

*Extratropical transition of tropical cyclones
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climate models*

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Extratropical transition of tropical cyclones in a multiresolution ensemble of atmosphere-only and fully coupled global climate models

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1 **Abstract**

2 Tropical cyclones undergo extratropical transition (ET) in every ocean basin. Projected
3 changes in ET frequency under climate change are uncertain and differ between basins, so
4 multimodel studies are required to establish confidence. We used a feature-tracking algorithm
5 to identify tropical cyclones and performed cyclone phase-space analysis to identify ET in an
6 ensemble of atmosphere-only and fully coupled global model simulations, run at various
7 resolutions under historical (1950–2014) and future (2015–2050) forcing. Historical
8 simulations were evaluated against five reanalyses for 1979–2018. Considering ET globally,
9 ensemble-mean biases in track and genesis densities are reduced in the North Atlantic and
10 Western North Pacific when horizontal resolution is increased from ~100 to ~25km. At high
11 resolution, multireanalysis-mean climatological ET frequencies across most ocean basins as
12 well as basins’ seasonal cycles are reproduced better than in low-resolution models. Skill in
13 simulating historical ET interannual variability in the North Atlantic and Western North
14 Pacific is ~0.3, which is lower than for all tropical cyclones. Models project an increase in ET
15 frequency in the North Atlantic and a decrease in the Western North Pacific. We explain
16 these opposing responses by secular change in ET seasonality and an increase in lower-
17 tropospheric, pre-ET warm-core strength, both of which are largely unique to the North
18 Atlantic. Multimodel consensus about climate-change responses is clearer for frequency
19 metrics than for intensity metrics. These results help clarify the role of model resolution in
20 simulating ET and help quantify uncertainty surrounding ET in a warming climate.

21

22 **1. Introduction**

23 The impacts of tropical cyclones are not confined to the tropics. Their post-tropical evolution
24 makes these storms an important natural hazard across the midlatitudes (Baker et al., 2021;
25 Bieli et al., 2019; Evans et al., 2017; Jones et al., 2003; Keller et al., 2019). The poleward
26 propagation of tropical cyclones and the occurrence of extratropical transition (ET) exposes
27 populous regions where risks to life and infrastructure are high—Northeast United States,
28 maritime and eastern Canada, Western Europe, and East Asia—to hurricane-force wind
29 speeds and extreme precipitation (Evans et al., 2017). In the North Atlantic, tropical-origin
30 systems reached Northeast North America and Europe almost every year since 1979 (Baker
31 et al., 2021), including recent intense landfalls. For instance, Hurricane Sandy (22nd–29th
32 October, 2012)—the fourth costliest (by inflation-adjusted losses) North Atlantic hurricane
33 yet recorded (Weinkle et al., 2018)—caused devastation across the Northeast United States
34 and eastern Canada (Blake et al., 2013). Ex-hurricane Ophelia (9th–15th October, 2017) led to
35 loss of life and severe wind damage across Ireland, the United Kingdom, and Scandinavia
36 (Rantanen et al., 2020; Stewart, 2018). At midlatitude landfall, both systems were post-
37 tropical, having begun ET, but possessed hurricane-like intensities, the human and economic
38 impacts of which were felt across substantial areas. In the Western North Pacific, Typhoon
39 Nabi (29th August–12th September, 2005) impacted two thirds of Japan’s prefectures as both
40 a tropical and transitioning cyclone before undergoing cyclolysis over Alaska (Harr et al.,
41 2008). These events, along with the current lack of consensus regarding ET in a changing
42 climate, heighten the urgency with which global studies of historical and near-future post-
43 tropical cyclone activity are needed.

44

45 Tropical cyclones undergo ET in every ocean basin (Hart and Evans, 2001; Studholme et al.,
46 2015; Wood and Ritchie, 2014; Zarzycki et al., 2017), but pronounced interannual variability
47 (Baker et al., 2021) and basin-to-basin differences (Bieli et al., 2019) exist. Transitioning
48 cyclones are also known to influence the large-scale circulation, such as Hurricane Debbie in
49 1982 (Laurila et al., 2019), and excite or amplify downstream Rossby waves (Evans et al.,
50 2017; Jones et al., 2003; Keller et al., 2019; Michaelis and Lackmann, 2019). These cyclone-
51 wave interactions influence downstream weather (Grams and Blumer, 2015; Keller et al.,
52 2019). Of those cyclones which undergo ET, an appreciable proportion reintensify under
53 favourable environmental conditions, where appropriate phasing between the transitioning
54 cyclone and the upper-tropospheric flow pattern enhances baroclinic instability (Keller et al.,

55 2019). During and after ET, baroclinicity (Evans et al., 2017) and diabatic heating (Rantanen
56 et al., 2020) may reintensify the post-tropical cyclone.

57

58 Over the period of 1979–2018, statistically significant positive trends in the frequency of
59 North Atlantic ET events exist in several, but not all, reanalysis datasets (Baker et al., 2021).
60 Existing climate model projections underline the plausibility of increased tropical and post-
61 tropical cyclone activity in the midlatitudes in response to anthropogenic warming. There is
62 evidence that more frequent ET events may occur in the future in the North Atlantic (Baatsen
63 et al., 2015; Haarsma et al., 2013; Liu et al., 2017; Michaelis and Lackmann, 2019) and
64 Western North Pacific (Bieli et al., 2020) ocean basins, but no consensus yet exists across
65 studies, modelling campaigns, and methodologies. Moreover, best-track data limitations,
66 which are well documented (Chang and Guo, 2007; Delgado et al., 2018; Hagen et al., 2012;
67 Vecchi and Knutson, 2008), engender substantial uncertainty in observed trends (Lanzante,
68 2019; Moon et al., 2019). Additionally, natural, multidecadal variability in tropical-cyclone
69 frequency is yet to be accounted for (Knutson et al., 2020). Although global climate models
70 project reduced frequencies of tropical cyclones, more intense tropical cyclones are expected
71 in response to twenty-first-century warming (Knutson et al., 2020), potentially allowing a
72 higher proportion of cyclones to survive cooler midlatitude sea-surface temperatures
73 experienced prior to and during ET (Michaelis and Lackmann, 2019). Other factors,
74 particularly changes in shear, will also be important, with current evidence suggesting that
75 these will undergo ET-favourable future changes (Jung and Lackmann, 2021; Liu et al.,
76 2017; Michaelis and Lackmann, 2021). Increased future ET event frequency is also
77 consistent with the projected expansion of tropical-cyclone genesis regions (Studholme et al.,
78 2022), potentially reducing the mean displacement cyclones must undergo prior to
79 midlatitude ET. Together, these changes imply an increase in post-tropical cyclone impacts
80 across populated midlatitude regions, and idealised experiments suggest an increase in ET-
81 related, high-impact weather across Europe (Jung and Lackmann, 2021), where our
82 understanding of historical risks is developing (Baker et al., 2021). Studies of historical and
83 future model simulations are therefore needed to assess both contemporary risk and future
84 changes more comprehensively.

85

86 One aspect of climate model evaluation important for both tropical and extratropical cyclones
87 is understanding the role of horizontal resolution in simulated climates, prompted by recent
88 developments in high-performance computing and data-management facilities. With

89 increases in model resolution to approximately 25 km, improved fidelity is anticipated for
90 many synoptic phenomena, particularly tropical and midlatitude cyclones, which ultimately
91 feed back onto the large scale. Recent studies have now firmly established that increasing
92 model resolution improves simulated tropical-cyclone frequency statistics across most ocean
93 basins (Manganello et al., 2019; Roberts et al., 2020a), leads to a more realistic global spatial
94 distribution (Roberts et al., 2020a; Roberts et al., 2015; Strachan et al., 2013), and results in
95 more realistic simulated warm-core vertical structures (Vannière et al., 2020). Moreover,
96 model resolution is a key constraint on the intensity which simulated cyclones may reach
97 (Davis, 2018). It is anticipated that atmospheric resolutions of ~ 50 km or finer ($\sim 0.25^\circ$
98 ocean-model resolution) will yield improvement in the simulation of post-tropical cyclones
99 and ET (Haarsma, 2021). However, no systematic multimodel studies of ET have been
100 undertaken, and the impact of increasing model resolution (atmosphere and ocean) on
101 simulated ET is also yet to be quantified. We address these issues in this paper using model
102 simulations from the 6th phase of the Coupled Model Intercomparison Project (CMIP6),
103 which follow an experimental protocol designed to isolate the impacts of changes in model
104 resolution.

105

106 In this study of the representation of tropical cyclones undergoing ET across a multimodel
107 ensemble, we focus on climatological statistics, interannual variability, and cyclone structure
108 and intensity. These analyses are centred around two questions. What is the impact of
109 increasing model atmospheric resolution on simulated ET? What changes in ET metrics
110 under climate change are consistent across models? This paper continues in section 2 with a
111 description of the model and reanalysis data as well as the cyclone-tracking and analysis
112 methodologies. Our results are presented in section 3 and our conclusions are summarised,
113 with further discussion, in section 4.

114

115

116 **2. Data and methodology**

117 *2.1 Reanalysis data*

118 Tropical cyclone best-track datasets are not well suited to analysis of cyclones undergoing ET
119 because there are known heterogeneities within individual datasets (Barcikowska et al., 2012;
120 Chu et al., 2002; Kossin et al., 2007; Vecchi and Knutson, 2008, 2011), especially for storms'
121 post-tropical stages, under-counting biases (Chang and Guo, 2007; Delgado et al., 2018;
122 Hagen et al., 2012), and differences between operational centres' data-collection

123 methodologies (Hodges et al., 2017; Schreck III et al., 2014). We therefore evaluated model
124 simulations against five global reanalyses (Table 1): the European Centre for Medium-Range
125 Weather Forecasts' Interim Reanalysis (ERA-Interim; Dee et al., 2011) and Fifth Reanalysis (ERA5;
126 Hersbach et al., 2020); the Japanese 55-year Reanalysis (JRA55; Kobayashi et al., 2015); the
127 National Aeronautics and Space Administration's Modern-Era Retrospective Analysis for
128 Research and Applications version 2 (MERRA2; Molod et al., 2015); and the combined
129 National Centers for Environmental Prediction Climate Forecast System Reanalysis and
130 Climate Forecast System version 2 dataset (NCEP; Saha et al., 2014)—the sole fully coupled
131 (atmosphere, ocean, land surface, and sea ice) reanalysis used herein. Between reanalyses,
132 differing forecast model formulations and resolutions (horizontal and vertical), as well as
133 data-assimilation schemes lead to differences in the representation of tropical-cyclone
134 vertical structure, which was examined by Hodges et al. (2017). Baker et al. (2021) found
135 that interannual variability in the number ET events is well correlated between reanalyses,
136 but the percentage of tropical cyclones undergoing ET agrees less well between reanalyses on
137 the interannual timescale. It is therefore necessary to consider multiple reanalyses as an
138 observation-based reference, against which models may be evaluated.
139

Reanalysis	Analysis period	Analysis grid	Model resolution (grid spacing)	Data assimilation	Sample sizes (n_{NH} , n_{SH})
ERA5	1979–2017	512x256	TL255L60 (80 km)	4D-Var.	35.4, 37.0
ERA5	1979–2018	1140x721	T639L137 (33km)	4D-Var.	35.4, 37.0
JRA55	1959–2017	288x145	TL319L60 (55 km)	4D-Var.	35.4, 37.0
MERRA2	1980–2016	576x361	Cubed sphere (50 km)	3D-Var. + GSI + IAU	35.4, 37.0
NCEP	1979–2016	720x361	T382L64 (38 km)	3D-Var. + GSI	35.4, 37.0

142 **Table 1.** Reanalyses. Atmospheric mesh spacing at 50 °N in units of km is given in brackets.
143 3(4)D-Var.: 3(4)D variational data assimilation; GSI: Grid-point Statistical Interpolation;
144 IAU: Incremental Analysis Update. The representation of tropical and post-tropical cyclones
145 in these reanalyses were evaluated by Hodges et al. (2017) and (Baker et al., 2021),
146 respectively. Annual-mean global sample sizes (cyclones year⁻¹) for all tropical cyclones
147 undergoing ET for each reanalysis are given as n_{NH} , n_{SH} .
148

149 2.2 The multiresolution PRIMAVERA model ensemble

150 We evaluated CMIP6 High-Resolution Model Intercomparison Project (HighResMIP;
151 Haarsma et al., 2016) historical and future atmosphere-only (Tier 1 and Tier 3, respectively),
152 including interaction with the land surface, and fully coupled (Tier 2) simulations from five
153 global climate models (Table 2): CNRM-CM6.1 (Voldoire et al., 2019), EC-Earth3P
154 (Haarsma et al., 2020), ECMWF-IFS (cycle 43r1; Roberts et al., 2018), HadGEM3-GC3.1
155 (Roberts et al., 2019; Williams et al., 2018), and MPI-ESM1.2 (Gutjahr et al., 2019). Each
156 model participated in the European Commission Horizon2020-funded project PRIMAVERA
157 (PProcess-based climate sIMulation: AdVances in high-resolution modelling and European
158 climate Risk Assessments; primavera-h2020.eu). Historical (1950–2014) and future (2015–
159 2050) atmosphere-only experiments are termed *highresSST-present* and *highresSST-future*,
160 respectively, and fully coupled experiments are termed *hist-1950* and *highres-future*,
161 respectively. Historical *highresSST-present* simulations were forced by HadISST2 daily sea-
162 surface temperature (SST) at a resolution of 0.25° interpolated to each model's grid (no
163 ocean mixed-layer model). Out to 2050, *highresSST-future* simulations were forced
164 according to Representative Concentration Pathway 8.5 (RCP8.5). (Use of RCP8.5 allowed
165 modelling centres to begin their model simulations before Shared Socioeconomic Pathways
166 scenarios became available.) In HighResMIP, future simulations were performed with all
167 models except ECMWF-IFS. The rate of projected sea-surface temperature (SST) warming
168 was derived from an ensemble mean of CMIP5, with interannual variability derived from the
169 historical period 1950–2014 (Haarsma et al., 2016).

170

171 Under the HighResMIP experimental protocol, minimal changes in model-tuning parameters
172 were made between low- and high-resolution integrations to ensure that resolution-sensitivity
173 studies were not confounded by substantial differences in model configurations between
174 resolutions (Haarsma et al., 2016). Between low- and high-resolution configurations, no
175 model-physics changes were made to the atmospheric components of CNRM-CM6.1 and
176 EC-Earth3P, but minor adjustments were made to a single parameter in ECMWF-IFS (related
177 to net surface energy balance), HadGEM3-GC3.1 (related to quasi-biennial oscillation
178 period), and MPI-ESM1.2 (related to numerical stability). For the ocean model in coupled
179 configurations, one key difference is the effects of mesoscale eddies are parameterised at low
180 resolution ($\sim 1^\circ$) but partially resolved at high resolution ($\sim 0.25^\circ$) (e.g., Roberts et al., 2018;
181 Roberts et al., 2019). For all models, shorter dynamical timesteps were used in the high-
182 resolution integrations to ensure numerical stability. The effective resolutions of the high-

183 resolution model configurations, measured by kinetic energy spectra, resolve synoptic-scale
184 dynamics (Klaver et al., 2020). Since this study concerns cyclone translation from the tropics
185 to the extratropics, resolutions are given as a model's regular mesh spacing at a latitude of 50
186 ° (Table 2). For convenience, we refer to resolutions nominally (i.e., 'low' or 'high') as well
187 as quantitatively, where necessary. A single ensemble member was analysed at each
188 resolution for both the atmosphere-only and fully coupled experiments.
189

Atmospheric model	Ocean model	Atmospheric dynamical core	Resolution nomenclature	Atmospheric resolution	Atmospheric mesh spacing
ARPEGE6.3	NEMO	Spectral (linear, reduced Gaussian)	LR; HR	TL127; TL359	142; 50 km
IFS cyc36r4	NEMO	Spectral (linear, reduced Gaussian)	LR; HR	TL255; TL511	71; 36 km
IFS cyc43r1	NEMO3.4	Spectral (cubic octahedral; reduced Gaussian)	LR; HR	Tco199; Tco399	50; 25 km
MetUM	NEMO	Grid point (SISL)	LM (LL); MM; HM (HH)	N96; N216; N512	135; 60; 25 km
ECHAM6.3	MPIOM1.63	Spectral (triangular; Gaussian)	HR; XR	T127; T255	67; 34 km

Model name	CNRM-CM6.1	EC-Earth3P	ECMWF-IFS	HadGEM3-GC3.1	MPI-ESM1.2
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190

191 **Table 2.** The PRIMAVERA (HighResMIP) model ensemble. NEMO: Nucleus for European
 192 Modelling of the Ocean. MPIOM: Max Plank Institute Ocean Model. SISL: semi-implicit,
 193 semi-Lagrangian. For fully coupled simulations, the LL and HH configurations of
 194 HadGEM3-GC3.1 were also included; LL denoting low-resolution atmosphere and low-
 195 resolution (1 °) ocean and HH denoting high-resolution atmosphere and high-resolution (1/12
 196 °) ocean. Atmosphere mesh spacing is given for 50 °N. Sample sizes for all tropical cyclones
 197 undergoing ET across this ensemble are given in Table 3. DOIs for each simulation are listed
 198 at primavera-h2020.eu/modelling/.

199

200 *2.3 Lagrangian tropical-cyclone tracking*

201 To identify and track the evolution of tropical cyclones, we used the objective feature-
202 tracking algorithm—TRACK—of Hodges (1995), a well-established tool for identifying
203 cyclones in reanalyses (Hodges et al., 2017) and model simulations (Roberts et al., 2020a).
204 The TRACK algorithm was applied to six-hourly relative vorticity, computed from the zonal
205 and meridional wind fields, which was vertically averaged over the 850-, 700- and 600-hPa
206 levels and spectrally filtered. (Upper-level vorticity is used in subsequent identification.)
207 Filtering to the T5–T63 spectral band removes both large, planetary scales (total
208 wavenumbers 0–5) and small-scale noise (total wavenumbers >63). Vorticity maxima
209 exceeding $0.5 \times 10^{-5} \text{ s}^{-1}$ (in the Northern Hemisphere; scaled by -1 in the Southern
210 Hemisphere) were identified, initialised into tracks using a nearest-neighbour approach, and
211 subsequently refined by minimising a cost function for track smoothness, subject to adaptive
212 constraints on track displacement and smoothness (Hodges, 1995, 1999). The use of
213 vertically averaged vorticity improves temporal coherence in instances where vorticity
214 maxima shift between levels (Hodges et al., 2017).

215

216 Cyclone-centred sampling of meteorological fields along cyclone tracks was performed to
217 detect warm-core structures and measure cyclone intensities, following Hodges et al. (2017).
218 For warm-core identification, the T63-truncated vorticity data on seven levels covering 850–
219 250 hPa were added to tracks by recursively searching for a vorticity maximum at each level
220 using the previous level’s maximum as the starting point for a steepest-ascent maximization
221 applied to the B-spline-interpolated field. A search radius of 5° was used, centred on each
222 level’s maximum. For the Southern Hemisphere, fields were scaled by -1 . To quantify
223 cyclone intensity, mean sea-level pressure minima within a radius of 5° and 925-hPa and 10-
224 metre wind speed maxima within a radius of 6° of the storm centre were sampled from
225 reanalysis or model-output fields at their native, non-truncated resolutions. (All radii are
226 geodesic.)

227

228 Following Hodges et al. (2017), objective identification of tropical cyclones adhered to the
229 following criteria:

- 230 ▪ cyclogenesis equatorward of 30°N
- 231 ▪ total cyclone lifetime must exceed two days
- 232 ▪ T63 relative vorticity at 850 hPa must exceed $6 \times 10^{-5} \text{ s}^{-1}$

- 233 ▪ T63 relative vorticity centre must exist at each level between 850 and 250 hPa to
234 indicate a coherent vertical structure
- 235 ▪ T63 relative vorticity decrease with increasing height between 850 and 250 hPa by at
236 least $6 \times 10^{-5} \text{ s}^{-1}$ to indicate the presence of a warm core

237 The three T63 relative vorticity criteria must also be jointly attained for at least four
238 consecutive time steps (i.e., one day) over ocean only. Together, these criteria minimise
239 inclusion of spurious short-lived or relatively weak vorticity features. The same criteria were
240 used for each reanalysis and model simulation and across all ocean basins.

241

242 Crucial to our analyses, vorticity-based tracking and post-tracking identification of tropical
243 cyclones yields longer cyclone lifecycles (compared with central-pressure-based algorithms
244 and methodologies where identification is performed during tracking), which allows for
245 objective analysis of post-tropical storm evolution (Hodges et al., 2017). A comparison of
246 TRACK results with results from a different tracking algorithm, which does not capture the
247 full lifecycle, demonstrates this advantage of vorticity-based tracking (section S1.1; Fig. S1).
248 In addition, filtering gridded data to a common spectral truncation, rather than tuning the
249 cyclone-tracking algorithm to a given dataset, allows both inter-model and inter-resolution
250 comparisons that are not complicated by methodological differences (Hodges et al., 2017).
251 Applying TRACK to a reanalysis globally, as described here, identifies ~30,000 tropical
252 vortices per year. Of these, ~8,000 per year have a lifetime that exceeds two days and are
253 retained; of these, ~120 per year exhibit the warm-core structure of a tropical cyclone
254 (Vannière et al., 2020). Our study is based on recently published tropical cyclone track
255 datasets, derived using a consistent methodology (Roberts et al., 2020a; Roberts et al.,
256 2020b). Sample sizes for all tropical cyclones undergoing ET are given in Table 3. Finally,
257 spatial track statistics—track and genesis densities—were computed using spherical kernel
258 estimators, following Hodges (1996).

259

Model name	Atmosphere-only		Fully coupled	
	<i>highresSST-present</i>	<i>highresSST-future</i>	<i>hist-1950</i>	<i>highres-future</i>
CNRM-CM6.1	42.3, 52.0	41.0, 47.5	43.4, 45.8	40.1, 39.4
CNRM-CM6.1-HR	47.9, 55.9	46.7, 51.5	50.0, 49.6	46.8, 42.3
EC-Earth3P	19.2, 29.3	20.1, 28.9	19.9, 27.6	19.1, 24.0
EC-Earth3P-HR	30.1, 32.1	29.1, 29.4	26.6, 28.8	26.8, 27.8
ECMWF-IFS-LR	34.7, 41.6	n/a	29.6, 41.5	n/a
ECMWF-IFS-HR	39.8, 44.6	n/a	34.5, 41.7	n/a
HadGEM3-GC3.1-LL	n/a	n/a	28.4, 38.7	28.6, 36.3
HadGEM3-GC3.1-LM	36.3, 50.0	36.5, 50.7	n/a	n/a
HadGEM3-GC3.1-MM	60.1, 68.8	60.9, 65.0	55.0, 56.0	53.2, 53.4
HadGEM3-GC3.1-HM	63.8, 69.0	63.1, 64.6	58.1, 56.4	58.9, 54.3
HadGEM3-GC3.1-HH	n/a	n/a	63.4, 56.2	60.1, 52.9
MPI-ESM1.2-HR	10.5, 16.0	9.4, 14.5	11.4, 16.9	10.4, 15.5
MPI-ESM1.2-XR	10.1, 17.0	9.6, 15.0	11.1, 17.4	10.1, 14.9

260

261 **Table 3.** Annual-mean global sample sizes (cyclones year⁻¹) for all tropical cyclones
262 undergoing ET in each model simulation, given as n_{NH} , n_{SH} .

263

264 2.4 Cyclone phase-space analysis

265 The temporal evolution of cyclone structure, including identifying ET, is quantifiable by
 266 analysis of a cyclone’s thermal wind fields (Hart, 2003; Hart and Evans, 2001). So-called
 267 cyclone phase-space analysis involves three parameters: the thermal axisymmetry of the
 268 cyclone (B ; Eq. 1) and the lower- (T_L ; Eq. 2) and upper-tropospheric (T_U ; Eq. 3) cyclone-
 269 relative thermal winds. In this study, these parameters were computed using 6-hourly data for
 270 all reanalyses and climate models. B is defined as:

$$271 \quad B = h \left(\overline{Z_{600} - Z_{925}} \Big|_R - \overline{Z_{600} - Z_{925}} \Big|_L \right) \quad (1)$$

272 where $h = 1$ for the Northern Hemisphere and -1 for the Southern Hemisphere, Z_p is
 273 geopotential height (m) at level p (isobaric; hPa), and R and L denote the right- and left-hand
 274 semicircles, respectively, relative to the cyclone’s displacement direction. In this study, we
 275 followed the majority of previous research (Bieli et al., 2019; Bieli et al., 2020; Dekker et al.,
 276 2018; Hart, 2003; Liu et al., 2017; Studholme et al., 2015) and defined thermal axisymmetry
 277 (i.e., non-frontal) as $B < 10$ and asymmetry (i.e., frontal) as $B \geq 10$ m. To compute T_L and T_U
 278 between isobaric surfaces, Hart (2003) used the slope of the linear regression between ΔZ and
 279 $\ln p$ as the derivative of ΔZ relative to $\ln p$ to determine the mean ΔZ over a given pressure
 280 range. However, to ensure consistency between phase-space parameters computed from
 281 reanalyses and model output, and to account for the different pressure levels on which
 282 reanalysis and model data are available, it was necessary to adopt a three-level procedure,
 283 following recent studies (Bieli et al., 2019; Bieli et al., 2020; Liu et al., 2017; Studholme et
 284 al., 2015). Here, T_L (925–600 hPa) and T_U (600–250 hPa) are defined as vertical derivatives
 285 of the horizontal geopotential height gradient:

$$286 \quad T_L \equiv -|V_T^L| = \frac{\partial(\Delta Z)}{\partial \ln p} \Big|_{925 \text{ hPa}}^{600 \text{ hPa}} \quad (2)$$

$$287 \quad T_U \equiv -|V_T^U| = \frac{\partial(\Delta Z)}{\partial \ln p} \Big|_{600 \text{ hPa}}^{250 \text{ hPa}} \quad (3)$$

288 where p is pressure and $\Delta Z = Z_{max} - Z_{min}$, where Z_{max} and Z_{min} are the maximum and minimum
 289 geopotential height, respectively, at a given level within a 5° radius of the cyclone centre.
 290 Positive T_L or T_U indicates the presence of a warm core in the upper or lower troposphere,
 291 respectively; negative values indicate a cold core. A deep warm- or cold-core structure is
 292 identified where T_L and T_U have the same sign. We performed phase-space analysis for all
 293 reanalyses (section 2.1) and all PRIMAVERA models (section 2.2). In our analysis, cyclone
 294 centres in reanalyses and model output are those identified objectively by TRACK, which

295 ensures dynamical consistency between cyclone positions and the geopotential height field.
296 This differs from Bieli et al. (2020), who centred reanalysis geopotential data on best-track
297 storm locations. The approach taken in our study avoids any potential inconsistencies
298 between reanalysis and best-track storm centres, which would need to be accounted for,
299 particularly at weaker intensities (Hodges et al., 2017).

300

301 Among existing studies, various phase-space thresholds have been employed to identify ET
302 (e.g., Bieli et al., 2019; Hart and Evans, 2001; Kofron et al., 2010; Liu et al., 2017; Zarzycki
303 et al., 2017). We defined ET onset as either cold-core development (i.e., $T_L < 0$) or
304 development of thermal asymmetry (i.e., $B \geq 10$), thereby allowing for either ET pathway. ET
305 completion is defined as the first occurrence of both $B \geq 10$ m and $T_L < 0$. These thresholds
306 are suitable for high-resolution gridded data (Michaelis and Lackmann, 2019) and are
307 supported by cluster analysis of observed ET events (Arnott et al., 2004). However, much of
308 the ET-identification literature has focussed on the North Atlantic, yet ET phase-space
309 pathways may differ between ocean basins (Bieli et al., 2019). To account for these
310 difficulties in our global study, ET was identified only where the completion criterion is
311 satisfied for at least four consecutive timesteps (i.e., one day). The use of this additional one-
312 day criterion identifies meaningful temporal changes in B and T_L and avoids counting any
313 spurious, high-frequency temporal variability in phase-space parameters as multiple core-
314 structure changes, following (Baker et al., 2021). An analysis of the sensitivity of ET location
315 to methodological choices is presented in section S1.2, showing a large spread in ET location
316 (Fig. S2). In this study, ET-completion latitude was identified after a warm-core structure
317 persisted for at least 2 days based on phase-space parameters (i.e., $T_L > 0$ and $T_U > 0$),
318 corresponding to ‘ w ’ in Fig. S2. As such, sample sizes (Table 1 and Table 3) remain
319 unchanged. This method avoids false positives in ET identification arising from tropical
320 depressions and other weak, precursor systems (Bieli et al., 2020), and is therefore more
321 appropriate to analysis of ET location (see section S1.2 for details).

322

323

324 *2.5. Identifying post-ET reintensification*

325 Instances of post-ET reintensification were defined as a post-ET increase in p_{min} of at least -4
326 hPa, a threshold that is based on published case studies (e.g., Zhu et al., 2018), but the
327 number of identified reintensification events is necessarily sensitive to this threshold. For
328 consistency, we applied a single threshold across all reanalyses and models; a higher

329 threshold will likely be appropriate for any future analysis of higher-resolution (i.e.,
330 convection-permitting) models. We used p_{min} to avoid any complications arising from inter-
331 model differences in how near-surface wind speeds are computed (e.g., related to surface
332 roughness).

333

334

335 *2.6 Eady growth rate*

336 Eady growth rate maxima (Eq. 4) were computed as (Hoskins and Valdes, 1990):

$$337 \sigma_{max} = 0.31 \frac{f}{N} \frac{\partial(u,v)}{\partial Z} \quad (4)$$

338 where f is the Coriolis parameter, N is the static stability parameter, Z is geopotential height,
339 and u and v are the zonal and meridional winds, respectively, which were used to compute
340 the magnitude of the horizontal wind (i.e., $\sqrt{u^2 + v^2}$). The vertical derivatives, $\partial(u,v)$ and ∂Z ,
341 were computed between the 850- and 250-hPa levels using 6-hourly data.

342

343

344 **3. Results**

345 In each of the following sections, we present historical results and model evaluation followed
346 by analysis of projected future changes out to 2050.

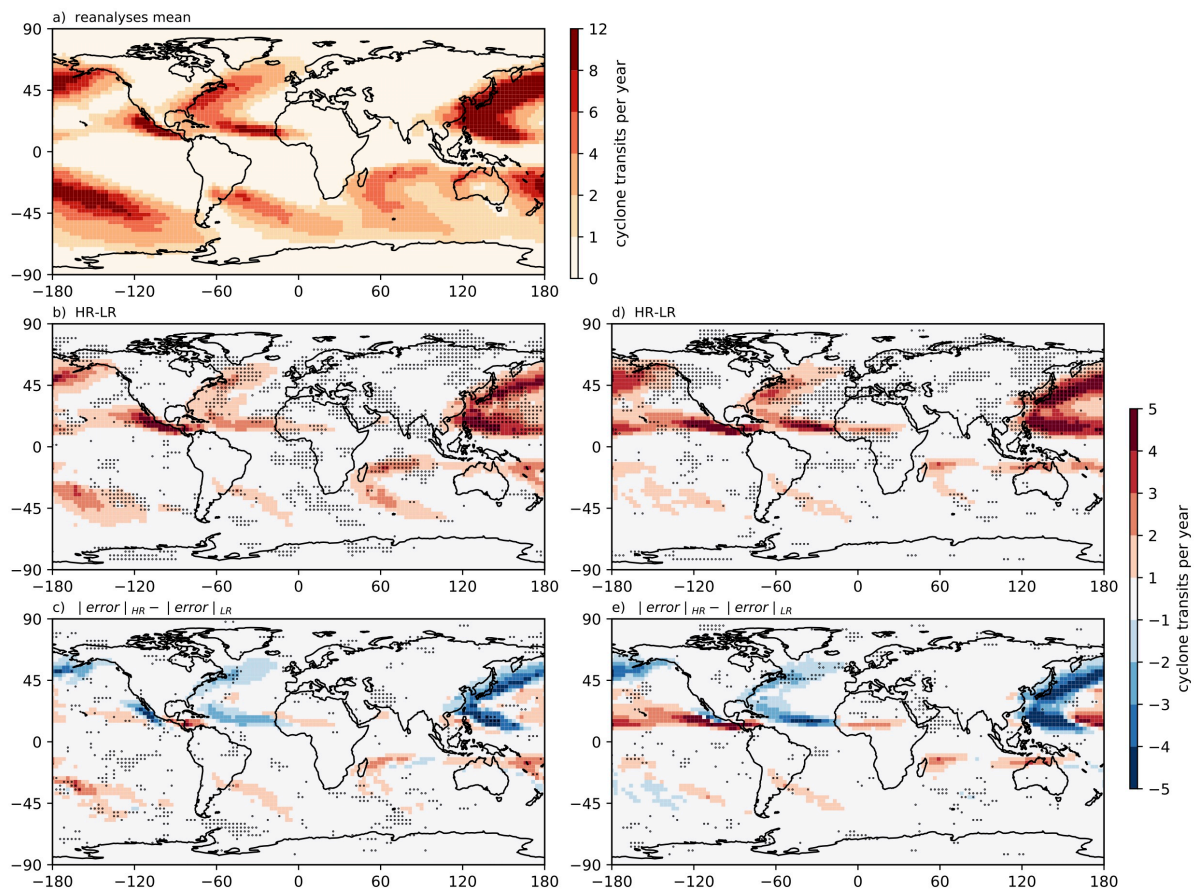
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348

349 *3.1 Spatial cyclone statistics*

350 We first present spatial track density patterns for tropical cyclones undergoing ET in
351 reanalyses and simulated across the PRIMAVERA ensemble. Reanalyses exhibit a high
352 degree of consistency for track density and demonstrate that tropical cyclones undergo ET in
353 all ocean basins. However, fewer ET events are identified over the Northern Indian Ocean
354 (Fig. 1a), where relatively low-latitude landfall either disrupts liminal ET events or averts
355 potential ET cases altogether, primarily via boundary-layer frictional effects (Bieli et al.,
356 2019). Overall, basins' climatological ET activity is proportional to their tropical cyclone
357 activity. The highest ET frequencies are identified in both the Western North Pacific and
358 South Pacific basins, with climatological mean values of ~ 12 cyclones year⁻¹. The North
359 Atlantic is the most active basin for ET outside the Pacific, and comparably low activity
360 occurs across the South Atlantic and South Indian basins (Fig. 1a).

361



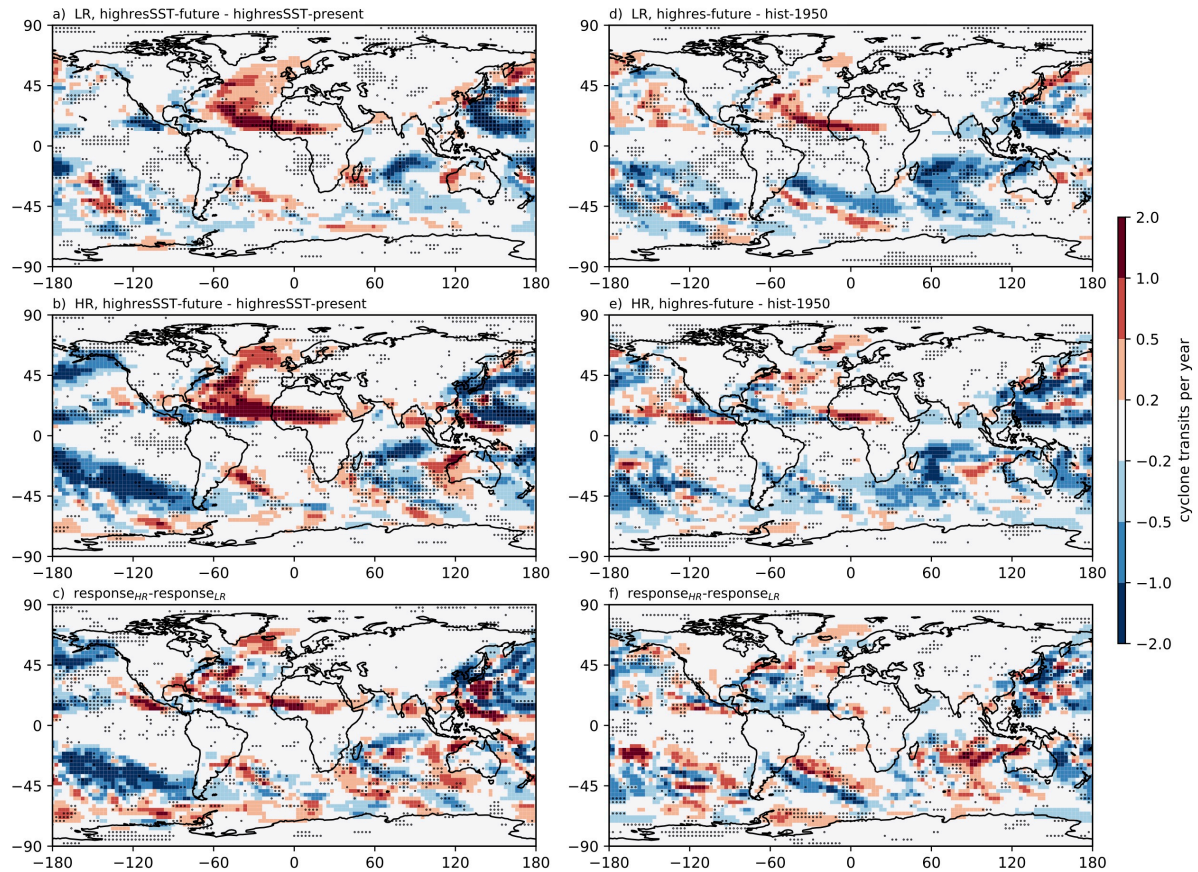
362

363 **Fig. 1.** Cyclone track density for all tropical cyclones undergoing ET. (a) Multireanalysis
 364 mean, (b–c) *highresSST-present* and (d–e) *hist-1950*. Track density was computed from
 365 complete tracks, including precursor stages, and is shown in units of cyclone transits per year
 366 per unit area (within a 5° geodesic radius of storm centres). All available reanalysis years
 367 (Table 1) are included in this analysis. (b, d) HR–LR denotes the ensemble-mean difference
 368 between high and low resolution. (c, e) $|error|_{HR} - |error|_{LR}$ denotes the ensemble-mean
 369 difference of the absolute error (model versus multireanalysis mean) between high and low
 370 resolution. The low-resolution (‘LR’) sub-ensemble includes CNRM-CM6.1-LR, EC-
 371 Earth3P-LR, ECMWF-IFS-LR, HadGEM3-GC3.1-LM(-LL), and MPI-ESM1.2-HR. The
 372 high-resolution (‘HR’) sub-ensemble includes CNRM-CM6.1-HR, EC-Earth3P-HR,
 373 ECMWF-IFS-HR, HadGEM3-GC3.1-HM(-HH), and MPI-ESM1.2-XR. In b)–e), stippling
 374 indicates where all five models agree on the sign of the difference.

375

376 The frequency of ET events simulated by PRIMAVERA models increases when resolution is
377 increased from ~ 100 km to ~ 25 km in all basins, both in the *highresSST-present* (Fig. 1b) and
378 *hist-1950* (Fig. 1d) experiments. Ensemble-mean climatologies are similar between both
379 experiments (Fig. S3). The North Atlantic and Western North Pacific basins are regions of
380 relatively widespread inter-model agreement on the sign of this resolution-sensitivity in track
381 density, again regardless of whether SST is prescribed. When prescribed, inter-model
382 agreement is also identified in the South Pacific and South Indian basins (Fig. 1b). This result
383 is consistent with a recent equivalent analysis of all tropical cyclones in PRIMAVERA
384 simulations (Roberts et al., 2020a), where increased frequencies were simulated at higher
385 model resolution across all ocean basins, for which the leading explanation is that finer
386 atmospheric resolution increases the conversion rate of precursor vortices (or ‘seeds’) to
387 tropical cyclones (Roberts et al., 2020a; Vecchi et al., 2019; Vidale et al., 2021). Tropical-
388 cyclone intensities simulated at model resolutions in the range 50–20 km are more
389 comparable with observational estimates (Roberts et al., 2020a), due in part to enhanced
390 surface latent heat flux (Vannière et al., 2020), implying that a more realistic proportion may
391 withstand midlatitude environmental conditions hostile to tropical cyclones prior to and
392 during the initial stages of ET. At low resolutions (typically ~ 100 km), PRIMAVERA models
393 simulate too few ET systems compared with reanalyses, particularly across the North Atlantic
394 and Western North Pacific, in both the *highresSST-present* (Fig. S4a) and *hist-1950* (Fig.
395 S4c) experiments. Increasing resolution to ~ 25 km leads to increased track density globally,
396 reducing negative biases in these basins but engendering positive biases in the Eastern North
397 Pacific and South Pacific (Fig. S4c, d). In *hist-1950*, this bias reduction is consistent with a
398 reduction in negative surface temperature biases at high resolution (e.g., ~ 1 °K reduction in
399 the North Atlantic; Moreno-Chamarro et al., 2022). In section 3.2, we examine ET frequency
400 and the percentage of tropical cyclones undergoing ET separately.

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Fig. 2. Climate-change response of track density for all cyclones undergoing ET. (a–c) *highresSST-future* minus *highresSST-present* and (d–f) *highres-future* minus *hist-1950*. Track density was computed from complete tracks, including precursor stages, and is shown in units of cyclone transits per year per unit area (within a 5° geodesic radius of storm centres). The low-resolution (‘LR’) sub-ensemble includes CNRM-CM6.1-LR, EC-Earth3P-LR, HadGEM3-GC3.1-LM(-LL), and MPI-ESM1.2-HR. The high-resolution (‘HR’) sub-ensemble includes CNRM-CM6.1-HR, EC-Earth3P-HR, HadGEM3-GC3.1-HM(-HH), and MPI-ESM1.2-XR. Stippling indicates where all models agree on the sign of the difference.

412 Overall, PRIMAVERA simulations indicate that increasing resolution improves the
413 representation of ET frequency, as measured by track density, particularly across the North
414 Atlantic and Western North Pacific (Fig. 1c, e). For these basins, reductions in ensemble-
415 mean absolute biases are found in both *highresSST-present* and *hist-1950*, and areas of bias
416 reduction across multiple models occur primarily over western boundary currents—the Gulf
417 Stream and Kuroshio, respectively. That these regions of resolution-dependence and reduced
418 biases overlap indicates that capturing the sharpness of SST fronts and associated
419 baroclinicity is important in simulating ET (Evans et al., 2017; Klein et al., 2002), and,
420 consistent with this, we find enhanced meridional SST gradients in both of these boundary-
421 current regions (Fig. S5). In the Southern Hemisphere, little difference in ensemble-mean
422 biases is found between resolutions, with a caveat that observational or reanalysis-based
423 climatologies for the Southern Ocean are themselves more uncertain (Hodges et al., 2017).
424 The PRIMAVERA ensemble provides evidence that atmospheric resolutions typical of
425 CMIP6 are too coarse to adequately capture basin-mean tropical-cyclone (Roberts et al.,
426 2020a) and ET statistics (this study). Increasing resolution to ~25 km partly addresses this
427 shortcoming.

428

429 The climate-change response of track density for tropical cyclones undergoing ET in high-
430 resolution simulations is basin-dependent, with differences between atmosphere-only and
431 fully coupled simulations also apparent. In *highresSST-future*, increased track density is
432 simulated across the North and South Atlantic (but decreased over the eastern United States)
433 and over the Maritime Continent; decreases are simulated over the Eastern and Western
434 North Pacific and South Indian basins; and an unclear, mixed response characterises the
435 North Indian Ocean (Fig. 2a–b). Inter-model agreement about the sign of these changes is
436 largely confined to cyclogenesis regions (e.g., equatorial West Africa) and over the Gulf
437 Stream and Kuroshio Current. In *highres-future* simulations, positive climate-change
438 responses are confined to the central and Eastern North Pacific. The spatial response pattern
439 over the North Atlantic—increased over central and eastern North Atlantic and decreased
440 along the United States’ east coast—is similar between *highresSST-future* and *highres-future*,
441 but the magnitude of the response is reduced in the fully coupled simulations (Fig. 2d–f).
442 This spatial pattern is supported by recent projections, with increases particularly apparent in
443 the eastern North Atlantic (Liu et al., 2017), consistent with the projected eastward and
444 poleward expansion of cyclogenesis within this basin (Haarsma et al., 2013).

445

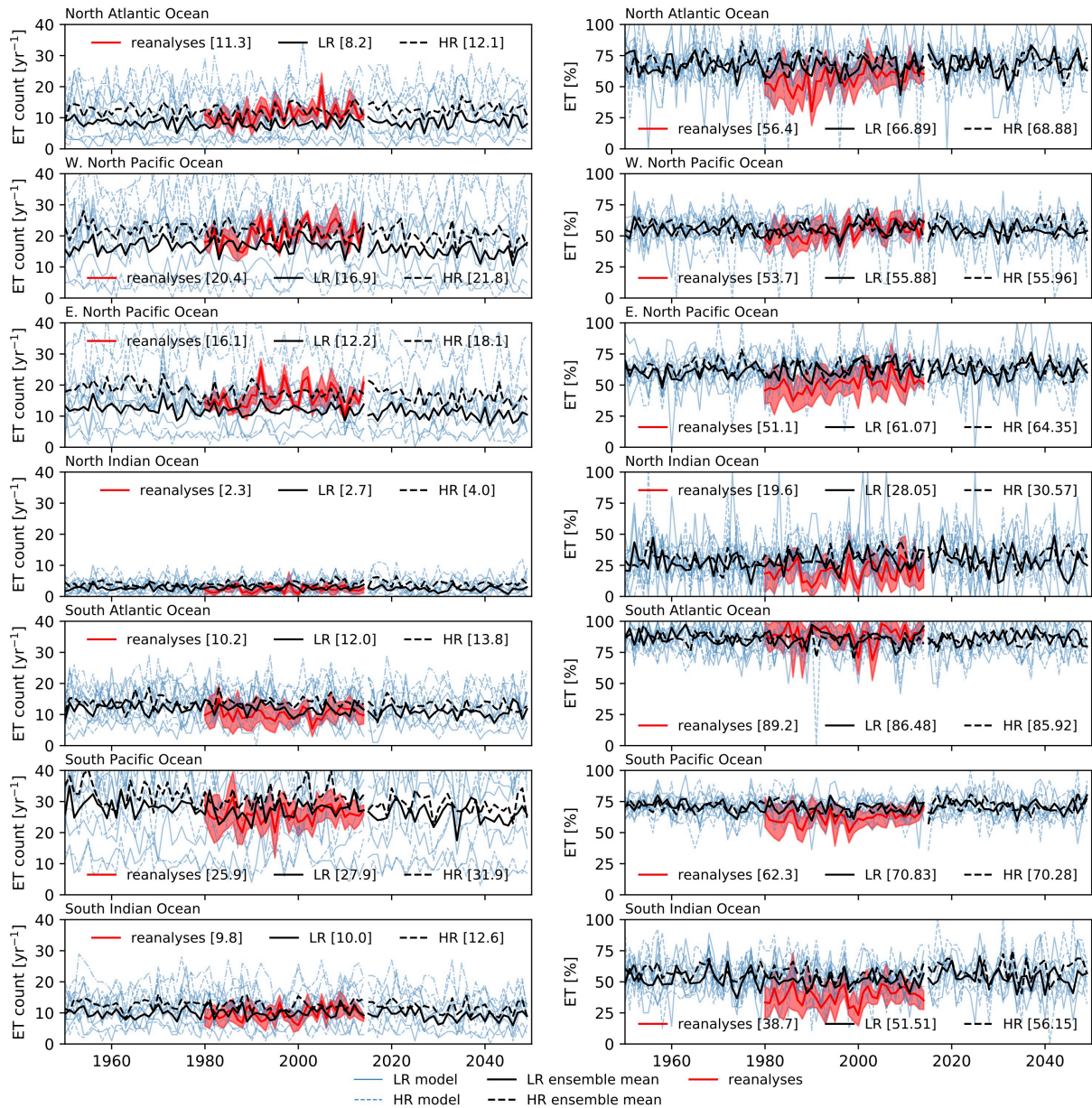
446 Increasing horizontal resolution has a localised effect on the climate-change response of track
447 density for ET (Fig. 2c, f). In *highresSST-future*, resolution-sensitive responses to climate
448 change, which are common across *all* models, are seen only over the central North Atlantic
449 and parts of the Southern Ocean. In *highres-future*, spatially coherent and resolution-sensitive
450 responses to climate change are seen over the South Atlantic and Eastern North Pacific
451 basins, where simulated track density maxima are shifted equatorward at high resolution.
452 However, the spatial patterns of resolution sensitivity over the North Atlantic and Western
453 North Pacific broadly resemble the spatial climate-change response patterns, which indicates
454 that these responses are enhanced at high resolution in most models. This is seen more clearly
455 in the atmosphere-only experiment (Fig. 2c) than in the fully coupled experiment (Fig. 2f).

456
457

458 *3.2 Interannual variability in ET*

459 Over the period 1979–2018, high-resolution *highresSST-present* simulations reproduce the
460 multireanalysis-mean climatological ET counts for Northern Hemisphere basins (Fig. 3, left),
461 except for the Northern Indian Ocean, a basin where few ET events occur. However, little
462 improvement with increased resolution is seen for Southern Hemisphere basins (Fig. 3, left).
463 Again, uncertainty is higher across the Southern Ocean, with greater inter-reanalysis spread
464 seen for Southern Hemisphere basins. These results are also true of the *hist-1950* simulations
465 (Fig. 4, left). The *highresSST-present* simulations appear to capture decadal variability in the
466 role of SST in sustaining tropical cyclones to ET. In certain basins, periods are apparent
467 where the *highresSST-present* ensemble mean and multireanalysis mean ET count match
468 well: e.g., 1985–2000 for the North Atlantic and 1990–2005 for the Western North Pacific
469 (Fig. 3, left). These periods coincide with observed positive phases in Atlantic Multidecadal
470 Variability and Pacific Decadal Oscillation, respectively. For ET %, differences between
471 low- and high-resolution ensemble means are small for most basins (Fig. 3, right). This
472 suggests that the large-scale environmental conditions conducive to ET are not substantially
473 different across the range of model resolutions considered here. This indicates that increased
474 ET frequency at high resolution is driven primarily by increased tropical cyclone frequency,
475 not by an increase in ET %. Similar mean values and variance in ensemble-mean ET count
476 and ET % are simulated in both *highresSST-present* (Fig. 3) and *hist-1950* (Fig. 4)
477 experiments.

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Fig. 3. Interannual variability in (left) the number of ET events and (right) the percentage of tropical cyclones undergoing ET in each ocean basin in reanalyses and simulated in *highresSST-present* and *highresSST-future* experiments. Shown are (red) the multireanalysis mean, with 1 standard deviation of the reanalysis spread indicated by red shading, and (solid black) low- and (dashed black) high-resolution ensemble means. Each panel's legend gives climatological-mean values of (left) ET count or (right) ET % for the reanalyses and historical simulations. Also shown are (blue) timeseries for individual simulations to indicate the ensemble spread for each basin.

489 In *highresSST-present*, models' skill in reproducing the multireanalysis-mean interannual
490 variability in ET count varies between basins (Table 4). Interannual variability in ensemble-
491 mean and multireanalysis-mean ET counts are significantly, positively correlated for three
492 basins at low resolution and four basins at high resolution. The North Atlantic and Western
493 North Pacific basins are significantly correlated at both resolutions; the South Atlantic and
494 South Pacific basins are significantly correlated only at high resolution; and the Eastern
495 North Pacific is significant only at low resolution. Only for the North and South Indian basins
496 is ensemble-mean variability uncorrelated with reanalyses at either resolution. (Correlation
497 coefficients for *hist-1950* simulations are not shown because it is not expected that fully
498 coupled models' internal year-to-year variability would mimic that of forced simulations or
499 reanalyses.) For ET %, fewer significant correlations are found between ensemble-mean and
500 multireanalysis-mean timeseries (Table 4). Positive correlations are seen in the Northern and
501 Southern Indian basins and in the South Pacific basin at high resolution. However, low- and
502 high-resolution ensemble-mean ET % timeseries covary in most basins in both *highresSST-*
503 *present* (Fig. 3) and *hist-1950* (Fig. 4), more so than for ET count. To explain this, we
504 hypothesise that the large-scale environment conducive to the baroclinic conversion of
505 tropical cyclones is less sensitive to model resolution, while ET count depends on tropical
506 cyclone count, which is sensitive to model resolution (Roberts et al., 2020a).

507

508 Recent analysis of an ensemble of HadGEM3-GC3.1 simulations, performed under
509 HighResMIP, demonstrated that mean skill in representing interannual variability in tropical
510 cyclone count improves with additional members (Roberts et al., 2020a). At present, the
511 required six-hourly geopotential outputs are available for too few ensemble members to
512 repeat such an analysis for tropical cyclones undergoing ET, but this would constitute
513 valuable future work when sufficient model output is obtainable. Nonetheless, quantifying
514 the level of skill that exists in capturing interannual variability in the subset of tropical
515 cyclones that undergo ET, while lower than that for all tropical cyclones, is important,
516 establishing the baseline for HighResMIP-class models. This prompts further examination of
517 ET seasonality in the historical and future atmosphere-only simulations, which is possible in
518 the continuous PRIMAVERA simulations.

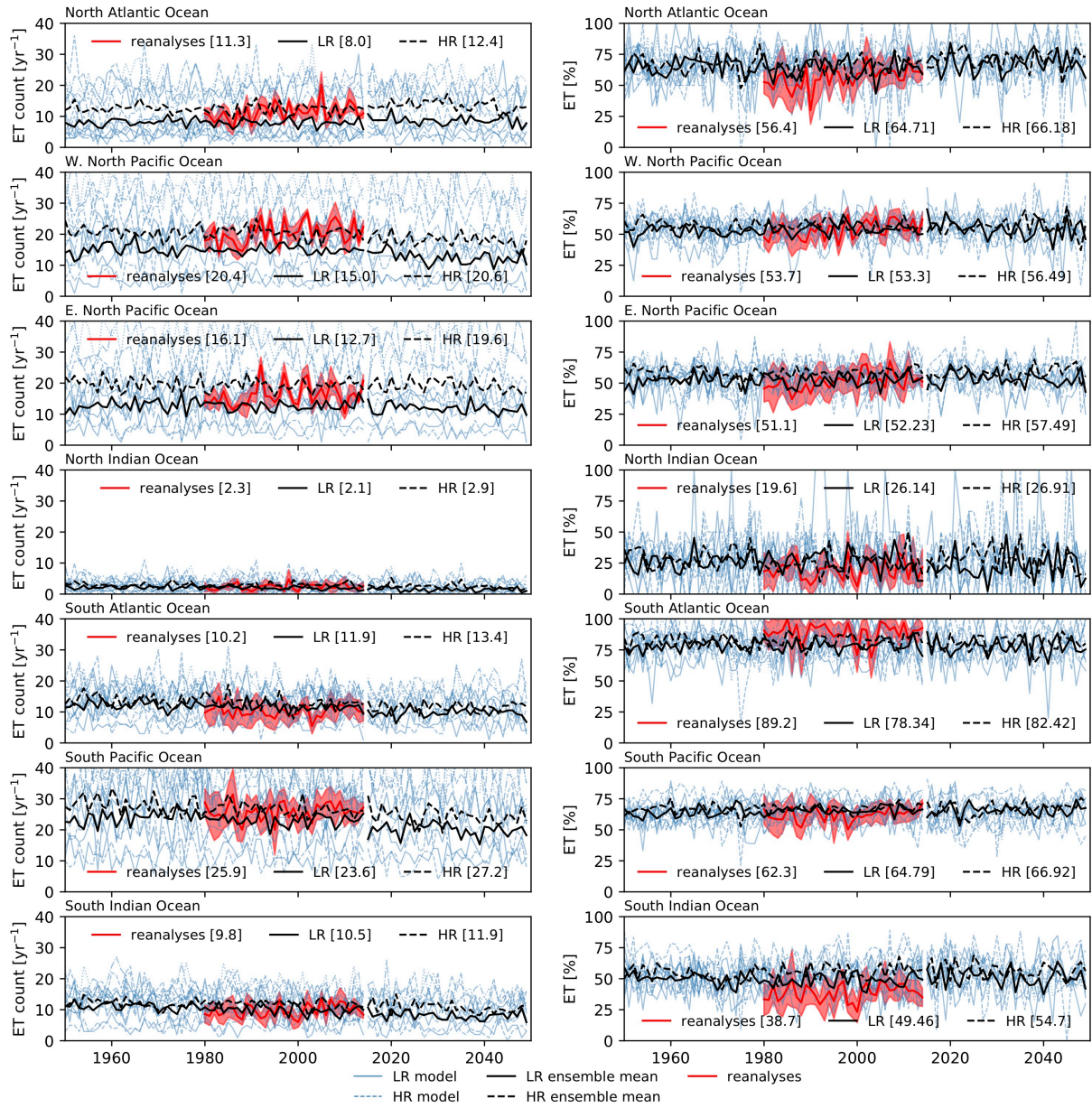
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Ocean basin	ET count		ET %	
	LR	HR	LR	HR
North Atlantic	0.31	0.30	0.24	-0.16
Western North Pacific	0.50	0.34	0.21	0.24
Eastern North Pacific	0.43	0.22	0.42	0.16
North Indian	-0.08	0.03	0.03	0.38
South Atlantic	0.07	0.34	0.12	0.27
South Pacific	0.08	0.50	0.17	0.34
South Indian	-0.04	-0.19	0.24	0.33

520

521 **Table 4.** Pearson’s r coefficients for correlations between low- (LR) or high-resolution (HR)
522 ensemble-mean and multireanalysis-mean interannual variability in ET count and ET % for
523 each ocean basin. Coefficients are shown only for *highresSST-present*; *hist-1950* simulations
524 are not shown because it is not expected that coupled models’ internal year-to-year variability
525 would mimic that of forced simulations or reanalyses. Significant ($p < 0.1$) correlations are in
526 bold type.

527



528

529 **Fig. 4.** As in Fig. 3 for fully coupled *hist-1950* and *highres-future* simulations.

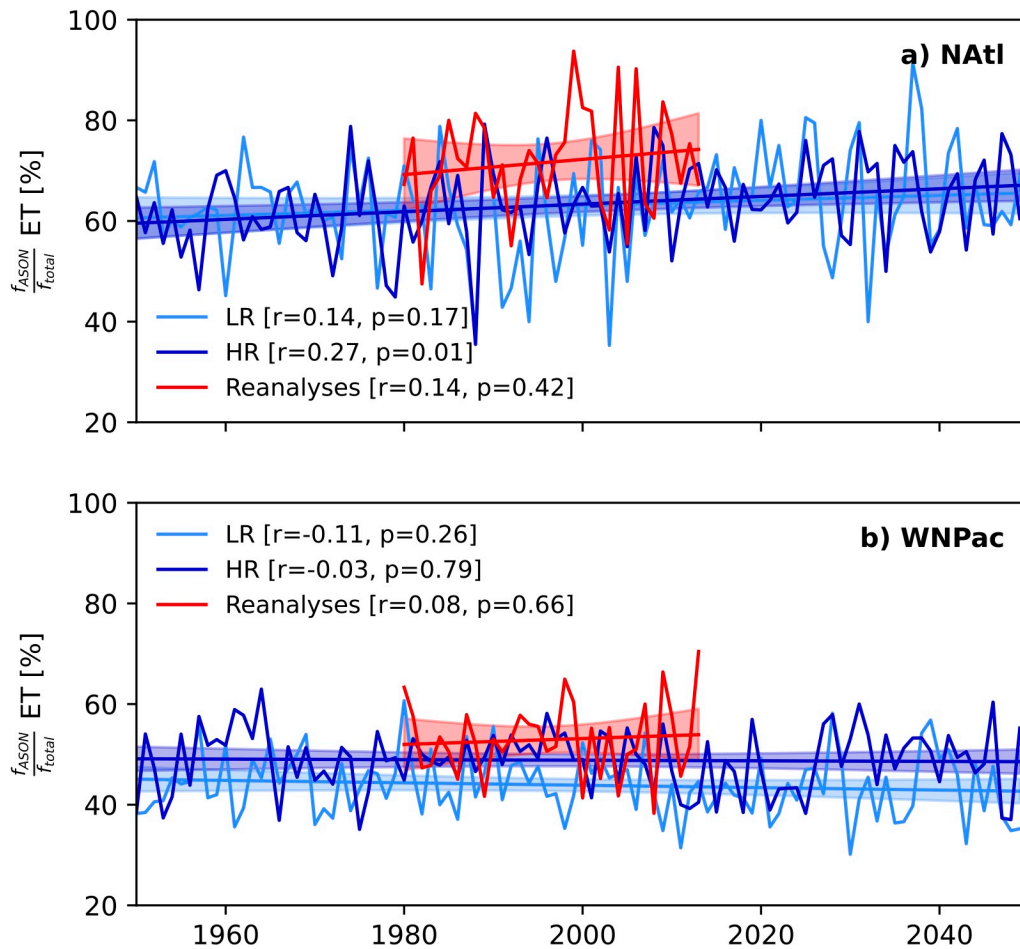
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531 *3.3 Historical and future ET seasonality*

532 We next evaluate the seasonal cycle of ET, focussing on the North Atlantic and Western
533 North Pacific basins for which both climatological ET statistics (Fig. 1) and interannual ET
534 variability (Table 4) are represented reasonably across models. In the North Atlantic,
535 reanalyses show ET % increasing from July to a peak in September before declining into
536 winter (Fig. S6a). In the *highresSST-present* experiment, most models reproduce this
537 seasonality, but the magnitude of the seasonal peak is overestimated by ~10 % at high-
538 resolution. There are indications that increased atmospheric resolution improves the
539 simulation of the timing of the seasonal ET % peak. Two models—CNRM-CM6.1 and EC-
540 Earth3P—simulate the seasonal peak too early (in August) at low resolution but simulate a
541 later peak (in September) at high resolution. Additionally, MPI-ESM1.2, the lowest-
542 resolution model in this ensemble, simulates comparably muted seasonality that also peaks
543 earlier than reanalyses at both resolutions. In the fully coupled *hist-1950* experiment, models
544 reproduce the multireanalysis-mean seasonal cycle, but HadGEM3-GC3.1 and CNRM-
545 CM6.1 simulate a broader seasonal distribution compared with reanalyses (Fig. S7a). In the
546 Western North Pacific, reanalyses show bimodal seasonality, with peaks in ET % in May and
547 September (Fig. S6b). Excepting the MPI-ESM1.2 model, which does not capture
548 bimodality, *highresSST-present* simulations also exhibit two seasonal peaks, but each occurs
549 one to two months later than in reanalyses in both low- and high-resolution integrations (Fig.
550 S6b), and this also holds true for *hist-1950* simulations (Fig. S7b).

551

552 To assess any potential future change in seasonality, ΔET %, we differenced the historical
553 and future seasonal cycles. For the North Atlantic, despite pronounced inter-model spread
554 throughout most of the annual cycle, there is an indication of more consistent model
555 behaviour during August–November, months for which most models simulate an increase in
556 ET % in both the *highresSST-future* (Fig. S6c) and *highres-future* experiments (Fig. S7c). To
557 quantify the degree to which this inter-model consistency represents secular change in ET
558 seasonality, the annual fraction of total annual ET events occurring during August–November
559 was computed. A significant, positive trend in this quantity over the period 1950–2050 is
560 found in the ensemble mean of high-resolution atmosphere-only simulations (Fig. 5a), but the
561 trend is not significant in reanalyses, which likely cover too short a period (1980–) to assess
562 secular change, and is significant in the low-resolution ensemble mean only at the 80 % level.
563 In fully coupled simulations, no significant trends are seen (Fig. S8a). Conducting a similar
564 analysis of the forthcoming extension of ERA5 back to 1950 is warranted, pre-satellite
565 observational uncertainty notwithstanding. For the Western North Pacific, the inter-model
566 spread during the annual cycle of ΔET % is similar between *highresSST-future* (Fig. S6d) and
567 *highres-future* simulations (Fig. S7d) and, in contrast to the North Atlantic, no significant
568 secular change in ET seasonality is found in either reanalyses or in PRIMAVERA
569 simulations out to 2050 (Fig. 5b and Fig. S8b). However, together with projected changes in
570 track density (Fig. 2a–b, d–e), these results provide further evidence that the future response
571 of ET to climate change across the North Atlantic differs from that of the Western North
572 Pacific and of other ocean basins. Therefore, we next investigate the role of cyclone structure
573 in explaining these distinct North Atlantic and Western North Pacific responses.
574



575

576 **Fig. 5.** Secular change in the proportion of ET events occurring during August–November in
 577 reanalyses (red) and low- (pale blue) and high-resolution (dark blue) atmosphere-only
 578 simulations (ensemble mean) for the (a) North Atlantic and (b) Western North Pacific basins.
 579 Shading shows the 95 % confidence interval for the linear fit. ECMWF-IFS is not included in
 580 this analysis because no future simulations were performed in HighResMIP for this model.

581

582 3.4 Response of cyclone structures to climate change

583 To examine the response of cyclone core structure to climate change, we computed
584 ensemble-mean bivariate frequency distributions of phase-space parameters, B , T_L , and T_U in
585 the high-resolution simulations. The T_L – B distribution exhibits a similar general structure in
586 the *highresSST-present* and *-future* experiments for both the North Atlantic (Fig. 6a–b) and
587 Western North Pacific (Fig. 6d–e) basins. This is also true for T_L – T_U distributions (Fig. 7a–b
588 and Fig. 7d–e). Generally, tropical cyclones undergoing ET occupy the lower-right
589 (symmetric, warm core) and upper-left (asymmetric, cold core) quadrants, with fewer
590 instances in either hybrid (transitional) quadrant. The phase-space parameter distributions
591 simulated across PRIMAVERA models are consistent with previous studies (Hart et al.,
592 2006; Michaelis and Lackmann, 2019). Historical ensemble-mean values of B and T_L for the
593 North Atlantic are consistent with recent analysis of observations (Studholme et al., 2015) as
594 well as reanalyses and Community Atmosphere Model simulations at resolutions of 55 and
595 28 km (Zarzycki et al., 2017). Ensemble-mean T_U values are also consistent with these
596 existing studies, except that deep warm-core structures are less frequent in PRIMAVERA
597 models than in recent 15-km-resolution simulations with the Model for Prediction Across
598 Scales–Atmosphere model (Michaelis and Lackmann, 2019), likely due to differences in
599 atmospheric resolution. For the Western North Pacific, model-simulated phase-space
600 parameters are consistent with reanalysis-based values (Kitabatake, 2011). In the fully
601 coupled simulations, T_L – B distributions for both basins are similar to those of the
602 atmosphere-only simulations (Fig. 8c, f), but differences in ensemble-mean T_U values are
603 seen, with warm-core responses to climate change occurring variously throughout the
604 troposphere (Fig. 9c, f).

605

606 Under climate change, models forced by prescribed SST simulate stronger warm-core
607 structures in the North Atlantic, indicated by a shift towards higher T_L for axisymmetric
608 tropical cyclones (Fig. 6c). Moreover, T_L – T_U distributions show that the future shift to
609 stronger warm-core structures is primarily confined to the lower troposphere (Fig. 7c, f).
610 (Here, ‘strong’ refers to ensemble-mean T_L values at the higher end of the historical
611 distributions, in which a range of model-simulated intensities are averaged.) These findings
612 are supported by a recent single-model study (Michaelis and Lackmann, 2019), albeit the
613 ensemble-mean signal we report is less pronounced, and are consistent with increased low-
614 level moisture and the potential for enhanced latent heat release in a warmer climate. Future
615 changes in core structures offer a partly mechanistic explanation of the projected increase in

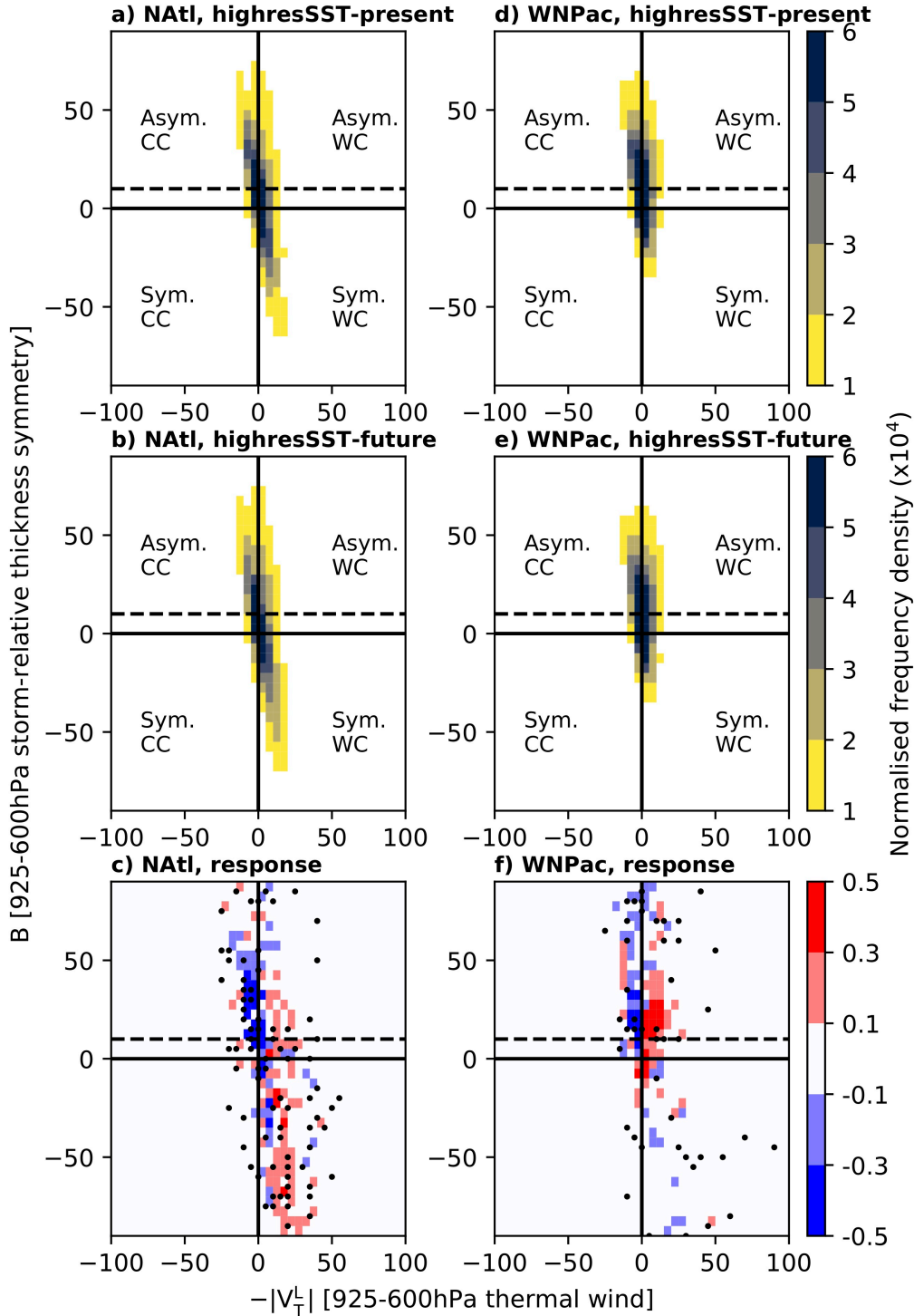
616 ET across the North Atlantic (Baatsen et al., 2015; Haarsma et al., 2013; Liu et al., 2017) as
617 well as the projected change in track density, which is largely unique to the North Atlantic
618 (Fig. 2a–b, d–e). The lesser energy of weak warm-core cyclones is more likely to dissipate
619 before ET may occur, but relatively strong warm-core structures make cyclones more
620 resilient to unfavourable midlatitude environmental conditions (primarily cooler SST and
621 increased vertical wind shear), prolonging their poleward propagation and making ET more
622 probable across the North Atlantic (Hart et al., 2006).

623

624 In the North Pacific, however, this future shift to stronger warm cores is not seen in
625 PRIMAVERA models (Fig. 6f), although more frequent asymmetric, warm-core hybrid
626 structures (upper-right quadrant) in the future are simulated. These instances of hybrid
627 structures show cyclones existing more frequently in the transitional quadrants, potentially
628 indicating a future elongation of ET time (Zarzycki et al., 2017) and an increase in warm-
629 seclusion occurrences, which involve multiple transitions (Baker et al., 2021; Dekker et al.,
630 2018). Also seen is a shift towards stronger upper-level, cold-core structures (Fig. 7f). The
631 Western North Pacific is therefore characterised by more mixed future changes in core-
632 structure frequencies, consistent with the projected response of track density, which generally
633 decreases across the basin but increases in localised areas (Fig. 2b, e). Broadly, these results
634 are also consistent with the lack of any consensus in published projections of ET frequency
635 across the Western North Pacific: both a less favourable future ET environment (Ito et al.,
636 2016) versus moderate future increase in ET frequency (Bieli et al., 2020) have been
637 suggested. For both basins, future phase-space changes in the fully coupled simulations
638 resemble those seen in the atmosphere-only experiments, but the North Atlantic climate-
639 change signal is comparably muted (Fig. 8c, f; Fig. 9c, f).

640

641 Overall, these results help clarify the potential role that the climate-change response of
642 cyclones' core structures have in determining future ET frequency changes, and quantifies
643 how this differs between basins. Differences in pre-ET structures potentially underpin basin-
644 specific responses of ET to climate change, and consistency exists among PRIMAVERA
645 models. However, to fully explain what drives disparate North Atlantic and Western North
646 Pacific responses, further studies of future changes in cyclogenesis and midlatitude large-
647 scale conditions are needed, based on models of higher resolution than those in
648 PRIMAVERA, which better simulate the most intense systems (Judt et al., 2021), and,
649 potentially, their interactions with the large-scale environment.



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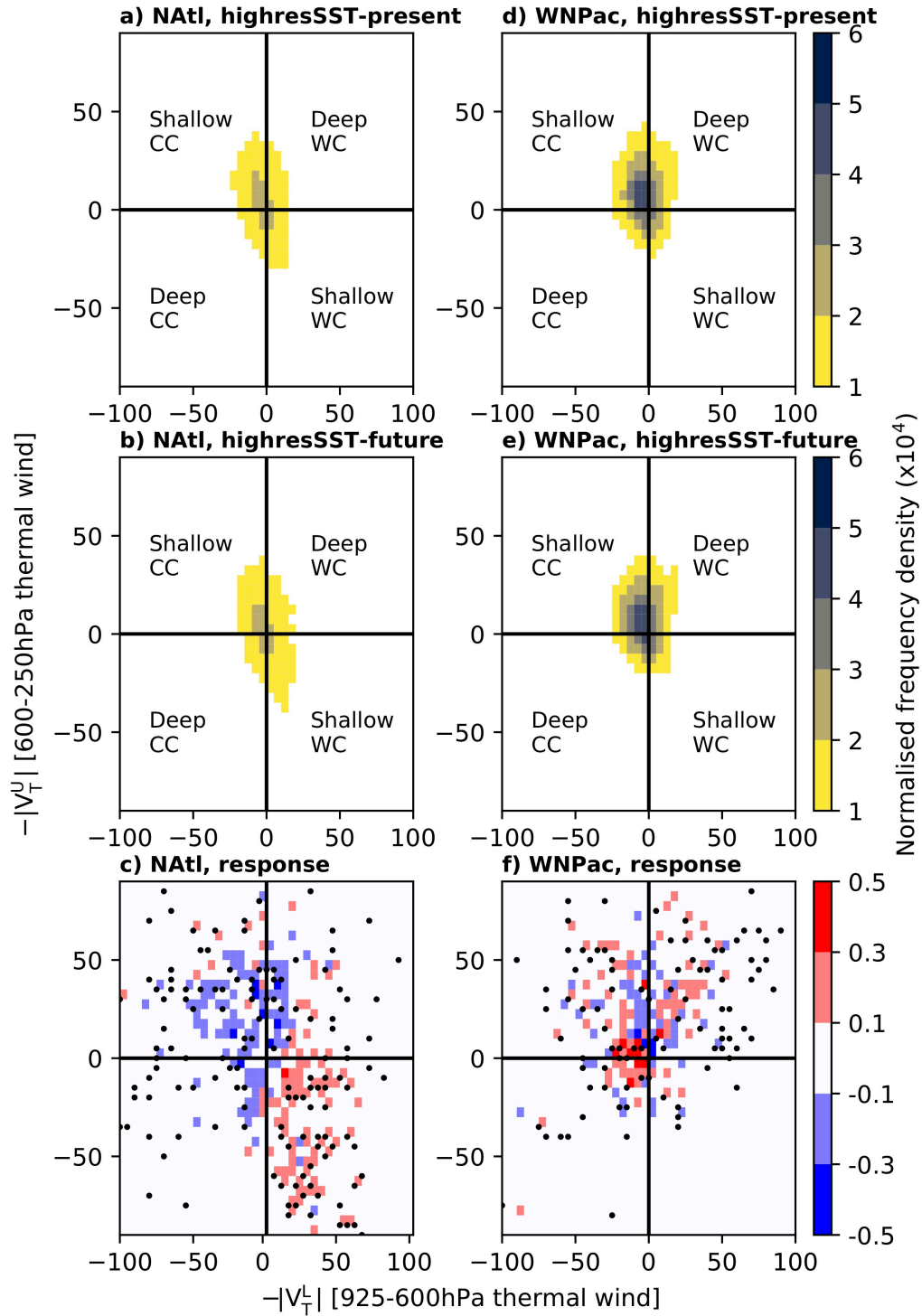
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Fig. 6. Ensemble-mean distributions of T_L versus B in high-resolution (a, d) *highresSST-present* and (b, e) *-future* simulations as well as (c, f) the climate-change response for the North Atlantic ('NAtl') and Western North Pacific ('WNPac'). Distributions are computed from every 6-hourly point during the entire lifetime of all storms undergoing ET, plotted as two-dimensional histograms, and normalised by the total number of cyclones (sample sizes for each model are given in Table 3). Values are scaled by 10^4 . Cyclone phase-space

658 categories are warm- ('WC') or cold-core ('CC') and either symmetrical (i.e., non-frontal;
659 'Sym.') or asymmetrical (i.e., frontal; 'Asym.'). The threshold of 10 m used to distinguish
660 thermally symmetric from asymmetric cyclones is indicated (dashed line). Stippling in c) and
661 f) indicates where all models agree on the sign of the difference.

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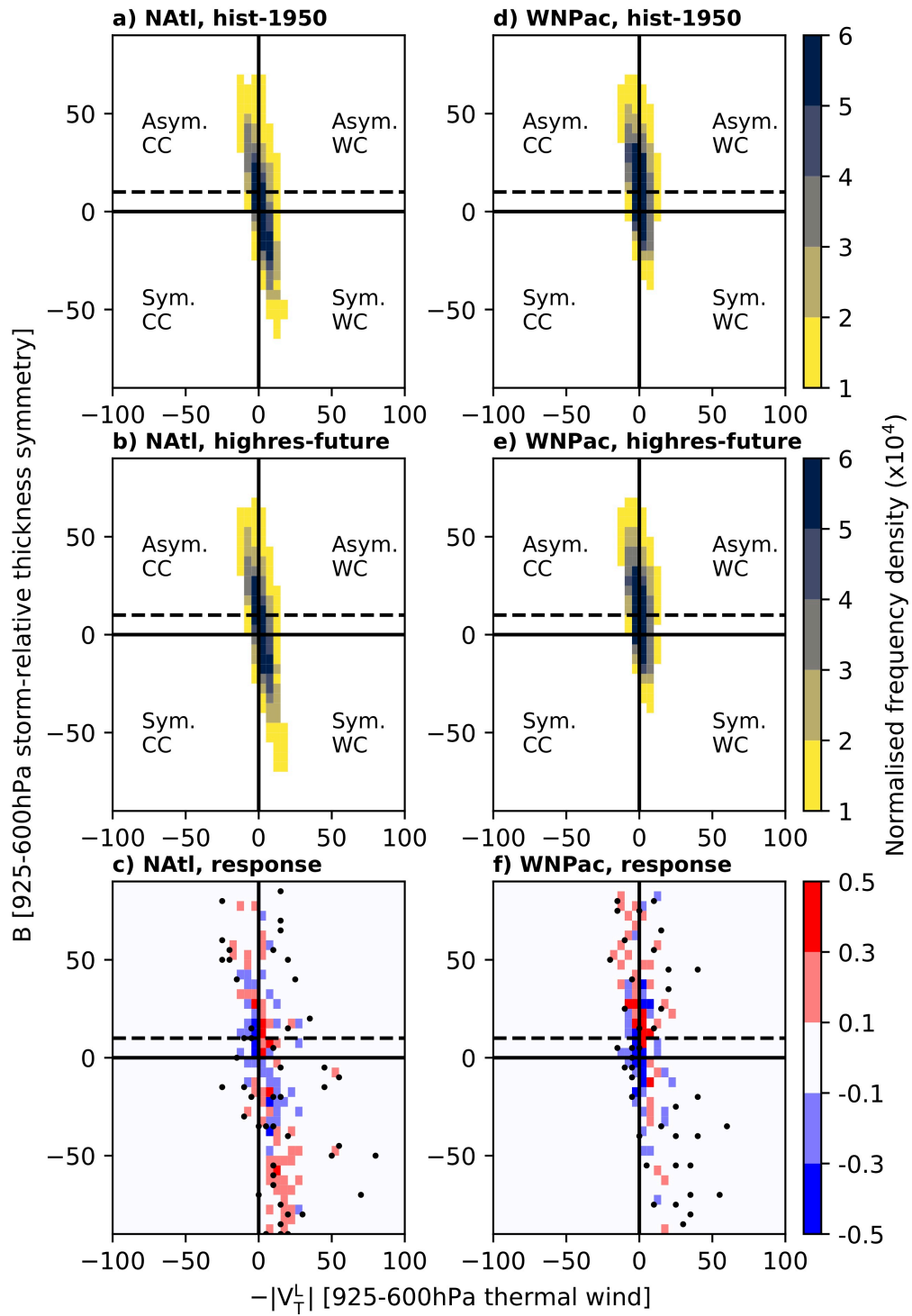
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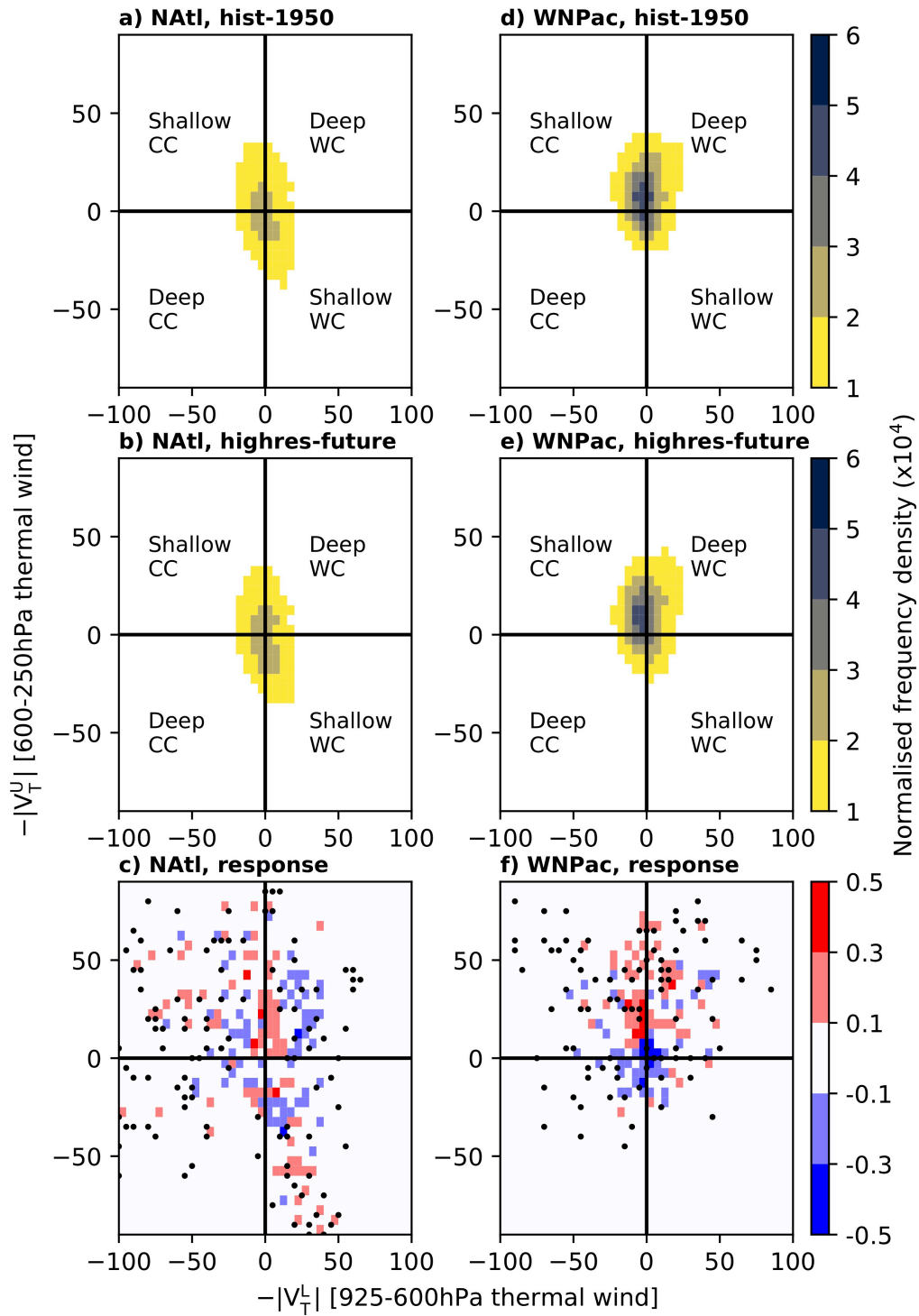
Fig. 7. Ensemble-mean distributions of T_L versus T_U in high-resolution (a, d) *highresSST-present* and (b, e) *-future* simulations as well as (c, f) the climate-change response for the North Atlantic ('NAtl') and Western North Pacific ('WNPac'). Distributions are computed from every 6-hourly point during the entire lifetime of all storms undergoing ET, plotted as two-dimensional histograms, and normalised by the total number of cyclones (sample sizes for each model are given in Table 3). Values are scaled by 10^4 . Cyclone phase-space

670 categories are shallow or deep warm- ('WC') or cold-core ('CC'). Stippling in c) and f)
671 indicates where all models agree on the sign of the difference.
672



673

674 **Fig. 8.** As in Fig. 6 for *hist-1950* and *highres-future* experiments.



675

676 **Fig. 9.** As in Fig. 7 for *hist-1950* and *highres-future* experiments.

677

678 *3.5 Pre- and post-ET cyclone intensity*

679 During ET, cyclones develop low-level frontal structures and their horizontal size increases
680 (Evans et al., 2017). As such, increasing model resolution is expected to impact the
681 simulation of cyclones pre- and post-ET differently, particularly in models whose effective
682 resolutions coarsen equatorward. However, performing a global analysis of the pre- and post-
683 ET stages of tropical cyclones' lifecycles is not trivial because ET pathways (i.e., the order in
684 which B and T_L changes occur) differ between ocean basins (Bieli et al., 2019). We therefore
685 separated cyclone tracks' warm- and cold-core stages about ET completion, when both B and
686 T_L satisfy ET criteria, following the definition first used by Hart (2003). Our additional 1-day
687 criterion (see Methods) helps increase confidence in the following inter-model comparison.

688

689 Compared with best-track intensity estimates, certain atmosphere-only models (particularly
690 CNRM-CM6.1) simulate realistic intensities at resolutions in the range 20–50 km (Roberts et
691 al., 2020a). However, best-track intensity estimates are not well suited to evaluating post-ET
692 systems (Velden et al., 2006), and the available primary cyclone wind-speed observations,
693 such as satellite scatterometry data, seldom include cyclones' post-tropical stages and span
694 too short a temporal range for climatological evaluation. We therefore turn to reanalyses,
695 which are constrained by observational data, to provide a homogeneous global reference. An
696 important caveat, however, is the underestimation of cyclone wind speeds in reanalyses
697 (Hodges et al., 2017; Murakami, 2014), although this underestimation is less marked at
698 higher latitudes (Sainsbury et al., 2020).

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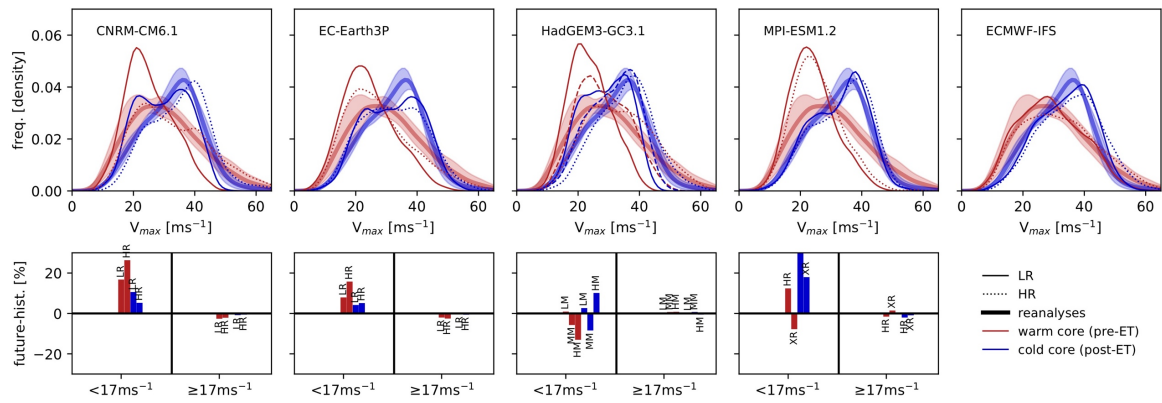
700 Considering all storms globally, PRIMAVERA models reproduce the reanalyses' cold-core,
701 post-ET intensity distributions at both low and high resolution and in both atmosphere-only
702 and fully coupled simulations (Fig. 10 and Fig. 11, top rows). However, models'
703 representation of warm-core, pre-ET distributions improve markedly with increasing
704 resolution, especially for CNRM-CM6.1 and HadGEM3-GC3.1, but more clearly so in the
705 atmosphere-only than in the fully coupled simulations, wherein cold-wake feedbacks reduce
706 upper-ocean temperatures and weaken subsequent tropical cyclones (Balaguru et al., 2014).
707 Sensitivity to resolution is similar in the fully coupled CNRM-CM6.1 and HadGEM3-GC3.1
708 simulations (Fig. 11, top row). These results show that horizontal resolutions typical of
709 CMIP6 appear sufficient to simulate cold-core (post-ET) intensity distributions, including the
710 relatively high-intensity tail—resolutions at which large-ensemble studies to quantify
711 multiannual variability of the strongest post-ET storms are computationally feasible.

712 However, among high-resolution PRIMAVERA models, the high-intensity tail of the warm-
713 core distribution is reproduced only by CNRM-CM6.1.

714

715 For *highresSST-future*, several models project decreasing warm-core and increasing cold-
716 core intensities for weaker storms ($<17\text{ms}^{-1}$) but simulate opposite warm- and cold-core
717 responses for stronger storms ($\geq 17\text{ms}^{-1}$) (Fig. 10, bottom row). This warm-core response is
718 consistent with projections of intensified tropical cyclones under anthropogenic warming
719 (Knutson et al., 2020). However, these responses are not replicated by fully coupled models
720 (Fig. 11, bottom row), in which intensity changes are weak (Roberts et al., 2020b). In the
721 fully coupled simulations, the responses of pre- and post-ET intensity distributions to climate
722 change are equivocal, with substantial inter-model differences. We speculate that the climate-
723 change forcing out to 2050 in the HighResMIP experimental protocol is insufficiently strong
724 (i.e., the future simulation period is too short) for a clear signal to emerge. However, it is
725 unclear whether intensity changes would be seen. For tropical cyclones overall, Roberts et al.
726 (2020b) found a weak future intensification in these simulations, and Bieli et al. (2020) found
727 equivocal ET climate-change responses in many basins out to 2100 under the weaker RCP4.5
728 scenario. If a clear climate-change signal were to emerge with further increases in model
729 resolution, which would increase the relative difference between the weakest and strongest
730 simulated tropical cyclones, this would suggest that processes important for intensity change
731 are not adequately captured at ~ 25 km resolution.

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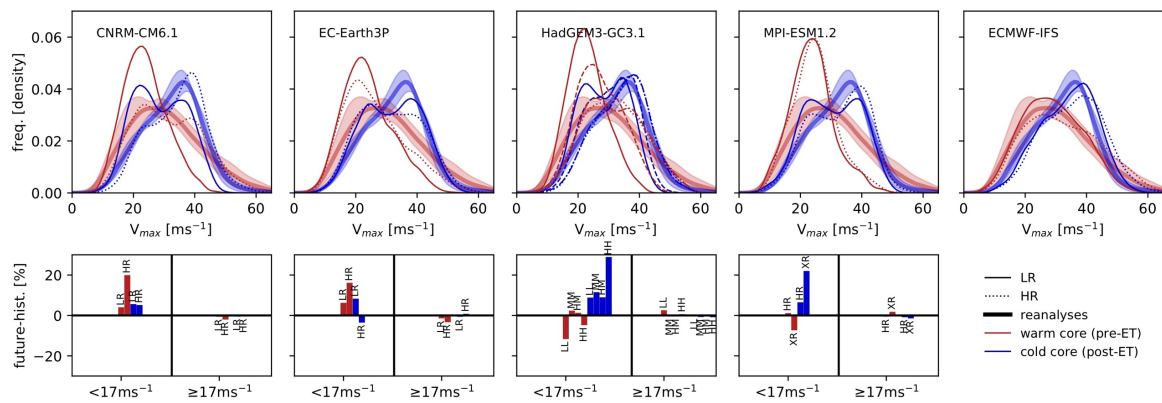
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Fig. 10. Intensity (v_{max} at 925hPa) distributions in atmosphere-only simulations for all cyclones undergoing ET globally. For each model, historical simulations are shown in the top row and future simulations in the bottom row. Multireanalysis-mean curves (thick, solid lines) are shown with 1 s.d. (shading). Both low- (thin, solid lines) and high-resolution (thin, dashed lines) simulations are shown. Climate-change responses (i.e., *highresSST-future* minus *highresSST-present*), computed as integrated differences, are shown as percentages for storms whose lifetime-maximum intensity is $<17 \text{ ms}^{-1}$ or $\geq 17 \text{ ms}^{-1}$ for each atmospheric model resolution (ordered left to right).



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Fig. 11. As in Fig. 10 for fully coupled simulations.

747 3.6 Post-ET reintensification

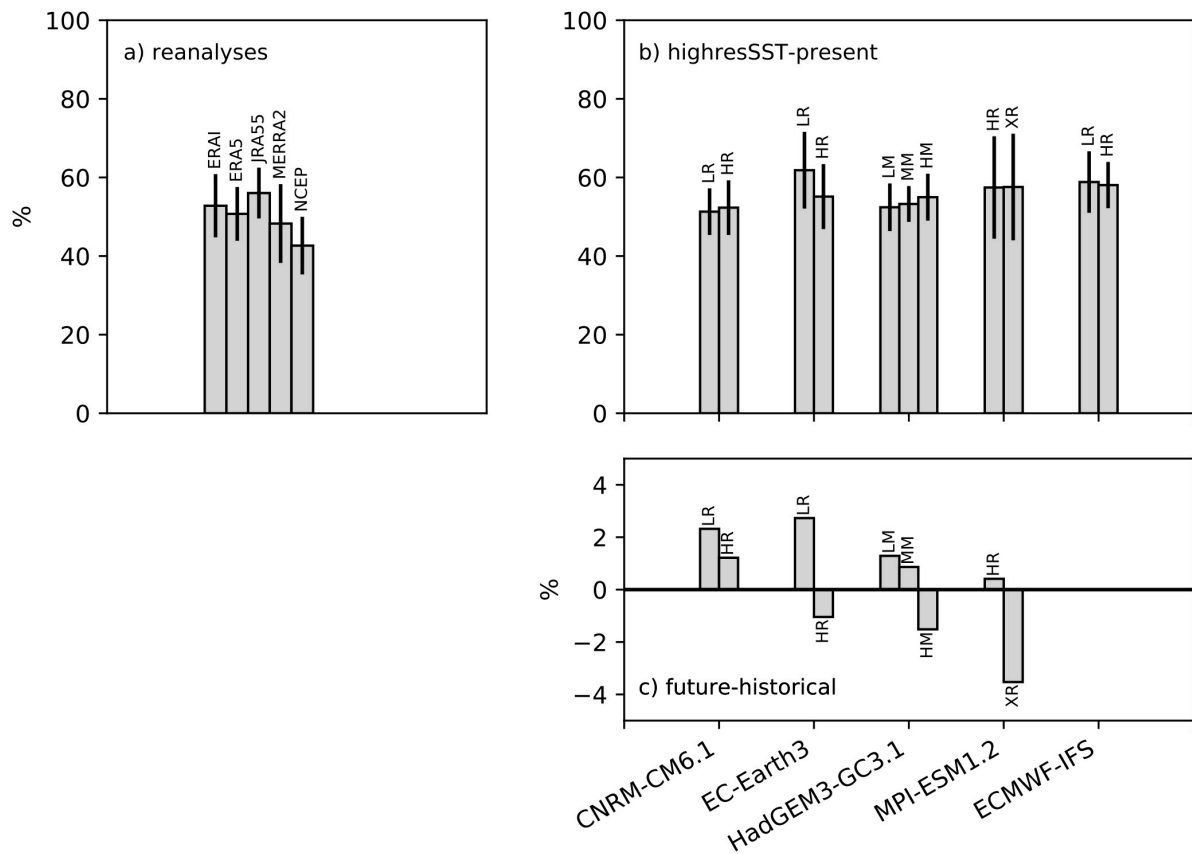
748 The lifetime-maximum intensity of transitioning tropical cyclones typically occurs during the
749 warm-core, tropical phase. However, the addition of a baroclinic energy source and cyclone-
750 wave interactions induce post-ET reintensification (Evans et al., 2017). We quantified the
751 frequencies of reintensifying versus non-reintensifying cyclones in reanalyses and in the
752 PRIMAVERA ensemble. Globally, reanalyses indicate that approximately 50 % of tropical
753 cyclones that undergo ET undergo post-ET reintensification (Fig. 12a). For the North
754 Atlantic and Western North Pacific basins, ~55 and ~45 %, respectively, reintensify (not
755 shown), consistent with Hart and Evans (2001). These results are not significantly different
756 when reintensification is defined using 925-hPa wind speed (not shown). Globally,
757 PRIMAVERA models generally overestimate climatological reintensification frequency at
758 low resolution, but increasing resolution decreases the proportion of reintensifying systems
759 (and increases the proportion of non-reintensifying systems) in all models except MPI-
760 ESM1.2, which better matches reanalyses (Fig. 12b and Fig. 13b). This result potentially
761 reflects improved simulation of the interactions between cyclones and the large-scale
762 circulation, which acts to reintensify systems (Keller et al., 2019), at high resolution. Which
763 processes facilitate such improvement should be a focus of future research because these
764 processes will be important for risk assessments of reintensification. However, it is also
765 possible that post-ET reintensification arises in models whose effective resolution increases
766 with increasing latitude (e.g., HadGEM3-GC3.1), allowing stronger simulated winds at
767 higher latitudes, but the impact of this artifact will be reduced at higher resolutions.

768

769 In HadGEM3-GC3.1, for an atmospheric resolution of 25 km (at 50 ° latitude), increasing
770 ocean resolution from 1/4 ° to 1/12 ° (-HM and -HH, respectively) does not impact the
771 proportion of reintensifying cyclones (Fig. 13b). An increase in the proportion might be
772 expected because increasing ocean resolution and therefore more sharply resolving SST
773 fronts (around western boundary currents; Fig. S5) is likely to enhance baroclinicity and
774 provide atmospheric conditions conducive to post-ET reintensification. That no increase is
775 seen implies that atmospheric resolution, to which simulated tropical-cyclone frequency and
776 intensity are sensitive, acts as a constraint on reintensification statistics, at least for this
777 particular model. Further investigation with multiple ocean models would establish more
778 robustly whether this is the case.

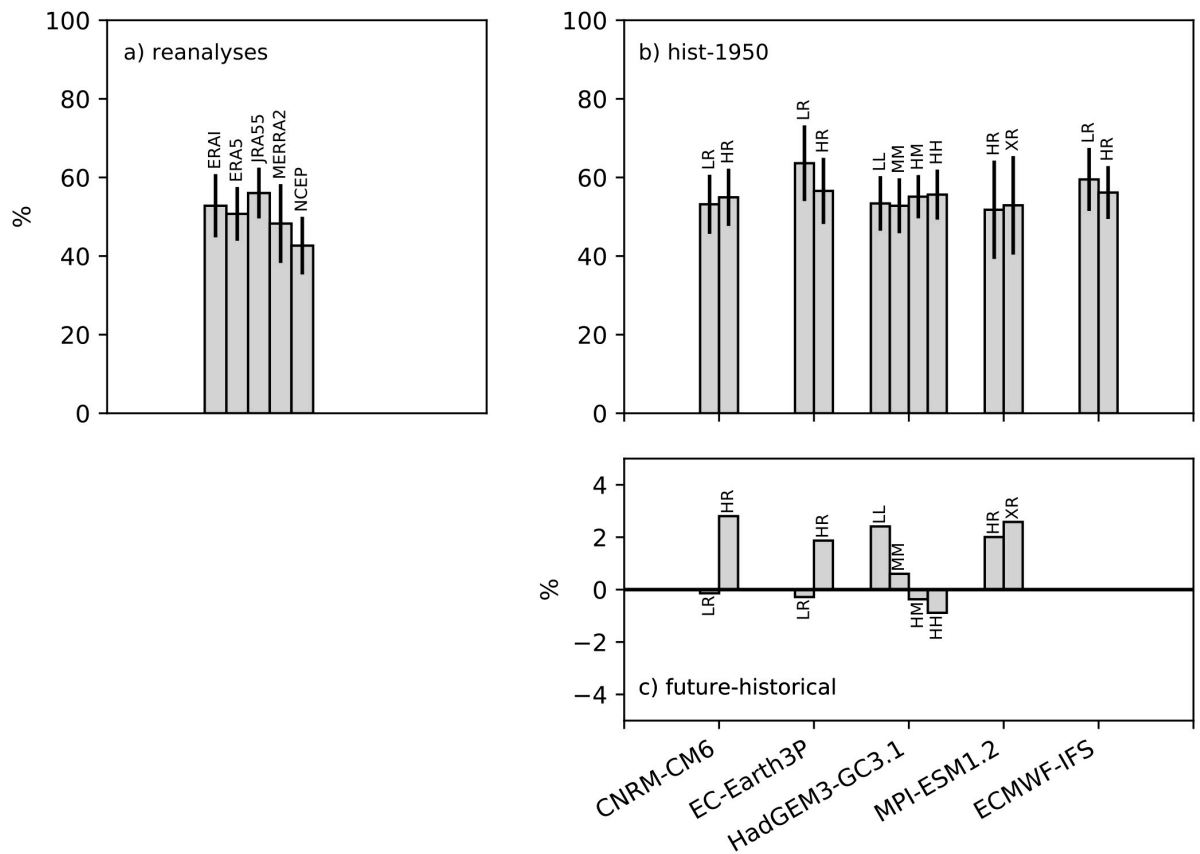
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780 In both the atmosphere-only and fully coupled simulations, future changes in the proportion
781 of post-ET reintensifying systems are small and generally within one standard deviation of
782 historical interannual variability (Fig. 12c and Fig. 13c), again suggesting that any climate-
783 change response under RCP8.5 emerges after 2050. In atmosphere-only simulations, low-
784 resolution models all simulate an increase the proportion of reintensifying cyclones, but high-
785 resolution models simulate a decrease (Fig. 12c), except for CNRM-CM6.1. Fully coupled
786 models typically simulate a future increase across resolutions (Fig. 13c).
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Fig. 12. Global analysis of the percentage of transitioning storms that undergo post-ET reintensification in (a) reanalyses and (b) *highresSST-present* simulations, and (c) the percentage change simulated for *highresSST-future* experiments. One standard deviation of interannual variability is indicated for each reanalysis and historical model simulation (black lines).



795

796 **Fig. 13.** As Fig. 12 but for *hist-1950* and *highres-future* simulations.

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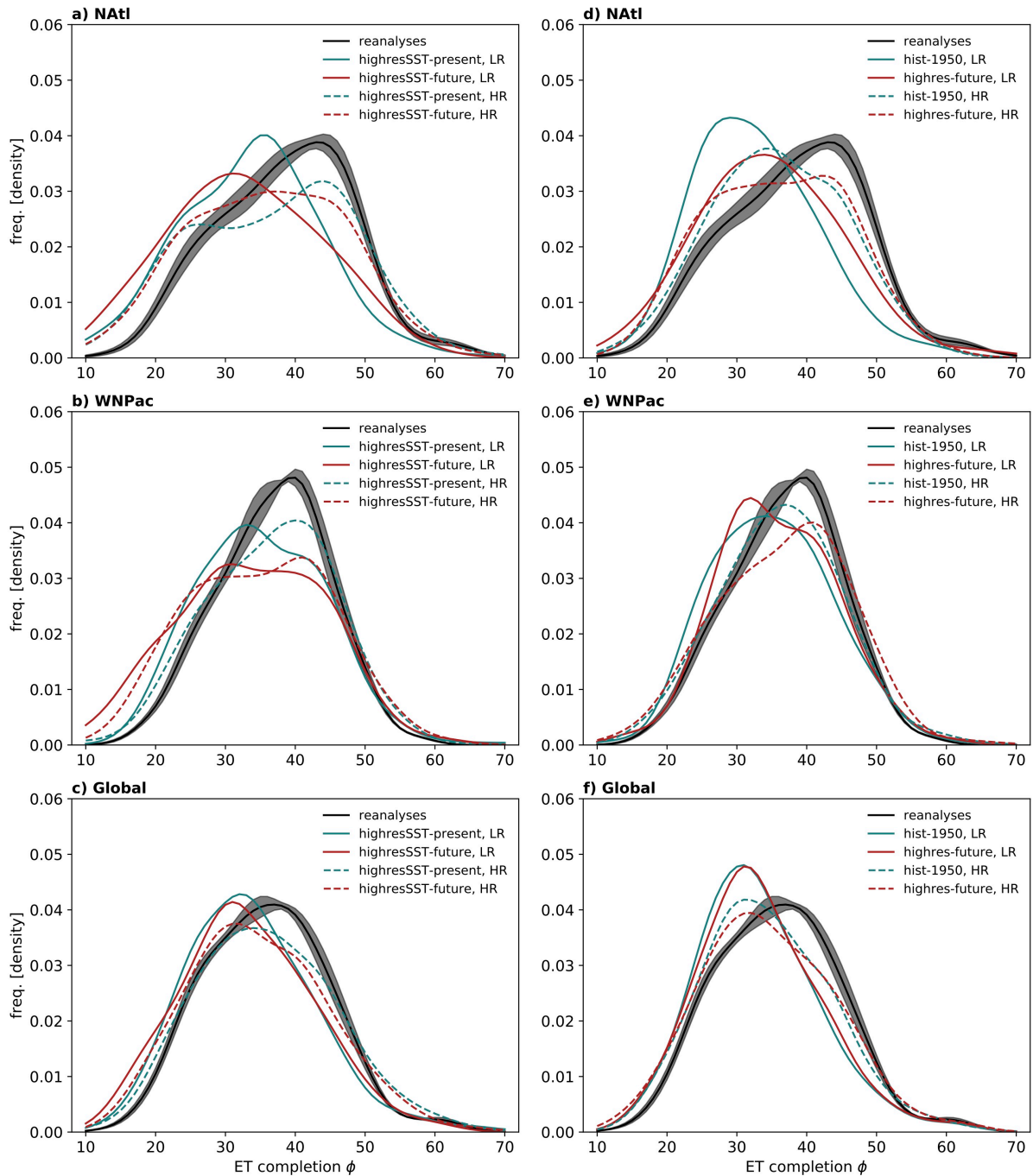
798 3.7 ET latitude

799 Finally, we assess how ET location responds to both increased resolution and to climate
800 change out to 2050. Distributions of ET-completion latitude were computed from reanalyses
801 and all PRIMAVERA experiments globally as well as well as separately for the basins where
802 models exhibit the best performance: the North Atlantic and Western North Pacific basins
803 (Fig. 14). For *highresSST-present*, model-simulated ET completion occurs at lower latitudes
804 than in reanalyses (Fig. 14a–c). At high resolution, this is partially rectified: peak frequency
805 occurs at a similar latitude to reanalyses in both the North Atlantic (Fig. 14a) and Western
806 North Pacific (Fig. 14b), but the magnitudes of both peaks are underestimated and
807 occurrences of low-latitude ET (i.e., 10–20 °) remain too frequent. Globally, an equatorward
808 bias in peak frequency across resolutions indicates that ET-completion latitude is less well
809 simulated in other basins (Fig. 14c). These results hold true for *hist-1950* simulations (Fig.
810 14d–f), except there are fewer instances of low-latitude ET (i.e., 10–20 °), likely reflecting
811 slower development of warm-core structures and subsequent ET in the fully coupled
812 experiments.

813

814 In response to climate change, the ensemble-mean distribution of ET-completion latitude
815 exhibits an equatorward shift in the North Atlantic in the atmosphere-only experiment (Fig.
816 14a), but a poleward shift in the fully coupled simulations (Fig. 14d), with an increased
817 frequency of ET completion particularly between 45–55 °N. In the Western North Pacific, a
818 poleward shift is seen in the latitude of the peak frequency, from ~30 to ~40 °N, in both
819 experiments, but little change is simulated at higher latitudes (i.e., > 45 °N). Globally, a small
820 equatorward shift of ~2 ° is simulated in atmosphere-only (Fig. 14c) and no meridional shift
821 is seen in coupled simulations (Fig. 14f). Previously, we showed stronger low-level warm-
822 core structures are simulated in future (Fig. 6 and Fig. 8), which potentially allow tropical
823 cyclones to propagate farther poleward prior to ET, with the most pronounced signal seen in
824 the North Atlantic. While coupled PRIMAVERA models provide evidence for a poleward
825 shift of ET, climate-change responses globally are equivocal out to 2050.

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Fig. 14. Ensemble-mean frequency distributions of ET-completion latitude for (solid lines) low- and (dashed lines) high-resolution simulations, for both 1950–2014 (teal) and 2015–2050 (red). Results are shown for (a–c) atmosphere-only and (d–f) fully coupled experiments for the North Atlantic basin (“NATl”), Western North Pacific basin (“WNPac”) and all global basins combined. ‘LR’ and ‘HR’ denote low- and high-resolution distributions, respectively. Also shown is the multireanalysis-mean distribution with shading indicating the standard error for the five reanalyses. Note that frequency is plotted as a function of absolute latitude (ϕ) to combine Northern and Southern Hemisphere results in c) and f).

836

837 **4. Summary and discussion**

838 This paper presents an analysis of ET across five reanalysis datasets and climate simulations
839 performed with five atmosphere-only and full coupled global models participating in CMIP6
840 HighResMIP, focussing on (i) the effect of increased model resolution on the representation
841 of ET and (ii) the response of ET to climate change.

842

843 For all tropical cyclones undergoing ET, we find an increase in the climatological track
844 density simulated at high resolution (~25 km) compared with low resolution (~100 km) in all
845 ocean basins and in both atmosphere-only and fully coupled model configurations (Fig. 1b,
846 d), particularly over Northern Hemisphere western boundary currents. Model error in
847 simulated track density (compared with the multireanalysis-mean track density) is reduced at
848 high resolution in the North Atlantic and Western North Pacific (Fig. 1c, e). The simulated
849 climatological annual-mean count of ET events is closer to that of reanalyses in the ocean
850 basins where ET activity is highest—the North Atlantic and the Western and Eastern North
851 Pacific—in both atmosphere-only (Fig. 3) and fully coupled (Fig. 4) experiments. In these
852 basins, atmosphere-only simulations exhibit skill of ~0.3 in capturing interannual variability
853 in just the subset of tropical cyclones that undergo ET (Table 4), demonstrating that the skill
854 of these models in simulating all tropical cyclones does not remain throughout the complete
855 cyclone lifecycle. Additionally, this level of skill in atmosphere-only simulations is lower
856 than that found for similar-resolution initialised seasonal forecasts (Liu et al., 2018). For the
857 other basins—the Northern Indian and Southern Hemisphere—frequencies simulated by
858 high-resolution models overestimate reanalyses. ET %, however, is similar between low- and
859 high-resolution simulations, indicating that the resolution sensitivity of ET is driven by that
860 of tropical cyclone frequency, not by an enhancement of environmental conditions conducive
861 to ET. The seasonal cycle of ET is reproduced by most models, with both the seasonal timing
862 and the magnitude of the seasonal peak simulated more correctly at high-resolution, but the
863 impact of increased atmospheric resolution is model-dependent.

864

865 In general, PRIMAVERA models show clearer inter-model agreement on the climate-change
866 response of ET frequency than on the response of intensity-related metrics. For most basins,
867 models simulate a frequency decrease in response to climate change, except over the North
868 Atlantic, where an increase is projected (Fig. 2). The magnitude of the North Atlantic
869 response is larger in atmosphere-only simulations than in fully coupled integrations and is

870 enhanced by increasing atmospheric model resolution, although interannual variability is
871 pronounced (Fig. 3 and Fig. 4). A significant positive trend in the ensemble-mean fraction of
872 North Atlantic ET events occurring during August–November is found over the period 1950–
873 2050 at high-resolution, indicating long-term change in ET seasonality in this basin, but no
874 secular seasonality change is simulated in the Western North Pacific (Fig. 5). North Atlantic
875 seasonality change may result in a higher proportion of tropical cyclones encountering the
876 midlatitude environment during the part of the seasonal cycle when, climatologically,
877 baroclinicity is highest (Hoskins and Hodges, 2019). Opposing future ET responses between
878 the North Atlantic and Western North Pacific are potentially underpinned by changes in low-
879 level, pre-ET warm-core structures, which strengthen in response to climate change in the
880 North Atlantic but undergo little change in the Western North Pacific (Fig. 6 and Fig. 7).
881 Comparing atmosphere-only with fully coupled simulations, the North Atlantic track density
882 response to climate change is more muted in the fully coupled experiment, which is
883 consistent with a less pronounced climate-change response of pre-ET structures simulated by
884 coupled models. Simulations with higher-resolution, storm-resolving models will open
885 opportunities to further study realistically deep warm-core cyclones.

886

887 Globally, simulated warm-core, pre-ET intensity distributions improve with resolution in
888 most models in both atmosphere-only and fully coupled experiments, better resembling
889 reanalyses (Fig. 10 and Fig. 11). Simulated cold-core, post-ET intensity distributions exhibit
890 little sensitivity to resolution across models. Globally, models simulate no clear climate-
891 change response of pre- or post-ET intensity distributions, suggesting that, if a signal exists,
892 extending simulations beyond 2050 may be required. Under *highresSST-future* forcing, some
893 models show decreasing warm-core and increasing cold-core intensities for storms $<17\text{ms}^{-1}$,
894 but the opposite response for storms $\geq 17\text{ms}^{-1}$. However, this is not reproduced by fully
895 coupled models. Globally, increasing resolution increases the proportion of simulated post-
896 ET reintensifications to approximately match reanalyses, but not in all models. Climate-
897 change responses are not significant with respect to historical interannual variability and are
898 model-dependent (Fig. 12 and Fig. 13).

899

900 The role of model resolution is become clearer, but uncertainties remain. Recent analysis of
901 tropical cyclones the PRIMAVERA simulations (Roberts et al., 2020b) has shown that the
902 high-resolution atmosphere-only models, which typically have lower wind-speed biases,
903 show either reduced future wind speeds or no change. Fully coupled models with the smallest

904 historical biases simulate either no change in future wind speeds or increases of only a few
905 percent. These models therefore project weaker intensity responses to climate change
906 compared with other studies (Knutson et al., 2020). One potential factor is the simplifying
907 aspects of the HighResMIP protocol that are necessary to isolate the role of model resolution,
908 particularly the standardised aerosol forcing and use of a single set of SST and sea-ice
909 boundary conditions shared across models (Haarsma et al., 2016). For ET, the climate-change
910 responses of pre- and post-ET intensity analysed in this study are largely model-dependent,
911 with models exhibiting little systematic change between atmospheric resolutions of ~100 and
912 ~25 km. This suggests that these disparate responses are due to differences in model
913 formulation, but a larger ensemble of models is likely needed to assess this fully. For post-ET
914 reintensification, increasing atmospheric resolution appears to result in more consistent
915 model behaviour, but resolution remains a key research issue because several models still
916 underestimate tropical cyclone intensities at ~25 km grid spacing (Roberts et al., 2020a) and
917 further improvements are anticipated by increasing resolution to at least 10 km (Haarsma,
918 2021; Judt et al., 2021). To obtain samples of ET events comparable to this study, however,
919 running sufficiently long simulations (and / or a sufficiently large ensemble) at these storm-
920 resolving resolutions, even without coupling to an ocean model, remains a significant
921 computational challenge (Roberts et al., 2020b).

922

923 Additional outstanding questions and uncertainties remain. A poleward expansion of Hadley
924 circulation termini is projected in a warmer climate (Lu et al., 2007), which implies
925 meridional shifts in tropical storm tracks (Sharmila and Walsh, 2018; Studholme and Gulev,
926 2018). However, the impacts of this large-scale change on the spatial distribution and
927 frequency of ET are equivocal. The poleward expansion of regions conducive to tropical
928 cyclone genesis and development that results from an increase in Hadley cell width will
929 reduce the mean displacement required for tropical cyclones to reach the midlatitude
930 baroclinic zone, increasing the likelihood of ET. However, a poleward shift of the midlatitude
931 storm track in response to warming has been projected (Bengtsson et al., 2006), which in turn
932 shifts environmental conditions conducive to extratropical transition poleward, potentially
933 offsetting Hadley-driven changes. Here, we find minimal changes in ET-completion latitude
934 out to 2050 (Fig. 14), suggesting cancellation in the net effect of these competing large-scale
935 changes. Further work is needed to establish the time of emergence of any meridional shift
936 and will require dedicated studies, exploring a range of climate-change scenarios with models

937 run at resolutions sufficiently high to adequately represent both tropical cyclones and ET—at
938 least 25 km, according to our results.

939

940 This study provides evidence that pre-ET cyclone intensity and warm-core strength exert
941 influence over future changes in ET statistics and seasonality. Analysis of higher-resolution
942 and storm-resolving models (at least 10 km) will help establish whether these results hold
943 true for models able to reproduce more realistic tropical-cyclone maximum intensities,
944 including rapidly intensifying systems. Additionally, there is a need to contextualise future
945 projections of ET, accounting for natural variability, and in particular the roles of regional
946 (e.g., Atlantic Multidecadal Variability) and global (i.e., El Niño–Southern Oscillation)
947 modes of variability on ET frequency. Dedicated sensitivity experiments will be required,
948 and such a study is forthcoming for the North Atlantic, where this work has identified future
949 changes that are important and often unique to this basin. Finally, investigation of secular
950 change in ET seasonality, as seen in the North Atlantic in this study, will be important
951 globally because future modification to the interval between the seasonal maximum of ET
952 occurrence and wintertime storminess may engender considerable changes in risk for
953 populous midlatitude regions.

954

955 **Data and code availability**

956 All reanalysis data for tropical-cyclone tracking (vorticity, wind fields, and sea-level
957 pressure) and cyclone phase-space analysis (geopotential) are available from rda.ucar.edu or
958 disc.gsfc.nasa.gov. Model data are available from Earth System Grid Foundation nodes
959 (esgf.llnl.gov). TRACK is available for download at gitlab.act.reading.ac.uk/track and the
960 track datasets used in this paper may be downloaded from
961 catalogue.ceda.ac.uk/uuid/e82a62d926d7448696a2b60c1925f811. Data analysis and
962 visualisation code is available from the lead author upon request (hrcm.ceda.ac.uk/contact).

963

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974

975 **Author contributions**

976 AJB, PLV, RJH and MJR conceived the study. Simulations were performed by MJR, ET,
977 KL, CDR, and LT. Output data were managed by JS. MJR performed the cyclone tracking.
978 BV computed Eady growth rate. AJB undertook cyclone phase-space analysis and all other
979 data analyses, figure preparation, and wrote the manuscript. All authors provided input in
980 interpreting results and approved the final manuscript.

981

982 **Competing interests**

983 The authors declare no competing interests.

984

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Supplementary information

Extratropical transition of tropical cyclones in a multiresolution ensemble of atmosphere-only and fully coupled global climate models

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S1. Methodological considerations

Two important methodological considerations in ET studies are discussed in this section: (i) the cyclone-tracking algorithm and (ii) the sensitivity of ET location to how ET is identified.

S1.1 Cyclone-tracking algorithm

Recent studies of tropical cyclones in reanalyses and simulated by climate models (e.g., Roberts et al., 2020a; Vanni re et al., 2020) compared results obtained using TRACK with TempestExtremes, a sea-level-pressure-based tracking algorithm (Ullrich and Zarzycki, 2017), to show that, broadly, their results are robust to algorithm choice. However, tracks output by TempestExtremes represent only cyclones' warm-core stages, and as such few identified systems undergo ET (Fig. S1). Therefore, supplementary algorithms are required to extend cyclone tracks generated using TempestExtremes into the midlatitudes (e.g., Michaelis and Lackmann, 2019; Zarzycki et al., 2017), but the extent to which results are sensitive to the additional methodological choices necessary in this approach is unclear. In this study, use of TRACK, a vorticity-based algorithm that satisfactorily yields complete cyclone lifecycles based on a single set of identification criteria, is clearly advantageous in our analysis of ET statistics. Once comparable whole-lifecycle tracks, including post-tropical evolution, from multiple, independent algorithms are available, sensitivity analysis should be a research priority.

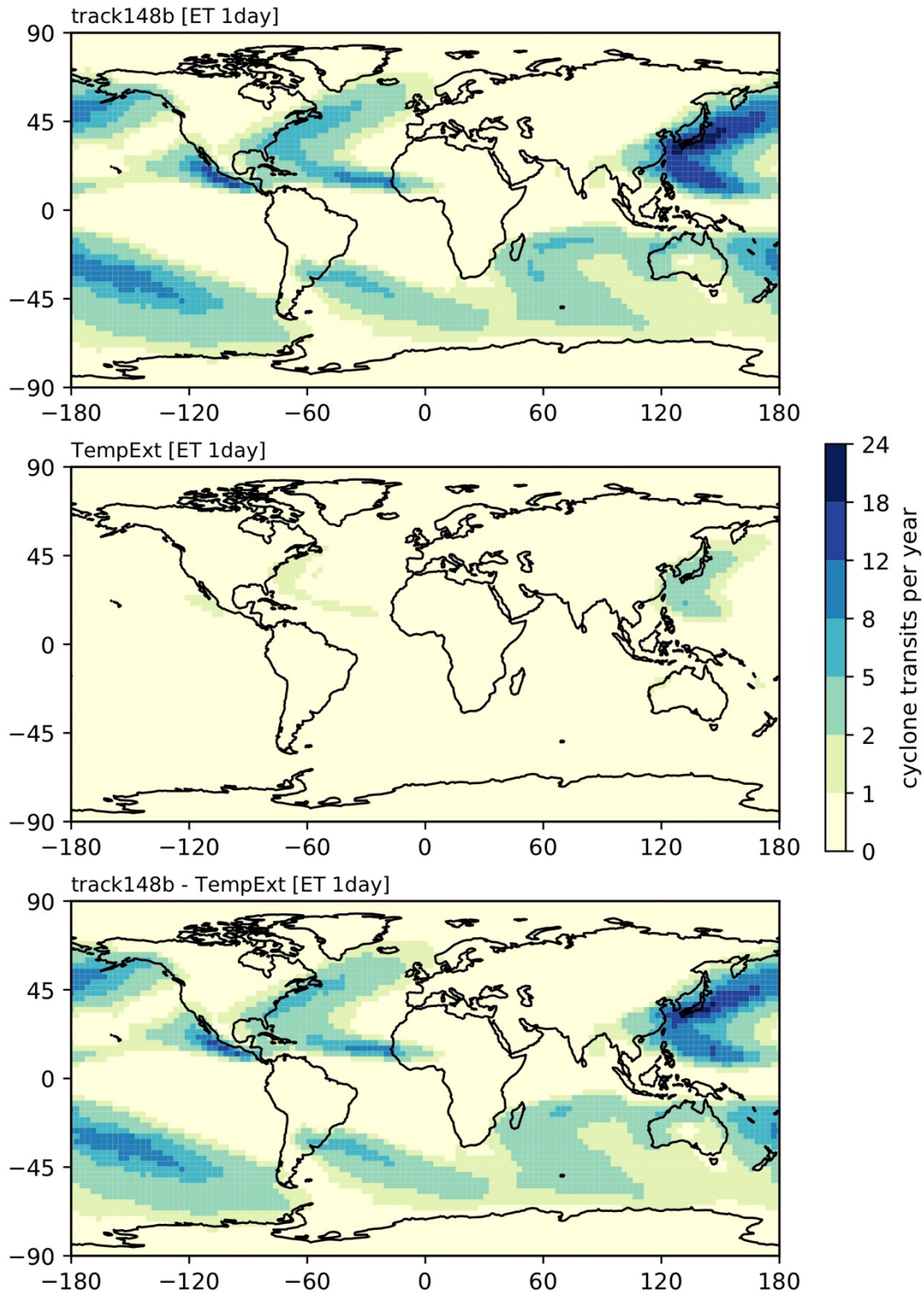


Fig. S1. Multi-reanalysis-mean track density in TCs undergoing ET identified by two feature-tracking algorithms: (top) TRACK (“track148b”) and (middle) TempestExtremes (“TempExt”). We also show (bottom) the inter-algorithm difference (i.e., TRACK minus TempestExtremes).

S1.2 Sensitivity of ET location to identification method

Recent analysis of ERA5 shows that phase-space-based identification of post-tropical structures compares well with other methods in terms of the number of ET events identified (Sainsbury et al., 2020) and phase-space methodologies are the most common across studies of ET. However, no consensus approach to identifying ET onset and completion based on phase-space parameters exists. Previous studies have applied differing absolute thresholds to identify changes in cyclone thermal symmetry and employed additional criteria, such as intensity thresholds and temporal smoothing of phase-space series. These modifications have little impact on the number of identified events, but the location of ET may be sensitive to how phase-space parameters are treated. We conducted an overview assessment of this in reanalyses by mapping the mean ET-completion locations for various identification approaches (Fig. S2). In the North Atlantic and Western North Pacific, a definition of ET where both ET onset and completion are identified by single-timestep B or T_L changes (‘conventional’) yields the lowest-latitude ET completion (Fig. S2). In contrast, applying a prior warm-core test, as in the previous section (4.2), yields ET completion in the range of 30–40 ° latitude, coinciding with the known centres of baroclinicity associated with western boundary currents. Other proposed modifications of ET identification—imposing a v_{max} threshold, applying a temporal smoothing, and requiring B and T_L criteria are met for consecutive timesteps—yield locations in between these two approaches, with greater overlap seen in the Western North Pacific (Fig. S2). These results help quantify the sensitivity of ET location to various methodological choices, and the results presented in Fig. 16 are necessarily sensitive to such choices, as are other published analyses. The method should fit the research question. When ET completion is identified post-warm core, it is broadly co-located with climatological, basin-high values of Eady growth rate (Fig. S2), indicating that this approach may be preferable for analyses of ET location, particularly in the North Atlantic.

In addition, application to climate models presents additional concerns. Bieli et al. (2020) identified grid-scale convective updrafts in 50-km-resolution simulations with the Forecast-oriented Low Ocean Resolution (FLOR) version of the GFDL CM2.5 that triggered erroneous diagnoses of warm- and cold-core cyclone structures. These were rectified by computing storm-centric 95th-percentile geopotential (rather than local maxima) and by applying a temporal smoothing to phase-space trajectories. Although these issues are not pertinent to all models, understanding the effect of convection-parameterisation schemes on geopotential maxima and phase-space results, particularly T_U , requires a systematic investigation across

multiple high-resolution models, contrasting simulations run using parameterised versus explicitly resolved convection.

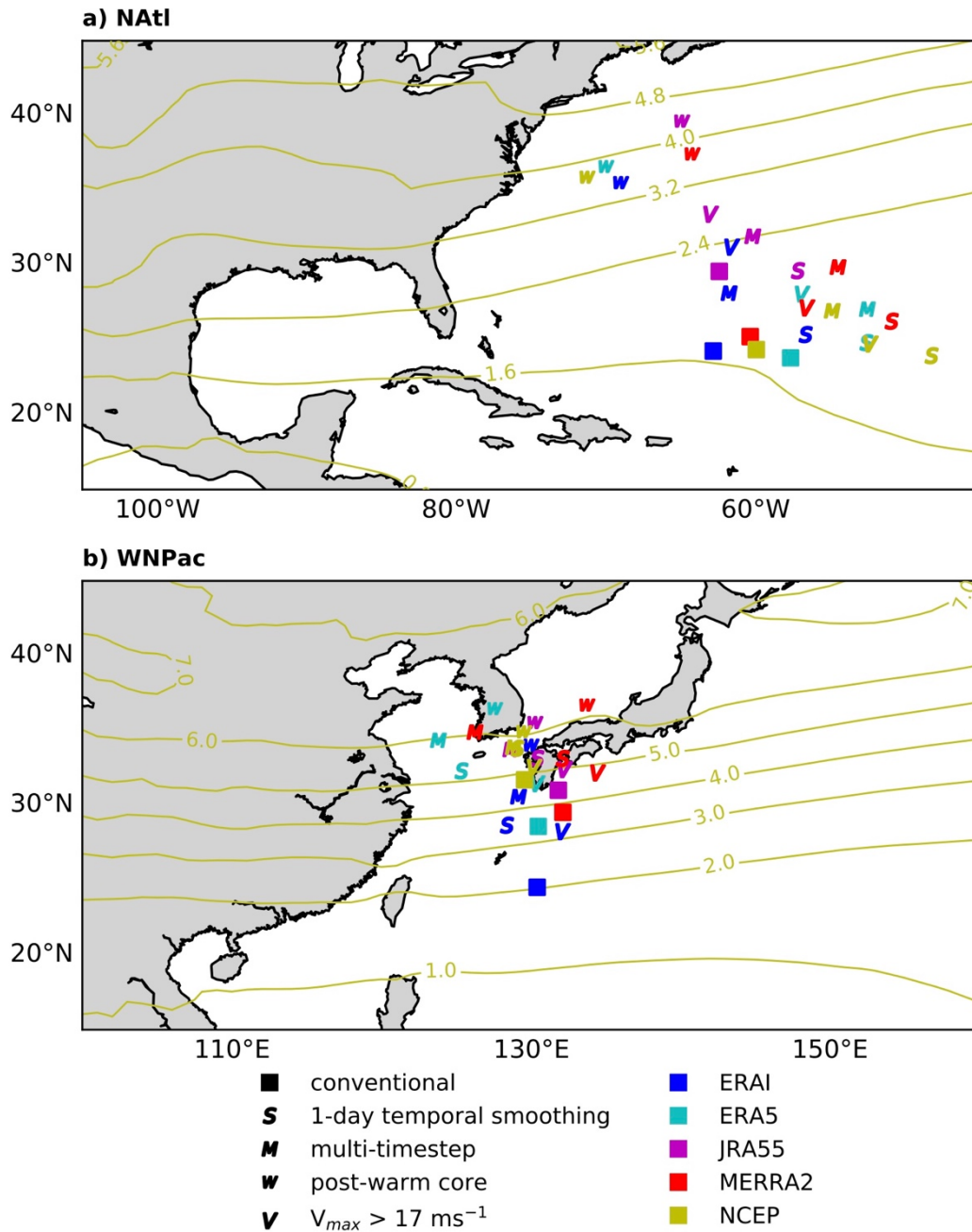


Fig. S2. Sensitivity of ET-completion location in reanalyses to published methodological approaches. Results are shown for (a) the North Atlantic ('NAAtl') and (b) the Western North Pacific ('WNPac'). 'Conventional' (square markers) refers to the commonly used definition of ET completion: the first timestep at which both B and T_L indicate an extratropical structure. The other markers indicate a single modification of this definition. 'S': a 24-hour temporal smoothing of B and T_L trajectories was applied. 'M': ET completion is only identified where B and T_L criteria are satisfied for four consecutive timesteps. 'V': ET completion is only

identified for storms whose lifetime-maximum intensity exceeds 17 ms^{-1} . ‘WC’: a warm core lasting for at least two days is first identified for each storm and ET completion is identified thereafter. Overlain are contours of climatological-mean Eady growth rate maxima for August–November in units of day^{-1} , computed using ERA5 wind and geopotential data using Eq. 4.

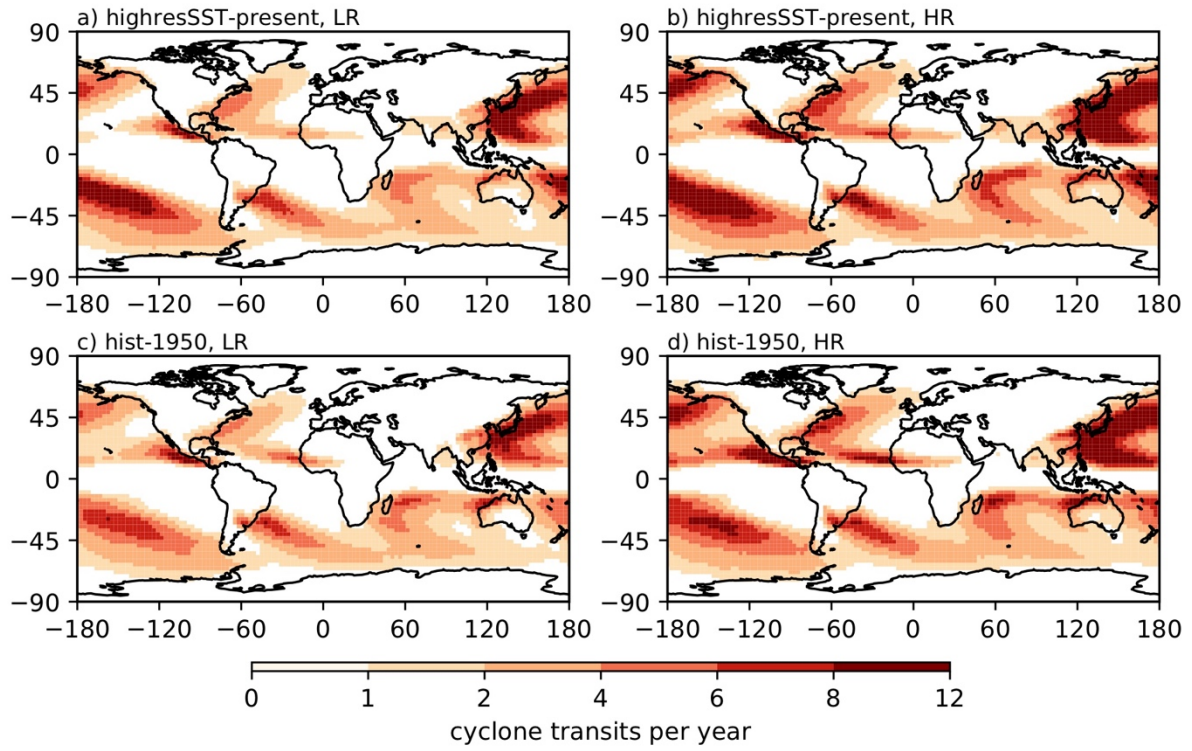


Fig. S3. Historical ensemble-mean track density simulated in low- and high-resolution (a–b) *highresSST-present* and (c–d) *hist-1950* experiments. Unit is cyclone transits per year per unit area (within a 5° geodesic radius of storm centres). Colour scale is the same as in Fig. 1a.

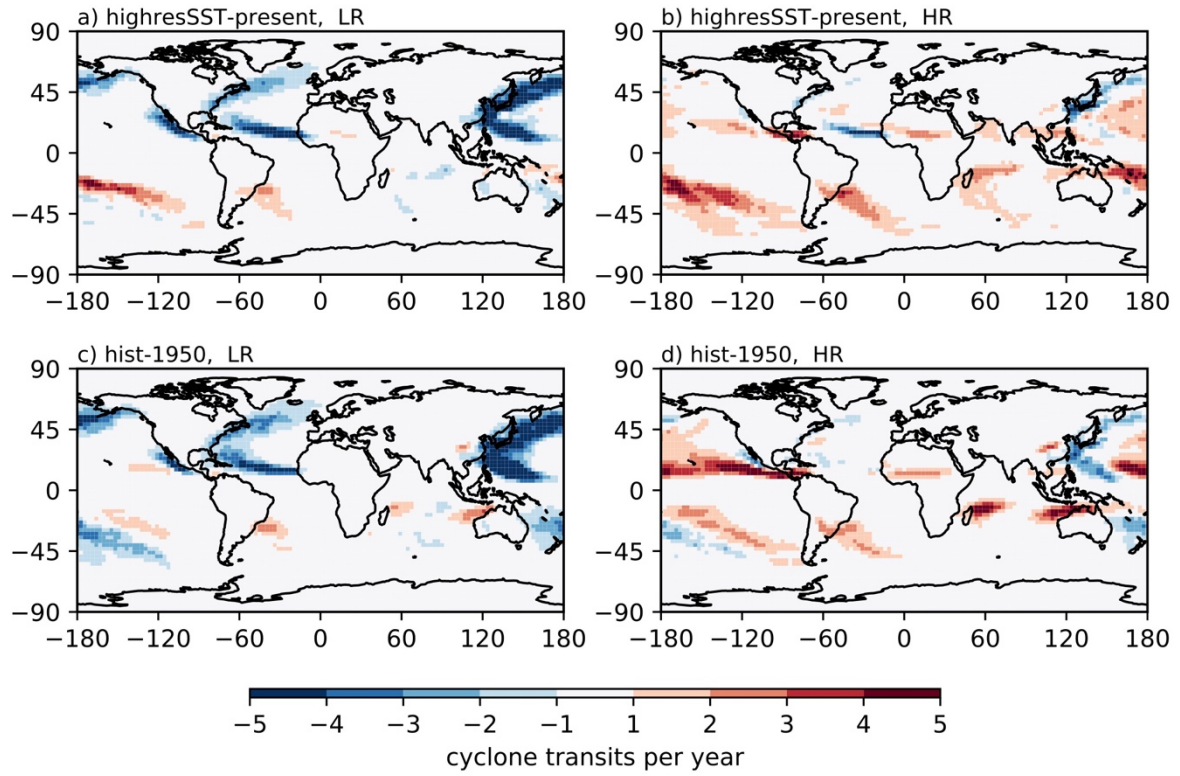


Fig. S4. Historical ensemble-mean track density biases (compared with multireanalysis-mean track density) simulated in low- and high-resolution (a–b) *highresSST-present* and (c–d) *hist-1950* experiments. Unit is cyclone transits per year per unit area (within a 5° geodesic radius of storm centres). Colour scale is the same as in Fig. 1c, e.

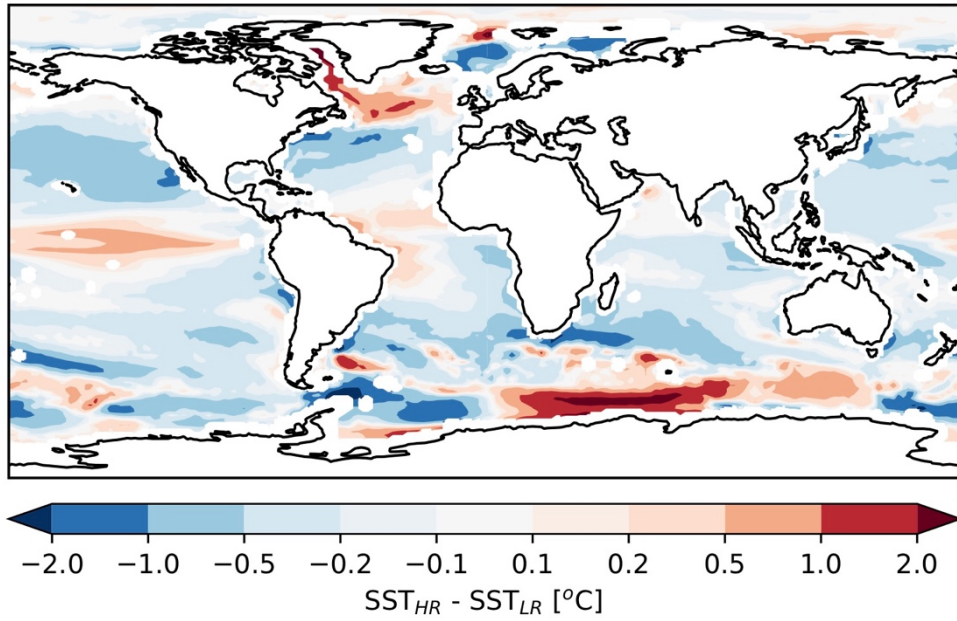


Fig. S5. Historical ensemble-mean August-November SST difference between low- and high-resolution *hist-1950* simulations. The low- and high-resolution sub-ensembles correspond to those of Fig. 1. Note the non-linear colour scale.

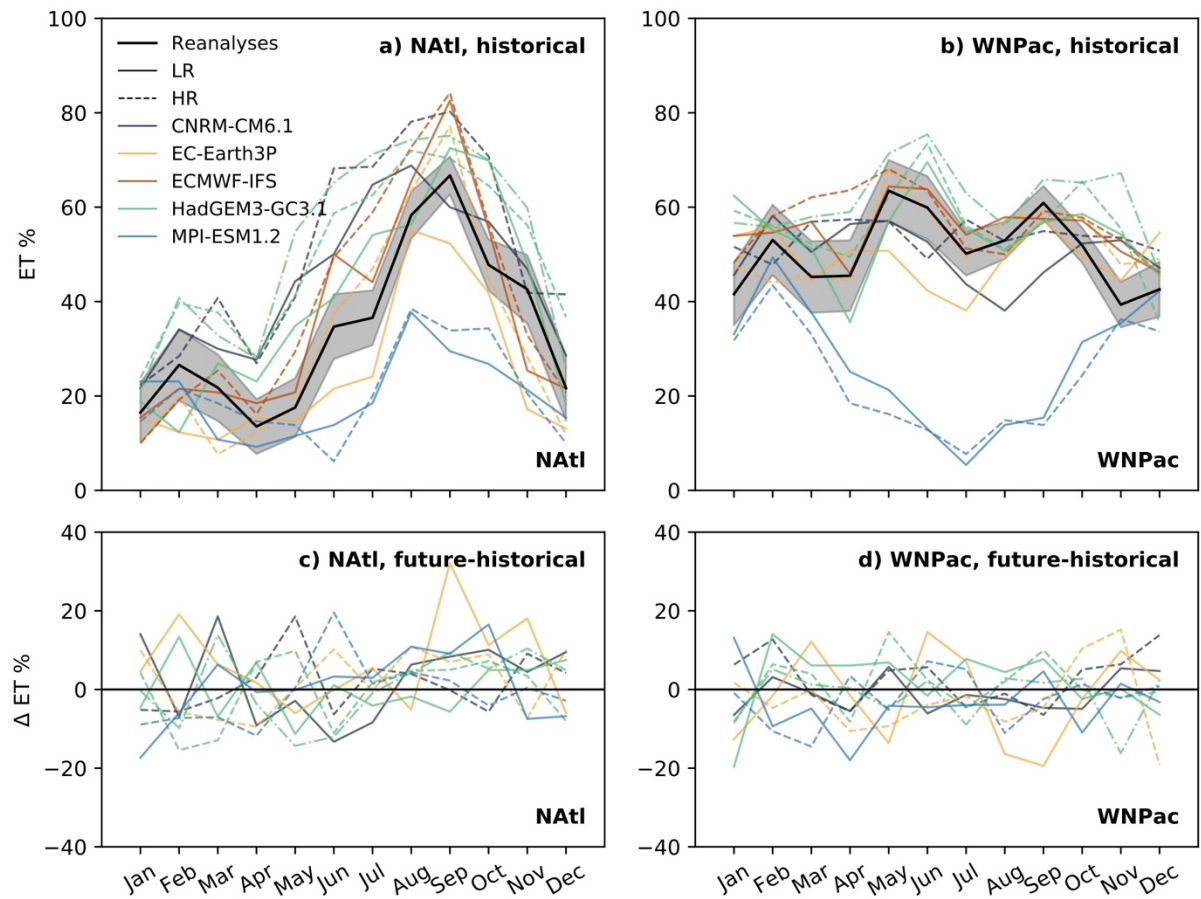


Fig. S6. Historical seasonal cycle of ET % in the (a) North Atlantic and (b) Western North Pacific basins. Shown are the multireanalysis mean (black with shading indicating standard error) and low- (solid) and high-resolution (dashed) *highresSST-present* simulations. (c–d) The difference between the future and historical seasonal cycles in ET % (i.e., *highresSST-future* minus *highresSST-present*). HadGEM3-GC3.1-MM is indicated by the dot-dashed line.

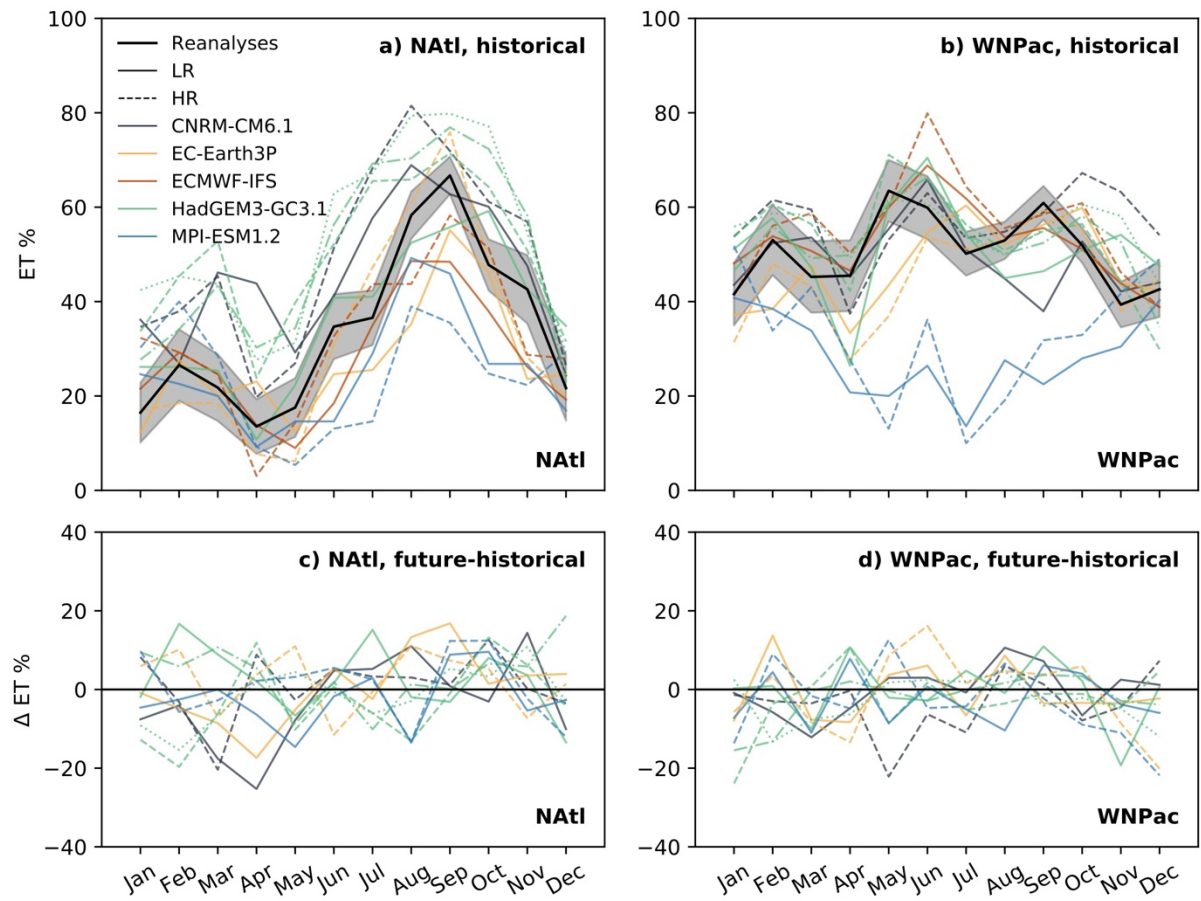


Fig. S7. Same as Fig. S6 but for fully coupled simulations. HadGEM3-GC3.1-MM and -HH are indicated by the dot-dashed and dotted lines, respectively.

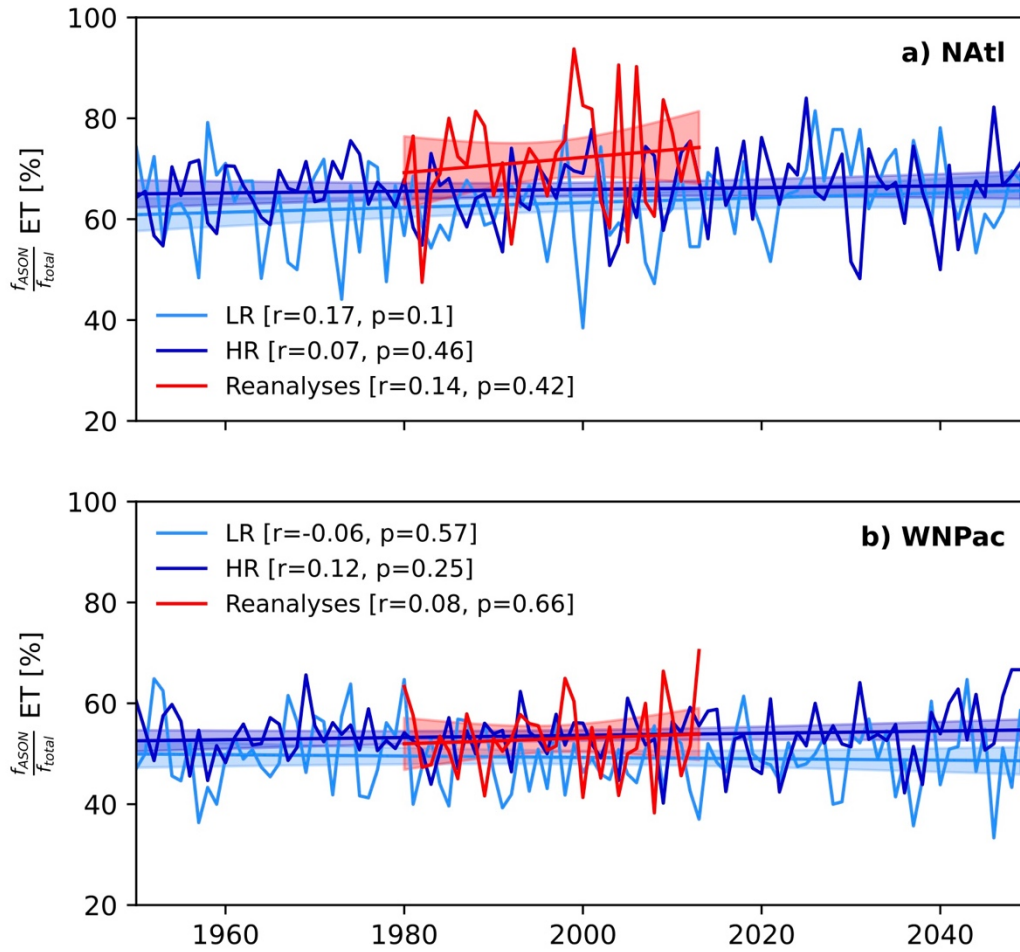


Fig. S8. Secular change in the ensemble-mean proportion of ET events occurring during August–November in reanalyses (red) and in low- (pale blue) and high-resolution (dark blue) fully coupled simulations for (a) the North Atlantic and (b) the Western North Pacific basins. Shading shows the 95 % confidence interval for the linear fit. ECMWF-IFS is not included in this analysis because no future simulations were performed in HighResMIP for this model.