

## Response to Reviewer 1 (James Done)

We'd like thank Dr. Done for his insightful review of our manuscript, including his positive comments and suggestions for improvement. Responses to his suggestions are included below.

1. The claim is made that the experiment assesses future thermodynamic environments. I understand that the frequency distribution of ENSO phases is the same in current and future climate. But MPAS will permit some large-scale circulation change in response to the future SSTs so I'm not sure circulation change is small compared to the thermodynamic change. Please clarify the contributions of thermodynamic change and circulation change permitted by your experimental setup. Perhaps checking the magnitude of circulation difference fields would help. More generally, can you provide more guidance on how the future change results should be interpreted? How should we interpret future changes based on fixed ENSO phase frequencies but variable atmospheric circulation response?

*It is true that we are assessing more than just thermodynamic changes since MPAS is a global model allowing for circulation changes to occur. We have removed "thermodynamic" from L8 on page 1, L11 on page 2, L3 and L29 on page 6, L9 and L34 on page 8, and L2 on page 13. We have also added L12–15 on page 2 to further clarify our methods.*

*In regard to the circulation changes, please see Figs. 1 and 2 below. Fig. 1 shows a stronger wintertime Pacific jet and a deeper Aleutian Low in the future simulations. Changes in the Atlantic are less severe, but show a weaker jet over the northeast US and a slightly weaker Icelandic Low. The cross-sections in Fig. 2 show an overall stronger jet in the future simulations, with a poleward shift in the maximum.*

*Given GCM difficulties in representing ENSO, we felt that the approach taken was the most defensible. Future research will need to determine how ENSO phase frequencies will change with warming, but we agree that this is another assumption in our methods and have added a sentence acknowledging this on page 7, L27.*

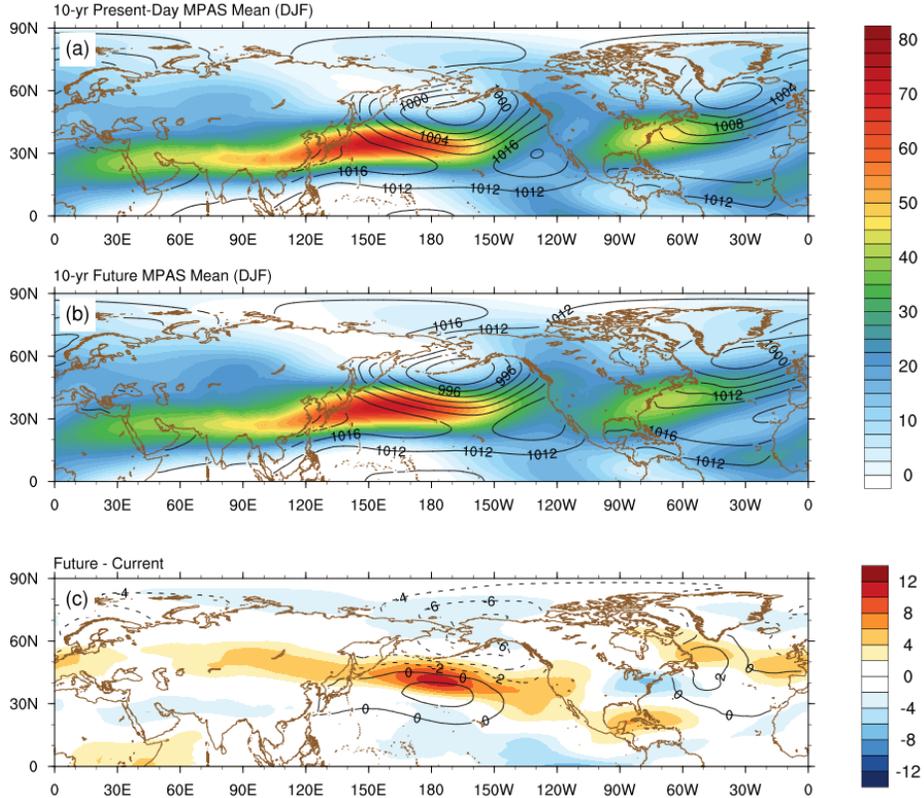


Figure 1. Average wintertime (DJF) 250-hPa zonal wind speed ( $\text{ms}^{-1}$ ; shaded) and SLP (hPa; contoured) over the ten MPAS simulation years for (a) present-day, (b) future, and (c) future minus present-day 250-hPa wind speed (shaded) and SLP (contours). Wind contours are shaded every  $2.5 \text{ ms}^{-1}$  in (a), (b) and every  $2 \text{ ms}^{-1}$  in (c). SLP is contoured every 4 hPa in (a), (b) and every 2 hPa in (c).

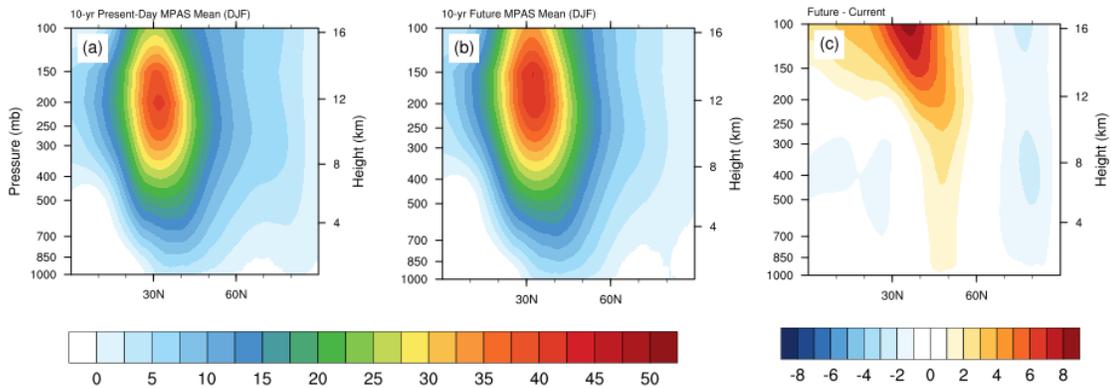


Figure 2. Average wintertime (DJF) cross-section of zonally averaged zonal wind ( $\text{ms}^{-1}$ ; shaded) over the ten MPAS simulation years for (a) present-day, (b) future, and (c) future minus present-day. Contours are shaded every  $2.5 \text{ ms}^{-1}$  in (a), (b) and every  $1 \text{ ms}^{-1}$  in (c).

2. A related issue is that the experimental approach assumes that SSTs in all phases of ENSO change by the same magnitude and spatial pattern. Can you comment on how realistic this assumption is and any implications for the results?

*We agree that patterns of SST change may differ as a function of ENSO phase, and applying a single “delta” in our experiments is a limitation. However, we felt that this was the most straightforward starting point, given limited computational resources. We have added mention of this limitation, along with justification for our approach to L24–26 on page 7.*

*We are aware of the “remote SST effect” on tropical cyclones, and that TC activity can exhibit considerable sensitivity to the patterns of SST change (e.g., Vecchi and Soden, 2007; Xie et al., 2010). Mizuta et al. (2014) have identified three distinct patterns of SST change in the CMIP5 CGCM output, and if resources were available, it would be quite interesting to run additional future experiments with these patterns in order to gauge the resulting sensitivity. One could also compute CGCM SST delta fields as a function of ENSO phase, and apply the deltas accordingly (perhaps what you are suggesting). We have added mention these possible future research directions.*

*References:*

*Mizuta, R., O. Arakawa, T. Ose, S. Kusunoki, H. Endo, and A. Kitoh, 2014: Classification of CMIP5 future climate responses by the tropical sea surface temperature changes. SOLA, 10, 167-171, <https://doi.org/10.2151/sola.2014-035>*

*Vecchi, G. A. and B. J. Soden, 2007: Effect of remote sea surface temperature change on tropical cyclone potential intensity. Nature, 450, p.1066.*

*Xie, S.-P., C. Deser, G. A. Vecchi, J. Ma, H. Teng, and A. T. Wittenberg, 2010: Global warming pattern formation: Sea surface temperature and rainfall. J. Climate, 23, 966–986, doi:10.1175/2009JCLI3329.1*

3. Please explain why this study chose to sample ENSO phases and not phases of some other interannual or decadal mode of climate variability?

*ENSO is a global forcing phenomenon, and is often used as a basis for seasonal forecasts of tropical cyclone activity (e.g., Klotzbach and Gray, 2004; Camargo et al., 2007; Klotzbach, 2007) due to its strong connection with global tropical cyclone activity (e.g., Gray, 1984; Chan, 1985; Lander, 1994; Chu and Wang, 1997; Kossin et al., 2010). Since our initial motivation for conducting these simulations was to examine TCs and extratropical transition events, we wanted to sample a mode of variability with a large influence on global tropical cyclone activity, and therefore, chose ENSO. We have added this reasoning and associated references to L15–17 on page 5.*

*References:*

*Camargo, S. J., Barnston, A. G., Klotzbach, P. J., and Landsea, C. W.: Seasonal tropical cyclone forecasts, World Meteorological Organization (WMO) Bulletin, 56, 297–309, 2007.*

Chan, J. C. L.: Tropical cyclone activity in the Northwest Pacific in relation to the El Niño/Southern Oscillation phenomenon, *Mon. Weather Rev.*, 113, 599–606, doi:10.1175/1520-0493(1985)113<0599:TCAITN>2.0.CO%3B2, 1985.

Chu, P.-S., and Wang, J.: Tropical cyclone occurrences in the vicinity of Hawaii: Are the differences between El Niño and Non-El Niño years significant?, *J. Climate*, 10, 2683–2689, doi:10.1175/1520-0442%281997%29010<2683%3ATCOITV>2.0.CO%3B2, 1997.

Gray, W. M.: Atlantic season hurricane frequency. Part I: El Niño and 30 mb Quasi-Biennial Oscillation influences, *Mon. Weather Rev.*, 112, 1649–1668, doi:10.1175/1520-0493(1984)112<1649:ASHFPI>2.0.CO;2, 1984.

Klotzbach, P. J.: Recent developments in statistical prediction of seasonal Atlantic basin tropical cyclone activity, *Tellus A*, 59, 511–518, doi:10.1111/j.1600-0870.2007.00239.x, 2007.

Klotzbach, P. J., and Gray, W. M.: Updated 6–11 month prediction of Atlantic basin seasonal hurricane activity, *Mon. Weather Rev.*, 19, 917–934, doi: 10.1175/1520-0434%282004%29019<0917%3AUMPOAB>2.0.CO%3B2, 2004.

Kossin, J. P., Camargo, S. J., and Sitkowski, M.: Climate modulation of North Atlantic hurricane tracks, *J. Climate*, 23, 3057–3076, doi:10.1175/2010JCLI3497.1, 2010.

Lander, M. A.: An exploratory analysis of the relationship between tropical storm formation in the Western North Pacific and ENSO, *Mon. Weather Rev.*, 122, 636–651, doi:10.1175/1520-0493(1994)122<0636:AEAOTR>2.0.CO;2, 1994.

4. The discussion of missing cool-wakes in the simulations on page 8 is well stated. But that is only half the story. The reanalysis SSTs will contain the cool wakes of observed TCs, so this could unphysically dampen simulated TCs that cross these ‘phantom’ cool wakes.

*This is a good point; however, the cold wakes in the OSTIA SST analysis field do not appear to be very strong. For example, Figs. 3 and 4 show a sample of times during the 2010 Atlantic TC season when two major hurricanes (Fig. 3: Danielle from 21–30 August and Fig. 4: Earl from 25 August–4 September) were present. While these are just two instances, it does illustrate that the phantom cold wakes in the SST analysis may not substantially impact the simulated TCs. The documentation of OSTIA does not explicitly state how SSTs are handled during TC conditions, but their interpolation to remove diurnal variability could be a factor in smoothing out the effects of cold wakes from TCs. Nevertheless, thank you for raising a valid point; we have added L18–20 and L26–27 on page 8 to acknowledge this limitation.*

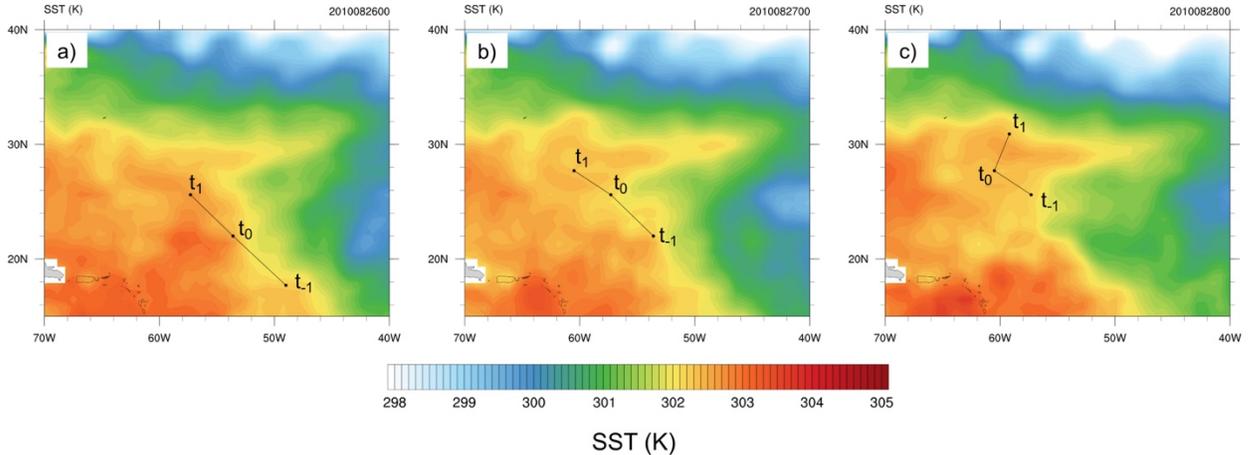


Figure 3: SST (K; shaded every 0.1 K) for (a) 00 UTC 26 August 2010, (b) 00 UTC 27 August 2010, and (c) 00 UTC 28 August 2010 when Hurricane Danielle was present in the North Atlantic. A subset of the storm track is shown in black with  $t_0$  in each panel marking the position at the time in the top right corner, and  $t_1$  and  $t_{-1}$  marking the positions 24-h before and after, respectively. Danielle was classified as a Major Hurricane in panel (b).

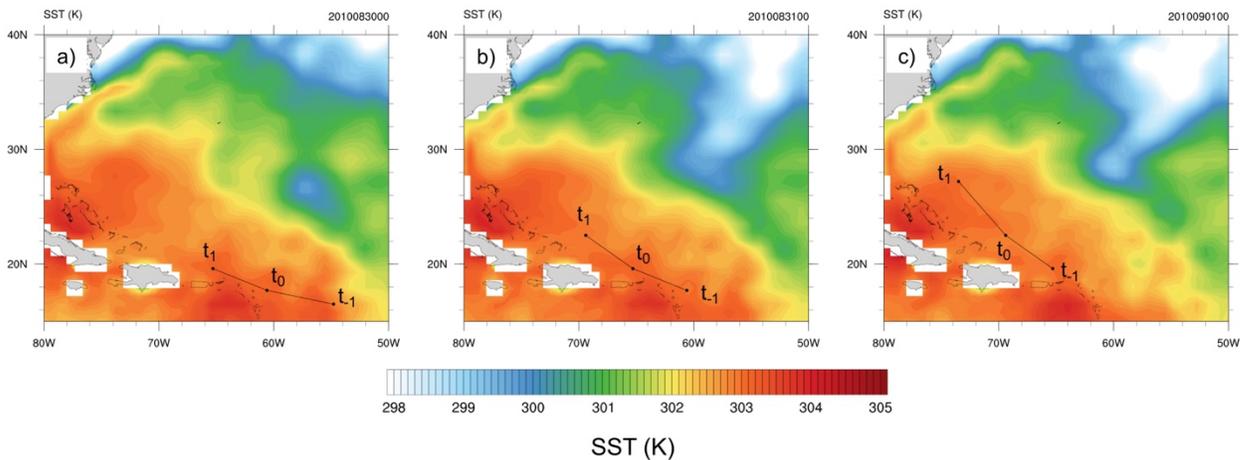


Figure 4: SST (K; shaded every 0.1 K) for (a) 00 UTC 30 August 2010, (b) 00 UTC 31 August 2010, and (c) 00 UTC 01 September 2010 when Hurricane Earl was present in the North Atlantic. A subset of the storm track is shown in black with  $t_0$  in each panel marking the position at the time in the top right corner, and  $t_1$  and  $t_{-1}$  marking the positions 24-h before and after, respectively. Earl was classified as a Major Hurricane in panels (b) and (c).

- The evaluation shows MPAS does a reasonable job at capturing the climatological spatial distribution of TCs. Given that your approach emphasizes ENSO phases, can you also check whether MPAS captures the observed TC response to ENSO phase in the Atlantic and Pacific.

Histograms of TC count by year are shown in Fig. 5. For the North Atlantic (Fig. 5a–b), MPAS does simulate a negative trend in TC activity indicating, in general, higher levels of TC activity during La Niña years and lower levels during El Niño years, consistent with expectations and the IBTrACS trend. For the Western North Pacific (Fig. 5c–d), however, MPAS simulates a

negative trend in TC activity, contrary to expectations and the positive trend exhibited by the IBTrACS. The trends in both data sets for the Western North Pacific, though, are not particularly strong.

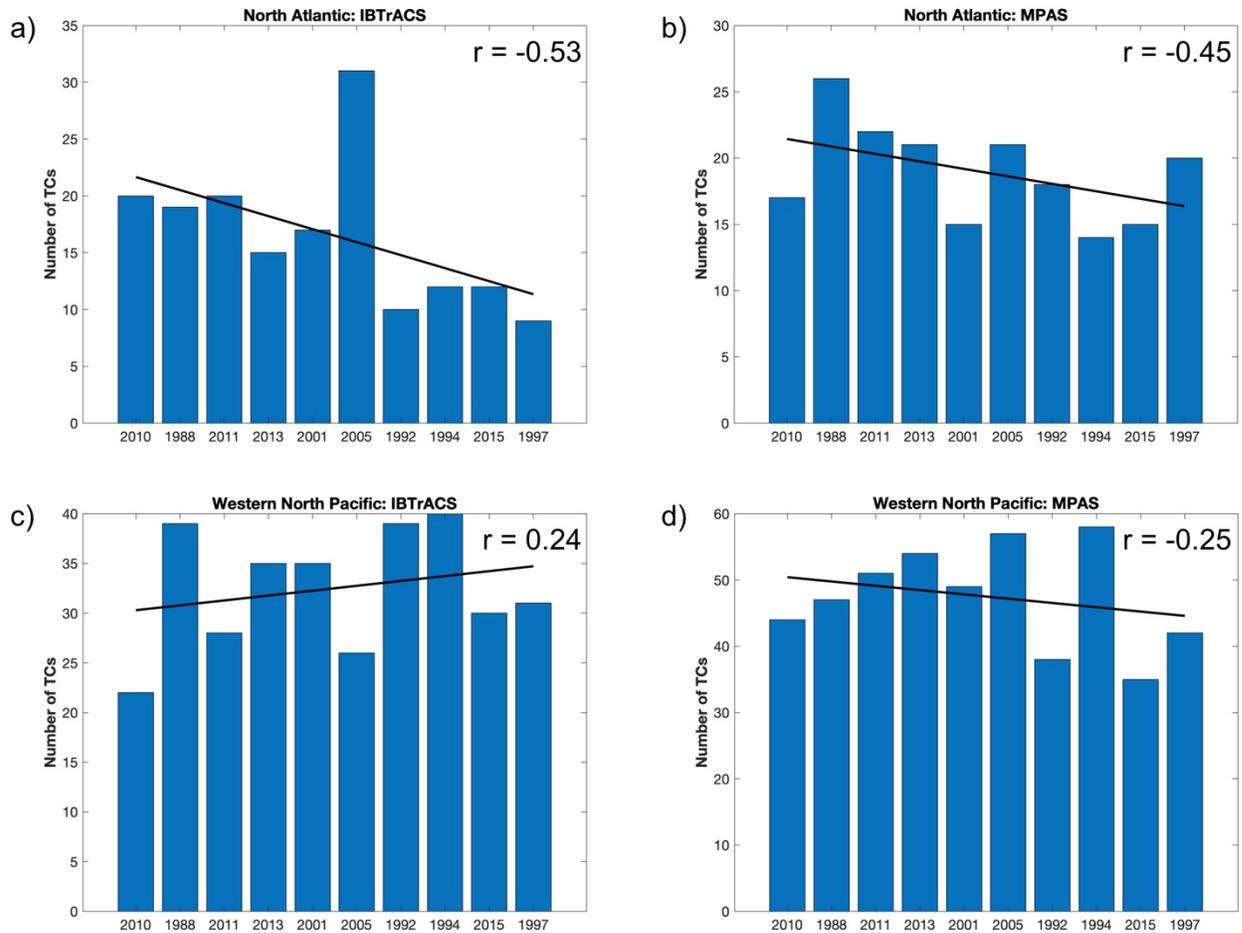


Figure 5: Histograms of TC count as a function of year for the North Atlantic basin for the (a) IBTrACS and the (b) MPAS simulations and for the Western North Pacific basin for the (c) IBTrACS and the (d) MPAS simulations. The years on the abscissa are ordered from the strongest La Niña year (2010) to the strongest El Niño year (1997). Linear regression lines are shown in black with the correlation coefficient in the top right corner in all panels.

- MPAS misses TCs that develop from easterly waves over the eastern North Atlantic. I agree that one possible reason is enhanced shear. Another reason could be that the westward shift in the African Easterly Jet means that the wave energy accumulation zone (as discussed in Done et al. 2011) is also shifted westward.

Done, J.M., Holland, G.J. and Webster, P.J., 2011. The role of wave energy accumulation in tropical cyclogenesis over the tropical North Atlantic. *Climate dynamics*, 36(3-4), pp.753-767.

*Thank you for this additional reference and reason for our displaced Atlantic TCs, this information has been added to L30–32 on page 11.*

7. The purpose of Fig. 10 is to compare seasonal cycles of TCs. I suggest plotting the normalized distributions to remove the effect of differences in absolute numbers.

*Thank you for this suggestion. Fig. 10 has been updated with normalized probability histograms.*

8. The motivation for this experimental design is to assess future changes in high- impact weather. But no results on future changes to high impact weather are presented. I read in the final section that results will be forthcoming in a separate publication. To set readers expectations I suggest adding a note earlier in the manuscript, perhaps at the end of the introduction.

*Thank you for this suggestion. A version of L34–35 from page 13 has been added to L21–23 at the end of the Introduction on page 4.*

## Response to Anonymous Reviewer 2

We'd like thank Reviewer 2 for their insightful comments and suggestions on our manuscript. Responses to their suggestions are included below.

1. The authors show the TC characteristics for the present climate, but not for the future climate, whereas they show in section 5 plots of the future climate. I was disappointed that nothing was said about TCs in the future climate. I assume that it will be analysed in a future paper, but for the reader it feels very disappointing. Why not postpone the climate change simulations altogether to a following paper? The added value of section 5 is very minor. Concentrating on the present climate and showing the quality of the model as in sections 3 and 4 should be fine for presenting the model and the experimental set-up. If you want to discuss climate change I would like it to be more than what is in the paper and include TC changes, or make it explicitly that they will be discussed in a future paper.

*Since we plan for climate change effects on TCs, and other high-impact weather events, to be the subject of future publications (several of which are either submitted or in preparation), we feel this manuscript should focus solely on the credibility of the current and future simulations. Therefore, we will retain Section 5 in this manuscript in order to avoid repeating it in each future publication. However, as you've suggested, we have added L21–23 at the end of the Introduction on page 4 and L21–22 on page 11 to help make this point as explicit as possible.*

2. The experimental set-up is new in certain aspects, but I had the feeling that the authors overstate a bit the uniqueness. They present it as global PGW simulations, whereas to me they are a clever way of time slice experiments with prescribed SSTs, with a spin-up to let the model come into equilibrium with the SST and CO<sub>2</sub> concentrations. I found the PGW term therefore confusing and suggest to remove it.

*You are correct, our simulations are better described as time-slice experiments rather than global PGW experiments. We will retain L8–20 on page 3 in the Introduction as it is relevant background information. We also opt to retain L28–30 on page 6 as we feel comparing and contrasting our methods to the PGW technique is necessary. However, we had added L34 on page 6 to explicitly state that our future simulations are not global PGW experiments, but rather time-slice experiments. Additionally, our experiments are described as time-slice runs in L31 on page 3. It is our application of MPAS for climate-change purposes that we feel is truly unique, rather than the time-slice experiment method.*

3. Ordering the simulations by El-Nino strength is clever, but anomalies outside El-Nino can occur, which can be substantial, and still have to be adjusted. I am not sure if one month spin-up is sufficient for these regions.

*We acknowledge that after one month, the spin-up to the new boundary conditions may be incomplete; however, during our initial evaluation, no obvious "shocks" appears at these inter-year boundaries after a couple of weeks, leading us to believe that one month was a sufficient amount of spin-up time. Additionally, other users who wish to use the output could*

*discard more of the beginning portions of the simulations if they feel more spin-up time is needed; discarding the first 2.5 months would still yield full yearly simulations.*

4. The use of high resolution SST data and computation of future SST by taking the delta of an ensemble of low resolution GCMs is a good approach. However, the delta SST can be influenced by the biases in SST. This can be particular large at the western boundary currents, that are incorrectly represented by low-resolution GCMs.

*Because CMIP5 models do not resolve ocean fronts (e.g., associated with western boundary currents and eddies), there are likely to be systematic errors in their depiction of future changes in SST. They offer, however, the best estimates of such changes that are currently available, and our method retains realistically strong SST gradients.*

## Response to Executive Editor David Ham

1. The code and data availability section refers to MPAS, and to various data sets. However this is insufficient to allow the reader to reproduce the results in the manuscript. For this, the reader needs the configuration files or run scripts used to run each experiment, and any scripts which were used to post process or analyse the model data. The exact version of these should be persistently and publicly archived (for example on Zenodo) and cited from the manuscript.

*The initialisation files and surface update files containing sea-surface temperature and sea ice information are too large to upload to Zenodo given the 50GB limit (these files are ~30GB for each run and there are 20 experiments in total). Sample run scripts and sample post-processing scripts, however, have been uploaded to Zenodo; the “Data and code availability” section has been updated to reflect this new citation.*

2. The reference to the MPAS code used is on GitHub. This is both impermanent, as MPAS might move off GitHub in the future, and fails to identify the exact version of the code which was used (the model text refers to version 5.1, but was this the 5.1 release, or just a version of the master branch taken when 5.1 was the release number?) To remedy this, the exact version of MPAS used should be publicly and persistently archived. Since MPAS is developed on GitHub, the GitHub-Zenodo integration may be the easiest way to accomplish this. See: <https://guides.github.com/activities/citable-code/>. Similar issues apply to Tempest Extremes.

*Because we do not host the MPAS or TempestExtremes GitHub repositories, we are unable to use the GitHub-Zenodo integration to make these repositories citable. However, we have updated the MPAS website in the “Data and code availability section” to point directly to the version used in our study, and specified that we used the release version 5.1. We have also updated the citation of TempestExtremes in the “Data and code availability” section following the TempestExtremes source manuscript (Ullrich and Zarzycki, 2017).*

3. The result data is only available "on request". If at all possible, this data should be persistently and publicly archived so that the reader who wishes to investigate a result in the paper can do so directly. However, if this is not possible for licence reasons or because of the data volume, the data needs to be sufficiently precisely identified that it will still be retrievable if the authors have moved on from the host institution.

*The data volume is too large (~50TB) to host on a public repository like Zenodo. We have spoken with our host institutions at North Carolina State University and University of California San Diego, but they are unable to provide a permanent volume with a DOI for data storage. We will seek to store these data on the U.S. National Center for Atmospheric Research (NCAR) Computational and Information Systems Lab (CISL) archive. We welcome suggestions on how the data can be better identified in the manuscript.*

4. The external data sets used are incorrectly cited, and in some cases poorly identified. Specifically:  
ERA-Interim data. The NCAR data archive provides DOIs for this data, and even tells you what to write in the bibliography. Please cite this in accordance with their instructions.

*This citation has been updated in the references and the link in the “Data and code availability” section has also been updated as noted by NCAR.*

TRMM data. Similarly, there is a data citation tab which tells you how to cite this data correctly.

*This citation has been updated as prescribed by GES DISC, and the link in the “Data and code availability” section has been updated to point directly to the version used.*

PRISM appear not to have precise data citation instructions, however they do have a precise convention for identifying the exact data set(s) used.

*Reference to this data set is a remnant from a previous version of the manuscript, and has therefore been removed from the “Data and code availability” section.*

# Evaluation of a Unique Approach to High-Resolution Climate Modelling using the Model for Prediction Across Scales-Atmosphere (MPAS-A) version 5.1

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**Abstract.** We present multi-seasonal simulations representative of present-day and future environments using the global Model for Prediction Across Scales-Atmosphere (MPAS-A) version 5.1 with high resolution (15 km) throughout the Northern Hemisphere. We select ten simulation years with varying phases of El Niño-Southern Oscillation (ENSO) and integrate each for 14.5 months. We use analysed sea surface temperature (SST) patterns for present-day simulations. For the future climate simulations, we alter present-day SSTs by applying monthly-averaged temperature changes derived from a 20-member ensemble of Coupled Model Intercomparison Project Phase 5 (CMIP5) general circulation models (GCMs) following the Representative Concentration Pathway (RCP) 8.5 emissions scenario. Daily sea ice fields, obtained from the monthly-averaged CMIP5 ensemble mean sea ice, are used for present-day and future simulations. The present-day simulations provide a reasonable reproduction of large-scale atmospheric features in the Northern Hemisphere such as the wintertime midlatitude storm tracks, upper-tropospheric jets, and maritime sea-level pressure features as well as annual precipitation patterns across the tropics. The simulations also adequately represent tropical cyclone (TC) characteristics such as strength, spatial distribution, and seasonal cycles for most of Northern Hemispheric basins. These results demonstrate the applicability of these model simulations for future studies examining climate change effects on various Northern Hemispheric phenomena, and, more generally, the utility of MPAS-A for studying climate change at spatial scales generally unachievable in GCMs.

**Plain Language Summary.** We expect that high-impact weather events will change in a warmer climate. Computational constraints limit global climate models to resolutions that are too coarse to fully capture many societally significant weather events, such as tropical cyclones and flooding rains in middle-latitude low-pressure systems. While these global models often provide reliable projections of changes in mean temperatures and global circulation patterns, they cannot tell us how intense, high-impact events may be altered in response to climate change. Here, we present a novel set of atmospheric simulations designed to address changes in high-impact weather events. The model covers the globe, but has higher resolution in the Northern Hemisphere. We simulate ten years sampling a range of tropical climate conditions, as represented by observed ocean surface temperatures, and we carry out simulations for current and projected late 21st-century climate conditions. The future runs include increased heat-trapping gases, higher ocean temperatures, and reduced polar sea ice as projected by conventional global climate models. Our model reasonably replicates present-day climate features, such as large-scale precipitation patterns,

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mid-latitude cyclone activity, the jet stream, and tropical cyclone characteristics. It reproduces features of climate change that are expected from global climate models, but it also captures smaller scale, high-impact weather events. We anticipate that these simulations will have great value in understanding changes in high-impact weather events.

## 1 Introduction

5 We present a novel approach to high-resolution climate modelling with the intent of examining the effects of climate change on high-impact Northern Hemispheric weather phenomena. It is nearly certain that rising global greenhouse gas concentrations over the next century will result in significant changes to the Earth's climate system (IPCC, 2014). Further understanding of how climate change will affect global and regional weather is essential to informing the scientific community, stakeholders, and policymakers on what actions should be taken to prepare for the future. Here, we present a unique set of high-resolution, multi-seasonal, global, atmosphere-only simulations conducted with the Model for Prediction Across Scales-[Atmosphere \(MPAS-A](#); Skamarock et al., 2012) in present and future environments for this purpose: to study climate change effects on Northern Hemispheric weather phenomena, including extreme events. [Present-day conditions are simulated using the current atmospheric composition and observed lower boundary conditions; climate change is represented by modifying the atmospheric composition to values appropriate for the late 21st century, and applying consistent changes in lower boundary conditions derived from CMIP5 coupled global climate models.](#) Through its variable resolution grids, [MPAS-A](#) offers the possibility of investigating local weather phenomena at high resolution in the context of a global model, while avoiding the prohibitive demands on computational resources entailed by running a model globally at high resolution. To our knowledge, however, climate change experiments at long integration times are a novel application of [MPAS-A](#). Therefore, in order to demonstrate their utility for addressing climate change effects on high-impact weather events, it is necessary to evaluate how large-scale circulations and responses to warming are represented in such simulations, thus defining the objective of this paper.

With simulations spanning several centuries, multiple ensemble members, and the inclusion of atmosphere-ocean coupling, the latest generation of general circulation models (GCMs) from the Coupled Model Intercomparison Project Phase 5 (CMIP5) are common tools for determining the effects of climate change. Due to current computational limitations, however, the grid spacing of these simulations is largely restricted to  $\sim 1^\circ$  ( $\sim 100$  km) or greater. While this coarse resolution is suitable for representing large-scale atmospheric features such as the polar amplification of global warming and teleconnections, it is insufficient for resolving weather extremes, especially those associated with smaller-scale systems such as tropical cyclones, mesoscale features within extratropical cyclones, and convective storms (e.g., Mizieliński et al., 2014 and references therein; Small et al., 2014; Prein et al., 2015; Haarsma et al., 2016; Roberts et al., 2018). These smaller-scale systems often result in significant socioeconomic impacts; therefore, in order to fully ascertain the societal impacts of climate change, it is essential to complement existing GCM simulations with simulations at resolutions sufficiently fine to capture these high-impact phenomena. The ongoing High Resolution Model Intercomparison Project (HighResMIP; Haarsma et al., 2016) associated with the CMIP6 will also be highly beneficial in understanding how weather extremes respond to climate change.

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To date, several model downscaling techniques have been developed for this purpose. For example, regional downscaling (e.g., Wang et al., 2004; Giorgi et al., 2009) computationally allows for finer grid spacings by employing a smaller domain, thus circumventing the resolution deficiency of traditional GCMs. Using a regional domain, however, presents the issue of how to specify lateral boundary conditions, and two-way interactions with larger-scales cannot be fully incorporated (Small et al., 2014). Global models eliminate the constraints of lateral boundaries, but are expensive to run for long periods at high resolutions. Incorporating nests within a global domain, or using mesh refinement grids, however, can help alleviate this expense.

Another useful method for assessing climate change effects is the “pseudo-global warming” (PGW) method, initially called “surrogate global warming” (e.g., Schär et al., 1996; Frei et al., 1998; Kimura and Kitoh, 2007; Hara et al., 2008; Rasmussen et al., 2011; Mallard et al., 2013; Lackmann, 2013, 2015; Trapp and Hoogewind, 2016). In PGW experiments, high-resolution control simulations are conducted, typically replicating an observed weather event. The high-resolution initial and boundary conditions are then modified with “delta” fields derived from GCMs, and the event is re-simulated, allowing assessment of changes in the characteristics of the event as a function of larger-scale environmental change. An important advantage of the PGW method is that realistic, high-resolution synoptic-scale and mesoscale settings are guaranteed. This method is consistent with the “storyline” approach described by Shepherd (2016), Hazeleger et al. (2015), and Trenberth et al. (2015). A limitation of PGW case studies is the inability to study the frequency of occurrence of such events. To alleviate this limitation, some investigators have conducted long-duration regional PGW simulations (e.g., Ban et al., 2014; Willison et al., 2015; Liu et al., 2017), which allow for analysis of statistical changes extending beyond the case-study of a single event. All regional PGW experiments, however, are limited by the need to impose lateral boundary conditions, which reduces the dynamical freedom of the simulations.

Given recent advances in computational power and data storage, several modelling groups have performed long-term high-resolution global simulations, both with atmosphere-only and with coupled atmosphere-ocean configurations (e.g., Small et al., 2014; Komada et al., 2015; Murakami et al., 2015; Roberts et al., 2015 and references therein; Haarsma et al., 2016; Yamada et al., 2017). Models that include coupling between the atmosphere and ocean have the advantage of two-way communication, allowing the possibility of realistic atmosphere-ocean interactions. At long integration times, however, climatologies of coupled models have been known to suffer from biases due to the drift in sea surface temperatures (SSTs), which can negatively affect regional climate projections (e.g., He and Soden, 2016). Previous studies have determined that resolution is an important factor for a more accurate representation of synoptic and mesoscale phenomena in the atmosphere (Willison et al., 2013; Small et al., 2014; Prein et al., 2015), and thus should be maximized whenever possible.

The present global simulations use a 15 km grid in the Northern Hemisphere, relaxing to a 60 km grid in the Southern Hemisphere to reduce computational expense (Fig. 1). These simulations are conducted with [MPAS-A](#) as individual time-slice runs, selected to span a range of El Niño-Southern Oscillation (ENSO) states. We use high-resolution SST analyses that capture oceanic eddies and fronts, which have been shown to exert an important influence on atmospheric variability (e.g., Kirtman et al., 2012; Siqueira and Kirtman, 2016; Ma et al., 2017; Parfitt et al., 2017). Our novel modelling approach aims to eliminate several limitations from the previously discussed methods. For one, using a global model circumvents issues related to lateral

boundary conditions, thus improving upon limited-area simulations; here, higher-resolution in the Northern Hemisphere was obtained at the expense of reduced resolution in the Southern Hemisphere. Our future simulations are similar to PGW in the treatment of SST in that we apply a GCM-based delta to analysed SST fields; this incorporation of high-resolution SSTs precludes issues of the type noted by He and Soden (2016), and represents a potential improvement to coupled atmosphere-ocean model configurations. The atmospheric resolution of 15 km is sufficiently high to represent strong tropical and extratropical cyclones as well as flooding rainfall, which is a considerable improvement compared to GCM simulations. Furthermore, the sample size is sufficient to allow statistical comparisons of features such as Northern Hemisphere storm tracks and tropical cyclone activity, thus improving upon the traditional PGW case study approach.

We present our simulations with the intention of providing an additional realization of a complex system in order to improve our understanding of potential climate change effects on Northern Hemispheric high-impact weather phenomena. For such a modelling system to be useful for this purpose it is necessary that:

- It reproduces the present-day climate and global circulation of the atmosphere.
- It demonstrates the benefits of enhanced resolution in simulating high-impact weather phenomena.
- It provides simulations of a future climate consistent with expectations derived from GCMs.

Thus, in the present paper, following a discussion of the model and how our simulations were conducted (section 2), we offer analyses of its representation of the present-day climate (section 3), and its simulation of Northern Hemisphere tropical cyclones and their climatology within the present-day climate (section 4). We focus on tropical cyclones as an exemplar of high-impact weather phenomena that are challenging to represent accurately in models, and for which successful simulation demands high resolution (e.g., Davis, 2018). In section 5 we examine the model representation of climate change in response to global warming boundary conditions. Last, section 6 presents a summary of our findings and discusses future applications of our simulations for investigating how high-impact weather may change in a warmer climate. [These applications include ongoing research efforts investigating climate change effects on the extratropical transition of TCs, TC seasonality, and persistent anomalies and blocking, all of which will be subjects of future publications.](#)

## 2 Models, Experiments, and Performance

### 2.1 Model Configuration

We conduct our simulations using the atmospheric component of [MPAS-A](#), version 5.1 (Skamarock et al., 2012). [MPAS-A](#) is a non-hydrostatic global, atmosphere-only model that uses unstructured Voronoi meshes (Du et al., 1999) to create variable-resolution grids. This grid structure permits localized areas of high resolution to transition gradually to lower resolutions, thus alleviating the boundary issues associated with sharp transitions between domains in traditional nesting approaches (Park et al., 2014). The focus of the simulations presented here is on Northern Hemispheric phenomena; therefore, we use a variable-resolution mesh with 15 km grid spacing over the Northern Hemisphere, expanding out to 60 km in the Southern Hemisphere (Fig. 1).

The [MPAS-A](#) atmospheric physics suite includes a subset of schemes adapted from versions of the Weather Research and Forecasting (WRF) model (Skamarock et al., 2008). Our simulations employ the following physics parameterizations: WRF Single-Moment 6 class (WSM6; as in WRF 3.8.1) microphysics, Yonsei University (YSU; as in WRF 3.8.1) representation of the planetary boundary layer, Tiedtke (as in WRF 3.3.1) sub-grid scale convective parameterization, 5 Community Atmosphere Model (CAM; as in WRF 3.3.1) shortwave and longwave radiation, and the Noah land surface model (as in WRF 3.3.1) for surface processes. We selected the Tiedtke convective parameterization scheme because it includes convective momentum transport (CMT), which has been shown to be important for reducing model biases in surface winds and tropical cyclone (TC) intensity (Zhang and McFarlane, 1995; Han and Pan, 2006; Hogan and Pauley, 2007; Richter and Rasch, 2008). CMT also improves the representation of features such as the Intertropical Convergence Zone (Zhang and Wang, 10 2006; Kim et al., 2008). We completed a series of preliminary tests using a quasi-uniform 60 km mesh to further refine our physics choices (not shown).

## 2.2 Present-Day and Future Climate Simulations

We selected ten simulation years with varying phases of ENSO based on the Multivariate ENSO Index (MEI) and the Oceanic Niño Index (ONI) over the TC season (Table 1). These years were also chosen to sample a range of TC activity in the North 15 Atlantic, Eastern North Pacific, and Western North Pacific basins. [We chose to sample phases of ENSO rather than other modes of climate variability due to its strong connection to global TC activity \(e.g., Gray, 1984; Chan, 1985; Lander, 1994; Chu and Wang, 1997; Kossin et al., 2010\).](#) Each simulation is integrated for 14.5 months, from 1 March of the first year through 14 May of the following year, with the first month [of each simulation](#) discarded as spin-up; output is recorded every 6 h.

20 We used the ECMWF Interim Reanalysis (ERA-I; [European Centre for Medium-Range Weather Forecasts, 2009; Dee et al., 2011](#)) with a spectral T255 resolution ( $\sim 0.7^\circ$  horizontal grid spacing) for present-day initial conditions. SST and sea ice fields are updated daily throughout the simulations. The configuration of these surface fields is discussed further in Section 2.3. For the future climate simulations, we modify the ERA-I initial and lower boundary conditions by adding monthly-averaged temperature changes derived from a 20-member ensemble of CMIP5 GCMs (Table 2). [A similar change was applied](#) 25 [to the deep-soil temperature.](#) These temperature changes are calculated by subtracting the 1980–1999 average temperature from the 2080–2099 average temperature following the Intergovernmental Panel on Climate Change (IPCC) Fifth Assessment Report (AR5) Representative Concentration Pathway (RCP) 8.5 emissions scenario, interpolated to the ERA-I grid, and added to the existing temperature data at all atmospheric pressure and soil levels. Geopotential height and specific humidity are adjusted by the model based on the imposed temperature changes; relative humidity is held constant at the initial time [in the](#) 30 [spin-up run \(see following paragraph\).](#) We set carbon dioxide (CO<sub>2</sub>) concentrations in the future climate simulations to 936 ppm, the level projected by the RCP8.5 emissions scenario for 2100 (Meinshausen et al., 2011). Present-day CO<sub>2</sub> concentrations are based on analysed values set according to the year.

Rather than running the simulations in chronological order, the simulation years are sorted from the strongest La Niña year (i.e., the year with the smallest MEI and most negative ONI; Table 1) to the strongest El Niño year (i.e., the year with the

largest MEI and most positive ONI; Table 1). This design aims to minimize model spin-up in response to changes in SST. With the present-day and future initial conditions set, we conduct full simulations for a neutral ENSO year (e.g., 2013) [for both present-day and future environments](#). These [single-year](#) simulations are used as spin-up and are, therefore, excluded from our analysis. While we took this precaution to allow the model atmosphere to come into equilibrium with the imposed warming and adjusted CO<sub>2</sub> for the future climate experiment, we repeated this process for the present-day simulation to maintain consistency. We then used the output from 1 March, towards the end of the initial spin-up simulation, to initialise the first simulation year. This method continues for both the present-day and future experiments by using the output from the latter part of one simulation (e.g., 1 March) to initialise the next (Fig. 2). Applying this unique “daisy-chain” technique avoids the need for excessive spin-up times for each year; instead, we discard only the output from the first month, which allows any discontinuities arising from the change in SST to equilibrate.

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### 2.3 Lower Boundary Conditions

As with the atmospheric initial conditions, the SST fields used in the simulations are taken primarily from the ERA-I. The SSTs in the ERA-I have, however, been derived from several different datasets over the years (Dee et al., 2011). For reanalysis times after February 2009, ERA-I surface fields originate from the Operational Sea Surface Temperature and Sea-Ice Analysis (OSTIA; Donlon et al., 2012). To maintain consistency between all simulations, the OSTIA SST, interpolated from its native 0.05° horizontal grid spacing to the ERA-I grid, is used for simulation years prior to 2009. Therefore, we effectively use OSTIA SST for all simulations. For present-day soil temperature and moisture, we use the ERA-I fields.

The SSTs for the future climate simulations are altered in the same manner as the initial condition atmospheric and soil temperatures (e.g., Fig. 3a–b). The same technique of adding a GCM delta field onto existing data cannot, however, be used for sea ice. Instead, similar to Mizielinski et al. (2014), monthly-averaged CMIP5 ensemble mean sea ice fields are temporally interpolated to create daily sea ice fields for both present-day (1980–1999) and future (2080–2099 under the RCP8.5 emissions scenario) time periods. An example of these sea ice fields is shown in Figure 3c–d. We then replaced the analysed sea ice in the ERA-I with these climatological fields for use in all model simulations. While the climatological present-day sea ice does not entirely match the analysed field in the ERA-I (e.g., the sea ice edge is much more diffuse), handling the sea ice in this manner ensures that it is plausibly represented in the future climate simulations. The presence of an overly diffuse ice edge could result in unrealistically weak lower tropospheric baroclinicity during warm seasons in these locations.

Our technique for simulating a future climate is similar to the PGW approach in the sense that (1) the analysed initial and lower boundary conditions are altered by adding projected temperature changes from GCMs to represent future conditions, and (2) analysed high-resolution SST fields are used to preserve realistic representation of ocean eddies and SST gradients. High-resolution SST is of demonstrated importance for midlatitude cyclone development and other regional climate changes (e.g., Brayshaw et al., 2011; Booth et al., 2012; Kirtman et al., 2012; He and Soden, 2016; Siqueira and Kirtman, 2016). By using a global model, however, one of the main limitations of PGW, the constraint of the lateral boundary conditions, is alleviated. [Therefore, our future climate simulations are best described as time-slice experiments with prescribed high-](#)

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[resolution SSTs](#). The UPSCALE experiments described by Mizieliński et al. (2014) use a similar time-slice technique for simulating a future climate. By simulating a small ensemble of 26 years, UPSCALE samples a broad range of interannual variability and ENSO states; however, 25 km grid spacing is insufficient for resolving full strength tropical cyclones (Davis, 2018). Therefore, our simulations complement UPSCALE by offering sufficiently high resolution to better capture the atmospheric mesoscale, specifically tropical cyclones.

#### 2.4 Computational Performance

We conducted the [MPAS-A](#) simulations on the [U.S.](#) National Center for Atmospheric Research (NCAR) supercomputer, Cheyenne (Computational and Information Systems Laboratory, 2017). Cheyenne is a 5.34 petaflop SGI ICE XA Cluster with 145,152 Intel Xeon processor cores and 313 terabytes (TB) of memory. Each 14.5-month simulation was run on 1152 cores and consumed roughly 92,000 CPU hours, including resources needed for post-processing, leading to a total of ~1.9 million core hours used for these experiments.

We post-process model output to vertically interpolate fields to selected isobaric levels and horizontally interpolate from the native unstructured mesh to a 0.15° x 0.15° latitude-longitude grid. Due to storage constraints, we saved a limited number of variables for the Northern Hemisphere only; however, monthly restart files are archived, enabling replication of a particular period of time or event as needed. The post-processed output occupies approximately 50 TB of storage space for the output for all 20 simulations, and is currently stored on Cheyenne's High Performance Storage Space (HPSS) and at North Carolina State University.

#### 2.5 Assumptions and Limitations

For our future climate simulations, we computed temperature delta values using the mean IPCC AR5 RCP8.5 emissions pathway. While other plausible scenarios exist, we selected a high emission pathway to maximize the signal of climate change in our simulations. Using the GCM ensemble mean temperature changes to alter our initial and lower boundary conditions diminishes the considerable amount of variability in the temperature changes projected by individual GCMs. Computing an ensemble mean from a set of simulations using temperature change fields from each GCM is, however, unlikely to produce significantly different results (Hill, 2010; Lackmann, 2015; Marciano, 2014). [An alternate strategy to using the GCM ensemble mean SST change would be to apply changes computed as a function of GCM ENSO phase. However, given uncertainties in the GCM SST change fields, we felt that the use of the ensemble mean was best for our initial set of experiments. Likewise, our analysis does not address possible changes in the distribution of ENSO phases.](#)

The adjustment of geopotential height based on the imposed temperature changes for the future climate simulation introduces some degree of imbalance between the model mass and wind fields. In previous studies, we utilized the Digital Filter Initialization (DFI) capability of the WRF model to reduce these imbalances. Since this feature is not available in [MPAS-A](#), we conducted a full 14.5-month spin-up simulation to allow time for the dynamics of the model atmosphere to restore balance. We also maintain constant relative humidity between the present-day and future simulations in the initial conditions. While this assumption may be appropriate over ocean basins, it does not necessarily hold true over land areas (e.g., Sherwood

and Fu, 2014). With no constraints on the lateral boundaries, however, this constraint is [only applied once, at the beginning of our spin-up simulation, and therefore, is not enforced during the subsequent simulations](#). The relative humidity within our model domain evolves freely through the duration of the simulations; by the end the [initial year-long](#) spin-up period, we expect the distribution of water vapour is fully equilibrated with the simulated future climate.

5 While our treatment of sea ice in the model allows for a plausible representation of future conditions, we use identical sea ice fields in each member of our present-day simulation set, and similarly for the future set. We therefore exclude the effects of interannual variability in sea ice. Several studies have highlighted the connection between sea ice variability and atmospheric circulations in the Northern Hemisphere (e.g., Deser et al., 2000; Overland and Wang, 2010); our intention here, however, is to minimize this influence and instead, focus on changes due to altered [temperatures and atmospheric composition](#).

10 Another limitation inherent in our methods is the assumption that future patterns of SST variability will remain similar to what they are today. Nevertheless, we believe the benefits of using high-resolution SST analyses to preserve realistic SST gradients and alleviate regional biases associated with atmosphere-ocean coupling (e.g., He and Soden, 2016) outweigh this limitation.

Many previous studies have shown that neglect of SST cooling due to cyclone passage results in TCs that are too strong (e.g., Schade and Emanuel, 2009). Use of analysed SST fields in our simulations does not allow for TC-generated cold wakes, which could contribute towards a positive bias in TC intensities and could lead to unrealistic temporal clustering of TCs. The use of convective parameterization, however, particularly the Tiedtke scheme which adjusts momentum, tends to weaken TCs through momentum adjustment in a warm-core cyclonic structure, an effect opposite to that resulting from the neglect of SST cooling. [Additionally, the presence of pre-existing cold wakes in the OSTIA SST field could erroneously weaken TCs that occur in their path; however, these pre-existing cold wakes in the OSTIA do not appear to be particularly strong in magnitude \(not shown\), and therefore, are unlikely have a substantial impact on simulated TC strength](#). Ideally, a grid length of 4 km or less would be used to fully capture TC structure and intensity (e.g., Gentry and Lackmann, 2010), but computational expense does not allow this for the Northern Hemisphere region of interest for the simulation durations necessary to obtain statistically meaningful results regarding the impacts of climate change. A benefit of our configuration is that the resolution is sufficiently high to capture nearly the full range of TC intensity; preliminary testing highlighted the capability of our 15 km grid to replicate realistic TC structures, including spiral rain bands and a defined eye (not shown). We acknowledge that the neglect of sea-surface cooling, the use of parameterized convection, [and potential effects of pre-existing cold wakes in the SST analysis data](#) are limitations to our approach. These limitations are, however, consistent between present-day and future simulations, allowing any differences found in TC intensity to remain meaningful (Patricola and Wehner, 2018).

25 We recognize that the methods employed in this study account only for projected changes due to increased anthropogenic greenhouse gases and therefore, do not represent other external climate forcings. Changes in other aspects of the climate system, such as changes in aerosols, deep soil moisture, and vegetation, are not represented. Despite the limitations discussed, our method alleviates limitations associated with regional PGW and coarse GCMs, and is much more computationally efficient than running high-resolution global models for long integration periods (e.g., centuries); the result is a set of controlled simulations suitable for examining the effects [of climate change](#) on high-impact weather events.

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### 3 Model Climate: Precipitation and Midlatitude Features

#### 3.1 Extratropical Storm Tracks

There are two primary midlatitude storm track regions in the Northern Hemisphere, the North Pacific and the North Atlantic, where baroclinic waves form over regions of enhanced temperature contrast linked to warm western boundary currents off the east coasts of Asia and North America and propagate eastward through downstream development (Chang et al., 2002). The extratropical cyclones in these regions play an essential role in the Earth's climate system and contribute to everyday weather, including high-impact events. Therefore, it is important that they are well represented in model simulations.

As suggested by Chang and Fu (2003), variance in daily-mean fields can be used as proxies for storm track activity. Here, we use the 24-h variance of daily-mean sea-level pressure (SLP), calculated using Eq. (2) from Chang et al. (2013):

$$SLP \text{ variance} = \overline{[SLP(t + 24h) - SLP(t)]^2}, \quad (1)$$

where the overbar indicates the quantity is averaged over time, in this case over the winter season (December–February; DJF) when storm activity in the Northern Hemisphere is maximized (Chang et al., 2002; Brayshaw et al., 2009). Figure 4 shows the wintertime SLP variance for the [MPAS-A](#) simulations compared to the ERA-I; the ERA-I climatology in Figure 4b is computed using only the ten years corresponding to our simulations. The North Pacific and North Atlantic storm track regions are clearly evident in the model simulations; the overall spatial correlation coefficient is greater than 0.98, indicating that general patterns of SLP variance are well reproduced in the [MPAS-A](#) simulations. As evident by the positive biases  $\sim 35^\circ$  N,  $165^\circ$  W and  $\sim 40^\circ$  N,  $30^\circ$  W (Fig. 4c), both storm track regions are shifted equatorward, and the North Pacific storm track is more zonally oriented in the [MPAS-A](#) simulations. Comparison with 100 random samples of 10-year means from the ERA-I record indicates that these biases, primarily the shift in the North Pacific, likely represent true differences between the [MPAS-A](#) simulations and the real atmosphere. Negative biases in simulated storm activity occur east of Greenland, over Scandinavia, and throughout central North America; these differences, however, fall within the range of observed variability (Fig. 4c).

#### 3.2 Northern Hemispheric Jet and Sea-Level Pressure Features

Corresponding to the Northern Hemispheric extratropical storm tracks are the midlatitude jet features, represented by the wintertime average zonal wind speed in Figure 5. As in Figure 4, the ERA-I climatology in Figure 5b includes only the ten simulation years. As indicated by a pattern correlation coefficient of  $\sim 0.99$ , the orientation and spatial extent of the North Pacific and North Atlantic jets at the 250-hPa level are well replicated by [MPAS-A](#) (Fig. 5a–b). While the North Atlantic jet maximum is slightly stronger in the [MPAS-A](#) simulations, the general strength of both features compares well between the simulations and reanalysis (Fig. 5a–b). Furthermore, examination of a cross-section of zonally averaged zonal wind shows the jet maximizes at roughly the same altitude and latitude ( $\sim 200$  hPa and  $\sim 30^\circ$  N) in both the [MPAS-A](#) simulations and the ERA-I climatology, albeit the [MPAS-A](#) maximum is moderately weaker (Fig. 5c–d). Additionally, semi-permanent maritime SLP features, such as the Aleutian Low over the Bering Sea, the North Pacific Subtropical High, and the Icelandic Low, are well

captured in the [MPAS-A](#) simulations (Fig. 5a–b). The Bermuda High in the North Atlantic, however, while evident in the [MPAS-A](#) simulations, is comparatively weaker than analysed.

### 3.3 Large-scale Precipitation

Average precipitation over the ten simulation years compared to the 19-year (1998–2016) climatology from the Tropical Rainfall Measuring Mission (TRMM; Huffman et al., 2007; [Tropical Rainfall Measuring Mission, 2011](#)) 3B42 product is shown in Figure 6. Because four of our ten simulation years occur before the TRMM record began, we opted to use the full TRMM climatology for comparison. A pattern correlation coefficient of  $\sim 0.95$  indicates that [MPAS-A](#) simulates the general spatial pattern of tropical precipitation well; the Intertropical Convergence Zone (ITCZ) in the equatorial Pacific and maxima along the west coast of India, over the Himalayas, and throughout northern South America are all well-represented by the model. The primary difference between the two precipitation fields is the overproduction of precipitation by [MPAS-A](#) in many areas (Fig. 6c), an issue common among other high-resolution modelling studies (e.g., Bacmeister et al., 2014; Small et al., 2014). The overestimation in the subtropical Pacific basin (Fig. 6c) is primarily due to overproductions of summer and fall precipitation. The summer season is also responsible for the overproduction of precipitation through the Bay of Bengal and Gulf of Thailand, suggesting an overactive summer monsoon in the [MPAS-A](#) simulations. Another notable difference between [MPAS-A](#) and TRMM annual average precipitation is the westward shift of the heaviest precipitation along the ITCZ in the Atlantic basin. This shift in precipitation is likely related to a westward shift of the summertime African Easterly Jet (AEJ; not shown).

### 4 Model Climate: Tropical Cyclones

Tropical cyclones epitomise the high-impact weather phenomena that our simulations are designed to address; therefore, we consider tropical cyclones as an appropriate exemplar of high-impact weather systems, in order to explore the usefulness of our simulations for examining the effects of climate change on such phenomena. To analyse how TCs appear in our model, we track simulated Northern Hemisphere tropical cyclones using the TempestExtremes objective, feature-based tracking algorithm (Ullrich and Zarzycki, 2017; Zarzycki and Ullrich, 2017). TCs are initially detected as minima in SLP, and then retained as candidate cyclone centres if certain criteria are met. Here, we require that TCs must have a 2 hPa closed SLP contour within  $2^\circ$  of the storm centre and a 300–500 hPa geopotential thickness maximum within  $6^\circ$  of the storm centre to ensure the presence of a warm core. Additionally, TCs must not travel more than  $6^\circ$  within a 6-h period, must have a lifetime of at least two days, must be located over water for at least 12 h, must have at least two days of 10-m wind speed of at least 14 m/s, and are required to have a genesis latitude south of  $45^\circ$  N. Trajectories that end and begin within 12 h of each other are merged together to prevent broken tracks from being counted twice. Once TC tracks have been obtained, TCs are separated into basins (Fig. 1) based on their genesis location.

#### 4.1 Strength

We compare simulated TC characteristics to the International Best Track Archive for Climate Stewardship (IBTrACS; Knapp et al., 2010). Only the IBTrACS for the ten simulated years (Table 1) are considered for comparison. Consistent with similar studies (e.g., Murakami et al., 2015; Roberts et al., 2015; Yamada et al., 2017), model storms are generally weaker than observed in terms of maximum 10-m wind speed; several simulated storms do, however, attain a minimum SLP of less than 900-hPa (Fig. 7). Therefore, as in Roberts et al. (2015), storm intensity for the simulated TCs is measured by the minimum lifetime SLP of the storm in addition to maximum 10-m wind speed as defined by the Saffir-Simpson scale. Using the minimum SLP, categories are defined as >994 hPa for tropical storms ( $TS_p$ ) and 980–994 hPa, 965–979 hPa, 945–964 hPa, 920–944 hPa, and <920 hPa for category ( $Cat_p$ ) 1–5 equivalent tropical cyclones, respectively (Roberts et al., 2015). The subscript p is used to discriminate the SLP-based categories from those defined by the Saffir-Simpson wind speed thresholds. For IBTrACS, the maximum 10-m wind speed and minimum SLP across all reporting centres are used as the observed storm intensity for categorization.

Figure 8 shows the average TC frequency over the ten simulation years for the Northern Hemisphere as a whole, in addition to each basin. Our [MPAS-A](#) simulations generate excess TC activity in the Northern Hemisphere, primarily due to the over-activity in the Western North Pacific basin. Simulated TC frequencies for the North Atlantic, Eastern North Pacific, and Northern Indian basins are within the observed range. Across all basins, when categorizing TCs by minimum SLP (Fig. 8a), [MPAS-A](#) generally underestimates the number of weak systems (those with strengths less than  $Cat_p1$ ), and overestimates the number of  $Cat_p1$  and  $Cat_p2$  storms. The frequencies  $Cat_p3$  TCs and stronger, on the other hand, are simulated reasonably well. With regard to TC categorization by maximum 10-m wind speed (Fig. 8b), [MPAS-A](#) simulates the frequency of TS strength TCs quite well in all basins. Strong TCs ( $Cat4$  and  $Cat5$ ), however, are universally underestimated by the model in favour of  $Cat1$ – $Cat3$  TCs. [In the future simulations, there is an increase in TC activity in both the North Atlantic and Western North Pacific basins \(not shown\); further investigation into these future changes in TCs will be the subject of a future paper.](#)

#### 4.2 Location

Spatially, the model simulated TC track density compares reasonably well with observations; the pattern correlation coefficient is about 0.7 (Fig. 9). The most prominent difference is the lack of TC activity in the eastern portion of the North Atlantic basin, which is common among several similar modelling studies (e.g., Bell et al., 2013; Strachan et al., 2013; Small et al., 2014; Roberts et al., 2015). TC genesis in this region typically occurs during August and September (Kossin et al., 2010; Daloz et al., 2015); comparison between the simulated atmosphere and ERA-I monthly-averaged 850–200 hPa vertical wind shear for the ten simulation years during these months shows a strong positive bias in the model over the North Atlantic development region that is likely a primary factor in this lack of TC generation (not shown). [Additionally, the westward shift in the simulated summertime AEJ \(not shown\) is consistent with a westward shift in the wave accumulation zone, which is likely impacting the location of TC genesis in this area \(Done et al., 2011\).](#) Figure 9 does not show a strong track density bias in the Gulf of Mexico. Roberts et al. (2015) note that a steady supply of vorticity in the Caribbean contributed to their overestimation of track density in this area; thus, it is possible that tracking TCs as SLP minima, rather than maxima in 850 hPa relative vorticity helps

alleviate this bias. Unlike previous studies (e.g., Small et al., 2014; Murakami et al., 2015; Roberts et al., 2015; Yamada et al., 2017), we do not find a positive track density bias in the central North Pacific; instead, we see a slight underrepresentation of TC activity in that area around  $\sim 150^\circ$  W.

#### 4.3 Seasonal Cycle

5 Aside from the underestimation in August and September (likely attributed to the lack of TC genesis in the eastern portion of  
the basin), [MPAS-A](#) simulates the present-day seasonal TC cycle for the North Atlantic reasonably well; TC activity increases  
during the spring and summer seasons and reaches a maximum in the fall (Fig. 10a). For the Eastern North Pacific, [MPAS-A](#)  
10 produces too many storms in the springtime (April and May) and too few storms during the summer months (Fig. 10b). As  
defined by Camargo et al. (2008), cluster 2 type Eastern Pacific TCs form off the coast of Mexico, travel towards the northwest  
along the coastline, and have a bimodal seasonal distribution with peaks in late spring/early summer and early fall, similar to  
the modelled cycle in Fig. 10b. Compared to the ERA-I ten-year climatology, enhanced westerlies at the 500- and 850-hPa  
15 levels between  $0\text{--}20^\circ$  N in the eastern portion of the Eastern Pacific basin for April and May (not shown) suggest that our  
simulations may be in a regime more conducive to these cluster 2 storms. For the Western North Pacific (Fig. 10c), [MPAS-A](#)  
correctly simulates the fall peak in TC activity; there is, however, a secondary peak in April that does not match observations.  
Although there is a general overestimation of storm activity in the Northern Indian basin, the model does replicate the shape  
of the seasonal cycle with both the early summer and mid-fall peaks represented, albeit the fall peak occurs one month earlier  
than observed (Fig. 10d).

#### 5 Climate Change Representation

To ensure that our simulations are useful in studying climate change effects on weather phenomena, we compare temperature  
20 change fields with large-scale warming patterns generated by a subset of IPCC GCMs. Previous theoretical and modelling  
studies demonstrate that the Arctic region will continue to warm at a faster rate than the rest of the globe in response to an  
increase in greenhouse gases (IPCC, 2013, §12.4.2.2). This polar amplification effect is captured in our simulations with  
portions of the Arctic experiencing temperature changes greater than 16 K compared to differences  $\leq 10$  K elsewhere (Fig.  
11a). However, we note that Arctic temperatures in our present-day simulations were colder compared to ERA-I, while the  
25 future simulations resulted in Arctic temperatures comparable to those produced by GCMs (not shown). As a result, the [MPAS-  
A](#) simulations produce a larger magnitude of warming in the Arctic compared to the GCM ensemble (Fig. 11a–b). Another  
result consistent with theory and previous modelling studies is the development of a warming maximum in the tropical upper-  
troposphere (IPCC, 2013, §12.4.3.2). This area of warming, which occurs between the  $\sim 400\text{-}$  and  $\sim 150\text{-hPa}$  levels and which  
maximizes around the 250-hPa level (IPCC, 2013, §12.4.3.2), has been shown to partially mitigate projected increases in TC  
30 intensity associated with warming (e.g., Knutson and Tuleya, 1999; Shen et al., 2000; Hill and Lackmann, 2011). As shown  
in Fig. 11c, this warming signature is replicated in the [MPAS-A](#) simulations.

## 6 Summary and Conclusions

We present a novel set of model simulations produced using [MPAS-A](#) in current and future environments that is designed to maximize our ability to analyse changes in high-impact weather systems; such changes will be reported in future studies. Our use of a global model eliminates the lateral boundary constraints of regional models, while inclusion of high-resolution, analysed SSTs preserves realistic SST gradients throughout the duration of the simulations. Furthermore, a grid length of 15 km offers an advantage over coarser modelling studies to better represent the atmospheric mesoscale. The future climate simulations employ a technique that combines methods associated with PGW and time-slice experiments; this allows for the inclusion of high-resolution SSTs, plausible future sea ice fields, and seamless simulation of non-consecutive years without excessive spin-up time. While the primary purpose of our simulations is to study climate change effects on Northern Hemispheric high-impact weather, to achieve this it is necessary to first evaluate the model climate in regard to present-day large-scale circulations as well as the large-scale responses to warming in our climate change experiments; reasonable representation of these aspects is essential to justify moving forward to investigate smaller-scale, high-impact phenomena.

Key results from these simulations include the ability of [MPAS-A](#) to reproduce Northern Hemispheric wintertime midlatitude storm tracks (Fig. 4) along with semi-permanent maritime SLP and upper-tropospheric jet features (Fig. 5). Tropical characteristics, such as precipitation along the ITCZ in the equatorial Pacific, are also well simulated ( $r \sim 0.95$ ), although the ITCZ representation in the Atlantic does not compare as favourably to observations (Fig. 6). In regard to TC strength, [MPAS-A](#) is able to produce several tropical cyclones of Cat4 strength, as defined by traditional maximum 10-m wind speed thresholds of the Saffir-Simpson scale (Fig. 7 and Fig. 8b). Categorizing TCs using the minimum SLP thresholds of Roberts et al. (2015), on the other hand, shows simulated TCs across the full intensity spectrum, including Cat<sub>5</sub> storms (Fig. 7 and Fig. 8a).

While [MPAS-A](#) overestimates TC activity in the Western North Pacific, TC frequency in other Northern Hemispheric basins is within the range of observations (Fig. 8). The largest discrepancy in the simulated spatial distribution of TCs is the lack of TC genesis in the eastern North Atlantic (Fig. 9), likely due to a positive bias in vertical wind shear (not shown). Otherwise, TC density patterns match observations reasonably well ( $r \sim 0.7$ ). Additionally, with the exception of the Eastern North Pacific, the seasonal cycles for the Northern Hemispheric basins are well reproduced (Fig. 10). Last, our future simulations replicate two key warming signatures produced by GCMs: Arctic amplification and the warming maximum in the tropical upper-troposphere (Fig. 11).

With our modelling approach, we strive to contribute to the intersection of weather and climate modelling, and aim to fill a gap between GCMs, which are unable to simulate small-scale weather phenomena, and high-resolution limited area models, which are constrained by lateral boundaries, to provide the possibility of studying high-impact events in a consistent global context. We anticipate these simulations, in conjunction with similar efforts, will have great value in projecting and understanding changes in high-impact weather phenomena for which dynamics on sub-synoptic scales are important. Beyond the tropical cyclones described here, this could include flooding rains and damaging winds associated with extratropical cyclones, flooding monsoon rains, and localized droughts and heat waves. Research involving these simulations is currently underway investigating climate change effects on the following phenomena:

Deleted: thermodynamic

- Extratropical transition of TCs
- TC seasonality
- Midlatitude precipitation extremes and windstorms embedded in extratropical cyclones
- Persistent anomalies and blocking

5 Many more aspects of these simulations, however, remain to be explored. Therefore, we make the simulation output available to the research community as detailed in the Acknowledgements section in the hope that it will be useful to the broader scientific community for studying various meteorological phenomena, as well as for conducting model comparison studies.

#### Data and code availability

10 The source code for the model used in this study, [MPAS-A](https://mpas-dev.github.io), is freely available from <https://mpas-dev.github.io>. [The archived source code for the version of MPAS-A used in this study \(release version 5.1\) is available from https://github.com/MPAS-Dev/MPAS-Model/releases/tag/v5.1](https://github.com/MPAS-Dev/MPAS-Model/releases/tag/v5.1). The [MPAS-A](https://github.com/MPAS-Dev/MPAS-Model/releases/tag/v5.1) model output from the simulations presented in this manuscript is located on the Cheyenne High Performance Storage System (HPSS) and on the [North Carolina State University \(NCSU\)](https://www.ncsu.edu) Henry2 Cluster. Please contact the corresponding author for additional details on accessing this data. [The initialisation files can also be accessed by contacting the corresponding author. Sample run and post-processing scripts are available from](#)

15 <http://doi.org/10.5281/zenodo.3244401> (Michaelis 2019). The TempestExtremes tracking algorithm used in this study is available [under the Lesser GNU Public Licence \(LGPL\) and can be accessed from](#)

20 <https://github.com/ClimateGlobalChange/tempestextremes> as detailed in Ullrich and Zarzycki (2017). ECMWF Interim Reanalysis can be obtained from <https://doi.org/10.5065/D6CR5RD9> (European Centre for Medium-Range Weather Forecasts, 2009). TRMM 3B42v7 data can be accessed from [https://disc.gsfc.nasa.gov/datasets/TRMM\\_3B42\\_V7/summary](https://disc.gsfc.nasa.gov/datasets/TRMM_3B42_V7/summary) (Tropical Rainfall Measuring Mission, 2011). IBTrACS version v03r10 can be obtained from <https://www.ncdc.noaa.gov/ibtracs>.

#### Author contributions

All authors contributed equally in designing the model experiments. A. Michaelis conducted and analysed the simulations. A. Michaelis wrote the manuscript with editorial modifications from G. Lackmann and W. Robinson.

#### 25 Competing interests

The authors declare they have no conflicts of interest.

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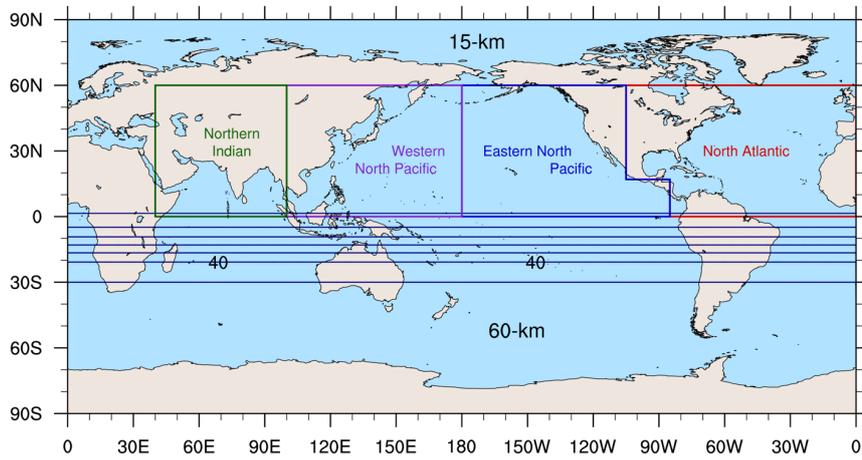
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**Table 1: Average Multivariate ENSO Index (MEI), Oceanic Niño Index (ONI), and Corresponding ENSO Phase During the TC Season (June–November) for the Chosen Simulation Years.**

Year	Multivariate ENSO Index (MEI) Rank:	Oceanic Niño Index (ONI):	ENSO phase
	JJ–ON average	JJA–SON average	
2010	3.8	-1.2	Strong La Niña
1988	6.6	-1.2	Strong La Niña
2011	16.2	-0.7	Weak La Niña
2013	26.8	-0.2	Neutral
2001	31.8	-0.1	Neutral
2005	34.2	0.0	Neutral
1992	47.5	0.3	Neutral
1994	57.1	0.5	Weak El Niño
2015	64.8	1.7	Strong El Niño
1997	66.0	1.8	Strong El Niño

**Table 2: List of 20 CMIP5 GCMs Used to Compute Ensemble Mean Temperature “Deltas” and Sea Ice Fields.**

Model	Modeling Center/Group	Grid Length
ACCESS1-0	Commonwealth Scientific and Industrial Research Organization (CSIRO) and Bureau of Meteorology (BOM), Australia	1.25° x 1.875°
ACCESS1-3		
CanESM2	Canadian Centre for Climate Modeling and Analysis	2.8° x 2.8°
CMCC-CM	Centro Euro-Mediterraneo sui Cambiamenti Climatici (Euro-Mediterranean Center on Climate Change)	0.8° x 0.8°
CNRM-CM5	National Centre of Meteorological Research, France	1.4° x 1.4°
GISS-E2-H	NASA Goddard Institute for Space Studies	2° x 2.5°
GISS-E2-H-CC		
GISS-E2-R-CC		
HadGEM2-AO	Met Office Hadley Centre	1.25° x 1.875°
HadGEM2-ES		
INMCM4	Institute for Numerical Mathematics	1.5° x 2.0°
IPSL-CM5A-MR	Institut Pierre Simon Laplace, France	1.25° x 2.5°
IPSL-CM5B-LR		1.8° x 2.75°
MIROC-ESM	Japan Agency for Marine-Earth Science and Technology, Atmosphere and Ocean Research Institute (The University of Tokyo), and National Institute for Environmental Studies	2.8° x 2.8°
MIROC-ESM-CHEM		
MPI-ESM-LR	Max Planck Institute for Meteorology	1.8° x 1.8°
MPI-ESM-MR		
MRI-ESM1	Meteorological Research Institute, Japan	1.1° x 1.1°
NorESM1-M	Norwegian Climate Center, Norway	1.9° x 2.5°
NorESM1-ME		



5 **Figure 1: Variable-resolution mesh for MPAS simulations and geographical regions of the tropical cyclone basins defined in this study.**

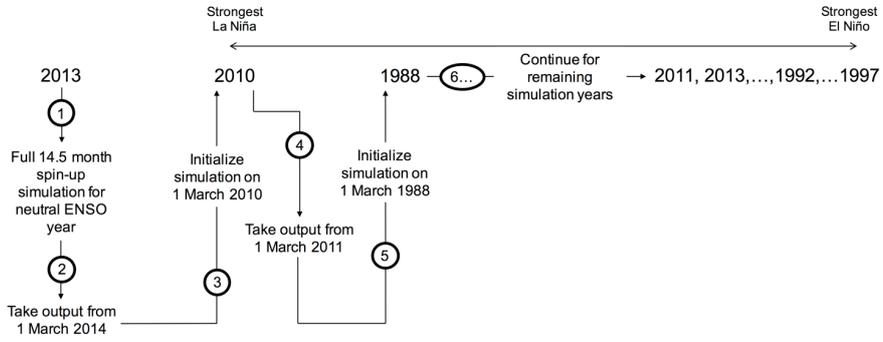
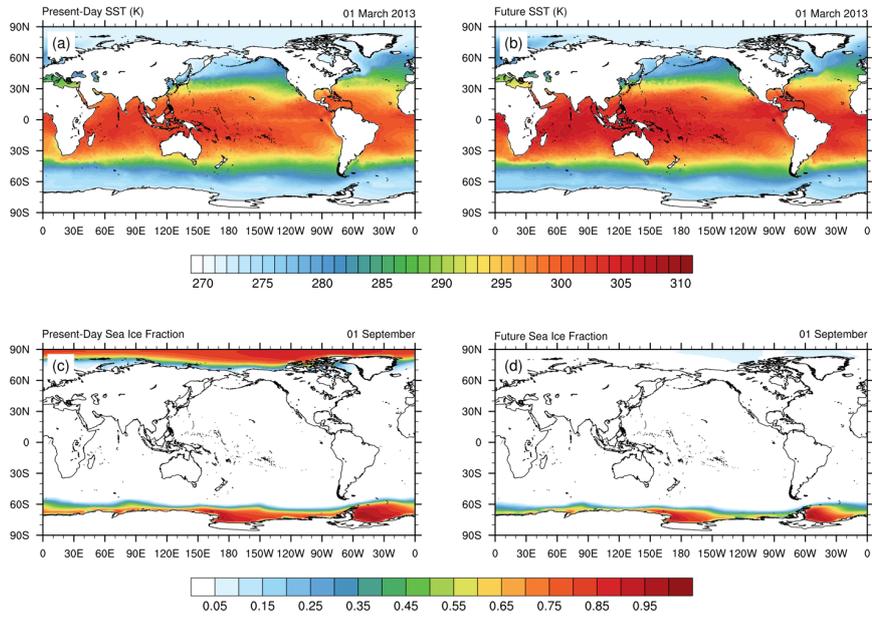
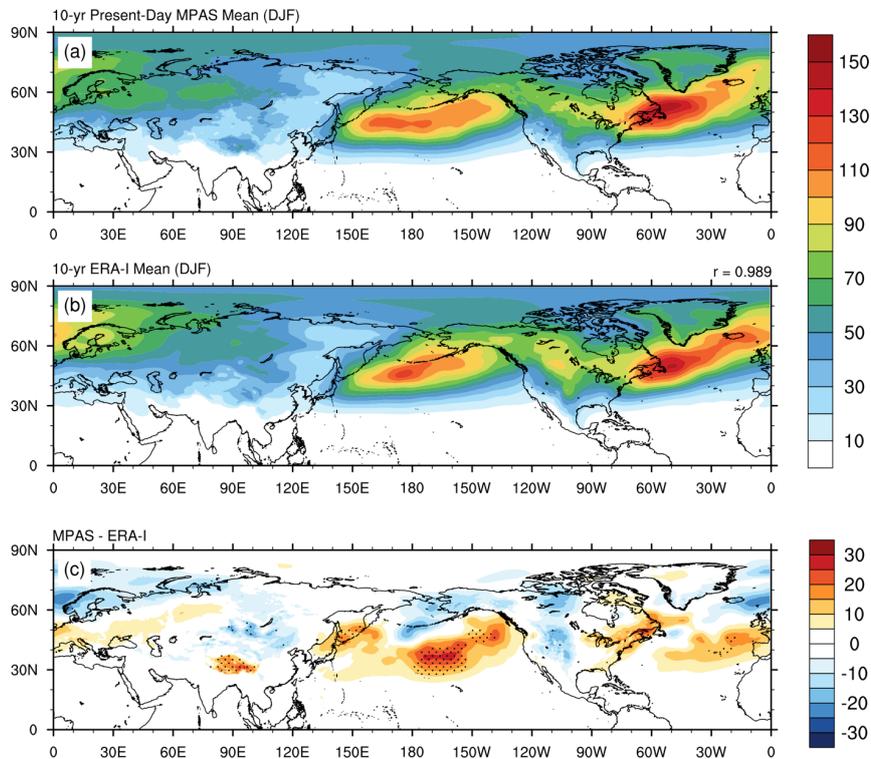


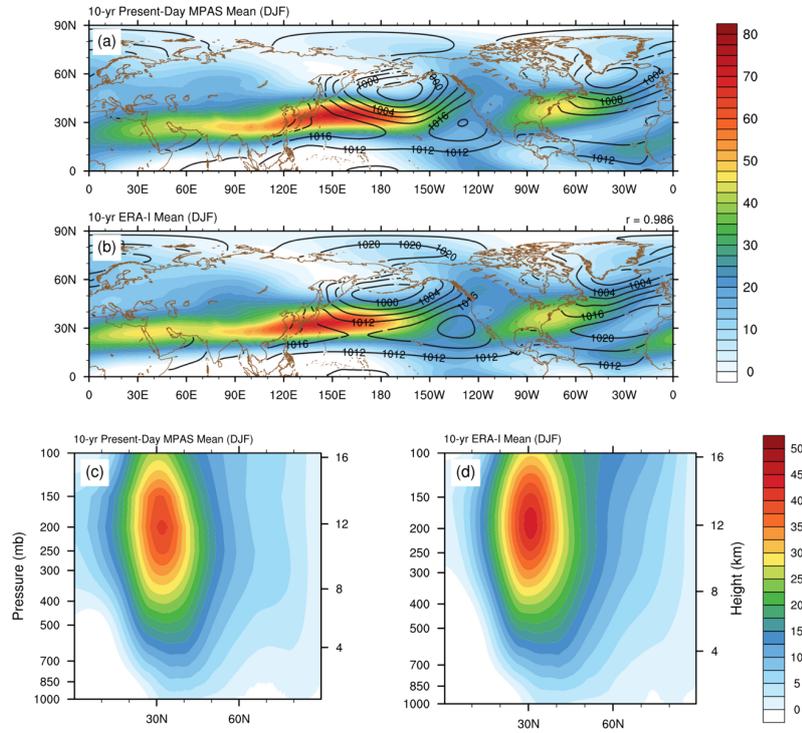
Figure 2: Flow chart depicting the “daisy-chain” simulation method.



**Figure 3: Example SST (K) on 1 March 2013 for (a) present-day and (b) future MPAS simulations and example sea ice fraction on 1 September for the (c) present-day and (d) future MPAS simulations. Contours are shaded every 1 K in (a) and (b) and every 0.05 units in (c) and (d).**



5 **Figure 4:** Average wintertime (DJF) SLP variance ( $\text{hPa}^2$ ) over the ten simulation years for (a) present-day MPAS simulations, (b) ERA-I 10-yr climatology, and (c) the model bias (MPAS minus ERA-I). Contours are shaded every  $10 \text{ hPa}^2$  in (a) and (b) and every  $5 \text{ hPa}^2$  in (c). MPAS output were linearly interpolated to the ERA-I grid for point-to-point comparison. The pattern correlation coefficient is reported at the top-right of (b). Stippling in (c) indicates locations where the MPAS 10-yr mean exceeds the range computed from 100 random samples of 10-yr means from ERA-I by more than 5% ( $\sim 7 \text{ hPa}^2$ ).



5 **Figure 5: (Top row) Average wintertime (DJF) 250-hPa zonal wind speed ( $\text{ms}^{-1}$ ; shaded every  $2.5 \text{ ms}^{-1}$ ) and SLP (hPa; contoured every  $4 \text{ hPa}$ ) over the ten simulation years for (a) present-day MPAS simulations and (b) ERA-I 10-yr climatology. (Bottom row) Average wintertime (DJF) cross-section of zonally averaged zonal wind speed ( $\text{ms}^{-1}$ ; shaded every  $2.5 \text{ ms}^{-1}$ ) for (c) present-day MPAS simulations and (d) ERA-I 10-yr climatology. The pattern correlation coefficient for 250-hPa zonal wind speed is reported at the top-right of (b). SLP contours in (a) and (b) are masked over land owing to noise in areas of complex terrain.**

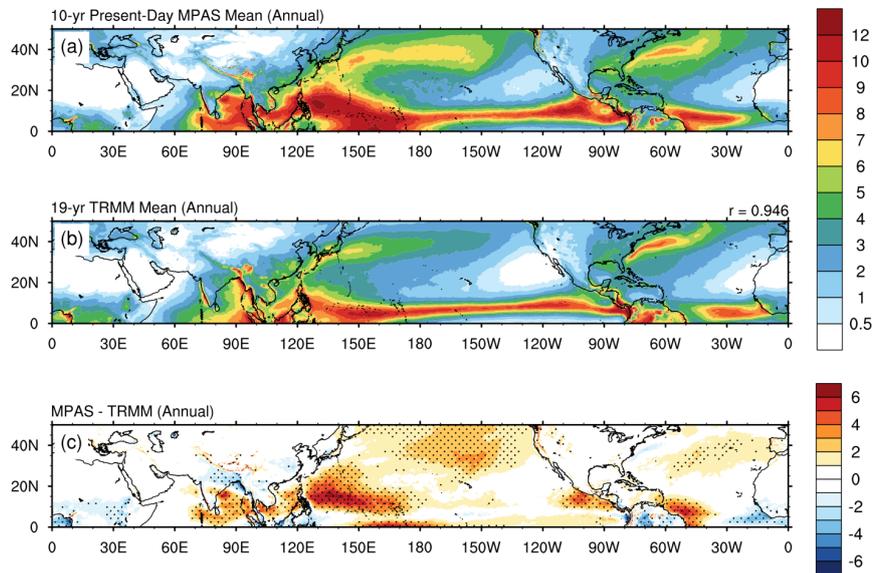


Figure 6: Average total annual precipitation (mm day<sup>-1</sup>) for (a) present-day MPAS simulations, (b) TRMM 19-yr climatology, and (c) the model bias (MPAS minus TRMM). MPAS output were linearly interpolated to the TRMM grid for point-to-point comparison. Pattern correlation coefficient is reported in the top-right of (b). Stippling in (c) indicates locations where the MPAS 10-yr mean exceeds the range computed from 100 random samples of 10-yr means from TRMM by more than 5% ( $\sim 1.2$  mm day<sup>-1</sup>).

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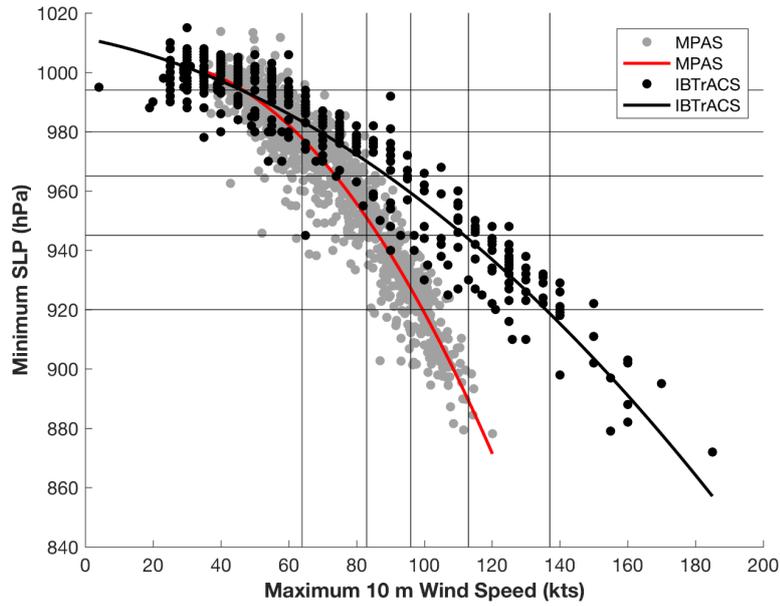


Figure 7: Scatter plot of maximum 10-m wind speed (kts) versus minimum SLP (hPa) for IBTrACS (black) and present-day MPAS (grey) Northern Hemispheric TCs. The lines of best fit for each (IBTrACS in black and MPAS in red) were computed using a second-order polynomial. The wind (SLP) category thresholds are indicated by the vertical (horizontal) lines.

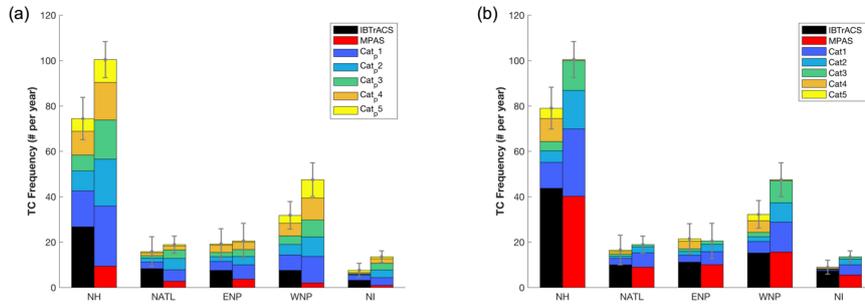


Figure 8: Average number of TCs over the ten simulation years for the Northern Hemisphere and each Northern Hemisphere basin. Columns are coloured by intensity categories based on (a) minimum lifetime SLP and (b) maximum lifetime 10-m wind speed. The bottom colour represents intensities of tropical storm strength or less for IBTrACS and MPAS in the first and second columns, respectively. Categories 1–5 are shaded for both data sets according to the legend. The error bars indicate the interannual standard deviation. The number of TCs for IBTrACS varies based on strength metric due to the lack of SLP records for a select number of storms.

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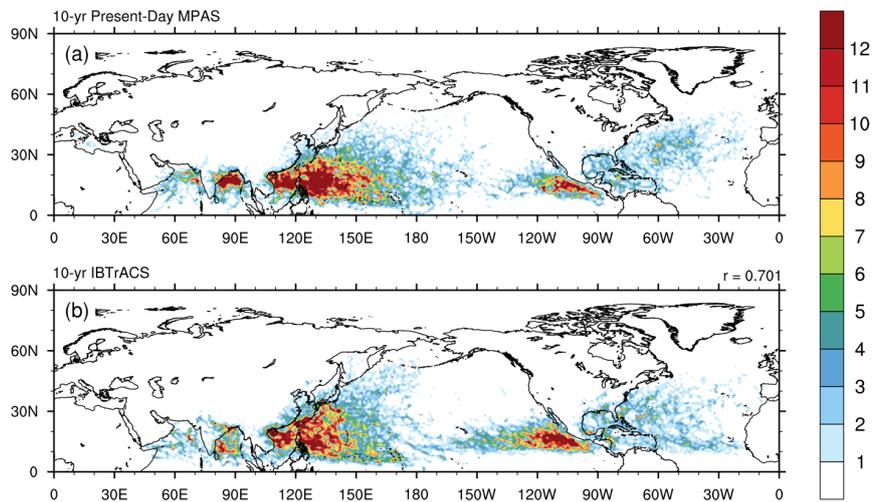
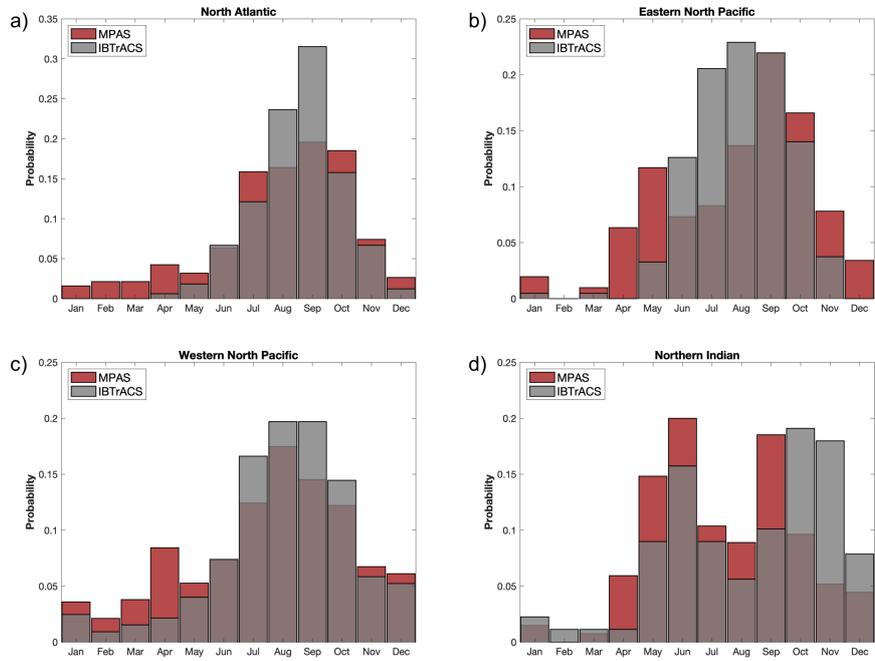
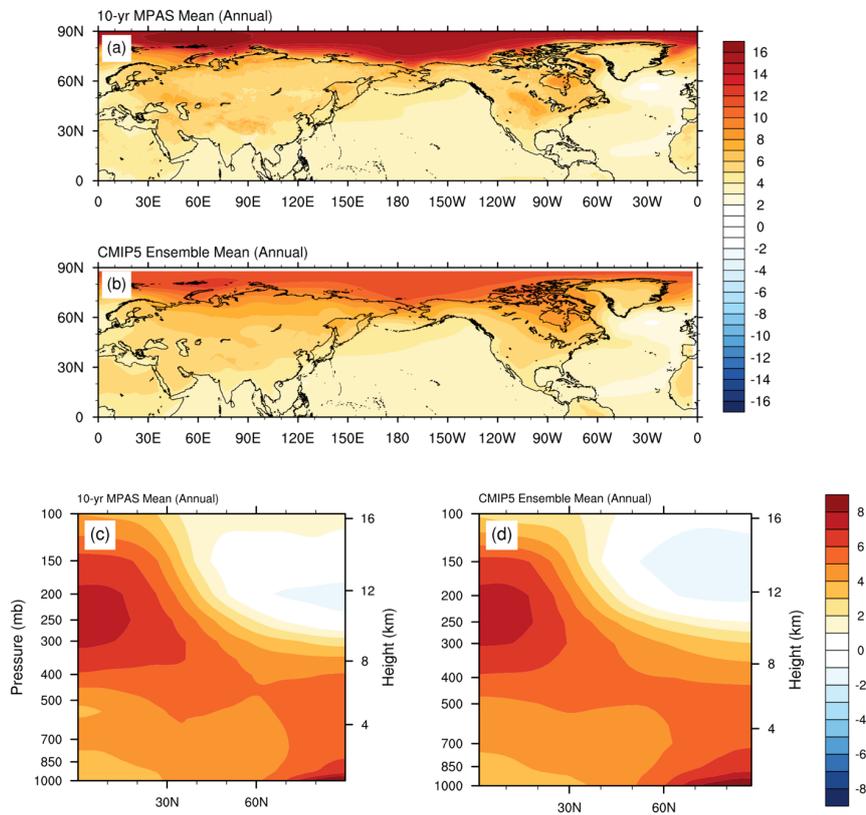


Figure 9: Track density (number of cyclone tracks per 1° x 1° area) over the ten simulated years for (a) present-day MPAS simulations and (b) IBTrACS. Contours are shaded every 1 count.



**Figure 10: Monthly average TC frequency over the ten simulated years for the (a) North Atlantic, (b) Eastern North Pacific, (c) Western North Pacific, and (d) Northern Indian basins. The frequencies for IBTrACS (MPAS simulations) are shown in the grey (red) bars.**



**Figure 11: (Top row) Average annual 2 m temperature difference (K; future minus current) for the (a) MPAS simulations and (b) CMIP5 GCM ensemble mean. (Bottom row) Average annual difference cross-section of zonal mean temperature (K; future minus current) for the (c) MPAS simulations and (d) CMIP5 GCM ensemble mean. Contours are shaded every 1 K in all panels.**