Changes in Cyclone Characteristics in Response to Modified SSTs

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ABSTRACT

The impact of changes in sea surface temperature (SST) on the statistics of extratropical cyclones is investigated. The cyclones were identified in an atmospheric general circulation model (AGCM) using an objective Lagrangian tracking algorithm, applied to the 850-hPa relative vorticity. The statistics were generated for several 20-yr simulations, in which the SSTs were warmed or cooled by 2 K in latitudinal bands. The response was studied in both hemispheres, during summer and winter.

Changes in the position of the storm tracks are largely consistent with those seen in previous studies. Increasing SSTs uniformly or increasing the midlatitude SST gradient results in a poleward shift in the storm tracks, with the clearest trends seen in the Southern Hemisphere (SH). Here it is demonstrated that the SST modifications alter the cyclone characteristics as well. When the warming includes the low latitudes and/or the midlatitude gradient is increased, there are more short-lived cyclones. These are also on average more intense and translate faster, both poleward and eastward.

The poleward displacement is correlated with cyclone intensity, so that stronger cyclones translate to higher latitudes. This is suggestive of vortex self-advection in the presence of a mean potential vorticity (PV) gradient. The increased eastward translation is correlated with the depth-averaged zonal velocity, and so is likely related to an increase in the steering-level velocity. These changes in cyclone translation probably contribute to the changes in the storm tracks seen previously.

1. Introduction

Storm tracks are regions of heightened cyclone activity, located at midlatitudes in both hemispheres and intensified over the major oceans. The Northern Hemisphere (NH) contains two main storm tracks, over the North Atlantic (NA) and the North Pacific (NP). These stretch from the east coast of North America and Asia across the ocean basins. The NA storm track, in particular, displays a characteristic poleward tilt. An additional track, associated with the Mediterranean Sea, is present during winter. In the Southern Hemisphere (SH), a single storm track encircles Antarctica at approximately 50°S. This is more zonally symmetric than its NH counterparts, but its intensity does vary somewhat with longitude.

The storm tracks are linked to regions of strong surface temperature contrast, such as between the land and ocean in the NH entrance regions (e.g., Hoskins and Valdes 1990) and in oceanic frontal zones (e.g., Nakamura and Shimpo 2004; Nakamura et al. 2008; Sampe et al. 2010; Hotta and Nakamura 2011; Ogawa et al. 2012). The latter occur both near land, as with the Gulf Stream and Kuroshio, and over the open ocean, as with the Antarctic polar frontal zone (APFZ) in the southern Atlantic and Indian Oceans. The frontal zones exert a restoring effect on low-level atmospheric baroclinicity, maintaining the meridional temperature gradients against the mixing induced by the storms. They therefore act as an “anchor” for the storm tracks (Nakamura and Shimpo 2004; Nakamura et al. 2004, 2008; Hotta and Nakamura 2011; Sampe et al. 2010). Model simulations show that the storm tracks weaken and shift equatorward when frontal zones, for instance based on Indian Ocean climatology, are removed (e.g., Nakamura et al. 2008; Sampe et al. 2010).

The literature suggests that the storm tracks have shifted poleward in recent years, both in observations and reanalysis data (e.g., McCabe et al. 2001; Fyfe 2003; Chang 2007; Wang et al. 2006; Ulbrich et al. 2009; Vilibić and Šepić 2010; Bender et al. 2012). Similar shifts have been found in models forced under global warming scenarios (e.g., Yin 2005; Bengtsson et al. 2006; Ulbrich...
et al. 2008; Schuenemann and Cassano 2010; Wu et al. 2010; Chang et al. 2012). The shift is particularly clear in the SH (e.g., Chang et al. 2012).

A number of mechanisms have been proposed to explain the shift. Among these is increased atmospheric heating by the ocean. That the ocean is warming is becoming clear (e.g., Levitus et al. 2005; Bindoff et al. 2007). Increasing sea surface temperatures (SSTs) in models warms the overlying atmosphere, changing tropospheric baroclinicity and thus affecting the storm tracks (e.g., Caballero and Langen 2005; Kodama and Iwasaki 2009; Lu et al. 2010; Graff and LaCasce 2012, hereafter GL12). When the meridional SST gradients are altered, the resulting change in storm-track position depends on the location and the sign of the change (e.g., Brayshaw et al. 2008; Chen et al. 2010; GL12). Increasing the gradient at midlatitudes causes the tracks to shift poleward, whereas decreasing the gradient at midlatitudes or increasing it in the tropics produces an equatorward shift.

The storm tracks, and their changes, are usually diagnosed in two ways. The first involves using bandpass-filtered fields, such as geopotential height or eddy heat and momentum fluxes, to isolate variability in the 2–7-day band (e.g., Blackmon 1976; Blackmon et al. 1977; Yin 2005; Sampe et al. 2010; GL12). The second focuses on changes in the cyclones themselves, their positions and regions of cyclogenesis and cyclolysis (e.g., Hoskins and Hodges 2002, 2005; McCabe et al. 2001; Bengtsson et al. 2006; Wang et al. 2006; Bengtsson et al. 2009; Catto et al. 2011). The first approach is an Eulerian analysis, based on averaging fields by location and height; the second, which involves tracking individual storms, is closer to a Lagrangian one.

In addition to knowing how the storm tracks will shift, it is also of interest to know how the cyclones themselves will change in the future climate. While the total number of cyclones has been projected to decrease (e.g., Bengtsson et al. 2006; Pinto et al. 2007; Ulbrich et al. 2009), projections regarding their intensities differ substantially, with some indicating an increase (e.g., Lambert and Fyfe 2006; Mizuta et al. 2011) and others finding weak or no changes (e.g., Bengtsson et al. 2006, 2009; Pinto et al. 2007; Catto et al. 2011). Indeed, trends differ substantially between regions, with increases in intensity in some places and decreases in others. The picture is further complicated by the use of different tracking algorithms and measures of intensity (Ulbrich et al. 2009, 2013), although the results for the strongest cyclones are more consistent (Ulbrich et al. 2013).

The present study examines how cyclone characteristics change under controlled SST forcing in an atmospheric general circulation model (AGCM). The cyclones are identified using an objective Lagrangian tracking routine (Hodges 1994, 1995, 1999) applied to anomalies in relative vorticity on the 850-hPa surface ($\zeta_{850\text{hPa}}$). The AGCM is the Community Atmosphere Model, version 3 (CAM3), and the runs are the same as those analyzed by GL12, who examined changes in bandpass-filtered statistics during boreal winter. The cyclone-based results verify the changes described by GL12, but additionally reveal changes in the cyclone characteristics. Furthermore, while GL12 focused on the zonally averaged response in the latitude–pressure plane, we concentrate on the horizontal (latitude–longitude) plane and describe regional (e.g., Atlantic versus Pacific in the NH) variations. We also expand the analysis to include both the summer and winter seasons in both hemispheres.

The data and methods are described in section 2. The results are presented in section 3, and are summarized and discussed in section 4.

2. Data and methods

a. Model and reanalysis data

For the analysis, we use 6-hourly data from the simulations presented in GL12. These were carried out for a 20-yr period\(^1\) using the National Center for Atmospheric Research (NCAR) CAM3, with climatological input data from the 1980s and 1990s (Collins et al. 2004). The model was run with a spectral T42 horizontal and 26-level vertical (T42L26) resolution, and was integrated with an active land surface (the Community Land Model), a thermodynamic sea ice model, and the Data Ocean Model (DOM). The latter reads and interpolates prescribed time-varying SST data, in this case the built-in climatological dataset derived from the Smith/Reynold EOF dataset for 1981–2001 (Reynolds et al. 2002; Collins et al. 2004, 2006). The dataset includes 12 monthly samples that are cycled throughout the integration period. Note that the sea ice coverage is prescribed when using DOM. As such, the primary function of the sea ice model is to compute the surface fluxes between the ice and atmosphere.

To validate the CAM3 results, we used 6-hourly reanalysis data from the European Centre for Medium-Range Weather Forecasts (ECMWF) Interim Re-Analysis (ERA-Interim; Dee et al. 2011). ERA-Interim is simulated using the 2006 version of the ECMWF Integrated Forecast System (IFS), including fully

\(^1\) Control simulations were also performed over a 30-yr period, and for a 20-yr period with and without random initial perturbations to the initial temperature field, yielding similar results. Thus the 20-yr integration was deemed sufficient.
coupled components for the atmosphere, land surface, and ocean waves. The IFS is integrated with a four-dimensional variational data assimilation (4DVAR) scheme at spectral T255 horizontal and 60-level vertical (T255L60) resolution. Although data are available for a longer time period, we consider the same 20-yr period from which the climatological input data in CAM3 were constructed. We also interpolated the ERA-Interim data to the CAM3 grid, to simplify comparisons.

b. SST perturbations

We consider five CAM3 runs, one control and four sensitivity runs. The SSTs from the control run are shown in Fig. 1. The left and right panels show the field averaged over the NH winter and summer periods in the three-month blocks December–February (DJF) and June–August (JJA), respectively. In the sensitivity runs, a 2-K modification was imposed to the climatological SSTs in different regions (see Figs. 1 and 2 in GL12). In the 2K run, the SSTs were increased uniformly over the oceans, while in the three remaining runs the SSTs were increased/decreased equatorward or poleward of 45° (indicated by the dark horizontal lines in Fig. 1). The modifications were tapered to zero over an 8° range. Thus the three latter forcings alter the SST gradients at 45°, where the climatological SSTs have a largely zonal orientation.

In the 2K-lowlat run, the low latitudes were warmed equatorward of 45°. Thus both the low-latitude SSTs and the midlatitude SST gradients were increased. In the ±2K-highlat runs, the SSTs were warmed and cooled poleward of 45°. Cooling the high-latitude SSTs, in the −2K-highlat run, allows investigating the effect of stronger midlatitude SST gradients without increased low-latitude heating. The opposite effect is examined in the 2K-highlat run, as the midlatitude gradient is weakened. Note too that only the SSTs were altered; the sea ice distribution was unchanged in all cases.

The modifications to the midlatitude SST gradients in the 2K-lowlat and ±2K-highlat runs actually vary between ocean basins, depending on the climatology (not shown). During DJF, the zonally averaged unperturbed midlatitude SST gradient in the SH has one large maximum at about 43°S, which strengthens and broadens toward the pole in the 2K-lowlat and −2K-highlat runs. In the 2K-highlat run, the gradient weakens between 42° and 55°, and shifts equatorward by about 3°.

The NA winter gradient has two minima, one at 43°N (reflecting the Gulf Stream) and a weaker one at 51°N (reflecting the North Atlantic Current). The general picture is similar to SH DJF: the strongest gradient intensifies and broadens in the 2K-lowlat and −2K-highlat runs; in the 2K-highlat run the gradient generally weakens but also shifts somewhat equatorward.

In the NP, the winter gradient is strongest at 40°N. The minimum intensifies and broadens in the 2K-lowlat and −2K-highlat runs; in the 2K-highlat run the gradient generally weakens but also shifts somewhat equatorward.

c. Cyclone tracking

We computed the individual cyclone trajectories using the objective tracking routine of Hodges (1994, 1995, 1999). This employs a two-step process: first, the feature points (the positions of the cyclone centers) are identified;
second, they are linked into trajectories by minimizing a cost function that restricts variations in direction and speed between consecutive time steps. Off-grid feature points are identified through interpolation/smoothing and treated as well.

Different fields may be used for tracking, such as geopotential height, mean sea level pressure (MSLP), or relative vorticity (e.g., Hoskins and Hodges 2002, 2005; Bengtsson et al. 2006, 2009; Catto et al. 2011). These reflect different aspects of the flow field. MSLP is more influenced by large-scale features like the Icelandic low and strong mean flows (e.g., Hoskins and Hodges 2002). Relative vorticity better captures small-scale features and is less influenced by the background flow (e.g., Hoskins and Hodges 2002). It also allows for identifying the cyclones earlier in their life cycle and better reflects the centers of rotation (Hodges 1996).

We track cyclonic vortices in the $\zeta_{850\text{hPa}}$ field in both hemispheres, during their respective winter and summer seasons. We neglect weak, short-lived, and slow-moving systems, requiring that the cyclones 1) have a $\zeta_{850\text{hPa}}$ exceeding $10^{-5}$ s$^{-1}$ in magnitude, 2) last for at least 2 days, and 3) move more than 10° geodesic. The $\zeta_{850\text{hPa}}$ field was spatially filtered prior to the tracking process, removing scales with planetary wavenumbers less than 5 and larger than 42 (e.g., Hoskins and Hodges 2002). All $\zeta_{850\text{hPa}}$ values reported in the following stem from the filtered fields.

We focus initially on the track density fields, which indicate the concentration of cyclone trajectories (Hodges 1996). This facilitates the comparison with GL12 and other studies. Thereafter we concentrate on the cyclones themselves, examining quantities averaged among a given population (e.g., from a selected storm track). This is effectively a more Lagrangian approach.

**d. Statistics**

We make use of probability density functions (PDFs) to illustrate the range of values present, and PDFs from different runs are compared. To determine where changes with respect to the control run are significant, we use two different tests: the two-sample Kolmogorov–Smirnov test (KS test) and the Wilcoxon–Mann–Whitney rank-sum test for two independent samples (WMW test) (e.g., Wilks 2006). The former determines the probability that the largest difference between two cumulative density functions (CDFs) can occur. The CDF is the cumulative integral of the PDF. An advantage with the KS test is that it is nonparametric; that is, it makes no assumptions about the underlying distribution.

The WMW test, like the Student’s $t$ test, is used to determine whether one sample population tends to have significantly larger values than the other. It determines the probability of the sum of the ranks of the two samples being significantly different, with the ranks taken from the pooled data. The WMW test is favorable to the more traditional $t$ test as it is nonparametric (like the KS test) and less influenced by outliers. But even when the assumptions of the $t$ test are met, the WMW test is almost as good (Wilks 2006).

The probability in the KS test depends on the numbers of degrees of freedom in the two datasets. The WMW test moreover assumes that the data constituents are statistically independent. Most of the samples considered are assumed to be independent as they include only one data point from each cyclone trajectory (such as lifetime or lysis latitude). However, we also evaluate samples of feature-point latitudes. These are retrieved at 6-h intervals, and as such the adjacent points are not necessarily independent. Calculating the autocorrelation of the cyclone speeds along the tracks, we found that these decay approximately exponentially, with decay times from 1–4 days in the different basins. So we reduced the number of degrees of freedom assuming the feature points were independent after 3 days.

3. **Results**

a. **Changes in baroclinicity**

The atmospheric response to changes in SST was explored by GL12, but it is useful to re-examine the effect on baroclinicity, for reference with the later results. GL12 also presented zonally averaged boreal winter fields, but we will consider the response in the SH and in the two separate NH basins (the NA and NP), in both summer and winter.

Shown in Figs. 2 and 3 is the Eady parameter (Lindzen and Farrell 1980), defined as

$$\sigma_{B1} = 0.31 \frac{g}{NT} \left| \frac{\partial T}{\partial y} \right|. \quad (1)$$

where $N$ is the Brunt–Väisälä frequency (the remaining notation is standard). This is a common measure of baroclinicity and indicates the maximum growth of baroclinic disturbances. We consider the zonally averaged parameter as a function of pressure. Plotted in the figures are the differences in $\sigma_{B1}$ between each of the four sensitivity experiments and the control run. The result for the latter is indicated by the solid contours (with values of 4, 6, and 8 days$^{-1}$ for NH winter, and of 3, 5, and 7 days$^{-1}$ for summer), and the former by the color shading. The results from three regions (the SH, NA, and NP) are shown from left to right, separated by black vertical lines.

The strongest response is in the 2K-lowlat run (in which SSTs are increased poleward of 45°), as seen in Figs. 2a
In DJF (Fig. 2a), the maxima in baroclinicity shift poleward, all three regions. The shift is evident at both the jet level and at low levels. In JJA (Fig. 2b), a poleward shift is still seen in the SH. The changes occur over a larger meridional extent, and subsequent results will show that the storm track broadens similarly. This is consistent with the SH being less zonally symmetric in winter, exhibiting a characteristic spiraling toward the pole (e.g., Hoskins and Hodges 2005). A poleward shift is also evident in the NA during JJA, but the maximum shifts equatorward in the NP (Fig. 2b). This was not noted by GL12, who only examined the zonally averaged fields in DJF.

The response in the 2K run (in which SSTs are increased uniformly over all oceans) is similar (cf. Figs. 2c,d with Figs. 2a,b). The main difference is that the low-level changes are weaker. Excess latent heating in the tropics increases the temperature gradient aloft, but the surface gradient is largely unaffected by the uniform warming. The exception is again the NP track (Figs. 2c,d) where the changes resemble those seen in the 2K-lowlat run at the jet level and at low levels.

Many similar features are also seen with the −2K-highlat run, albeit with smaller amplitude and a more barotropic structure. There is a poleward shift in baroclinicity in all three regions in the DJF period (Fig. 3a), and similar, though less pronounced, shifts in JJA (Fig. 3b). The −2K-highlat run lacks the large changes near the jet level seen in the 2K-lowlat run, as there is no excess heating in the tropics. Furthermore, there is little change in the NP in JJA, indicating that the polar cooling has little effect on the (relatively high latitude) baroclinicity there. The difference from the previous runs is presumably due to the lack of low-latitude heating.

The changes in the 2K-highlat run resemble those in the −2K-highlat run, but with the opposite sign. There is an equatorward shift in the SH and NA baroclinicity maxima in DJF (Fig. 3c) and in the SH in JJA (Fig. 3d). However, the response in the NP differs, particularly in
DJF. Then, there is a clear poleward shift in the \(-2\)K-highlat run and little to no response with the 2K-highlat perturbation.

To summarize, the forcings generate two distinct changes in baroclinicity. In the runs where the warming includes the low latitudes, the baroclinicity increases mostly aloft, at the jet level. In the cases where the surface gradients are altered, the changes in baroclinicity are more barotropic, extending up to about 300 hPa. Increasing either the low-latitude warming or the midlatitude SST gradient results in a poleward shift in baroclinicity. The exception is the NP in summertime, where the response is not always so clear. As shown by GL12 for DJF, the changes in the storm tracks (in terms of bandpass-filtered geopotential height variability and eddy fluxes and heat and momentum) largely mirror those seen in baroclinicity.

b. Changes in storm-track position by track density

We now consider changes in the storm tracks as seen by tracking individual cyclones. First, we examine the track density, indicating the concentration of cyclone trajectories (Hodges 1996). As this is a geographical mapping of the storm paths, it most closely resembles the Eulerian fields discussed in GL12. We consider track density fields from the control run, from ERA-Interim for validation, and from the 2K-lowlat run. As will be shown, the latter case presents the clearest response of the sensitivity runs, and we therefore focus more on this run. The remaining sensitivity runs are considered later.

The control run fields are shown in Fig. 4. As always, the SH storm track is much more zonal than its NH counterparts, particularly in summer (DJF). It is centered at about 50°S and exhibits only weak longitudinal variations (Fig. 4a). There are greater such variations in winter though, with the largest concentration of storms southwest of Australia (Fig. 4b). In the NH, there are three distinct tracks in wintertime, in the NA, NP, and Mediterranean region (Fig. 4c). The first two display a characteristic southwest–northeast tilt, with the NA track for example bending toward the United Kingdom. In summer, (Fig. 4d) the tracks are narrower and weaker, and the Mediterranean track is absent. In addition, the NA and NP summer tracks are poleward of the winter positions.
It is useful to compare with the track densities obtained with the ERA-Interim data (Fig. 5). The ERA-Interim fields are very similar to those from CAM3, both in location and magnitude, but there are differences as well. The maximum southwest of Australia in the SH winter is much weaker in ERA-Interim than in CAM3 (Figs. 5b and 4b); however, a corresponding inconsistency is not evident when considering the feature-density field (not shown). There are also more storms off the Antarctic coast, near Wilkes Land and over the Amundsen and Bellinghausen Seas. The latter is likely related to differences in sea ice coverage between CAM3 and ERA-Interim. The structure of the SH summer track is more similar (Figs. 4a and 5a), although the CAM3 track has greater magnitudes at midlatitudes and somewhat smaller magnitudes near the Antarctic coast.

In NH winter, the NP storm track and the entrance region of the NA track are very similar between CAM3
However, the exit region of the NA track lies farther north in ERA-Interim, being intensified off the southern tip of Greenland. Correspondingly, ERA-Interim exhibits more storms over the Nordic Seas and into the Arctic. That the storm tracks are too zonal is a known issue with CAM3 (Hurrell et al. 2006). Perhaps the most striking difference, though, is that the NP track lies much farther south in summer in ERA-Interim than in CAM3 (Figs. 5d and 4d). The NA summer track, however, is more similar, albeit somewhat too zonal. Beyond the mentioned differences the agreement is reasonably good, particularly in light of the differences between the two datasets; ERA-Interim has dramatically higher spatial resolution (T255L60 in ERA-Interim versus T42L26 in CAM3) and uses observed SST and sea ice data varying from year to year, whereas CAM3 was run with a climatological dataset.
The changes in storm-track position induced by the 2K-lowlat forcing are shown in Fig. 6. The response is again clearest in SH summer (Fig. 6a), with a decrease in track density equatorward of 50°S and an increase near 60°S. The changes are consistent with those in the Eady parameter at low levels (Fig. 2a), except that the track changes occur about 10° poleward. As noted in GL12, the changes in bandpass-filtered variability likewise occur poleward of those in baroclinicity. The difference field for the winter tracks (Fig. 6b) is more spread out latitudinally, in line with the control run (Fig. 4b), but the poleward shift is clear nevertheless. Note too there is a decrease in storms over the Ross Sea, but this is in a region where CAM3 differs with reanalysis (Fig. 5b).

The changes in the NH, as expected, vary more spatially. Nevertheless, the NP and NA winter tracks clearly shift poleward, the Mediterranean track weakens, and...
there are more storms occurring over northern Europe (Fig. 6c). The changes in the NH summer (Fig. 6d) are less clear. There is a poleward shift downwind of the Rockies and in the entrance region to the NA track, but the NP track on the other hand shifts equatorward, in line with the Eady parameter (Fig. 2).

c. Changes in feature-point latitude

The track changes can be seen succinctly in PDFs of feature-point latitude. This reflects the cyclone-center concentration (and only includes feature points that have been assigned to a trajectory). We subdivide the results according to the genesis regions shown in Fig. 7, corresponding to the SH, NA, and NP storm tracks. Note that the regions differ between winter and summer, in line with the changes in the tracks described above. In keeping with our focus on extratropical storms, tropical cyclones are excluded (details are given in the caption). We also exclude a small group of cyclones that form in the low subtropics (25°–35°S) in the SH summer. These exhibit distinctly different behavior than the midlatitude storms, as discussed in section 4.

The PDFs for the control run are shown in Fig. 8 (solid lines), with the PDFs from ERA-Interim for comparison (dotted lines). In the SH in DJF (Fig. 8a), the majority of storms in the control run occur between 45° and 65°S, with a peak near 53°S. The distribution from ERA-Interim is similar, albeit slightly broader and shifted poleward. The agreement is better in SH winter (Fig. 8b), and both PDFs cover a broader range of latitudes. The agreement is similarly good in NH winter, in both the Atlantic (Fig. 8c) and Pacific (Fig. 8e) tracks. It is less convincing in NH summer (Figs. 8d,f). The CAM3 distribution is more sharply peaked, with the mode near 55°N in both basins, while there are more storms toward the equator in ERA-Interim and the modes lie to the south.

Also shown in the figure are the differences between the PDFs from the 2K-lowlat and control runs, indicating latitudinal shifts in the storm concentrations (orange lines with gray shading). The significance of the differences is assessed using the KS test, described in section 2. In the SH summer (Fig. 8a), the difference PDF is positive south of 55°S and negative to the north, indicating a poleward shift. The differences are more significant by the KS test. In the SH winter (Fig. 8b), the changes are more spread out, like the PDFs themselves, but the differences are significant and indicate a poleward shift. The change in the NA winter (Fig. 8c) is likewise poleward and significant. The PDF shifts poleward too in the NP in winter (Fig. 8d) and in the NA in summer (Fig. 8e), but the differences in both cases are not significant. The PDF is significantly different in the NP in summer, though (Fig. 8f), showing an equatorward shift. The results of the WMW test (section 2) were largely consistent with those from the KS test (Tables 1 and 2).

The changes in the mean feature-point latitudes are listed by storm-track region and season in Tables 1 and 2 for the 2K-lowlat and the other sensitivity runs.
Fig. 8. Feature-point latitude PDFs for the control run (solid black line), ERA-Interim (dotted black line), and 2K-lowlat difference with respect to the control run (solid orange line with light and dark gray shading). The three PDFs are shown for (a),(b) SH summer and winter cyclones and for NA and NP (c),(d) winter and (e),(f) summer cyclones. When the difference between the 2K-lowlat and the control PDFs are significant by the KS test at the 95% level, an asterisk is given in the bottom-left corner.
changes are generally modest. For example, the mean storm-track latitude in the 2K-lowlat run shifts poleward by 1.94% in the SH winter, corresponding to 1°. In winter, the track shifts by 0.66% (0.4°). The NA track shifts poleward by 2.01% (1.3°) in winter, while the NP track shifts by 1.47% (0.7°). In summer, the mean for the NA track shifts barely at all, while it shifts equatorward by 1.62% in the NP, corresponding to about 1°.

Thus, in most regions the storm tracks shift poleward under the 2K-lowlat warming. The exception is the NP, where the track shifts equatorward. But as noted, the track in the NP lies too far north in CAM3; the result might have been otherwise had it been in the proper position.

The results for all four sensitivity runs are shown in Fig. 9. Plotted are the changes in mean feature-point latitude as well as in the mean genesis and lysis latitudes.

### Table 1. Overview of the mean feature-point latitude (FP), genesis latitude (gen.), lysis latitude (lys.), 5-day latitudinal and longitudinal displacements (dlat and dlon), lifetimes ($T_{life}$), maximum intensity ($\zeta_{500\,hPa}$), and total cyclone count ($N_c$) for the SH summer and winter. For the control run, we show the full values. For the sensitivity runs, we show the percent of change relative to the control run value for the relevant region. When the change in location is significant at the 95% level according to the WMW test, values are given in boldface. The significance of the changes in the cyclone count is not assessed.

<table>
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<th>FP (°S)</th>
<th>Gen. (°N)</th>
<th>Lys. (°N)</th>
<th>Dlat (°S)</th>
<th>Dlon (°E)</th>
<th>$T_{life}$ (days)</th>
<th>$\zeta_{500,hPa}$ (10⁻² s⁻¹)</th>
<th>$N_c$</th>
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<td>0.87%</td>
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</table>

### Table 2. As in Table 1, but for the NA and NP storm tracks.

<table>
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<tr>
<th></th>
<th>FP (°N)</th>
<th>Gen. (°N)</th>
<th>Lys. (°N)</th>
<th>Dlat (°N)</th>
<th>Dlon (°E)</th>
<th>$T_{life}$ (days)</th>
<th>$\zeta_{500,hPa}$ (10⁻² s⁻¹)</th>
<th>$N_c$</th>
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Encircled points indicate significant changes by the WMW test (see also Tables 1 and 2). In the 2K-lowlat case (left column of Fig. 9a), the mean latitude is seen to shift poleward in all regions, except in NP summer. The response in the −2K-highlat and 2K runs (second and third columns) is similar, with poleward shifts of up to 0.5°–1.1° in most cases. The summer tracks in the SH and the NP shift equatorward in the 2K and −2K-highlat runs, respectively, but the changes in both cases are insignificant. Interestingly though there is essentially no change in the 2K-highlat run, indicating that the high-latitude warming has little impact on the low-level storm tracks.

The changes in genesis latitude (Fig. 9b) are generally consistent but less pronounced. A 0.5°–1° poleward shift is seen in many basins in the 2K-lowlat, 2K, and −2K-highlat runs. The exception is again the NP summer cyclones, which form 1° equatorward in the two former cases. The changes in the 2K-highlat run are again weak and inconsistent. The shifts in lysis latitude (Fig. 9c), on the other hand, are more pronounced. In many cases, the shift is like that seen in the feature-point latitude.

Thus the runs with stronger midlatitude SST gradients (the 2K-lowlat and −2K-highlat runs) or uniform warming (the 2K run) show poleward shifts in mean storm-track and lysis latitude, while the changes in genesis latitude are somewhat less pronounced. Next we consider how the perturbations affect the characteristics of the cyclones themselves.

d. Cyclone characteristics

Figure 10a shows the mean cyclone intensity in the control run as a function of time (since formation), for the three track regions during the two seasons. Figure 10b shows the mean trajectories relative to the cyclones’ genesis positions. Corresponding results from the ERA-Interim data are shown in Figs. 10c and 10d. Note that the cyclone population is decreasing in time, so the error bars (not shown) increase in time. Generally the first 6–7 days represent the most robust response.

The cyclones intensify during the first 2–3 days, weakening or maintaining strength thereafter (Fig. 10a). The strongest growth occurs in the NP and NA tracks in winter, whereas the weakest cyclones occur in NH summer. The SH winter and summer cyclones lie between and are comparably strong. The general picture is the same with the ERA-Interim cyclones (Fig. 10c), although they are somewhat stronger than in CAM3. The largest difference is in NH summer, when the ERA-Interim cyclones are nearly twice as strong. The qualitative picture is nevertheless similar.
The cyclones move poleward and eastward in all cases (Fig. 10b). The largest poleward displacements occur with the NH winter cyclones. This is perhaps unsurprising given the southwest–northeast tilt of the NH winter tracks. The trajectories in the other cases are similar, at least over the first week. The result for the ERA-Interim cyclones is very similar (Fig. 10d).
PDFs of different characteristics for the SH DJF cyclones are shown in Fig. 11. Shown in the figure are cyclone intensity, 5-day latitudinal and longitudinal displacements, and cyclone lifetimes. We define the intensity here as the maximum absolute along-track relative vorticity. The 5-day latitudinal and longitudinal displacements are the degrees traveled toward the pole and eastward, respectively, during the first 5 days of the cyclone’s lifetime. The lifetime reflects the duration of each trajectory. As in Fig. 8, each panel displays the PDF from the control run, ERA-Interim, and the 2K-lowlat difference.

The PDFs for CAM3 and ERA-Interim compare well. The agreement is particularly good for the lifetimes (Fig. 11c). The displacement PDFs (Figs. 11a,b) have similar shapes and peak at the same latitude, but the ERA-Interim PDFs are slightly shifted toward the left, indicating somewhat greater translation for the CAM3 cyclones. The maximum intensity PDFs are also comparable (Fig. 11d), but there are more strong storms in ERA-Interim.

The cyclone characteristics exhibit clear changes in response to the 2K-lowlat forcing. The mode of the 5-day latitudinal displacement is 10° in the control run, and most cyclones travel from 0° to 20° poleward (Fig. 11a). The cyclones travel farther in the 2K-lowlat run. Indeed, the mean 5-day latitudinal displacement has increased by nearly 20%, amounting to an additional displacement of 1.6° toward the pole (Table 1). The greater displacement occurs over the same 5 days, the cyclones are moving faster poleward in the 2K-lowlat run. Furthermore, most cyclones in the control run travel about 80° eastward during the first

**FIG. 11.** As in Fig. 8, but for the SH summer cyclones and their (a),(b) 5-day meridional and zonal displacements (dlat and dlon), (c) lifetimes (T_life), and (d) intensity. The 5-day meridional displacement is defined as the latitudes covered by a cyclone during the first 5 days of its lifetime. Similarly the 5-day zonal displacement is the longitudes covered. Both measures exclude cyclones living less than 5 days. The intensity is defined as the absolute value of the most extreme along-track relative vorticity.
5 days (Fig. 11b). In the 2K-lowlat run the mean 5-day longitudinal displacement increases by about 15%, corresponding to $10^6$ of longitude (Table 1).

The SST perturbation also affects the cyclone lifetimes (Fig. 11c). Most cyclones live about 3 days in the control run, although the extended tail of the PDF indicates that there are cyclones that last up to two weeks. In the 2K-lowlat run, there are more short-lived and fewer long-lived cyclones. As such, the mean lifetime decreases by about 4% (corresponding to 0.2 days; Table 1).

The cyclones are also stronger in the 2K-lowlat run (Fig. 11d). In the control run the maximum intensities are up to and exceeding $10^{-2} s^{-1}$, with a mode of about $4 \times 10^{-5} s^{-1}$. In the 2K-lowlat run there is a decrease in the number of cyclones weaker than $8 \times 10^{-5} s^{-1}$ and an increase in the number of stronger cyclones, so the mean maximum intensity increases by roughly 3%, equivalent to $1.6 \times 10^{-6} s^{-1}$ (Table 1).

The results from the other perturbation runs are shown in Fig. 12. Once again, the changes in the 2K-highlat and 2K runs are broadly consistent with those seen in the 2K-lowlat run: the cyclones have shorter lifetimes, and they travel faster and are stronger. The changes in the 2K-lowlat run are the most pronounced, while the changes in the 2K run are the least. Evidently the increase in the midlatitude gradient is the most important here. And in line with the relatively weak response seen previously for the 2K-highlat run, the changes in the displacements and lifetimes are generally inconsistent in this case; the only clear aspect is that the
storms are weaker. However, considering only the SH tracks, the response to the 2K-highlat forcing is opposite to the other runs. There are more long-lived cyclones, they travel less, and the intensity is smaller.

Last, there is the number of cyclones in the different runs. Previous studies suggest that the number of cyclones will decrease in a warmer climate (e.g., Bengtsson et al. 2006; Lambert and Fyfe 2006; Pinto et al. 2007; Ulbrich et al. 2009; Catto et al. 2011). The number of identified cyclones is shown in the last columns of Tables 1 and 2. In the 2K-lowlat run, there is a decrease in the number of storms of between 2% and 5% in the SH and in NA winter. The results are largely consistent for the 2K and −2K-highlat runs. At the same time, though, there is an increase in the number in NA summer and in the NP in both seasons. Furthermore, the number of storms in the 2K-highlat run decreases in the SH and increases in the NH, during both seasons. So the results regarding cyclone number are less conclusive than the other characteristics.

e. Changes in translation speed

In addition, there are indications that the changes in cyclone properties are related to one other. Shown in Fig. 13 is the time evolution of the mean latitudinal displacements (relative to the initial position) vs the mean intensities. Results are shown from all five runs (the control and the sensitivity runs) for the SH (a) DJF and (b) JJA storm-track regions. As in Fig. 10, the intensity is the absolute values of $f_{500 \text{Pa}}$, and the bullets superimposed on the lines indicate time with 1-day intervals.

![Fig. 13. The time evolution of the mean latitudinal displacements (relative to the initial position) vs the mean intensities. Results are shown from all five runs (the control and the sensitivity runs) for the SH (a) DJF and (b) JJA storm-track regions. As in Fig. 10, the intensity is the absolute values of $f_{500 \text{Pa}}$, and the bullets superimposed on the lines indicate time with 1-day intervals.](image-url)
A number of studies have addressed how cyclones move poleward. The phenomenon was originally studied in the context of tropical cyclone migration, and early studies focused on the motion of an isolated vortex on the $\beta$ plane (Adem 1956; Chan and Williams 1987; Sutyrin and Flierl 1994). A cyclonic vortex advects fluid to the north on its eastern side and to the south on its western side, and this generates vorticity anomalies (\(\beta\) gyres), which then act back on the vortex. The result is

correlation coefficient in this case is 0.66. On average, an increase of \(2 \times 10^{-6} \text{s}^{-1}\) in vortex intensity corresponds to an additional 1° poleward displacement in 5 days.
a drift that is poleward and, in the absence of a mean flow, westward. Smith (1993) proposed a scaling relation for the drift velocity:

\[ v_d = \frac{C \zeta R}{\beta R} = C \frac{3}{2} R^{1/2} \beta^{-1/2}, \]

where \( R \) is the vortex radius, \( \zeta \) is its relative vorticity, and \( C \) is a dimensionless constant. The drift scales with the vortex intensity and size, and is inversely proportional to the square root of \( \beta \). So stronger/larger vortices drift faster than weaker ones, but the drift is inhibited by having a stronger potential vorticity (PV) gradient.

Recently, others have studied the poleward drift of extratropical cyclones. These studies suggest that baroclinic effects are important. In particular, advection by the main cyclone generates an oppositely signed vortex at higher levels, and the resulting baroclinic dipole can self-adveect poleward (Rivièr et al. 2012; Oruba et al. 2013). In these studies, the strength of the upper anomaly depends on the PV gradient, and a stronger gradient yields a stronger dipole, resulting in more rapid translation.

The studies are thus inconsistent with regards to the effect of the PV gradient, as a stronger gradient inhibits self-advection for barotropic vortices but enhances it for baroclinic vortices. We examined the zonally averaged PV gradients in the present runs but found no consistent relation with the changes in vortex translation. However, where the previous studies concur is with respect to vortex intensity. In both the barotropic and baroclinic studies, having stronger vortices implies more rapid self-advection. This is consistent with the present results.

The relation between the 5-day longitudinal displacement and the maximum intensity is less clear (Fig. 14b). The strong winter cyclones in the NP, for example, travel less far eastward than do the winter cyclones in the SH, which are substantially weaker. However, within a given storm track the two quantities appear to be correlated. Moreover the changes in the 5-day longitudinal displacement are correlated with changes in maximum intensity (Fig. 14d). Here the correlation coefficient is 0.6. Increasing the mean vortex intensity by \( 2 \times 10^{-6} \) \( \text{s}^{-1} \) corresponds to 5° of additional longitudinal displacement in 5 days.

While an increase in cyclone intensity can produce a more rapid poleward drift, as noted, it is not so obvious that it would also produce more rapid longitudinal displacement. A stronger barotropic vortex on the \( \beta \) plane

\[^3\text{Smith's results pertain to the total (zonal and meridional) drift velocity. However a similar scaling is expected to apply to the meridional drift alone.}\]
are less clear in winter (Fig. 13b), and this is also the case in similar plots for the NH (not shown). The reason is that the near-zonality of the SH summer storm track facilitates differentiating poleward and longitudinal translation. As self-advection presumably occurs across the mean PV gradient, the motion need not be poleward if the jet is tilted. And a tilting or spiraling jet will obviously impact poleward drift. So differentiating the two effects is less straightforward in the NH and the SH winter.

4. Summary and discussion

We have examined changes in cyclone characteristics induced by changes in SST in the AGCM simulations of GL12. The latter focused on changes in storm-track position and intensity by using bandpass-filtered statistics. Here we focused instead on properties associated with the cyclones’ life cycle. The cyclones were identified using the objective vortex tracking routine of Hodges (1994, 1995, 1999) applied to the $\zeta_{850 \text{ hPa}}$. We studied the statistics of the cyclones’ positions, lifetimes, intensities, and translational properties over the 20-yr period of the simulations. Whereas GL12 considered zonally averaged fields during boreal winter only, we expanded the analysis to include the summer and winter seasons in both hemispheres, and treat the NA and NP tracks separately.

In the control run, the SSTs were determined by a 20-yr climatology (Reynolds et al. 2002; Collins et al. 2004, 2006). In the perturbation experiments, SSTs were increased or decreased by 2 K equatorward and poleward of 45° (in the 2K-lowlat and ±2K-highlat runs) and uniformly over the oceans (in the 2K run). In the former perturbation runs the midlatitude SST gradients were increased or decreased, whereas these gradients were unaltered in the 2K run. These forcings in turn altered tropospheric baroclinicity (as indicated by the Eady parameter). It turns out that increasing either the mid-latitude gradient or the heating at low latitudes causes the maximum baroclinicity to shift poleward (Fig. 2 and 3).

The strongest response is in the 2K-lowlat run, where both effects are present. One can deduce changes in storm-track position by using the identified cyclone positions or “feature points” (Fig. 9 and Tables 1 and 2). Averaging the positions and comparing with the control run yielded track shifts that were consistent with the changes in baroclinicity and also with the results of GL12. The mean positions in the 2K-lowlat, 2K, and −2K-highlat runs shift poleward by 0.5°−1.1°. The warming in the 2K-highlat run, on the other hand, produces a relatively weak response.

An exception is the NP summer, where the baroclinicity maximum, storm track, and jet (latter is not shown) shift equatorward when the low latitudes are warmed (Fig. 9a). The shift occurs in the 2K-lowlat and 2K runs but is absent in the ±2K-highlat runs, so the effect evidently stems from low-latitude heating. However, this response may be peculiar to CAM3, as the NP storm track lies much farther north in summertime than it does for example in the ERA-Interim analysis.

Equatorward shifts in the storm tracks are also known to occur when changing the subtropical SST gradients (e.g., Brayshaw et al. 2008; Chen et al. 2010; GL12). In one run of GL12, a 2-K warming was imposed equatorward of 15°. This yielded an equatorward shift in the storm tracks, with similar changes in the jets, baroclinicity, Hadley cells, and eddy fluxes of heat and momentum. Applying the vortex tracking routine to that run, we found that the mean NP summer storm-track and genesis latitudes shift equatorward by more than 2° and 3°, respectively (not shown). So a warming confined to the tropics/subtropics can produce dramatic changes which differ from the typical global warming response. This point has been made previously in relation to El Niño (e.g., Lu et al. 2008; Chen et al. 2008).

The SST alterations also affect the cyclones themselves. When the storm tracks shift poleward, the cyclones are more intense. There are also more short-lived storms, so the mean lifetime is shorter. However, the results regarding the number of storms were inconclusive.

Previous studies generally find the number of cyclones to decrease in a warmer climate (e.g., Bengtsson et al. 2006; Lambert and Fyfe 2006; Pinto et al. 2007; Ulbrich et al. 2009; Catto et al. 2011), but the results regarding storm intensity are less consistent, with some finding an increase (e.g., Lambert and Fyfe 2006; Mizuta et al. 2011) and others finding relatively little change (e.g., Bengtsson et al. 2006, 2009; Pinto et al. 2007; Catto et al. 2011). The present results regarding changes in intensity are in most cases significant. However, it is worth emphasizing that we are only considering the effect of changes in the SSTs. Other effects may have been at play in the simulations cited above.

However, an increase in storm intensity may help explain the observed increase in bandpass-filtered geopotential height variability in GL12. Consider the SH summer track in the 2K-lowlat run. Here, the number of storms decreased by 3% while the intensity increased by roughly the same amount. GL12 found that the maximum bandpass-filtered height variability shifted poleward and also increased in magnitude. As the cyclones are a major contributor to the variability, it makes sense that the intensity should increase if the number decreases. In the other cases where GL12 observed increased variability, we also find stronger storms, and vice versa.
We also find that the cyclones translate faster, both poleward and eastward. The poleward translation moreover is positively correlated with cyclone intensity. That a stronger vortex can produce larger meridional displacements is to be expected from Kelvin’s circulation theorem (e.g., Pedlosky 1987). That stronger cyclones move poleward faster follows both from studies with barotropic vortices on the β plane (e.g., Smith 1993; LaCasce 1998) and in realistic models with baroclinic vortices (e.g., Rivière et al. 2012; Oruba et al. 2013). We have not established the exact cause of the increased translation in these runs, merely that the drift velocity is correlated with cyclone intensity. Changes in the mean PV gradient (not shown here) were inconclusive, but changes in the mean temperature gradient are clear. Thus it may be that the self-induced translation stems from an interaction with the temperature field. Further work, most likely with idealized models, is required to know for certain.

However, the change in poleward translation by itself cannot explain the shift in the storm tracks. The changes in baroclinicity suggest storm formation shifts with latitude, and the cyclone genesis statistics support this. But the changes in lysis were usually greater as the stronger storms move further poleward. This likely impacts the maximum in the bandpass-filtered geopotential variability, whose maximum lays poleward of the maximum in baroclinicity (GL12).

The increased eastward advection, on the other hand, stems from a strengthening of the mean zonal winds. The stronger temperature gradients imply greater vertical shear, so the jets are intensified. The eastward translation is accordingly strongly correlated with the depth-integrated zonal wind. A similar effect was suggested by Catto et al. (2011) in response to a quadrupling in CO₂, as cyclones formed over the Kuroshio during boreal winter were shown to translate farther downstream with a stronger jet.

As noted earlier, we omitted a set of cyclones which form between 25° and 35°S in the SH during summer. The changes associated with these systems are opposite to those of the mid- and high-latitude systems: with an increase in the mean SST or in the midlatitude gradients, these cyclones form and die closer to the equator and also travel less poleward and eastward. This set was exceptional (i.e., there is no corresponding set of storms in the NH in summer). The reason for the different behavior is not known but is likely related to changes in the mean circulation in the near tropics.

To isolate the effect of the SST changes, the sea ice distribution was left unaltered in the sensitivity runs. Realistically though, a cooling or warming of the high-latitude SSTs would also alter the sea ice cover. The storm-track response to such changes is, however, not straightforward, as can be seen from the body of idealized studies using AGCMs. For instance, Bader et al. (2011) found that the NA and NP tracks shift equatorward and poleward, respectively, in response to removing the Arctic ice sheet. Magnusdóttir et al. (2004) found a decreasing trend in the Labrador and Greenland Sea ice to cause an equatorward shift in the NA track. Kvamstø et al. (2004), on the other hand, found a poleward shift in the NA when reducing the Labrador Sea ice while Seierstad and Bader (2009) found projected sea ice loss to cause weakening of the NA track without a shift.

Similarly conflicting results have been found for the SH storm tracks. Removing the Antarctic ice sheet, Menéndez et al. (1999) found an equatorward shift in the SH track, and supporting results were obtained by Bader et al. (2013) using projected sea ice fields. In contrast, Kidston et al. (2011) only found weak changes with an ice reduction, but also a poleward shift in the tracks with a larger ice cover. Clearly more work is required to establish the dependence on ice cover.

The present SST forcings, while facilitating identification of regional responses, are idealized. Nevertheless, the track-density changes in the SH, NA, and NP storms induced by warming the low latitudes (the 2K-lowlat run) are similar to the response to global warming in Bengtsson et al. (2006). The latter used a coupled climate model and Hodges’ tracking routine to investigate storm-track response to the Intergovernmental Panel on Climate Change (IPCC) A1B scenario, and poleward shifts were seen in both hemispheres. As here, the changes were clearer in the SH than in the NH, and clearer in DJF than JJA. Bengtsson et al. (2006) furthermore observed an equatorward shift of the NP summer track. However, they did not find a consistent change in storm intensity. It is important to understand the difference between the two studies; besides translating faster, stronger storms have an obvious societal impact.

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REFERENCES


Bader, J., M. D. Mesquita, K. I. Hodges, N. Keenlyside, S. Osterhus, and M. Miles, 2011: A review on Northern Hemisphere sea-ice,


