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Simulating the Midlatitude Atmospheric Circulation: What Might We Gain From High-Resolution Modeling of Air-Sea Interactions?

A. Czaja[1] · C. Frankignoul[2] · S. Minobe[3] · B. Vannière[4]

Abstract

Purpose of Review To provide a snapshot of the current research on the oceanic forcing of the atmospheric circulation in midlatitudes and a concise update on previous review papers.

Recent Findings Atmospheric models used for seasonal and longer timescales predictions are starting to resolve motions so far only studied in conjunction with weather forecasts. These phenomena have horizontal scales of ~ 10–100 km which coincide with energetic scales in the ocean circulation. Evidence has been presented that, as a result of this matching of scale, oceanic forcing of the atmosphere was enhanced in models with 10–100 km grid size, especially at upper tropospheric levels. The robustness of these results and their underlying mechanisms are however unclear.

Summary Despite indications that higher resolution atmospheric models respond more strongly to sea surface temperature anomalies, their responses are still generally weaker than those estimated empirically from observations. Coarse atmospheric models (grid size greater than 100 km) will miss important signals arising from future changes in ocean circulation unless new parameterizations are developed.

Keywords Ocean-atmosphere interactions · High-resolution climate modeling · Midlatitude climate dynamics · Sea surface temperature anomalies · This article is part of the Topical Collection on *Mid-latitude Processes and Climate Change*

Introduction

Fluctuations in sea surface temperature (SST) in the extra-tropics have been documented since the 1970s. The available data at the time suggested that their spatial scale was large (i.e., on the scale of ocean basins), that their amplitude was on the order of 1 K, and that they were quite persistent (at least several months, i.e., much longer than the timescale associated with the weather). This led early investigations to hypothesize

that they exerted a significant impact on the extra-tropical atmospheric circulation, possibly enhancing predictive skill on weather, seasonal, and longer climate timescales.

As discussed in the reviews written on this issue since then by Frankignoul, Kushnir et al., Robinson, and Kwon et al. [1–4], evidence for a large impact of the extra-tropical SST anomalies on the atmospheric circulation had not been found. Rather, emphasis had been put on the fact that large-scale extra-tropical SST anomalies are forced by modes of atmospheric variability such as the North Atlantic Oscillation (NAO) and the Pacific North American pattern (PNA), and that the upper ocean provides a weak feedback on these modes. Large or weak refers here, as in [2], to whether the oceanic forcing, measured for example by the SST-induced geopotential height anomaly, is smaller or greater than that produced internally by the atmosphere. This quantification requires specifying a timescale since atmospheric spectra tend to be white and so the internal variability decreases with time averaging. Winter-to-winter changes in geopotential height of 50 m at upper levels are not uncommon so this number is taken in the following as a reference value. Quantifying the strength of the SST forcing also requires specifying a realistic amplitude for SST anomaly

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as there is little point in normalizing an atmospheric response by 1 K in a region where such SST amplitude is rarely reached or frequently exceeded. Hence, in this review, we will quote numbers as typical maximum geopotential height anomaly per maximum SST anomaly.

The purpose of this review is to provide a snapshot of the current state of research by offering a concise update on the review by Kushnir [2]. A central theme is to assess whether the "small oceanic forcing" regime suggested in that review (a number of 20 m at 500 hPa for a 1-K SST anomaly was quoted as a typical value) is relevant or not to the real world. This question is important in the context of climate change because the ocean circulation, especially in mid to high latitudes, is expected to respond significantly to increasing concentration of greenhouse gases and aerosols [5], and climate models need to be able to simulate accurately the associated impact on the weather patterns in midlatitudes and globally. Studies reviewed at the time in [2] used Atmospheric General Circulation Models (hereafter AGCMs) resolving adequately dynamics at the atmospheric deformation radius scale[1] (~ 1000 km) but not at the 10–100 km scale, which motivates a re-examination of this issue. Note that a focus is put here on ocean forcing at upper tropospheric levels because this is where teleconnection patterns such as the NAO and PNA are set [6]. We also focus on wintertime as this is the time of year when such patterns are the strongest, as are midlatitudes air-sea interactions.

Several recent results suggest that the "small oceanic forcing" regime might not be relevant to the real world. It is now known that AGCMs and the coupled seasonal forecasting systems based on them exhibit less predictability than nature in midlatitudes; for example, the correlation between predicted and observed NAO index is higher for the real atmosphere than for the atmosphere of the forecasting system, despite a realistic simulation of the noise level (e.g., [7–9]). This points to a lack of signal or sensitivity which could be related to an underestimation of boundary forcing in the current generation of atmospheric models, possibly from the midlatitude ocean. Additional support for this claim is that AGCMs are not able to simulate, even when forced with observed SST anomalies, the pronounced multidecadal variability displayed by the Jet Stream in the North Atlantic over the twentieth century [10]. In addition, it is clear that spatial fluctuations in SST on scales much smaller than ocean basins but closer to the oceanic deformation radius (10–100 km) affect the lower atmosphere (see for example the review in [11]). New studies start revealing mechanisms by which this oceanic variability might be "communicated upward" to the upper levels of the atmosphere if AGCMs have enough spatial resolution to see these features (e.g., [12–19]). Finally, and this might be the most

direct evidence for a stronger oceanic forcing than has previously been acknowledged and reviewed in [2], recent estimates of its strength in reanalysis data using statistical techniques are sometimes in excess of 60 m/0.5 K for geopotential height anomalies at upper levels (e.g., [20–22]).

We are rapidly moving away from AGCMs with horizontal resolution of a few 100 km (hereafter low res) toward AGCMs with a few tens of kilometers resolution (hereafter high-res). We refer throughout this paper to the "mesoscale" as the range of horizontal scale within 10–100 km which is starting to be resolved in global coupled seasonal forecasting systems but not yet in global coupled climate models used for climate change predictions. This corresponds to a dynamical regime where, in the extra-tropics, the Rossby number[2] is on the order of unity in the atmosphere ([23, 24], in agreement with their "advective regime"), but is smaller than unity in the ocean [25].

The review is structured as follows. In the "What Are the Atmospheric Motions and the Climatic Features Which Improve When Increasing an AGCM Resolution From ~ 100 to ~ 10 km in Midlatitudes?" section, relevant atmospheric features affected by spatial resolution are reviewed, while the "How Do the Extra-Tropical Oceans Affect the Storm Track/Jet Stream System?" section discusses mechanisms by which the ocean circulation might affect these features. The magnitude of the oceanic forcing is critically reviewed in the "Observational Constraints on the Extra-Tropical Oceanic Forcing" section, while the "Implications for Climate Change Predictions" section synthesizes the implications of all preceding sections for climate change predictions. Conclusions, suggestions for future research, and a tentative answer to the question posed in this review's title are offered in the "Conclusion" section.

What Are the Atmospheric Motions and the Climatic Features Which Improve When Increasing an AGCM Resolution From ~ 100 to ~ 10 km in Midlatitudes?

Increasing atmospheric model resolution from low res to high res has been shown to improve the representation of large-scale circulations in the midlatitudes: cyclones [26], the location and frequency of occurrence of blockings [26, 27], and the non-Gaussian nature of North-Atlantic weather regimes [28].

In addition, kinetic energy spectra have been used to show that additional contributions from smaller scales emerge when the resolution of AGCMs is increased [29–31]. This has led to the notion of an "effective resolution," defined as the

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[1] For the atmosphere and the ocean, the deformation radius is the horizontal scale at which rotation effects become as important as buoyancy effects.

[2] The Rossby number is a measure of the relative strength of inertia and Coriolis force.

wavenumber beyond which the slope of the spectrum steepens from − 5/3. Klaver [31] demonstrated, based on an ensemble of AGCMs spanning a large range of resolutions, that it is on average 3 times the size of the atmospheric grid cell. Thus, even a 25-km AGCM might only resolve the end of the mesoscale spectrum (∼ 100 km).

Midlatitude transient eddies in the atmosphere are shaped by the quasi-stationary flow but also affect it in return (e.g., [32]). The question naturally arises to which extent the increase in horizontal resolution could enhance this feedback. Held and Phillipps [33] suggested that resolution has little impact on eddy heat transport because, if the eddy heat flux is underestimated, a self-correcting feedback comes into play (the temperature gradient increases). This is in contrast with the eddy momentum flux which the model has little incentive to correct when it is underpredicted. Orlanski [34] demonstrated that the eddy momentum flux by synoptic eddies plays a different role on the mean climate in different frequency bands. Intermediate-frequency eddies (5–12 days) always flux momentum poleward, but a bifurcation exists for very high frequency eddies (1–5 days): eddy momentum is fluxed equatorward or poleward depending on whether their kinetic energy is high or low, respectively[3]. For example, the GFDL AM2 M90 model (at a 1° resolution) was shown to simulate correctly the momentum flux and energy of intermediate-frequency eddies but not those of very high frequency eddies (the variance of the latter was underestimated by ∼ 20% compared with ERA40 reanalysis [34]). These results have important implications for the oceanic forcing of the storm track/Jet Stream system on both the atmospheric and oceanic deformation radius scales which we will address in the "How Do the Extra-Tropical Oceans Affect the Storm Track/Jet Stream System?" section.

Despite the argument mentioned in [33], recent work has also suggested a non-negligible increase of the eddy heat flux (v′T′, where v′ and T′ refer respectively to the eddy meridional velocity and temperature) over western boundary currents of the ocean like the Gulf Stream or Kuroshio that could be caused by resolution. Willison et al. [35] found that v′T′ was increased by approximately 24% in the North-Atlantic when the resolution of their limited area model (LAM) was increased from 120 to 20 km. Smirnov et al. [14] found convincing evidence that the diabatic heating resulting from an anomalous SST in the Oyashio extension was balanced by mean flow cold advection at low resolution, but by an increased eddy heat flux in the high-resolution (such dynamics are further discussed in the "How Do the Extra-Tropical

Oceans Affect the Storm Track/Jet Stream System?" section). However, each of the two results comes with its own caveats. LAM studies, as used by Willison et al. [35], do not have a global energy constraint and as such might miss important feedbacks rectifying the mean state on which eddies develop. Conversely, the mean midlatitude atmospheric circulation of global AGCMs, as used in [14], is sensitive to resolution, though not always in the same way, and this might drive different storm track responses to a given anomalous heating [36, 37]. This sensitivity is illustrated in Fig. 1 for the wintertime mean zonal wind at upper levels in six state-of-the-art AGCMs. In Fig. 1a, we display in color the difference in this variable between a high- and a low-resolution version of the model (all models were forced with the same SST so as to emphasize changes arising from resolution as opposed to oceanic forcing, see [38]), superimposed on the high-resolution climatology (black contours; the dots indicate the 95% significance level). Changes on the order of 5–10 m/s (about 10–20%) are produced when increasing the resolution, sometimes inducing a poleward shift of the jet (e.g., Fig. 1a, top panel in the western Pacific), sometimes a southward shift (Fig.1a, bottom panel in the western Pacific). Indeed, as Fig. 1b quantifies, the larger the increase in resolution, the larger the change in the time mean zonal winds (a few m/s for each doubling of the resolution). Models for which the ratio R of resolution (high res/low res) is close to unity also display the least significant changes in panel (a), as expected. It is not clear however that even significant changes systematically bring the models closer to observations as both positive and negative changes in biases are found at the smallest values of R (∼ 0.2) in Fig. 1c.

Cold and warm fronts associated with extra-tropical cyclones have typically a width of ∼ 100 km and are characterized by strong uplift where most of the diabatic heating occurs. Willison et al. [35] convincingly showed that low-res models are unable to represent the narrow extent of the diabatic heating. Because the diagnostic relationship between heating and vertical motion in midlatitudes involves the horizontal Laplacian of the heating and thus filters out large spatial scales (the so-called "omega equation," see [39]), this also explains why the vertical velocity field is misrepresented in low-res models [40]. This is an issue because, as AGCMs become able to represent better the narrow nature of ascending motion, the compensating subsidence becomes more spread out and non-local. As a result, net upward motion can occur in the mean over western boundary currents (e.g., [41]), possibly acting as a vorticity forcing at upper levels [42]. The narrowness of the heating field also strongly impacts the covariance T′Q′ (where Q′ is the eddy diabatic heating) which is a source term of eddy potential energy in the Lorenz energy cycle [16, 35]. In their simulation, Willison et al. [35] showed that this term increased by ∼ 60% in the storm track region between low res and high res.

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[3] Note also that the eddy anisotropy is different for low and high frequency eddies. This is very important because derivatives of v′²-u′² dominate in the feedback on stationary waves (u′ and v′ denoting the zonal and meridional eddy velocity, respectively).
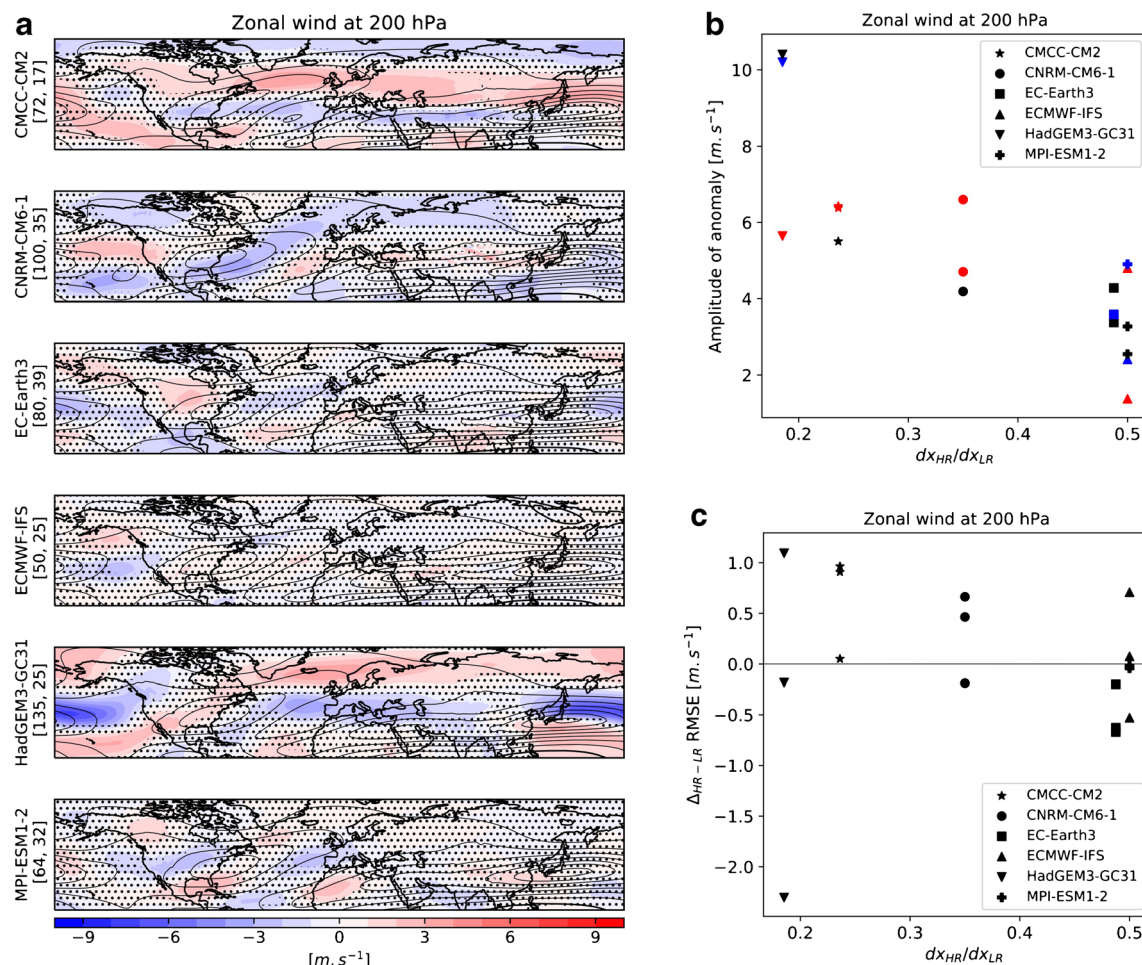
Fig. 1 **a** Difference in long time mean zonal wind between the high and low resolution (color) of six AGCMs forced with the same time varying SST. The high-res climatology is superimposed in black contours (ci = 10 m/s). Dots indicate where the change is not significant at the 95% confidence level. **b** Regional changes in jet strength (in m/s) and position as a function of the increase in resolution (i.e., the ratio of gridsize hi-res/low-res) for the Atlantic, Western, and Eastern Pacific sector (i.e., 3 points per model). Blue indicates equatorward shift, red poleward shift, and black no

shift. **c** Change in the rms error (in m/s) computed, for each basin, from a comparison with the ERA5 climatology. The approximate resolution of the models at 50 N is given in bracket (in km) for their high- and low-res versions in panel (**a**). All fields are for wintertime mean at 200 hPa over 1950–2014 except for (**c**) where the 1980–2014 mean is used to allow comparison with ERA5. More information about the simulations can be found in [38]

The conditional symmetric instability (CSI) is a type of moist instability which has received particular attention as it is the only instability occurring in the mesoscale range in the midlatitudes[4] [43–46]. It develops in regions of strong temperature gradient and large windshear, such as fronts. The occurrence of CSI can give rise to deep cross-frontal slantwise convection. This situation is difficult to observe as frontal regions are often close to neutrality[5] and the induced motion is difficult to distinguish from circulation induced by frontogenesis (see the review by Schultz and Schumacher [47] for a full discussion). Nevertheless, the slantwise moist instability

or neutrality means that the existing frontal circulation can intensify rapidly. Czaja and Blunt [48] used the moist Richardson number (Ri) as an indicator of CSI and found that situations favorable to slantwise convection (i.e., 0 < Ri < 1) were three times more frequent than for upright convection (Ri = 0) over the Gulf Stream. Similar qualitative results were obtained by Glinton et al. [49] in their analysis of slantwise and standard convective available potential energy (SCAPE and CAPE, respectively) over the North Atlantic. Resolution of at least 25 km has been suggested to represent CSI but increasing resolution allowed to simulate additional rain bands ([50]—the importance of vertical resolution must also be mentioned, even if only just briefly here, as the aspect ratio of the motion becomes closer to unity in frontal circulations). There have been several attempts at parameterizing slantwise convection (e.g., [51, 52]) which have been shown to produce a

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[4] CSI does not select a spatial scale but it occurs on the scale of the frontal updraft.

[5] That is, parcels ascending in a frontal region have a buoyancy close to that of the surrounding air.

more realistic organization of line convection and precipitation [53].

In summary, the eddy feedback on the mean flow, poleward heat fluxes, and frontal structures have all shown to be strongly affected by an increase in horizontal resolution in recent studies. Although we have emphasized positive results brought about by the increase in resolution, it must be kept in mind that increasing resolution can also sometimes degrade the performance of not only an AGCM (as Fig. 1c suggests) but also a coupled system. For example, although the GloSea5 forecasting system of the UK Met Office showed significant predictive skill of the NAO with the atmospheric component run at 60 km [7], that skill was found to decrease significantly (a factor of two in terms of NAO index correlation) when run at 25 km [54].

How Do the Extra-Tropical Oceans Affect the Storm Track/Jet Stream System?

Large-Scale Oceanic Forcing

The forcing of the atmosphere by the extra-tropical ocean on the scale of the atmospheric deformation radius (~ 1000 km) was reviewed by Kushnir et al. [2]. This study identified two mechanisms, the "linear baroclinic response" and the "storm track response." The first, according to the scale analysis of [55], is expected to cause descent over warm SST anomalies, as well as a surface low/upper level high geopotential height anomaly. This prediction is not expected to hold when the total horizontal wave number ($k^2 + l^2$ where k and l are the zonal and meridional wavenumbers, respectively) is large, which can occur either for oceanic fronts (one of k or l is large) or eddies (both k and l are large). If the linear baroclinic response causes upper level divergence or convergence, further perturbations can propagate to other regions via waves, causing remote responses to SST anomalies. A recent example is provided by Sato et al. [56] who suggested that the meridional shifts of the Gulf Stream cause linear Rossby wave responses in early winter over Europe, based on experiments with a linearized atmospheric model.

In practice, the ultimate fate of the linear baroclinic response is mediated by the interaction between the mean circulation and the eddies [57], and this is the second mechanism (storm track response) discussed by Kushnir et al. [2]. Often, but not in all AGCM studies, the upper level geopotential height anomaly is reinforced by this interaction, while the surface low weakens, leading, at equilibrium, to a column of the same sign geopotential height anomalies increasing with height in magnitude. Since that review was published, several studies have highlighted the role that mean SST structures play in shaping the North Atlantic eddy-driven jet [58] and blocking [59]. The latter two studies analyzed 20-year experiments using observed SSTs as well as smoothed SSTs in an AGCM, but different smoothing experiments using different AGCMs also reported storm track changes [60, 61]. It is important to emphasize that the prevailing paradigm is that the storm track response is well captured by dry quasi-geostrophic dynamics (hereafter QG, a theoretical framework adequate for motions of spatial scale comparable with the atmospheric deformation radius), since even very low–resolution QG models have been able to reproduce the salient features of the eddy mean flow feedback generated by low-res AGCMs [62, 63]. Nevertheless, as briefly reviewed in the "What Are the Atmospheric Motions and the Climatic Features Which Improve When Increasing an AGCM Resolution From ~ 100 to ~ 10 km in Midlatitudes?" section, more recent analyses of AGCMs show an important role for eddies whose dynamics is not captured by QG dynamics (the very high frequency eddies discussed in [34]), as well as an important role for latent heat release. Thus, this paradigm might need to be revisited. This is further underscored by the strong non-linearity seen in recent experiments (e.g., [64, 65]), while the QG paradigm in [62, 63] relies on linear storm track dynamics.

Another important line of research since the review by Kushnir et al. [2] has been to suggest that SST gradients are instrumental in setting cyclogenesis through diabatic effects in the lower troposphere [66–69]. Nevertheless, and although these studies have been phrased in the context of "oceanic frontal zones," the scales considered are comparable with the atmospheric deformation radius and so these studies may still be of too coarse resolution to represent oceanic fronts accurately. For example, in the study by Nakamura et al. [67], and more recently in [70], the "frontal" region is represented by 6 (8) grid points at 150 (180) km resolution, i.e., a width of 900 (1440) km. The impact of low-level baroclinicity[6] on cyclogenesis is typically evaluated using the Eady growth rate [71]. Piazza et al [60] (see also [72] in the context of sensitivity to ocean model resolution) showed that SST smoothing and non-smoothing experiments for the North Atlantic exhibited a substantial difference in this quantity associated with the horizontal air-temperature gradient, whereas Kuwano-Yoshida and Minobe [73] showed that it does not change much in their smoothing experiment over the Kuroshio region because of the compensating effect between the horizontal (destabilizing) and vertical (stabilizing) temperature gradients.

Surface latent heat fluxes and latent heat release can also contribute to the maintaining of atmospheric baroclinicity and the localization of storm tracks [71, 74]. A series of original studies by Ma et al. [15, 16] recently argued that the oceanic mesoscale eddy field modulates the amount of heat and moisture fluxed into the marine boundary layer as a little bit more heat is given to the atmosphere over a warm ocean eddy than

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[6] Baroclinicity is taken here to mean horizontal temperature gradients.

is lost over a cold ocean eddy (this effect involves both the non-linearity of the Clausius-Clapeyron relationship and the presence of enhanced winds over warmer waters—see the review by Small et al. [11]). Although this effect involves the oceanic mesoscale (10–100 km), it is spatially averaged over the larger scale of the cold and warm sectors of cyclones and thus does not represent a direct oceanic forcing on the scale of the oceanic eddies. Rather, it acts to shift the oceanic sources of heat and moisture a little bit poleward as ocean eddies typically develop over a region separating warm (further equatorward) and cold waters (further poleward). Ma et al. [16] showed in a LAM with a 27-km resolution that this induced a poleward shift of the North Pacific storm track in response but, interestingly, this did not happen when the resolution was decreased to ~ 160 km as the model was then not able to "see" the oceanic mesoscale. This striking result is further illustrated in Fig. 2. In an idealized experiment, Foussard et al. [19] were able to simulate the poleward shift of the storm track by simply adding a large-scale heat and moisture source centered on an oceanic front rather than explicitly representing the oceanic eddies developing on that front. An alternative explanation of these results is provided by Tamarin and Kaspi [75] who showed how increasing latent heating in cyclones, as a result of a stronger surface heat

source over the ocean, can also contribute to poleward migration of the storm tracks.

Small-Scale Oceanic Forcing

The forcing of the storm track/Jet Stream system at the scale of the oceanic deformation radius (10–100 km in midlatitudes) is only beginning to be investigated. It is well known from earlier studies that SST features on these scales are important to simulate the structure and movement of cyclones (e.g., [76, 77]), their deepening rate (e.g., [78, 79]), and maybe even the frequency of lightning [80, 81]. Nevertheless, the computational challenge of scaling up these results (weather) to long global simulations (climate), or ensemble of these, is very significant [82, 83] and only begins to be addressed in atmosphere-only set-ups (e.g., [84]).

A few stimulating ideas linking weather and climate are worth mentioning. First, the midlatitude ocean and its energetic mesoscale structures affect the rainfall distribution because the latter largely reflects the strength and movement of weather fronts [85] whose spatial scales match those of the oceanic mesoscale. The basic idea is illustrated in Fig. 3, which displays various possible configurations of oceanic and atmospheric temperature fronts. In some cases (Fig. 3b), warm
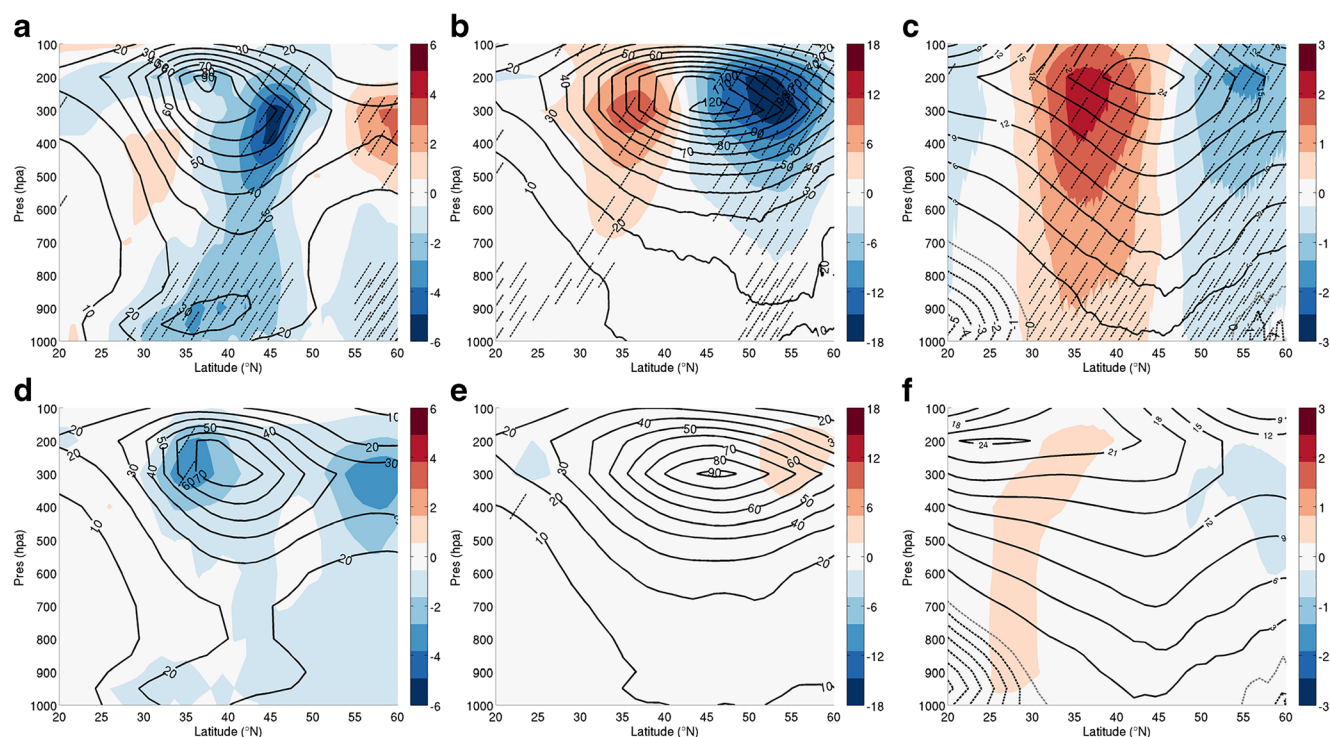

Fig. 2 **a** Vertical and latitudinal profile of winter mean (contours) storm track (v'^2 where v' is eddy meridional velocity) in the Western Pacific in a control simulation (CTRL) with realistic SSTs and 27-km resolution, and change in that quantity (color, in m^2/s^2) in an experiment where mesoscale SST features were removed (FILT), i.e., FILT-CTRL is plotted. **b** Same as (**a**) but for the Eastern Pacific. **c** Same as (**b**) but for the zonally averaged wind instead of v'^2, in units of m/s. Panels (**d**), (**e**), and (**f**) are the exact counterparts to panels (**a**), (**b**), and (**c**) but in a simulation where the horizontal resolution was degraded to 162 km. Note that in this case, no significant signals (hatched regions) are seen in the FILT-CTRL plots. Reproduced from [16]. ©American Meteorological Society. Used with permission
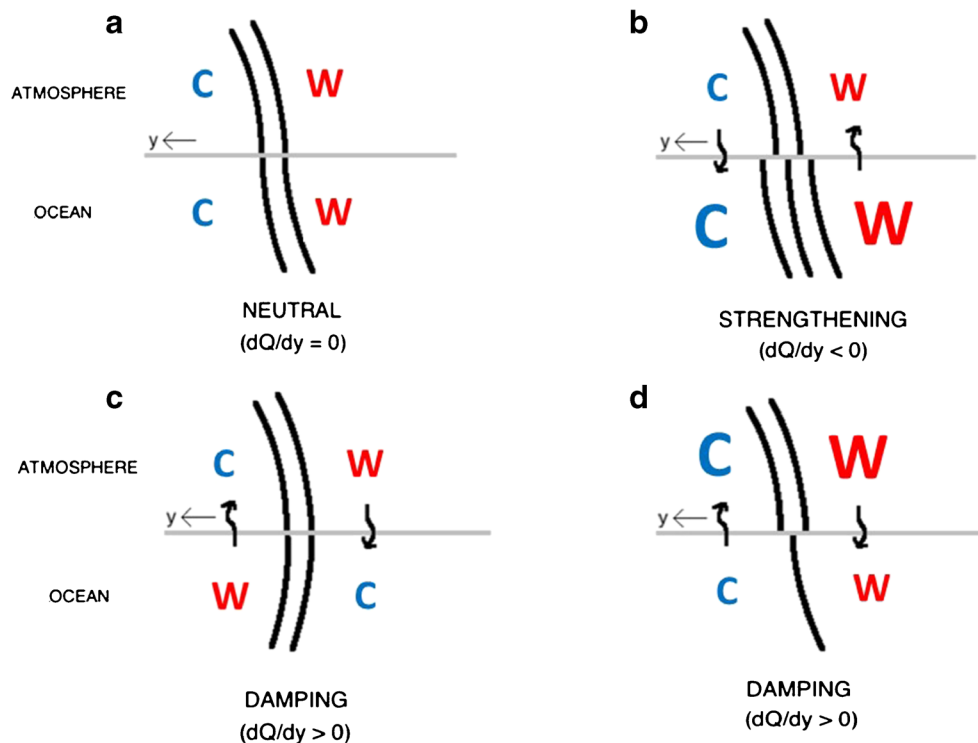
Fig. 3 Thermal damping and strengthening of atmospheric fronts. The relative orientation of oceanic and atmospheric fronts is indicated by the position of isentropic surfaces (black lines), with the warm and cold regions highlighted by the letters W and C respectively. The sensible surface heat flux Q, positive downward, is indicated by wavy arrows. Warm air over warm water and cold air over cold water in panel (**a**) is a thermally equilibrated situation ($Q = 0$). This contrasts with panel (**b**) where very warm (cold) ocean waters are found over warm (cold) air masses, leading to a horizontal gradient in Q and a strengthening of the frontal circulation (thermally direct, i.e., with warm air rising and cold air sinking). Panels (**c**) and (**d**) illustrate alternative situations where the horizontal heat flux gradient is reversed and the circulation weakened. Adapted from [17]

(cold) air lies above even warmer (colder) ocean temperature, and the resulting air-sea heat flux acts to strengthen the frontal circulation and the associated rainfall. Conversely, when warm (cold) air lies above cold (warm) water, as in Fig. 3c, the frontal circulation is weakened by the horizontal gradient in air-sea heat flux. Parfitt et al. [86] showed that these arguments could explain how the increase in SST resolution used in the ERA-interim reanalysis after 2001 caused an increase in precipitation by up to 30% in the North Atlantic in wintertime.

Another line of thought is that the more zonal or more tilted position of the Jet Stream is affected by the number of low potential vorticity (PV)[7] air masses brought to upper levels in the warm conveyor belt of cyclones. It was shown by Pfahl et al. [87] that 30 to 45% of air masses involved in blocking episodes in the North Atlantic underwent prior diabatic modification of their PV as they ascended from lower to upper levels of the troposphere. Here again oceanic mesoscale structures are likely key players in setting the production of these low PV air masses through air-sea interactions in the cold

sector of cyclones [69], and possibly in enhancing the ascent from low levels through a form of CSI (the "What Are the Atmospheric Motions and the Climatic Features Which Improve When Increasing an AGCM Resolution From ~ 100 to ~ 10 km in Midlatitudes?" section; [18]).

Finally, we note that the energy threshold emphasized in [34] and briefly discussed in the "What Are the Atmospheric Motions and the Climatic Features Which Improve When Increasing an AGCM Resolution From ~ 100 to ~ 10 km in Midlatitudes?" section in relation to momentum fluxes by very high frequency eddies is also likely to be affected by air-sea interactions on the scale of the oceanic deformation radius. Indeed, Ma et al. [16] showed that the presence of mesoscale oceanic features in SST enhances the production of eddy potential energy (the T′Q′ term introduced in the "What Are the Atmospheric Motions and the Climatic Features Which Improve When Increasing an AGCM Resolution From ~ 100 to ~ 10 km in Midlatitudes?" section), as well as the conversion of that energy to eddy kinetic energy (the w′T′ term in the Lorenz cycle, where w′ is perturbation vertical velocity). These two processes contribute constructively to overall enhance the kinetic energy of the eddies and thus, from the arguments in [34], to alter the momentum fluxes by synoptic waves.

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[7] PV is a scalar which contains dynamical and thermodynamical information about the atmosphere. Low values of PV are typically associated with anticyclonic flows and are produced above a region of heating.

Observational Constraints on the Extra-Tropical Oceanic Forcing

Observational Estimates of the Response of the Atmosphere to SST Anomalies

The atmospheric response to extra-tropical SST anomalies has been estimated from their relation to atmospheric fields lagging by more than atmospheric persistence, but less than SST persistence. The large-scale SST anomaly modes are fairly well observed since the twentieth century, and their impact has been extensively investigated since earlier studies by Ratcliffe and Murray [88], Palmer and Sun [89], and Czaja and Frankignoul [90, 91]. On the other hand, high SST resolution is needed to resolve SST anomalies associated with oceanic frontal variability and mesoscale eddies. A (nominal) ¼° resolution SST product is available since 1982 [92], but higher resolution only since 2005. Likewise, satellite observations of sea level only allow to link these features to ocean dynamics from 1992 onwards. Limited SST resolution is unlikely to substantially degrade the large-scale response to oceanic frontal variability in atmospheric reanalyses, as the tropospheric state is strongly constrained by observations, but the marine atmospheric boundary layer is poorly constrained, and local frontal impacts, including surface heat flux, surface-wind convergence, and cloudiness, are only well represented if the SST resolution is very high [93].

In the North Pacific, recent studies have focused on the cold season atmospheric response to the variability of the western boundary current extensions. Using the maximum monthly meridional SST gradient to locate the Oyashio Extension (OE) front in ¼° SST data, Frankignoul et al. [20] found that a northward shift of the OE was followed by a positive phase of the North Pacific Oscillation (NPO) with a northern lobe amplitude reaching − 35 m at 250 hPa (SLP of − 2 hPa) for a 0.45-K SST anomaly, stronger in fall and winter and when the front is displaced southward. Using a 1° resolution SST product to slightly extend the analysis period, Révelard et al. [22] found a smaller NPO response (-30 m/0.8 K) during fall, but a large one in late winter (− 60 to 80 m at 250 hPa, SLP of − 5.5 hPa/0.5 K). The corresponding SST and geopotential height anomaly at 300 hPa are shown in Fig. 4. There is some uncertainty in the estimated magnitudes, a point to which we return below, as they depend on an assumed 2-month response time to SST forcing (for monthly averages to reach full amplitude). This choice is roughly consistent with the 20–30 days needed to reach maximum amplitude in the observational analysis of [94, 95] and in high-resolution simulations [14, 64]. It also agrees with the 2-month response time for monthly averages in AGCM studies (e.g., [14, 96]), and the 2 to 3-month lag seen in the maximum co-variability between SST and large-scale atmospheric anomalies. To estimate the response to the OE fluctuations

in 1959–2006, Taguchi et al. [21] used the SST averaged in the subarctic frontal zone as OE index. They found that a frontal zone warming in November was followed by a strong strengthening of the Aleutian low (− 80 m, − 4 hPa/K) in January, but a smaller opposite response in February.

The Kuroshio Extension (KE) variability also influences the atmosphere in the cold season by modulating local ascending motions and shifting the zone of maximum baroclinicity (see the "How Do the Extra-Tropical Oceans Affect the Storm Track/Jet Stream System?" section), but in an asymmetric way, as the estimated response reaches 70 m/0.7 K at 250 hPa when the KE is in a stable (northerly) state, but is not significant in the unstable (southerly) state [97]. Wills and Thompson [95] used daily winter data and showed that a warm SST in the Kuroshio-Oyashio extension (KOE) region precedes a surface intensified low slightly downstream, which only reaches about − 12m at 1000 hPa/0.7 K after 30 days. Further empirical evidence for a direct relationship between the Kuroshio's large meander south of Japan and storm tracks and regional precipitation was obtained by Nakamura et al. [98].

In the North Atlantic, earlier analyses showed that a basin-scale "horseshoe" SST anomaly is followed in late fall/early winter by a large NAO-like signal [90, 91], primarily resulting from eastern subpolar SST forcing [99]. By explicitly modeling the SST anomaly decay and assuming again a 2-month response time, Frankignoul et al. [100] found that the NAO amplitude reached 45 m/0.45 K at 500 hPa, broadly consistent with [22].

Several studies investigated the influence of the latitudinal variability of the Gulf Stream (GS) front. An index with seasonal temporal resolution of the position of the GS north wall in 1979–2009 was used to show that a southward excursion of the GS front was followed a year later in winter by a storm track extension to the east [101] and, for small excursions, a negative NAO-like signal, with indication (in view of the limited sample) of large amplitude and strong non-linearity when the GS shifts far from its climatological value [65]. However, based on the same GS index, Joyce et al. [102] found that a southward GS shift leads by only 3 months enhanced Greenland blocking and a southward shift of the storm track over the Labrador Sea. They speculated that these changes may be linked to broader changes in the Atlantic Meridional Overturning Circulation (although synchronous SST variations were not explored) and also noted that the lag correlation was very different before the 1970s. It is noteworthy that Minobe and Maeda [103] suggested that the Gulf Stream migrates on multidecadal timescale in accordance with the Atlantic Multidecadal Oscillation (their Fig. 7). Wills et al. [94] used daily resolution during winter 1979–2013 to estimate the atmospheric response to the GS variability, as represented by the averaged SST in a broad GS region, finding that a low above the GS and a barotropic wavetrain extending

toward Europe and western Russia lags a positive SST anomaly by several weeks. The response seems primarily driven by anomalous vertical motion northeast of the GS path, but its amplitude is limited (about 23 m/K at 500 hPa).

In summary, large-scale SST anomalies and western boundary current frontal variability seem to drive a strong atmospheric response in the cold season, with evidence of asymmetry and nonstationarity. Larger atmospheric signals are not expected in higher resolution reanalyses, which are strongly constrained by observations, but higher oceanic resolution might lead to a more accurate assessment of their link with frontal variability and oceanic mesoscale structures.

Comparison With the Simulated Response of AGCMs to Prescribed Extra-Tropical SST Anomalies

Early response studies with low-resolution AGCMs found small responses to (broad) extra-tropical SST anomalies, on the order of 20 m per K at 500 hPa [2]. Similarly, Peings and Magnusdottir [104] found at 1° resolution a limited response to the Atlantic Multidecadal Oscillation (about 20 m at 500 hPa/1 K) and [105] a small response to SST in a broad region along the subarctic frontal zone (40 m at 300 hPa/3–4 K). Smirnov et al. [14] found that the (symmetric) cold season response to a meridional shift of the OE was much larger at 1/4° than at 1° resolution, but still smaller in magnitude and different in spatial structure compared with the NPO response estimated from observations and discussed in the "Observational Estimates of the Response of the Atmosphere to SST Anomalies" section. In an

attempt to provide a direct comparison between simulated and "observed" responses, we have focused on the OE region for which it was possible to find observational and modeling studies using consistent SST indices (in all the studies used in this figure—see the figure caption for details—the spatial pattern of the SST anomaly is qualitatively similar to that seen in Fig. 4). For compactness, the responses at upper levels are expressed in m/K and only the magnitude of the responses are compared in Fig. 5, not the sign or the actual pattern (the magnitude of the anomalies in geopotential height and SST are given separately in the figure caption). One first notices in this figure that the "observed" responses are larger than the simulated ones as all the points fall above the $y = x$ line (black dashed line). The high-resolution study by Smirnov et al. [14] shows the highest response magnitude but the latter is still about a factor of two smaller than the observational estimate by Frankignoul et al. [20]. The other interesting feature of this figure is the wide spread in the "observed" responses, from about 20 to 100 m/K. This emphasizes that although observational studies assess the statistical significance of a lagged relationship between SST and atmospheric variables, as well as, in more recent studies, the local significance of an atmospheric response (see the black lines in Fig. 4, right column), none to our knowledge has attempted to estimate an errorbar on the magnitude of the atmospheric response (nor have we in Fig. 5).

It must also be emphasized that while we have highlighted that smaller magnitudes are generally obtained with AGCMs compared with observational estimates, a few studies have reported more comparable responses. For example, a strongly
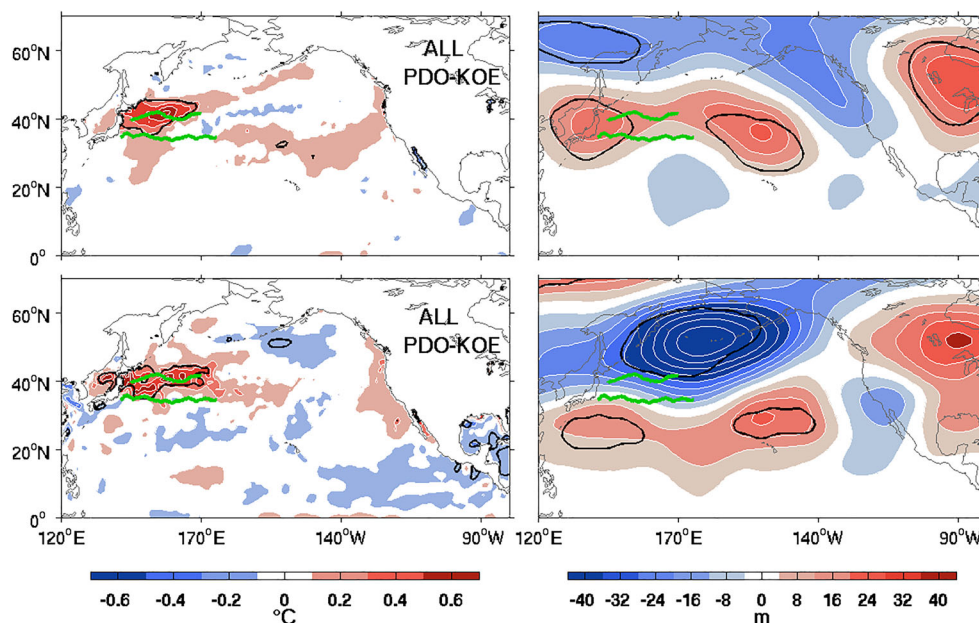

Fig. 4 The estimated response of the atmosphere (right column, geopotential height anomaly at 300 hPa, in m, CI = 8 m) to a northward shift of the OE (left column, SST anomaly in K, CI = 0.2 K) in the ERA-interim reanalysis (1979–2014). The early (October–November–December, top row) and late (February–March–April, bottom row) winter atmospheric responses are shown (the SST anomaly is shown 2 months earlier). The green curves (left panels) indicate the mean position of the OE and KE fronts and the black contours in all panels indicate the 10% significance level. Figure adapted from [22]. ©American Meteorological Society. Used with permission
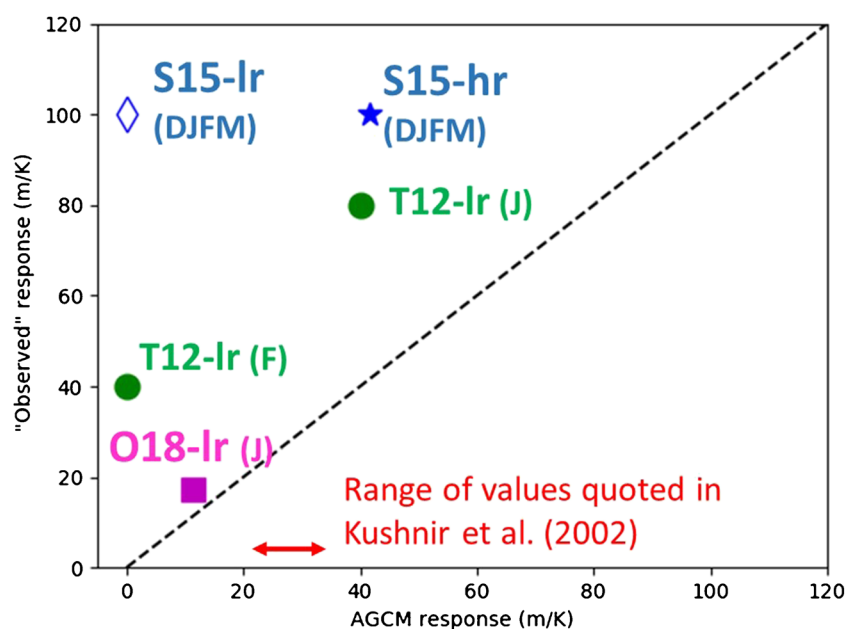
Fig. 5 A comparison of "observed" and simulated atmospheric response to a northward shift of the Oyashio extension. The responses are for the winter season and upper levels (the letter in parenthesis indicates the GCM month considered) and, for compactness, are expressed as geopotential height anomaly per degree SST anomaly (m/K). The studies were chosen as follows: S15 at 300 hPa ([14]; 50 m/1.2 K and 0 m/1.2 K for high and low resolution, respectively) compared with [20] at 250 hPa (45 m/0.45 K to account for seasonal enhancement); T12 [21] compared with itself as it presented both observed and modeled responses at 250 hPa (60 m/0.75 K in January, 30 m/0.75 K in February for the observations, and 40 m/1 K and 0 m/1 K for the coupled model in January and February respectively); O18 ([105], 40 m/3.5 K) compared at 300 hPa with [95] 200 hPa (12 m/0.7 K). The subscript "lr" and "hr" indicate low- and high-resolution (i.e., horizontal grid size smaller or larger than 100 km). The range of model responses quoted in [2] is given as a reference in red

asymmetric and, in addition, nonlinear remote response to a meridional GS shift was found by Seo et al. [65] with a regional model at 40-km resolution, showing a high sensitivity to SST forcing (in the weakest SST anomaly case considered, the response exceeds 20 m at 250 hPa for a strikingly small 0.4/9 = 0.04 K SST anomaly—see their Fig. 4i). In rather high-resolution (T213) simulations, Zhou et al. [106] found a large winter response to the extra-tropical PDO, reaching about − 75 m at 500 hPa (− 6.6 hPa for SLP)/0.7 K SST, when the daily global SST variability during 1981–1990 was included in the background, slightly smaller amplitude when the daily SSTs were only prescribed in the North Pacific, and no significant response in their absence or at lower resolution, suggesting that the small SST scales enhance the response to SST anomalies. However, the response was very weak with the 1991–2000 SST background, emphasizing its nonstationarity [107].

There is very little theoretical guidance beyond the arguments summarized in the "How Do the Extra-Tropical Oceans Affect the Storm Track/Jet Stream System?" section to explain this diversity of results. One attractive paradigm, reminiscent of earlier ideas by Shutts [108] and Peng and Robinson [109], is that as the resolution of an AGCM increases, the diabatic heating/cooling generated by an SST anomaly and its interaction with weather systems becomes nearly balanced by eddy heat transports (lateral and vertical). This was, for example, seen in the study by Smirnov et al. [14]

when moving from low to high resolution. In this limit, the response of the large-scale flow to an SST anomaly becomes closer to the "thermally equilibrated" mode of the AGCM, as the apparent heat source balancing the large-scale flow nearly vanishes. This dynamics leads to a weak anomalous surface heat flux (F′) and circulation over the SST anomaly because of the nature of the thermally equilibrated mode. In the opposite limit, possibly more likely to occur in a low-res AGCM, eddies come less efficiently into equilibrium with the diabatic heating and stronger F′ and surface circulation occur.

Alternatively, it might be that the differences in AGCMs' responses reflect the different history of natural variability experienced by the model atmospheres while being forced by the same SST anomaly. The study by Deser et al. [110] clearly illustrates how even ~ 100-year-long records do not fully constrain the response of the extra-tropical atmosphere to ENSO events owing to its large natural variability. It is possible that a similar effect is at work here, different AGCMs sampling preferentially different regions of the atmospheric phase space, even though their time mean might be in agreement with observations. If this view is correct, the nonlinearity and nonstationarity of some of the responses reported above should be taken with caution. As should then the more basic results from studies where the sampling of the atmospheric phase space is limited (e.g., only one winter in [16]).

Implications for Climate Change Predictions

The response of the midlatitude atmospheric circulation to anthropogenic forcing over the twenty-first century has been traditionally described as a tug of war between enhanced equator-to-pole temperature gradient at upper levels and reduced gradient at lower levels in the Northern Hemisphere (e.g., [111, 112]). Conversely, in the Southern Hemisphere, it has been viewed within the context of a competition between the impact of CO_2 forcing vs stratospheric ozone recovery (e.g., [113]). In addition, the global increase in moisture content of the atmosphere with increasing surface temperature results in a larger role of the moist dynamics discussed in the "How Do the Extra-Tropical Oceans Affect the Storm Track/ Jet Stream System?" section (e.g., [114]). In either Hemisphere, the role of the midlatitude oceans has only so far been discussed and demonstrated in relation to large-scale aspects of the oceanic circulation: retarding surface warming in the Southern Ocean (as a result of upwelling of cold deep water, e.g., [111, 115]) and weakening of the upper circulation and associated poleward heat transport in the North Atlantic (e.g., [116]).

Analysis of instrumental records over the twentieth century has suggested an enhanced surface warming over all western boundary currents compared with the global mean ocean [117]. This has been observed in AR5 models too, except for the Gulf Stream where models predict a weakening of the upper circulation [118, 119]. It is not clear at this stage if this pattern of future warming is robust, especially considering the large biases in the representation of the ocean circulation, western boundary currents, and their separation from the coast in coarse climate models. Indeed, recent studies [120, 121] have shown that very different SST responses develop in the Atlantic in response to increasing CO_2 concentrations as oceanic resolution increases. A similar behavior is suggested for the western North Pacific in Fig. 6.

Could these changes compete with the above-mentioned effects of equator-to-pole temperature gradients on the Jet Stream in the twenty-first century? A direct investigation of the impact of these SST changes in current high-resolution coupled simulations has not yet, to our knowledge, been conducted. Nevertheless, two lines of research offer some perspective in the North Atlantic basin. First, serendipitously, a proxy for the expected warming of the western boundary currents in this region is provided by the SST bias produced in most coupled models with oceanic resolution on the order of 25 km or coarser as a result of an inaccurate Gulf Stream separation[8]. The impact of this bias has been recently studied by Lee et al. [122] in the UK Met Office Unified Model during the cold season (about 60 km resolution in the atmosphere in midlatitudes). They showed that it induces a wavetrain across the Northern Hemisphere, with a positive geopotential height anomaly above the warm SST bias and a negative one over Iceland at 500 hPa (their Fig. 3e).

In addition, although the effect seen in [122] was rather weak (about 20 m at 500 hPa for a ~ 5 K SST anomaly), it is consistent with independent findings reported in the context of the observed and simulated decadal climate variability seen in the North Atlantic in wintertime. Indeed, both Sato et al. [56] and Siqueira and Kirtman [123] have shown a qualitatively similar wavetrain emanating from the Gulf Stream region in response to a surface warming on the North flank of the separated Gulf Stream. The anomalies suggested in the latter study are very significant as they are low-pass filtered (decadal timescale) and reach a magnitude of 30 m at 200 hPa for a 1-K SST anomaly (their Fig. 3b, e). It is also significant that such signals were simulated in the high res, but not the low res, version of the coupled model used in their study. The impact of warmer SST anomalies north of the separated Gulf Stream on the strength of the Icelandic Low, although with a different sign, was highlighted in earlier studies of decadal variability in the Atlantic in long instrumental records (e.g., [124, 125])[9]. The dynamics is certainly complex, as studies using high-resolution reanalysis dataset (ERA5, with 25 km resolution) have emphasized the presence of tropospheric mesoscale signals set by the Gulf Stream warm core [127], in addition to the non-linear effects suggested in some LAM studies (e.g., [65], see the "Comparison With the Simulated Response of AGCMs to Prescribed Extra-Tropical SST Anomalies" section).

In summary, uncertainties associated with the magnitude and structure of the atmospheric response in AGCMs (the "Observational Constraints on the Extra-Tropical Oceanic Forcing" section) warrant caution but the suggestion from recent studies with coupled models and AGCMs with an horizontal resolution on the order of 50 km, as well as the decadal variability seen in the instrumental records, is that midlatitude air-sea interactions associated with a warming of the western boundary currents will be an important factor setting the pattern of Jet Stream/storm track change in the twenty-first century.

Conclusion

This review has provided a snapshot of the state of the research at a very productive period, as many high-res simulations prepared for the 6th assessment report of the IPCC are beginning to be analyzed [128]. A whole spectrum of studies

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[8] A possible caveat here is that in the high-res experiments mentioned in the previous paragraph, not only is the ocean near the East coast of the US warming, but so does the land [121]. This does not occur in [122].

[9] A similar impact of decadal changes in the Kuroshio extension on the observed Jet Stream has been suggested by [126] in the North Pacific.
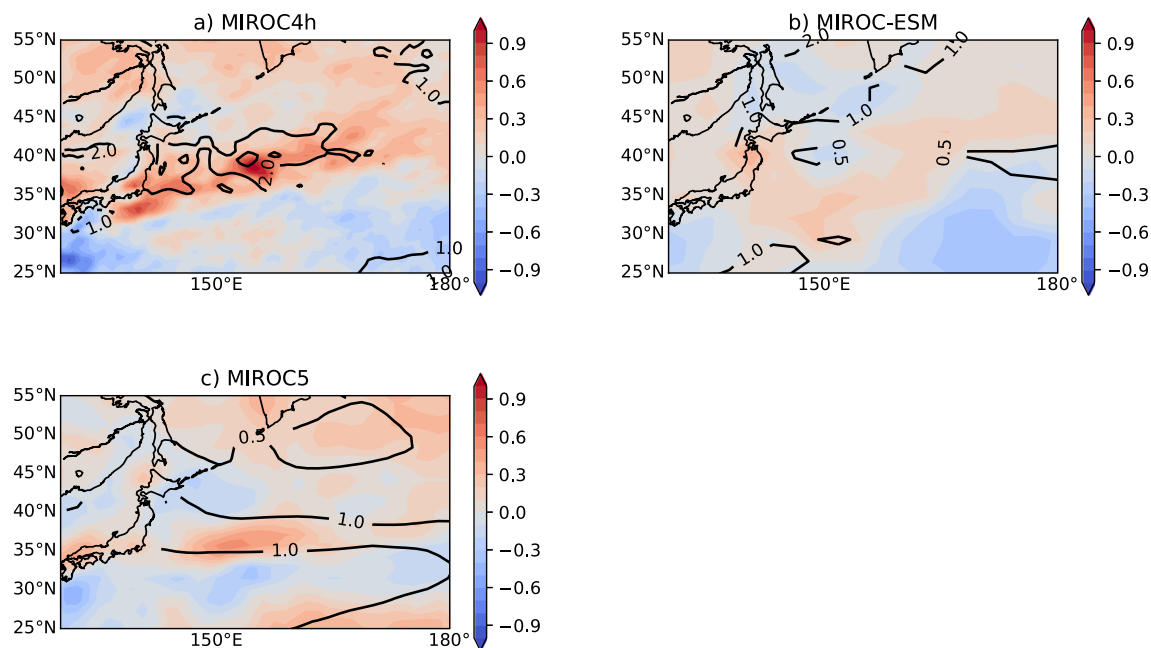
Fig. 6 Rain rate difference (in units of mm/day, color) between the first and the last 20 years (2016–2035 minus 1969–1950) of the historical-future (RCP4.5) simulation of MIROC4h (**a**), MIROC-ESM (**b**), and MIROC5 (**c**), along with SST differences in contours (in K). One observes a more localized and stronger SST and rainfall change in the higher resolution model (MIROC4h, 60 km atmosphere and eddy permitting ocean) compared with the other two (atmospheric resolution of 1.4° and 2.8° for MIROC5 and MIROC5-ESM, respectively). Figure courtesy of S. Minobe

has been discussed. Some emphasize the role of atmospheric resolution in setting different dynamical responses to large-scale SST anomalies and in hopefully converging to an accurate response of the atmosphere to an oceanic forcing on the scale of the atmospheric deformation radius (∼ 1000 km). Some studies, on the other hand, emphasize the impact of smaller spatial scales in the SST field on extra-tropical cyclones and, through as yet not fully understood mechanisms, on the storm track/Jet Stream system. This, we distinguished from the former forcing, as occurring on the scale of the oceanic deformation radius (10–100 km). We now tentatively attempt to answer the question raised in the title of this review.

First, we expect that further analysis of the new generation of coupled models will confirm the presence of stronger atmospheric responses to changes in extra-tropical SST anomalies, more in line with observational estimates. In addition, we expect that more variability simulated for the twenty-first century will be initiated at the oceanic deformation radius scale (10–100 km) where oceanic motions in the form of mesoscale eddies, quasi-permanent fronts, meandering, and shifts of the western boundary currents are strongest. Early studies investigating the response of the upper ocean circulation and SST field to increasing greenhouse gas concentrations have already demonstrated a large sensitivity to resolution, with previously unseen signals emerging in the highest resolution coupled models (the "Implications for Climate Change Predictions" section). Thus, we expect more surprises coming from the extra-tropical oceans to be unveiled, in addition to the

predicted weakening of the thermohaline circulation of the North Atlantic.

While the recent research provides more optimism for a source of predictability from the ocean than was available at the time of the review by Kushnir et al. [2], several outstanding issues remain:

Robustness Across Models The review by Kushnir et al. [2] emphasized that, although the remote response of a given AGCM to an extra-tropical SST anomaly might be significant, one cannot expect it to be reproduced by another AGCM. The limited number of studies currently available is such that there is no evidence yet to support or invalidate this conclusion for high-resolution models. To solve this problem, coordinated experiments should be conducted using high-resolution models [129, 130]. A promising example is HighResMIP [128] in which historical and future AGCM as well as coupled GCM experiments are conducted both with high-resolution and low-resolution models. Thus, how atmospheric responses to the ocean depend on resolution can be investigated systematically across models for the first time. In particular, HighREsMIP's optional experiment using SST smoothing will provide an opportunity to examine the robustness of recently reported findings based on single model experiments at high resolution. We have reviewed several mechanisms and diagnostics which could help in understanding model differences (e.g., thermal equilibration and surface heat flux anomaly at equilibrium, atmospheric noise, thermal damping at

fronts and anomalous horizontal surface heat flux gradient, production of low PV air masses, eddy kinetic energy thresholds). Our suggestion is to encourage readers to analyze what aspects of midlatitude air-sea interactions are robust across models and what aspects are not, and to test these mechanisms and apply these diagnostics to understand why. Data from the HighResMIP project is already beginning to be available for such effort (e.g., see Fig. 1).

Model Set-Ups For future coordinated experiments, we would like to suggest three different lines of research:

1. Slab experiments (or 1D mixed layer models) in which the mean SST pattern is controlled but which would allow a feedback of the atmosphere on the ocean. The motivation is that the atmospheric response to SST anomalies might overestimate the coupling on short time scale in AMIP style experiments because of this missing feedback and also underestimate or misrepresent the long term oceanic forcing [131].
2. Convection permitting experiments as some mechanisms such as CSI need grid-spacing finer than 25 km and might suffer from the parameterization of convection. Only a limited area model is possible here due to computational constraints but the domain needs to be sufficiently large to minimize the impact of lateral boundary conditions on the response to SST anomaly. The protocol should impose a given domain, initial and boundary conditions. The limitation of those experiments is that they would not include the feedback on the atmospheric mean state.
3. Global high-resolution AGCMs run in idealized geometries, stepping the complexity up bit by bit (i.e., building on what has been done for low-resolution AGCMs by Brayshaw et al. [68]).

Observational Benchmarks Model responses need to be tested against some observational counterparts. Unfortunately, empirical estimates of the magnitude of the observed response of the atmosphere to extra-tropical SST anomalies currently lack errorbars. The example given in Fig. 5 for the OE highlights the spread expected from different statistical methods, choice of time lag, and the subjectivity in putting a number on maps like those displayed in Fig. 4. Here again a coordinated project aiming at comparing the various methods and at constructing an objective metric is very much encouraged and needed.

Regional Differences The different continental geometries of the Northern and Southern Hemispheres may lead to differences in the dominant mechanism(s) of oceanic influence. For example, the North Atlantic western boundary is perfectly shaped for land sea temperature contrast and Gulf Stream induced SST contrasts to add constructively during the cold season, highlighting the likely importance of frontal circulations in this problem. Such dynamics is less clear for the Kuroshio region because of the buffering effect of the Sea of Japan and also less clear for the Agulhas region in the Southern Ocean where land sea contrasts are much weaker. A thorough comparison of the atmospheric response with SST anomalies in these regions would likely be insightful and could be tackled both using observations and global AGCM at high resolution.

This review has only focused on the Northern Hemisphere during the cold season because this is where, historically, most of the efforts of the community have been concentrated. Likewise, we have omitted discussing radiative and cloud effects. This choice was made to limit and focus this review but more and more research suggests that important signals from the oceans emerge in the summer season (e.g., [132]), and that the radiative effect of clouds on the storm track/Jet Stream system is significant (e.g., [133]). Conversely, few studies have investigated the role of ocean-atmosphere interactions at higher Southern latitudes, despite much research on the dynamics of the Antarctic Circumpolar Current. It is not clear that the dynamics highlighted in the "How Do the Extra-Tropical Oceans Affect the Storm Track/Jet Stream System?" section will be relevant there, as SSTs are generally lower and diabatic heating effects in storms are much weaker there [134].

A recent surprise in ocean modeling might be worth mentioning as a final thought. In a study with a high-res ocean model (about 7-km resolution), Renault et al. [135] found that, surprisingly, rather than increasing the resolution further to resolve the submesoscale dynamics (~ 1 km), adding a more realistic representation of the air-sea exchange of momentum on the scale of the oceanic eddies (10–100 km) improved the representation of western boundary current dynamics very significantly. Strikingly, the "missing physics" is actually well enough understood [136] that it can be parameterized for ocean-only or coupled integrations with a low-res atmospheric model. We wonder whether such "parameterisable missing physical processes" exist in the context of the oceanic forcing of the storm track/Jet Stream system. And if they do, whether they might soon be revealed, thereby alleviating the challenging and unsatisfactory prospect of ever increasing model resolution[10]. The slanted nature of convection in midlatitudes, which is not represented at all in any low-res AGCMs[11] although it can be parameterized (e.g., [53]), comes to mind as it allows a nearly local isentropic connection between upper tropospheric levels and the underlying sea surface [137]. It is

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[10] Considering the very large number of simulations required to match the demands of large ensemble (to separate signals from noise), multiple scenarios (to account for policy decisions) and multiple models (to account for model uncertainties), this is not even a practical strategy.

[11] Because of weak frontal circulations and the fact that model physics tests only stability to vertical displacements of air parcels.

hoped that further work with the CMIP6 ensemble including HighResMIP [128] and the latest reanalysis datasets such as ERA5, as well as the development of new theories, will answer this question.

Compliance with Ethical Standards

Conflict of Interest On behalf of all authors, the corresponding author states that there is no conflict of interest.

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