inconsistent with Lambert and Fyfe (2006). This disparity may partly arise from the methodology used to assess the intensity of extratropical cyclones (Sinclair and Watterson 1999; Ulbrich et al. 2009), which emphasizes the need for a more systematic assessment of extratropical cyclones in climate models. Furthermore, a decrease in the number of intense extratropical cyclones would appear to be at odds with the argument that in a warmer, moister climate, enhanced latent heat release would result in more intense extratropical cyclones. A possible explanation could be that, while the extra latent heat release is acting to intensify the cyclones, the reduction in baroclinicity due to meridional temperature gradient changes acts to reduce the intensity. Understanding why some climate models predict a decrease in intense extratropical cyclones in response to climate change will require further investigation.

The results from this study indicate that the response of the North Atlantic storm track is sensitive to a number of different processes that may change as CO$_2$ concentrations increase. First, the enhanced warming at low levels over the Arctic, which acts to decrease the equator-to-pole temperature gradient, is important for the low-level baroclinicity in the North Atlantic region. Although the models from the IPCC AR4 consistently predict this enhanced warming, they do so to different extents.

The second process is the enhanced warming of the tropical upper troposphere. This acts to shift the strongest upper-level temperature gradients poleward, thereby shifting the upper-level jet. This enhanced warming in the tropical upper troposphere is consistent between models of the IPCC AR4 (e.g., Yin 2005; Woollings 2008).

The third process that may have an impact on the response of the extratropical storm tracks in the North Atlantic region is the weakening of the MOC. The weaker transport of warm water associated with the slowdown of the MOC reduces the local warming of the SSTs in the North Atlantic. This influences the meridional temperature gradients and hence the position of the maximum zonal winds and the storm track. There is a large spread in the magnitude of the weakening of the MOC, and hence the North Atlantic SST response, between the models used in the IPCC AR4 (Meehl et al. 2007).

The prominence of each of these processes (and potentially others not considered here) may produce different regional responses in the extratropical storm tracks. Lainé et al. (2009) found that the response of the North Atlantic storm track to the $4 \times CO_2$ forcing in two climate models was sensitive to the model representation of the change in North Atlantic SSTs. In this study we find that the sensitivity of North Atlantic SSTs to different levels of CO$_2$ forcing gives rise to different responses in the North Atlantic storm track. Developing a deeper understanding of these processes, and how they interact, will be essential to producing more robust projections of extratropical climate change.

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