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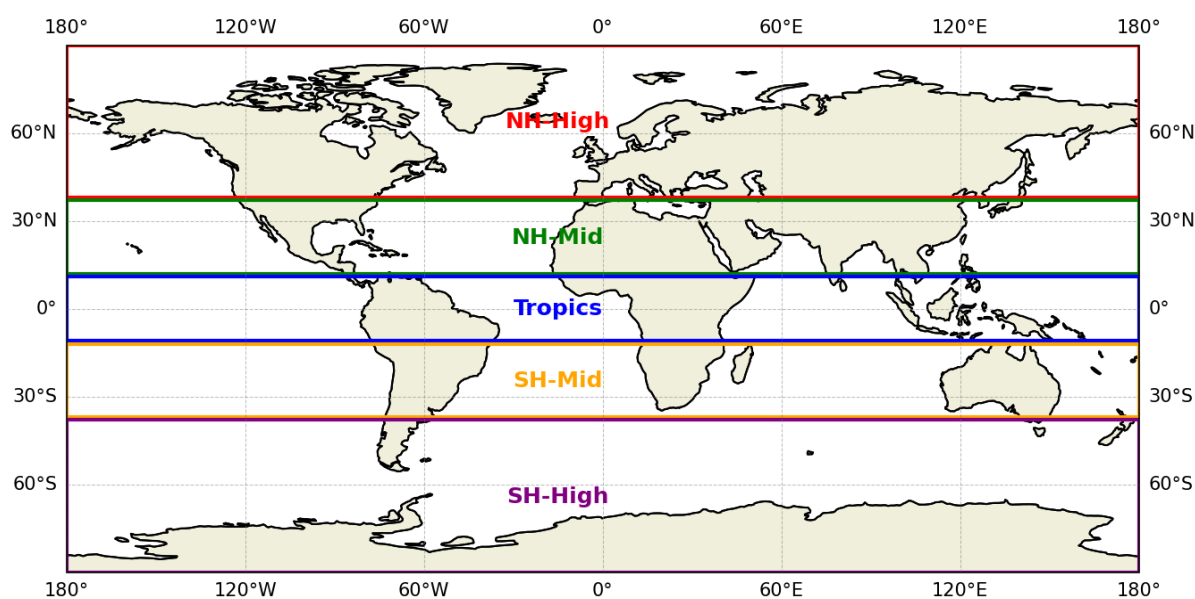
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Incorporating dynamic surface waves in NorESM

Alfatih Ali¹, Mats Bentsen², Thomas Toniazzo², Jens Debernard¹, Ana Carrasco¹, Clio Michel¹, Øyvind Breivik¹ and Thea Ellevold¹

Institutions: ¹ Norwegian Meteorological Institute; ² Norwegian Research Centre



Abstract

This report presents the integration of the prognostic wave model WAVEWATCH III (WW3) into the Norwegian Earth System Model (NorESM), as well as the wave coupling experiments performed within the framework of the MakingWaves project. Three wave-induced processes are investigated: (i) wave-modified Charnock parameter for wave-atmosphere coupling (ii) Langmuir Turbulence (LT) effects for wave-ocean coupling, and (iii) wave-ice radiation stress for wave-sea-ice coupling. The parametrization of LT mixing has been demonstrated to mainly deepen the ocean mixed layer (MLD) in both data-atmosphere-forced and fully coupled simulations, which in turn, reduces shallow MLD biases. The bias in the amplitude of sea surface temperature (SST) seasonal cycle in the southern hemisphere has significantly been reduced. Moreover, the SST exhibits cooling in summer and warming in winter, a pattern that is particularly prominent in the data-atmosphere forced case. However, in the active-atmosphere case, where atmospheric feedback is present, the average SST is generally found to be overall reduced, specially over the southern hemisphere. This reduction in SST is primarily attributed to the sea ice albedo feedback mechanism. The indirect LT mixing effects on sea ice are closely linked to those variations in SST. Coupling with WW3 involves a significant computational and development cost. To address this, a more efficient alternative is assessed, namely a wave-theory parameterization based on the 10m wind and an empirical wave spectrum. The LT mixing effects induced by this wave-theory parameterization are found to resemble those produced by the coupling with WW3. The impact due to wave-Charnock and wave-ice radiation stress on the model climatology is found to be minimal compared to the LT effects.



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1 Introduction

The goal of the MakingWaves project is to assess the effects of wind generated surface gravity waves on the global climate. To understand their climatological role through the analysis of simulations, we incorporated the wave model WAVEWATCH III (WW3) into the Norwegian Earth Model System (NorESM). NorESM is a derivative of the Community Earth System Model (CESM), with a different ocean model (Bergen Layered Ocean Model, BLOM), an interactive ocean biogeochemistry module (iHAMOCC) and modified atmosphere (CAM) and sea-ice (CICE6) components. The wave-ocean coupling consists of a Langmuir turbulence parametrization whereas wave-atmosphere coupling is in the form of the sea-state dependent Charnock parameter. Finally, the wave-ice coupling is done by including the wave-induced stresses.

Wave-atmosphere coupling has been explored mainly for forecasting intense storms. For instance, [Varlas et al. \[2018\]](#) applied Janssen’s modification of the wave-dependent Charnock parameter [[Janssen, 1991](#)] within the Weather Research and Forecasting model coupled with the wave model WAM to successfully predict a storm’s evolution over the Adriatic and Tyrrhenian Seas. They observed barometric weakening due to increased surface friction. The integration of WAM into ECMWF’s weather prediction system, which includes wave-field effects on momentum fluxes, has also led to reduced maximum storm intensities. Despite these advancements, extended impacts of wave-atmosphere coupling on timescales spanning months to years on the general atmospheric circulation remain to be addressed. In this study we analyze the impact of wave-atmospheric coupling, through wave-dependent Charnock parameter, on time scales of decades.

Wave-ice coupling has an evident but complex, two-way interaction. Waves can propagate into sea ice, impacting both dynamic and thermodynamic sea ice properties in the MIZ [[Asplin et al., 2012](#)]. Sea ice may be fractured by waves and waves also impact the properties of newly formed ice. Conversely, sea ice damps waves and also scatters waves, redirecting wave energy ([Cavaleri et al. \[2012\]](#); [Roach et al. \[2018\]](#)). Finally waves carry momentum and therefore, when they are attenuated in the MIZ part of their momentum goes into the sea ice [[Boutin et al., 2020](#)], through a wave-ice radiation stress on the ice. Changes in wind intensity together with a reduced sea ice cover in the Arctic summer, which gives longer wind fetch for the waves, facilitate wave growth propagating towards the sea-ice [[Thomson and Rogers, 2014](#)]. Unfortunately due to late model updates we were unable to simulate wave-induced ice fracturing. In this study we only analyze the

effects of wave-ice radiation stress on the sea ice.

Wave-induced effects in the ocean such as the Langmuir Turbulence (LT) [Craig and Leibovich, 1976; McWilliams et al., 1997] are recognized to play a significant role in enhancing the vertical mixing within the ocean boundary layer [e.g., Sullivan et al., 2007; Grant and Belcher, 2009; McWilliams and Sullivan, 2000]. The LT is produced when the mean current vorticity interacts with the Stokes drift, which forces overturning cell-like structures known as Langmuir circulation cells [Leibovich, 1983; Thorpe, 2004].

Recent studies have shown that including parametrization of the LT mixing effects in ocean general circulation models contributes to the deepening of the ocean mixed layer (MLD) and to the reduction of its shallow biases [e.g., Fan and Griffies, 2014; Li et al., 2016; Ali et al., 2019]. Accounting for the effects of non-breaking waves has also been shown to enhance upper-ocean mixing, deepen the mixed layer, cool SST during summer, and warm SST during winter [e.g., Li et al., 2025]. As waves propagate through the sea ice and are attenuated, their momentum is transferred to the sea ice, exerting a stress that drives the sea ice in the direction of wave propagation [Longuet-Higgins, 1977].

The parameterizations of LT mixing effects considered in this work are based on the approaches of Van Roekel et al. [2012] and Li and Fox-Kemper [2017], and are implemented by modifying the K-Profile boundary layer scheme [Large et al., 1994] as part of the CVMix community package [Griffies et al., 2015]. Due to the computational burden of running WW3, we have considered two approaches for the wave-ocean coupling. In the first approach, the Stokes drift profile is computed online from the full wave spectrum by WW3. In the second approach, the Stokes drift is approximated by a Wave-Theory parameterization Model (WTM) based on the 10-m wind and the Phillips spectrum [Phillips, 1958] following Breivik et al. [2016] and Li et al. [2017].

Along with introducing a new wave component in NorESM, a wave-diagnostic tool has been developed (Michel et al. [2026]) in order to facilitate the analysis of wave fields. In the current report, we don't present how the waves are affected by the couplings. This report is organized as follows. Section 2 provides a description of the data used in model forcing and evaluating the model's performance. Section 3 outlines the methods, including an overview of the model components, parameterization of wave-induced processes, and the experimental setup. The results are detailed in Section 4, while Section 5 concludes with a summary and discussion.

2 Data

In the experiments that were forced with prescribed atmosphere, the Japanese Atmospheric Re-analysis JRA-55 [Kobayashi et al., 2015] was used. The observed monthly mean mixed-layer climatology are taken from de Boyer Montégut et al. [2004] and used to evaluate the improvement in the predicted MLD as a result of the LT mixing enhancement. The sea surface temperature (SST) seasonal cycle is compared against observations from the World Ocean Atlas 2018 [Locarnini et al., 2019]. For the data-ocean forced experiments, the SST and sea-ice fraction are specified from a monthly climatology of a repeating annual cycle, obtained from observations (HadISST, Rayner et al. [2003]) a

3 Methods

3.1 Model description

The NorESM system used in this work is a development version consisting of the following components: CAM6 for atmosphere, BLOM for ocean and CICE6 for sea ice. In this work, the prognostic wave model WW3 has been added as a wave component. WW3 is forced by wind fields from the atmospheric component, ocean surface currents and boundary layer depth from the ocean component, as well as sea-ice concentration and thickness from the sea-ice component.

3.1.1 Coupling framework

The coupling framework used here is the National Unified Operational Prediction Capability (NUOPC), see <https://earthsystemmodeling.org/nuopc/>, which is based on the Earth System Modelling Framework (ESMF). This part of the model, also called "mediator", accepts variables from all model components on their native grids, and interpolates them to the target grid of other model components, as needed. It is called once per atmospheric model time-step; of which other model's time-steps are a multiple (2 for ocean and sea-ice), the coupling fields being accumulated in the mediator as required. For this work, we utilized the mediator base code tag `cmeps0.14.24_noresm_v1` on <https://github.com/NorESMhub/CMEPS.git>. A number of additional coupling fields were defined in the mediator in order to allow the coupling between the ocean surface wave components, WW3, and the atmosphere, ocean and sea-ice components.

As a rule, the computation and use of such fields occur within the model components themselves. The computation cost of additional couplings is entirely negligible; the coupled model cost in its entirety increased according to the additional computational cost of the new component WW3. We should like to note that this work was carried out while both the coupler and the individual model components were in development. As a consequence, the code and configurations used here are non-standard, reflecting additional, project specific developments on an intermediate development version of the base code itself. This situation has caused a very considerable delay in the actual project work, in particular for the completion of the required model integrations and for their evaluation.

3.1.2 CAM-Nor

The atmosphere component of NorESM is a derivative of CAM6 [Gettelman et al., 2019] with modification in both physics parameterizations [convection, cloud; cf Seland et al., 2020a] and dynamical core [Toniazzo et al., 2020] as used for the NorESM2 (CMIP6) model version, and on the same set of 32 hybrid sigma-p pressure levels. Consistently, we used a configuration using the hydrostatic primitive-equations finite-volume dynamical core [Lin, 2004; Toniazzo et al., 2020] on a regular, rectangular grid of 144 longitude points times 96 latitude points. For a better focus on the North-Atlantic region of the storm track (cf. Sect. 4.1 and Fig. 3), we also used a 288×192 grid in the data-ocean forced experiments. Our CAM-Nor computer code was originally based on CAM tag 6.3.109 https://github.com/ESCOMP/CAM/tree/cam6_3_109). Additional to the NorESM2 modification, numerical stability of the atmospheric PBL was ensured by calling the CLUBB parametrization [Bogenschutz et al., 2013] after the coupler, allowing an iteration on the surface friction velocity within CLUBB, and passing both the updated friction velocity and the updated surface wind to the coupler and the ocean model (with one atmospheric time-step delay with respect to the atmosphere) by means of a correction to the surface friction computed in the coupler. For the computation of the wave-Charnock parameter, which is performed within surface wave component, WW3, neutral instead of actual 10-meter winds consistent with the friction velocity were passed to WW3. CAM-Nor’s overall time-step was kept at the default of 1800 seconds.

3.1.3 BLOM

Compared to the version of Bergen Layered Ocean Model (BLOM) used in NorESM2 [Seland et al., 2020b], an upgraded version is used in this study. A major development is a complete implementation of the vertical Lagrangian-remap method [Griffies et al., 2020], providing great flexibility in the choice of vertical grid. Here a hybrid vertical coordinate is used where layer interfaces are nudged to approach reference densities at interfaces [continuous isopycnals, White et al., 2009], while a minimum layer thickness is enforced at the upper boundary to facilitate resolution in the planetary boundary layer (PBL) [Bleck, 2002]. With the hybrid vertical coordinate, vertical diffusivities are obtained using the non-local K Profile Parameterization [KPP, Large et al., 1994] as implemented in the CVMix community package [Griffies et al., 2015]. A deviation from the standard KPP is the non-local distribution of surface momentum flux and non-local distribution of brine from sea-ice freezing, where we absorb the brine flux towards the base of the PBL. The parameterization of wave-induced Langmuir mixing follows the implementation in CVMix as detailed in Sect. 3.2.2.

Horizontally, BLOM is discretized on the same tripolar grid as used for NorESM2, with a grid resolution at the Equator of 1° zonally and $1/4^\circ$ meridionally, gradually approaching more isotropic grid cells at higher latitudes. In the vertical 76 model layers are used with minimum layer pressure thickness at the surface of 2.5 dbar.

3.1.4 CICE

The present sea ice model is based on CICE6.4.1 (<https://zenodo.org/records/7419531>), with modifications for the CESM model. The actual CESM tag is `cice6_4_1_20230324` (https://github.com/ESCOMP/CICE/releases/tag/cice6_4_1_20230324). CICE uses the Elastic-Viscous-Plastic (EVP) rheology [Hunke and Dukowicz, 1997], and a sea ice thickness distribution formulation [Thorndike et al., 1975]. Vertical physics is modeled based on the icepack model (<https://github.com/CICE-Consortium/Icepack>) as an integrated part of CICE. NorESM uses the mushy-layer thermodynamics [Turner and Hunke, 2015], together with the Delta-Eddington short wave radiation [Briegleb and Light, 2007], and level ice melt ponds [Hunke et al., 2013], a similar configuration as in NorESM2 [Seland et al., 2020b].

3.1.5 WAVEWATCH III (WW3)

The prognostic model WW3 [Tolman, 1991; WAVEWATCH III Development Group, 2019] is a third generation wave model that calculates the evolution of the wave spectrum by solving the wave action equation (Komen et al. [1996]). The source terms of this equation represent several key processes involved in wave transformation. In this study, the source terms account for the following processes: wave generation by wind and wave dissipation by white-capping computed according to Ardhuin et al. [2010] (i.e. ST4 switch); nonlinear wave interactions modeled using the discrete interaction approximation of Hasselmann and Hasselmann [1985]; and the representation of wave attenuation by sea ice calculated following Yu et al. [2022]. The separation between wind-sea and swell regimes is done by using the wave-age criteria.

With the wave spectrum, integrated variables are calculated. We present some of them. The significant wave height, H_{m_0} , is obtained by calculating the total area under the wave spectrum curve. That is:

$$H_{m_0} = 4 \sqrt{\int_0^{2\pi} \int_0^\infty E(\omega, \theta) d\omega d\theta} \quad (1)$$

where $E(\omega, \theta)$ is the two-dimensional wave spectrum, ω is the angular frequency, θ is the direction in which the wave component is propagating.

The peak frequency f_p is the frequency at which the wave spectrum reaches its maximum value. In WW3, f_p is calculated from the one-dimensional frequency spectrum, $E(\omega)$, using a parabolic fit around the discrete peak.

The Stokes drift velocity profile, in deep water, is defined as:

$$\mathbf{u}^s(z) = \frac{2}{g} \int_0^{2\pi} \int_0^\infty \omega^3 \hat{\mathbf{k}} e^{2kz} E(\omega, \theta) d\omega d\theta, \quad (2)$$

where $\hat{\mathbf{k}} = (\sin(\theta), \cos(\theta))$ is a unit vector in the wave direction. The wave model WW3 provides only the surface Stokes drift, i.e $z = 0$ in Eq. (2).

The Stokes transport is the vertically integrated Stokes drift and its direction indicates the direction of the wave-induced horizontal momentum flux. In deep waters, the Stokes transport is calculated as:

$$\mathbf{V}^s = \int_0^{2\pi} \int_0^\infty \omega \hat{\mathbf{k}} E(\omega, \theta) d\omega d\theta \quad (3)$$

In our implementation, the wave model is set in a tripolar grid (named tnx1v4) equal to the ocean and ice grid, which only allows for first-order propagation. In contrast with previous studies [(e.g. Li et al., 2016), wave propagation is also permitted in the polar

regions. The tnx1v4 tripolar grid has a nominal spatial resolution of 360 by 387 grid points. The spectral resolution is 24 directions (with first directional bin at 7.5°) by 26 frequencies logarithmically spaced over the range 0.04118–0.4462 Hz with a standard incremental factor of 1.1. Although by construction, with the tripolar grid, the maximum number of processors allowed in WW3 is $624 = 26 \times 24$ (i.e., no of frequencies $\times$ no of directions), we only used 208 processors for the B experiments and 144 processors for the C experiments since the performance of the wave model, with respect to time, did not improve with more processors. When introducing WW3 in NorESM, the model cost increased by around 40% in both B and C experiments.

3.1.6 Wave-Theory Model (WTM)

Due to the significant computational expense associated with running WW3 and additionally computing the Stokes drift (Eq. (2)) at every ocean point, [Li et al. \[2017\]](#) proposed using a Wave-Theory parameterization Model (WTM) to reasonably approximate the surface-layer-averaged Langmuir number used in the parametrization of the LT mixing effects. Empirical wave spectra, such as the Phillips spectrum [[Phillips, 1958](#)] for the high-frequency range and the Pierson–Moskowitz spectrum [[Pierson Jr. and Moskowitz, 1964](#)] for fully developed seas, provide valuable representations of ocean waves dynamics. The [Li et al. \[2017\]](#) approach depends on the 10-m wind and the Phillips wave spectrum to reconstruct the Stokes drift profile. [Breivik et al. \[2014, 2016\]](#) demonstrated that the Phillips spectrum can be used to reconstruct a reasonable approximation of the full Stokes drift profile. This is justified under the assumption that the Phillips spectrum adequately represents the high frequency part of the real spectrum which accounts for the largest contribution to the Stokes drift near the sea surface.

The reconstructed Stokes drift profile is subsequently used to compute the surface-layer-averaged Langmuir turbulence number which is then utilized to calculate the Langmuir mixing enhancement factor. A detailed derivation of the WTM is found in [Li et al. \[2017\]](#). Here we present only a brief summary. The surface Stokes drift u_0^S is parameterized from the 10-m wind, U_{10} , by

$$u_0^S \approx c_2 U_{10},$$

where $c_2 = 0.016$ is a constant. The Stokes transport is derived from the wind speed based on a depth-integrated Stokes drift from the Pierson–Moskowitz spectrum [[McWilliams and Restrepo, 1999](#); [Webb and Fox-Kemper, 2011](#)]. The significant wave height H_{m0} and

the peak frequency f_p from the Pierson–Moskowitz spectrum [Pierson Jr. and Moskowitz, 1964; Bouws et al., 1998] are

$$H_{m_0} \approx 0.0246 U_{10}^2,$$

$$f_p \approx 0.816 \frac{g}{2\pi U_{10}}.$$

The Stokes transport is then estimated as

$$V^S \approx \frac{2\pi c_1}{16} \bar{f} H_{m_0}^2 \approx 2.67 \times 10^{-5} g U_{10}^3.$$

Here the coefficient $c_1 = 0.667$ accounts for the effect of swells and directional spreading. Based on the the Pierson–Moskowitz spectrum, the mean frequency $\bar{f}$ is related to the peak frequency [Webb and Fox-Kemper, 2011] as $\bar{f} \approx 1.296 f_p$. The values of the constants $c_1 = 0.667$ and $c_2 = 0.016$ are estimated based on the best agreement between the predictions from the WTM and the global wave hind-cast simulation with WW3 [RA13, Rascle and Ardhuin, 2013]. The general peak wavenumber which accounts for directional spreading is derived as

$$k_p \approx 0.176 \frac{u_0^S}{V^S}.$$

Finally, based on the Phillips spectrum, the Stokes drift averaged over a surface layer from $z = -H_{SL}$ to $z = 0$ is given by:

$$u_{SL}^S = u_0^S \left\{ 0.715 + \left(\frac{0.151}{k_p H_{SL}} - 0.840 \right) [1 - T_1(k_p, H_{SL})] - \left(0.840 + \frac{0.0591}{k_p H_{SL}} \right) T_2(k_p, H_{SL}) \right. \\ \left. + \left(\frac{0.0632}{k_p^* H_{SL}} + 0.125 \right) [1 - T_1(k_p^*, H_{SL})] + \left(0.125 + \frac{0.0946}{k_p^* H_{SL}} \right) T_2(k_p^*, H_{SL}) \right\}. \quad (4)$$

Here, the surface layer H_{SL} is determined as the 20% of the depth of ocean boundary layer h as $H_{SL} = h/5$, $k_p^* = 2.56 k_p$ and

$$T_1(k, z) = e^{2kz},$$

$$T_2(k, z) = \sqrt{2\pi k|z|} \operatorname{erfc} \left(\sqrt{2k|z|} \right),$$

where $\operatorname{erfc}(\cdot)$ is the complementary error function. The surface-layer-averaged Stokes drift given in Eq. (4) is used in the formulation of the Langmuir Turbulence enhancement factor in Sect. 3.2.2. Note that u_{SL}^S will be smaller for a deeper ocean surface boundary layer h , a typical situation during winter. The computational cost of the WTM is negligible compared to that of running WW3 [Li et al., 2017].

3.2 Wave coupling processes

In this section, the three wave-induced processes investigated in this work are described.

3.2.1 Wave-Atmosphere coupling through wave-Charnock

The Charnock parameter (Charnock [1955]), denoted here as α , is used in the formulation of the aerodynamical surface roughness z_0 , in the atmospheric model as:

$$z_0 = \frac{\alpha u_*^2}{g}$$

where g is the acceleration due to gravity and u_* is the friction velocity. When there is not coupling α is taken as a constant. When taking into account aerodynamic roughness as wave heights grow due to increasing surface stress α becomes the sea-state Charnock parameter [Janssen, 1991]:

$$\alpha = \frac{\beta}{\sqrt{1 - |\tau_w|/u_*^2}}$$

where $\beta = 0.01$ and τ_w is the wind to wave momentum flux calculated in WW3 as:

$$\vec{\tau}_w = g\rho \int_0^{2\pi} \int_0^\infty \frac{k}{\omega} S_{inp}(\omega, \theta) (\sin \theta, \cos \theta) d\omega d\theta$$

where ω is the radial frequency, k is the wavenumber, ρ is the air density and S_{inp} is the source term corresponding to wave growth computed according to Ardhuin et al. [2010].

3.2.2 Wave-ocean coupling through Langmuir turbulence effects

Parameterizations of the influence of LT in large-scale general circulation ocean models are derived using Large Eddy Simulations (LES) of the Craik-Leibovich vortex force, which demonstrate a significant enhancement of turbulent kinetic energy (TKE) [e.g., Skillingstad and Denbo, 1995; McWilliams and Sullivan, 2000; Harcourt and D'Asaro, 2008; Van Roekel et al., 2012]. The parameterization of the LT effects is achieved by applying an enhancement factor, within the ocean boundary layer, to the turbulence velocity scale w_x , as well as enhancing the boundary layer entrainment via modification of the unresolved turbulent velocity shear used in the calculation of the bulk Richardson number. In the KPP scheme [Large et al., 1994], the water column is divided into two regions: the ocean surface boundary layer and the ocean interior. The ocean surface layer is dominated by turbulence from wind, buoyancy fluxes and waves, while the ocean interior mixing is driven by processes such as internal wave breaking and shear instabilities.

The ocean boundary layer depth h is determined as the shallowest depth at which the bulk Richardson number $\text{Ri}_b(z)$ rises above a critical value Ri_c :

$$\text{Ri}_b(z) = \frac{(B_r - B(z))|z|}{|\mathbf{u}_r - \mathbf{u}(z)|^2 + V_t^2(z)}. \quad (5)$$

Here B_r is the buoyancy and $\mathbf{u}_r$ is the velocity, both averaged over a thin surface layer centered at z_r . Note that, in the CVmix implementation, the distance $|z_r - z|$ is used instead of $|z|$. The term V_t represents the unresolved turbulent velocity shear which is parameterized [Large et al., 1994] as :

$$V_t^2(z) = \frac{C_v N(z) w_s |z|}{\text{Ri}_c \kappa^2} \left(\frac{-\beta_T}{c_s \varepsilon} \right)^{\frac{1}{2}} \quad (6)$$

where $C_v = \max(2.1 - 200.0 \times \max(0, N), 1.7)$ [Danabasoglu et al., 2006], w_s is the turbulence velocity scale for scalar properties, $N(z)$ is the local buoyancy frequency, $\kappa = 0.4$ is the von Kármán constant, and $c_s = 98.96$ while $\varepsilon = 0.1$. The parameter $\beta_T \approx -0.2$ represents the ratio of the entrainment buoyancy flux at the base of the boundary layer to the surface buoyancy flux [Large et al., 1994]. In the KPP scheme, the turbulent eddy diffusivity/viscosity of a property X within the ocean surface boundary layer $0 < z < h$ is parameterized as

$$K_x(z, t) = h(t) w_x(z/h, t) G(z/h). \quad (7)$$

where G is a smooth shape function assumed to be a cubic polynomial, the turbulent velocity scale $w_x = \kappa u^* / \phi$, and u^* is the friction velocity and ϕ is the stability function.

Based on a LES study, McWilliams and Sullivan [2000] proposed that the LT effects to be parameterized through a mixing enhancement factor $\mathcal{E}(\text{La}_t) \geq 1$ applied to the turbulent velocity scale as

$$w_x = (\kappa u^* / \phi) \mathcal{E}(\text{La}_t), \quad (8)$$

This modification will enhance both the turbulent eddy diffusivity/viscosity (Eq. 7) and the unresolved turbulent velocity shear (Eq. 6) as both depend on the turbulent velocity scale w_x . The turbulent Langmuir number is defined as $\text{La}_t = \sqrt{u^* / u_0^S}$ [McWilliams et al., 1997], which measures the effects of the wind-driven shear given by the friction velocity u^* relative to the Stokes drift production [Grant and Belcher, 2009]. To account for the Stokes drift penetration depth, Harcourt and D'Asaro [2008] proposed a surface-layer averaged Langmuir number as

$$\text{La}_{\text{SL}} = [u^* / (\langle u^s \rangle_{\text{SL}})]^{1/2}. \quad (9)$$

The surface layer averaged Stokes drift is:

$$\langle u^s \rangle_{\text{SL}} = \frac{1}{H_{\text{SL}}} \int_{-H_{\text{SL}}}^0 u^s(z) dz,$$

where H_{SL} represents the upper 20% of the ocean boundary layer depth h .

The inclusion of the ocean boundary layer depth in the calculation of Langmuir number in Eq. (9) introduces seasonal variability to the Langmuir enhancement factor. Specifically, the enhancement factor becomes weaker in winter and stronger in summer due to the seasonal difference in the ocean boundary layer depth, which is typically deeper in winter than in summer. If this dependency is neglected, it could result in a vigorous Langmuir turbulence mixing during winter that overestimates the winter MLD.

Van Roekel et al. [2012] proposed the inclusion of potential mismatch in the direction between waves and wind, as this will suppress the Langmuir turbulence and hence reduce its vertical mixing enhancement. The combined effects from the the Stokes penetration depth and wave-wind misalignment gives a projected Langmuir number [Harcourt and D'Asaro, 2008; Van Roekel et al., 2012; Li et al., 2016] in the form:

$$\text{La}_{\text{SLP}} = \left[\frac{u^* \cos(\alpha)}{\langle u^s \rangle_{\text{SL}} \cos(\theta_{ww} - \alpha)} \right]^{1/2}, \quad (10)$$

where α represents the angle between wind and Langmuir cells, and θ_{ww} is the angle between Stokes drift and wind. Note that $\langle u^s \rangle_{\text{SL}}$ can be computed from WW3 using the full wave spectrum or approximated by the TWM using Eq. (4).

In this work, we have used three LT parameterization types, focusing on how the enhancement factor is formulated and how the unresolved turbulent velocity shear is modified, as well as which wave coupling source is used. These parameterizations are summarized in Table 1, with detailed descriptions given below.

(I) LT parameterization type ww3VR12:

In the ww3VR12 type, the enhancement factor is computed by WW3 and accounts for wind-wave misalignment. Therefore, it is based on the projected Langmuir number La_{SLP} [Van Roekel et al., 2012; Li et al., 2016] and expressed as:

$$\mathcal{E}(\text{La}_{\text{SLP}}) = |\cos \alpha| \sqrt{1 + (1.5\text{La}_{\text{SLP}})^{-2} + (5.4\text{La}_{\text{SLP}})^{-4}}. \quad (11)$$

This enhancement factor is applied to w_x in Eq. (8), consequently enhancing the eddy diffusivity/viscosity in Eq (7). In addition, the enhancement factor is applied to the turbulence velocity scale w_s appearing in the expression for the unresolved shear V_t . Thus the

enhanced unresolved turbulent velocity shear V_{tE} is given by:

$$V_{tE}^2(z) = \frac{C_v N(z) w_s \mathcal{E}(\text{LaSLP}) |z|}{\text{Ri}_c \kappa^2} \left(\frac{-\beta_T}{c_s \mathcal{E}} \right)^{\frac{1}{2}}. \quad (12)$$

The LT modification of the unresolved turbulent velocity shear V_t can increase its magnitude, which will decrease the bulk Richardson number Ri_b leading to a deepening of the MLD.

(II) LT parameterization type ww3LF17:

The ww3LF17 type is the same as ww3VR12 except that the unresolved turbulent shear is modified differently. Instead of applying the LT enhancement factor to the velocity scale as in Eq. (12), the ww3LF17 [Li and Fox-Kemper, 2017] proposed a scaling to the entrainment buoyancy flux $-\overline{w'b'_e}$ to account for the influence of LT on the boundary layer entrainment under destabilizing conditions. The unresolved shear given in Eq. (6) is re-derived [Li and Fox-Kemper, 2017] to incorporate the LT-enhanced entrainment buoyancy flux as:

$$V_{tL}^2(z) = \frac{C_v N(z) w_s(z) |z|}{\text{Ri}_c} \left[\frac{-\overline{w'b'_e} h}{w_s(z)^3} \right]^{\frac{1}{2}}, \quad (13)$$

where the LT-enhanced entrainment buoyancy flux, under unstable conditions, scales as

$$-\overline{w'b'_e} = \frac{u^{*3}}{h} \left(0.17 + 0.083 \text{La}_{\text{SL}}^{-2} - 0.15 \frac{h}{\kappa L} \right).$$

Here $L = u^{*3}/(\kappa B_f)$ is the Monin-Obukhov length.

Note that this modified turbulent shear by Li and Fox-Kemper [2017] is applied only under destabilizing surface buoyancy flux, because this is the regime in which their LES simulations were conducted. Under neutral and stabilizing regimes, we apply the default unresolved turbulent shear given in Eq. (6). As illustrated by Li and Fox-Kemper [2017], the modified unresolved turbulent shear velocity V_{tL}^2 will make the diagnosed surface boundary layer slightly shallower in the convection limit, as V_{tL}^2 becomes slightly smaller than the default V_t^2 . However, it will have deepening effects in the weak surface cooling limit. This means that the deepening effect will be constrained under pure convection conditions (dominant winter conditions for example in the Labrador Sea), while it will be enhanced under weak surface cooling and in the presence of wind and wave. In deriving expression (6) for the unresolved shear in the default KPP scheme, the entrainment buoyancy flux under pure convection is parameterized using the empirical relation $-\overline{w'b'_e} = \beta_T B_f$ under pure convection [Large et al., 1994]. Same as in ww3VR12, the LT enhancement factor and surface-layer averaged LT number are computed by WW3.

The ww3LF17 parameterization is used in Exps. XB_o, XC_o.

(III) LT parameterization type prmVR12:

The prmVR12 type is the same as ww3VR12 except that the wave coupling source is the WTM instead of WW3. In this type, the surface-layer averaged Stokes drift u_{SL}^S , estimated by Eq. (4) is utilized. However, since wave-wind misalignment is not explicitly accounted for, the LT enhancement factor is computed as:

$$\mathcal{E}(La_{SL}) = \sqrt{1 + (1.5La_{SL})^{-2} + (5.4La_{SL})^{-4}}, \quad (14)$$

where the surface-layer averaged Langmuir number

$$La_{SL} = [u^*/u_{SL}^S]^{1/2}, \quad (15)$$

and u_{SL}^S as in Eq. (4).

The prmVR12 type is used in Exps. XB_prmVR12_o and XC_prmVR12_o.

LT type	Subscript	Enhancement factor $\mathcal{E}(La_{SL})$	Turb. shear	Wave source
CNTL		1	V_t^2	No wave effects
ww3VR12	ww3VR12_o	$ \cos\alpha \sqrt{1 + (1.5La_{SLP})^{-2} + (5.4La_{SLP})^{-4}}$	V_{tE}^2	WW3
ww3LF17	_o	$ \cos\alpha \sqrt{1 + (1.5La_{SLP})^{-2} + (5.4La_{SLP})^{-4}}$	V_{tL}^2	WW3
prmVR12	prmVR12_o	$\sqrt{1 + (1.5La_{SL})^{-2} + (5.4La_{SL})^{-4}}$	V_{tE}^2	WTM

Table 1: List of the LT mixing parameterizations assessed in this work. The first column provides the parameterization type, while the second column shows the subscript used in the experiments names. The third column lists the formula for the LT mixing enhancement factor, followed by the modified unresolved turbulent velocity shear in the fourth column. The fifth column presents the wave coupling source. CNTL means no wave-induced effects are included.

3.2.3 Wave-Sea-ice coupling through wave-ice radiation stress

Waves transport momentum, and when they are attenuated by sea ice this momentum is transferred to the sea ice. This momentum loss acts as a stress that pushes the sea ice in the direction of the attenuated wave [Longuet-Higgins, 1977]. In WW3 the wave-ice radiation stress is computed following Williams et al. [2017] as:

$$\tau_{wi} = \rho_w g \int_0^{2\pi} \int_0^\infty \left(\frac{k}{\omega} \right) S_{ice}(\omega, \theta) (\cos \theta, \sin \theta) d\omega d\theta \quad (16)$$

where ρ_w is the water density; g is gravity; ω , θ and k are, respectively, the radial frequency, direction and wavenumber of waves; and S_{ice} , calculated following Yu et al. [2022], is the source term corresponding to wave attenuation by sea ice at a given position. Once estimated by WW3, this wave-ice radiation stress is sent to the sea-ice model.

Given the vertically integrated sea ice velocity $\vec{u}$, the force balance of sea ice per unit surface area can be written as

$$m \frac{\partial \mathbf{u}}{\partial t} + mf\mathbf{k} \times \mathbf{u} = -mg\nabla\zeta - \nabla \cdot \boldsymbol{\sigma} + \boldsymbol{\tau}_a + \boldsymbol{\tau}_w - \boldsymbol{\tau}_o \quad (17)$$

where m is the total mass of ice and snow, ζ is the ocean surface height, f is the Coriolis parameter, $\mathbf{k}$ is the vertical unit vector, $\boldsymbol{\sigma}$ is the internal stress tensor, and $\boldsymbol{\tau}_a$ and $\boldsymbol{\tau}_o$ are surface stresses from the air and water, respectively [Consortium, 2023]. In addition, we have added the wave-ice stress $\boldsymbol{\tau}_{wi}$. The surface stresses from air and water contain a factor proportional to the sea ice fraction a_{ice} to act over the sea ice covered part of the grid cell, only.

When solving Eq. (17), the surface elevation tilt term, air-ice stresses and wave-ice stresses are held constant, while Coriolis, internal stresses, and ice-ocean stresses are varying dependent on $\vec{u}$. Therefore, in the model implementation, the wave-ice and air-ice stress vectors are added together before applied in the force balance.

One important point to notice is that both wave-ice, and air-ice stresses, and the total mass m , are proportional to the ice fraction a_{ice} . Therefore, it is important that the wave-ice stress from Eq. (16) is calculated with respect to the same ice concentration as used in Eq. (17). If this requirement is not fulfilled, momentum will not be conserved, and the solution of Eq. (17) can be unstable. An alternative is to transfer the intensive stresses from the wave model (stress per unit ice area), with a value of zero if ice is not present. Then, the ice model can multiply this with the same ice fraction as used in the rest of Eq. (17). This approach should be stable, but does not conserve momentum between the models. Therefore, in the present experiments, we have opted to change the call sequence of the models, such that atmosphere, waves, and sea ice are called sequentially at every radiation time step (30 s) for the atmosphere model. Then the waves will use the same ice concentration field to calculate the wave-ice stresses, as is applied afterwards when solving Eq. (17). Unfortunately, due to the poor scaling of WW3 to many processors, this slows down the model considerably.

3.3 Experiments description

The experimental design and nomenclature used in this work are illustrated in Fig. 1, which also shows the coupling between the different components. Each experiment name starts by **X** followed by a primary: **A**, **B** or **C** denoting the forcing configuration, and an optional subscript indicating the type of wave coupling with the other components. Specifically, the primary **A** refers to experiments forced with prescribed ocean, **B** denotes experiments forced with prescribed atmosphere, and **C** indicates experiments where all components are active.

The subscripts _a refers to wave-atmosphere coupling using wave-dependent Charnock in CAM, _o refers to wave-ocean coupling using LT parameterization in BLOM while _i indicates wave-sea-ice interaction using wave-ice radiation stress in CICE. Besides the wave-ocean coupling with _o, additional experiments are conducted using alternative LT mixing parameterizations and wave sources, which are denoted with the subscripts _{ww3VR12_o} and _{prmVR12_o} and summarized in Table 1.

3.3.1 Data-ocean forced experiments: A

Experiments forced with prescribed ocean data are indicated by the primary **A**. Here the atmosphere and wave components are active and the wave component is forced by fields from the atmosphere component. The land component is still interactive, but SST and sea-ice fraction are specified from a monthly climatology of a repeating annual cycle, obtained from observations (HadISST, [Rayner et al. \[2003\]](#)) as an average over the decade 1995-2004. These experiment were run for 10 years from climatological initial conditions, and evaluated over the last seven years.

3.3.2 Data-atmosphere forced experiments: B

Experiments forced with prescribed atmosphere data are denoted by the primary **B** and utilize JRA-55 atmospheric reanalysis [[Kobayashi et al., 2015](#)] for forcing. Here the wave, ocean and sea-ice components are active. These experiments allow us to study the oceanic response without feedback from the atmosphere. Generally, **B** experiments were run for 62 years, from climatological (WOA) initial conditions. Only the control, XB, was run over several JRA cycles for a total of 240 years; and XB_o was initialized from year 180 of XB instead of WOA.

3.3.3 All active components: C

Experiments with all active components are identified by the primary **C**. Unlike the **A** and **B** cases, these experiments include an active atmosphere component. Therefore, this configuration will allow for the atmospheric feedback, for example when including wave-induced effects on the ocean or/and sea-ice. All **C** experiments were run for a total of 80 years. The initial conditions were provided by the end of a 100-year long spin-up integration of XC which is in turn initialized from WOA climatology. The spin-up integration has identical configuration to XC.

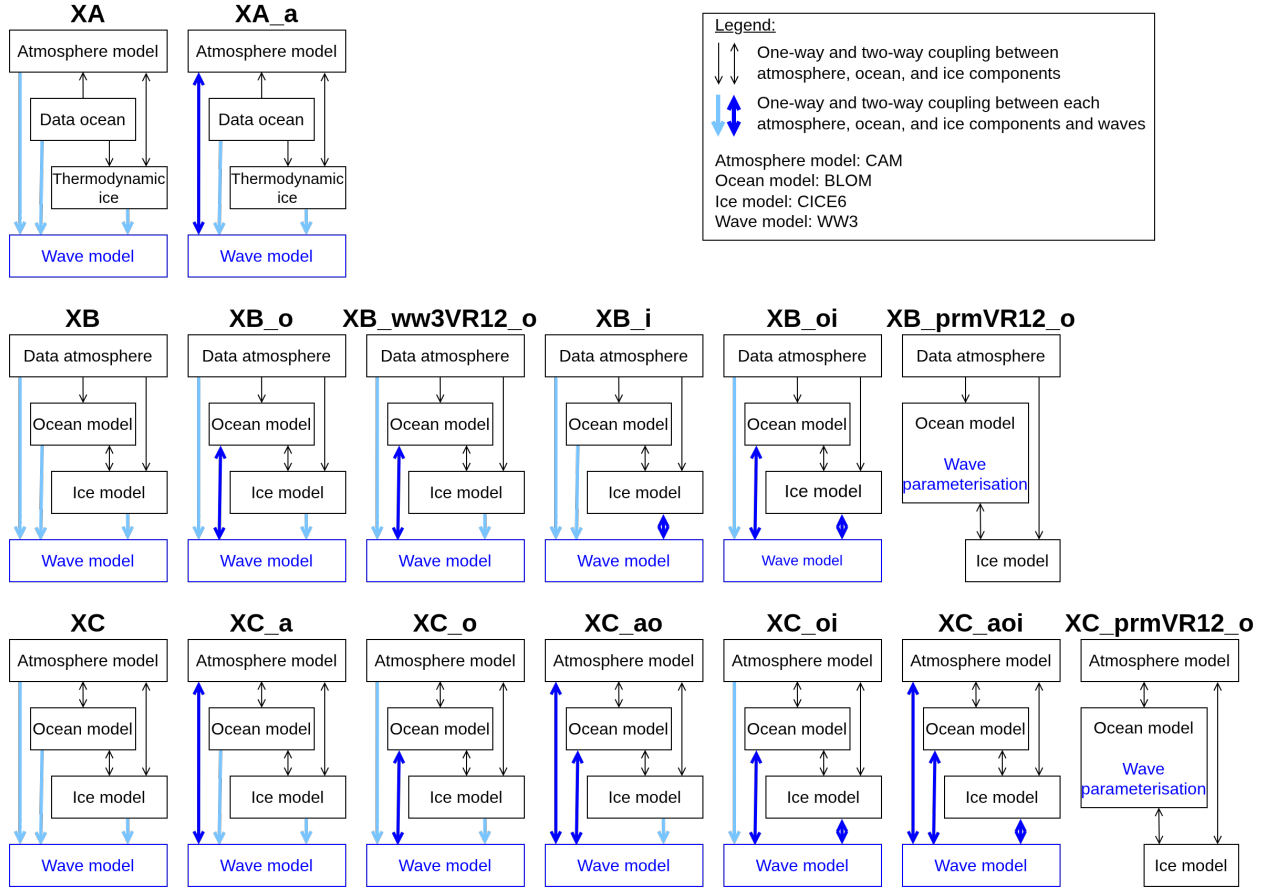


Figure 1: Schematic describing the experiments performed and highlighting in blue the connections with the waves: wave parameterization within the ocean model or one-way and two-way coupling with the wave model. The primary **A** refers to experiments with prescribed ocean, **B** denotes data-ocean forced experiments, and **C** indicates experiments with all active components. The subscript **_a** refers to wave-atmosphere coupling using wave-Charnock in CAM, **_o** refers to wave-ocean coupling using LT parametrization in BLOM while **_i** indicates wave-sea-ice interaction using wave-ice radiation stress in CICE. The subscript **prmVR12** points to the case where LT enhancement is diagnosed from WTM instead of WW3. For detailed LT parametrization see Table.1.

4 Results

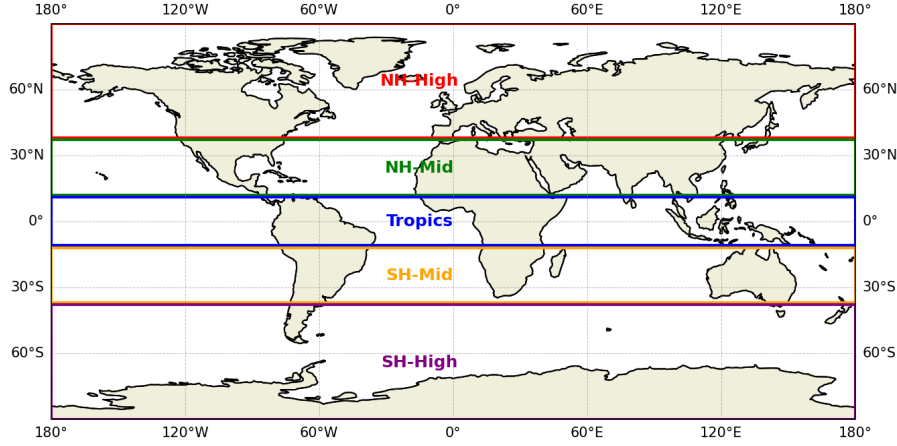


Figure 2: Subregions NH-High = (38°N to 90°N), NH-Mid = (12°N to 37°N), Tropics = (11°S to 11°N), SH-Mid = (37°S to 12°S), SH-High = (90°S to 38°S) representing high latitudes, extra tropics and tropics.

As will be shown, the results from the numerical experiments, defined in Figure 1, demonstrate that wave-induced effects yield changes in the atmosphere, ocean and sea-ice. A detailed presentation of these results is provided in this section.

To investigate the regional changes induced by the wave effects, the global domain is split into five equal-area meridional sub-regions as shown in Fig. 2, namely Northern Hemisphere - High Latitudes (NH-High: 38°N to 90°N), Northern Hemisphere - Mid Latitudes (NH-Mid: 12°N to 37°N), Equatorial Zone (Tropics 11°S to 11°N), Southern Hemisphere - Mid Latitudes (SH-Mid: 37°S to 12°S) and Southern Hemisphere - High Latitudes (SH-High 90°S to 38°S).

4.1 Wave-Atmosphere coupling

Initial experiment at storm-resolving resolution performed with the CAM-Nor model with the wave-Charnock parametrization (analogous to the XA and XA_a pair, but at higher resolution) indicated promising results from the use of the wave-Charnock parametrization with an interactive ocean surface wave field. Figure 3 shows that in the North Atlantic, and northern Europe, a pattern of low-high-low pressure emerges with systematic changes on surface temperature and precipitation. These are dynamically consistent with changes in the storm track that appear to be linked with increased latent heat fluxes over the North Atlantic during high-wind conditions. We found that these changes are not

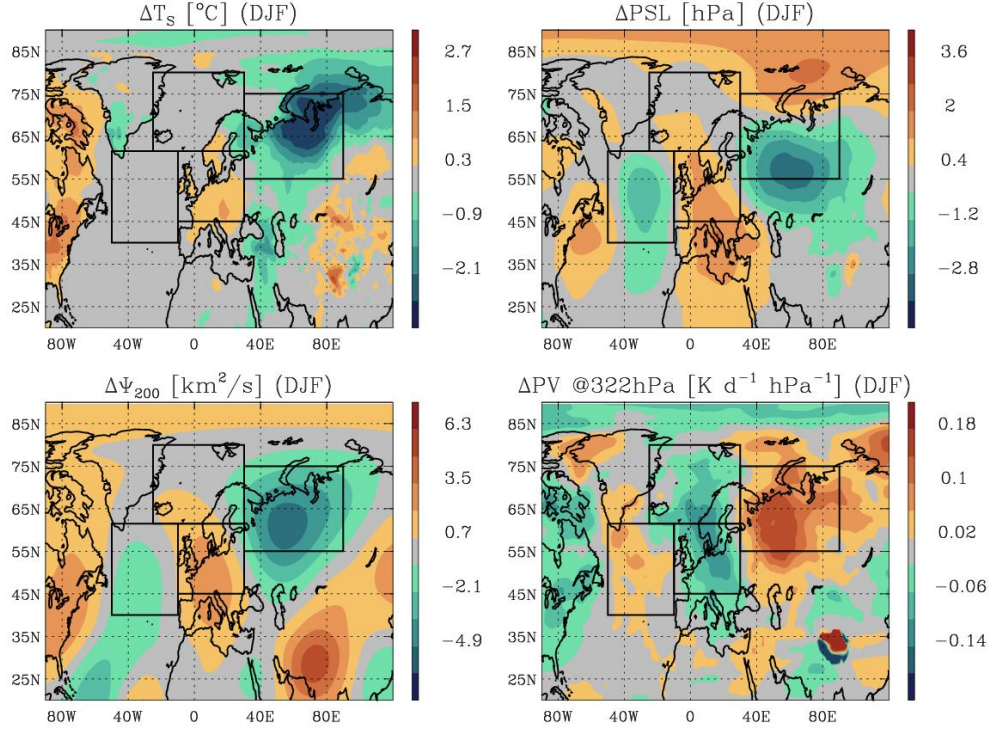


Figure 3: DJF climatological differences of atmospheric variables for an XA experiment, $\text{XA}_{\text{a}} - \text{XA}$ with observed SST and sea-ice climatology (HadISST), 1° resolution. The variables shown are indicated by the labels at the top of the panels: surface temperature (T_s , top left), pressure at mean sea level (PSL, top right), streamfunction at 200 hPa (Ψ_{200} , bottom left), and potential vorticity (PV) at the 322hPa model pressure level (bottom right).

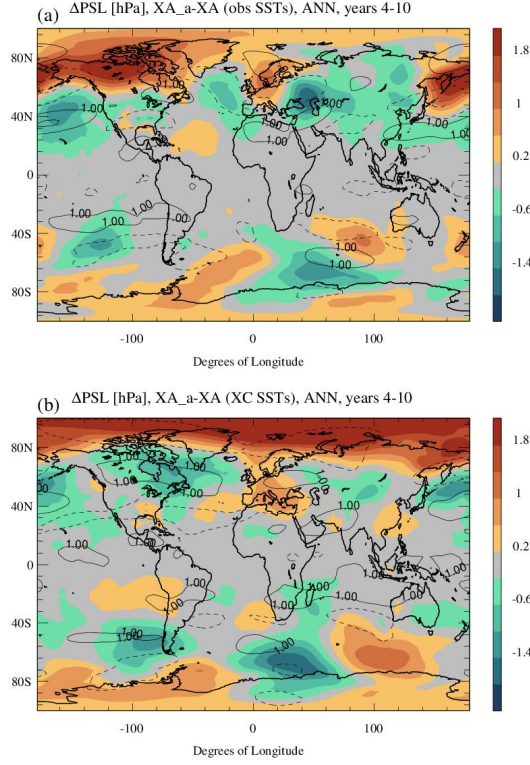


Figure 4: Annual mean climatology differences in XA_a – XA, standard (2°) resolution. (a) observed SST and sea-ice climatology (HadISST); (b) SST and sea-ice climatology from the XC (control, all active) experiment.

dependent on either teleconnections or on the details of the representation of the wave-Charnock parameter, so long as this is sensitive to high wind situations.

However, we also found that such changes, along with others in the climatology, are in fact very sensitive to the prescribed SST and sea-ice climatology in these experiments. Figure 4 illustrates this point on the basis of the standard XA_a experiment compared with XA and XA_a simulations using prescribed SSTs and sea-ice concentrations from out coupled all-active simulations, XC. The PSL pattern correlation (Pearson coefficient for the area-weighted grid-point arrays) between the two climatological differences is 0.16, i.e. not nearly significant. Similar results are found for other quantities. Furthermore, even if regionally there may still be similarities, perhaps governed by local dynamics, even these disappear when the SSTs are allowed to respond to atmospheric circulation changes. We note that, in general, unrealistic model configurations such as those used for the **A** or **B** experiments in this work, where heat is not allowed to modify a part of the model state, are only useful and informative to the extent that they provide explanatory

power for results from all-active simulations where such unphysical constraints are not applied. Moreover, a strong sensitivity of the effects of a certain physical mechanism, like wave-Charnock here, to the model state, implies compounding errors due to model systematic biases. Hence, even though some of the results from the **B** experiments are conceptually interesting, we must conclude that the use of wave-Charnock in climate simulations is, on its own, not useful.

4.2 Wave-Ocean coupling

The main impacts of accounting for wave-induced effects on the ocean through parameterization of Langmuir Turbulence effects are deepening of the ocean mixed layer depth, as demonstrated in the MLD difference timeseries in Fig. 5, summer-mean in Fig. 6 and winter-mean in Fig. 7. Moreover, these impacts are robust in both the **B** and **C** experiments, demonstrating that the MLD deepening due to the LT parameterization is generally not affected by the atmospheric feedback. Strong MLD deepening takes place in regions where high wave conditions are present, for example over the southern ocean. The seasonal averages are computed over a period of twenty years as given in Table ???. The summer-mean is calculated by averaging MLD over July through September (JAS) for the Northern-Hemisphere (NH) and over January through March (JFM) for the Southern-Hemisphere (SH). In contrast, the winter-mean is determined by averaging MLD over JFM for the NH and over JAS for the SH.

Figures 6 and 7 shows that the MLD deepening when coupling with WW3, ($XC_o - XC$), is mainly reproduced by using parametrized waves with the WTM, ($XC_{prmVR12_o} - XC$). A similar agreement can be observed in the **B** experiments, specifically between $XB_o - XB$ and $XB_{prmVR12_o} - XB$, except that there appear MLD decrease close to sea-ice during winter as shown in Fig. 7.

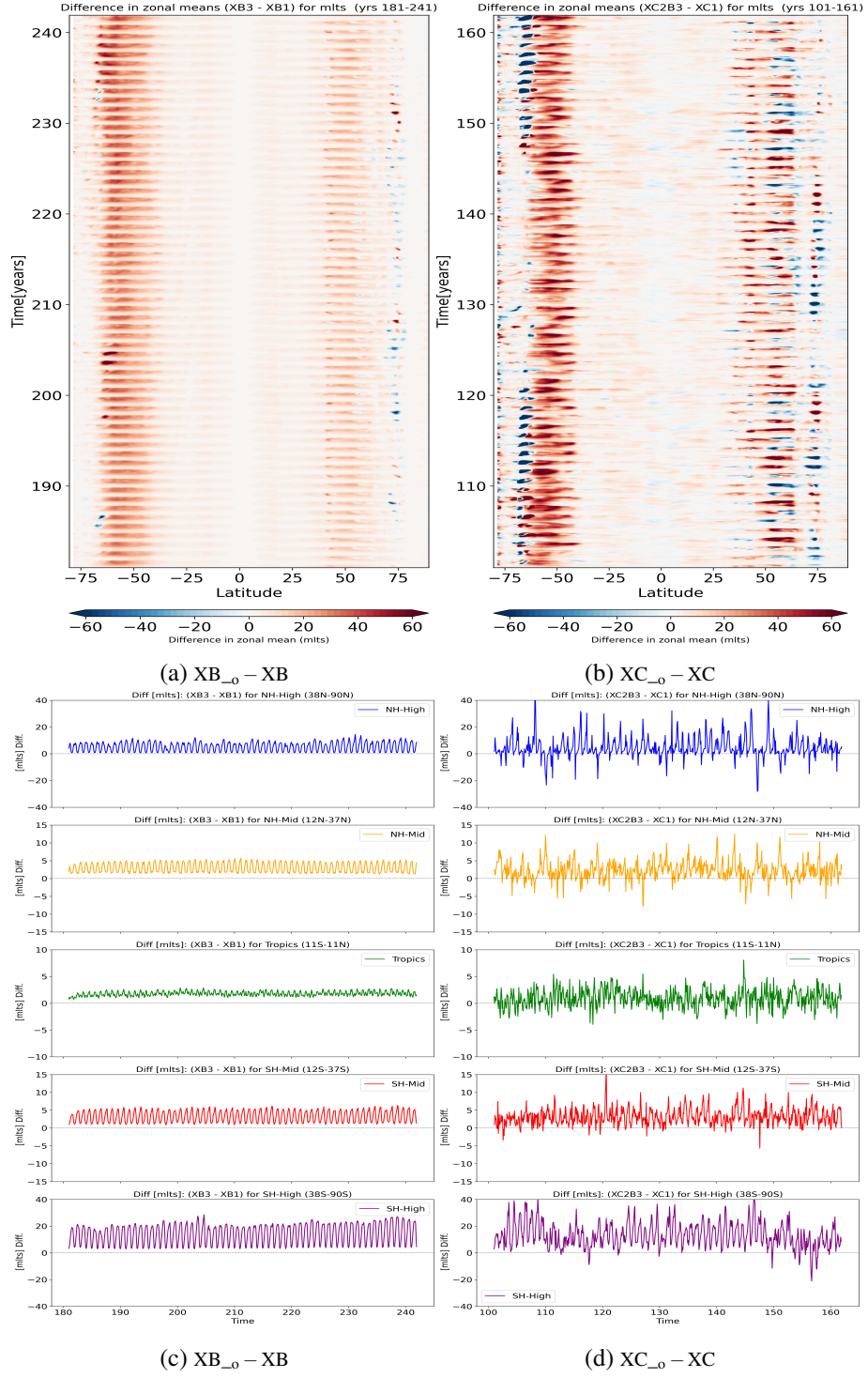


Figure 5: Timeseries showing the impacts of LT mixing enhancement on the MLD. Top panels: temporal evolution of the difference in the zonal-mean MLD, while bottom panels show timeseries of the difference in area-averaged MLD over subregions NH-High, NH-Mid, Tropics, SH-Mid and SH-High. Left panels: $XB_0 - XB$ and right panels: $XC_0 - XC$.

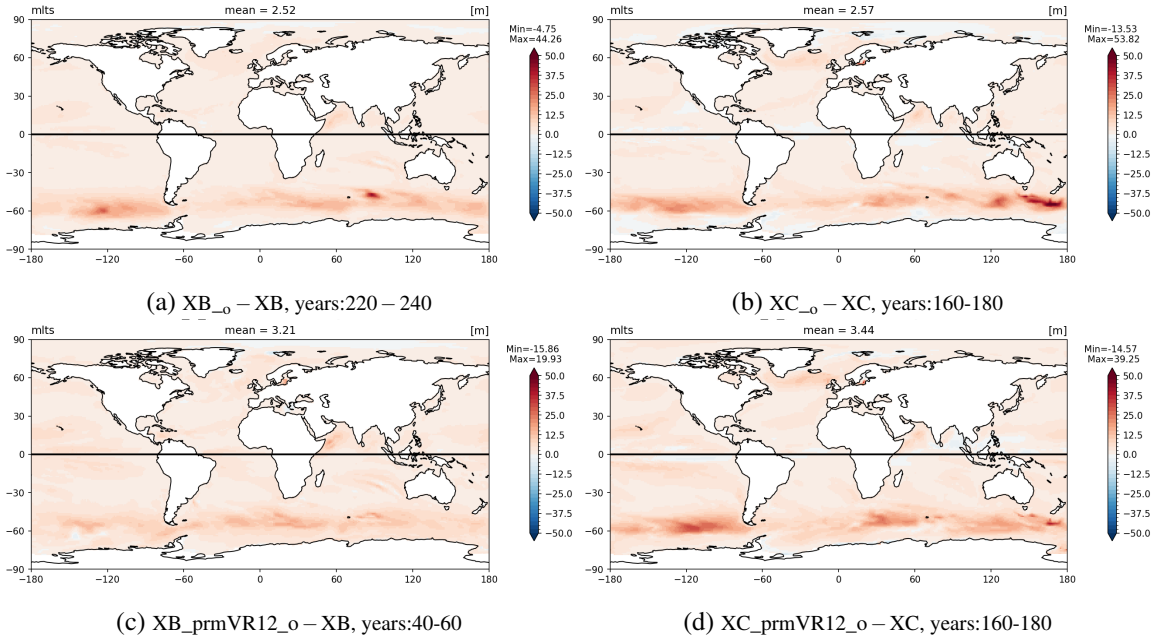


Figure 6: Maps depicting the impact on the summer-mean of the ocean mixed layer (MLD) due to parametrization of LT mixing enhancement for both hemispheres, where the summer-mean MLD is computed by averaging MLD over JAS for NH and over JFM for SH. Left) **B** configuration and right) **C** configuration. The top panels show the case when using the ww3LF17 type, and bottom panels show when using the prmVR12 type as defined in Table 1. Here the temporal averages from the final 20 years of the runtime for each experiment are utilized.

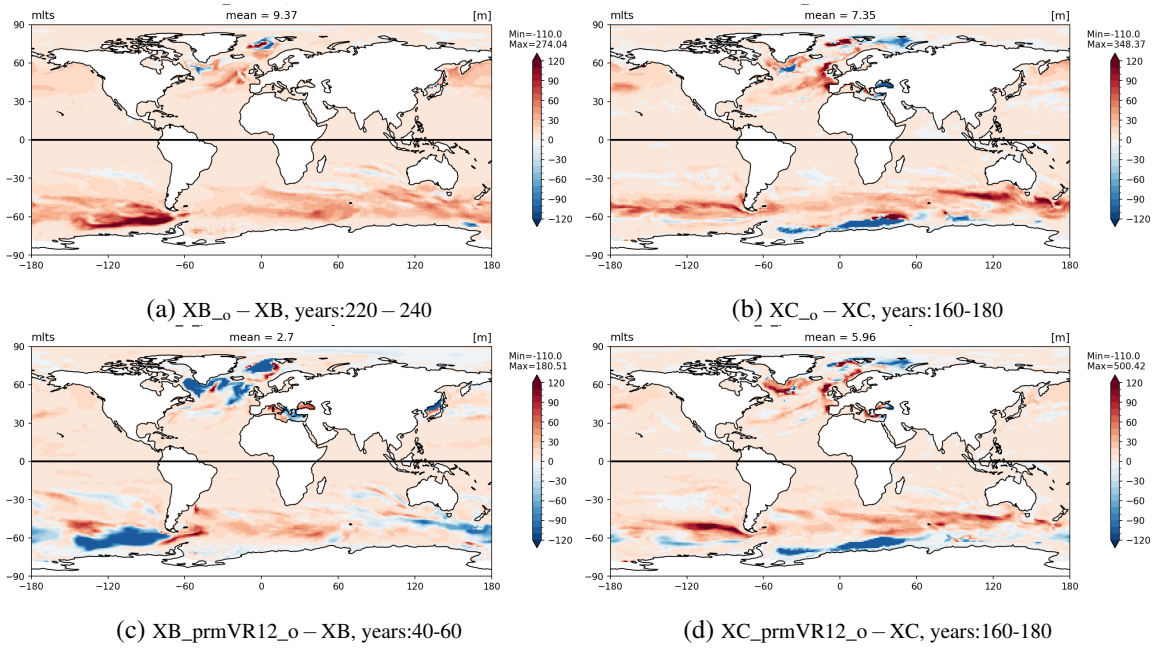


Figure 7: Same as in Fig. 6 but for the impact on winter-mean MLD for both hemispheres.

The zonal-averaged percentage differences in the summer and winter mean MLD relative to the control run are illustrated in Fig. 8. The MLD deepening due to the LT-induced effects is higher at high latitudes and it can reach 20%. Figure 8 illustrates that the MLD percentage increases in the cases prmVR12 (plotted in the dash line) and ww3LF17 (plotted in the solid line) are comparable.

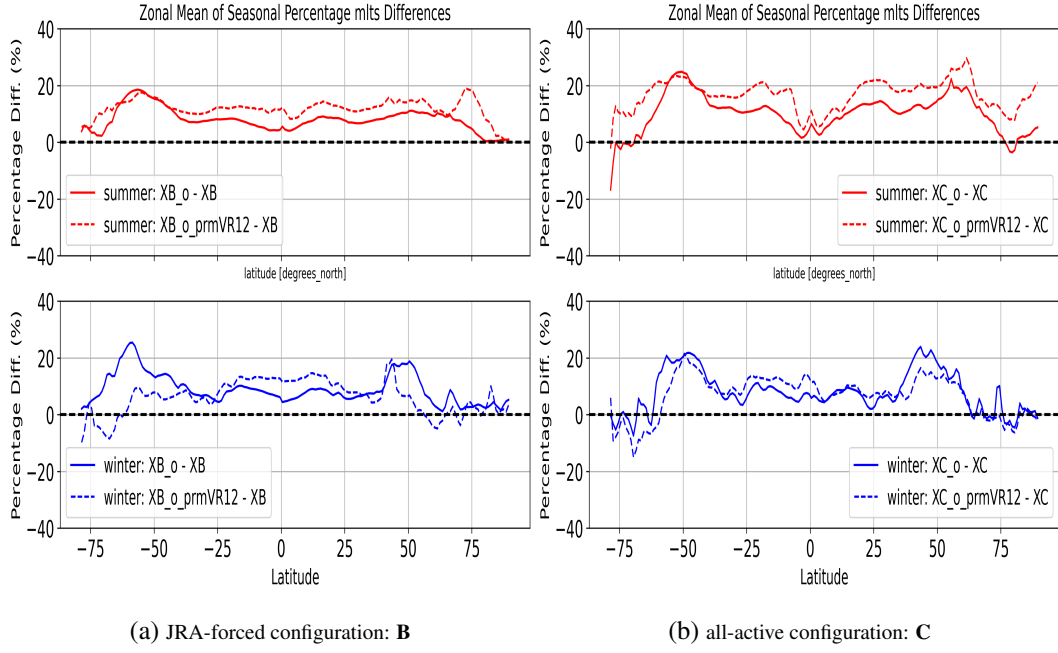


Figure 8: Zonal-averaged percentage differences in the summer and winter mean mixed layer depth (MLD) relative to the control run without wave coupling. Left) **B** experiments and right) **C** experiments. These computed from Figs. 6&7.

In agreement with previous results, for example from Li and Fox-Kemper [2017], including the LT effects deepens the MLD and hence reduces its MLD shallow biases, especially in the Southern Ocean during summer as shown in Fig. 9. This is more pronounced in the all-active experiments: **C**, where significant MLD improvements are demonstrated in Fig. 9(b). The changes on the winter-mean MLD biases, provided in Fig 10, are mainly small except too deep MLD in the SH. Since the winter MLD is generally deep and primarily dominated by convection, incorporating the Stokes penetration depth into the Langmuir enhancement factor—through the surface-layer-averaged Langmuir number—mainly serves to constrain the LT mixing effects during winter.

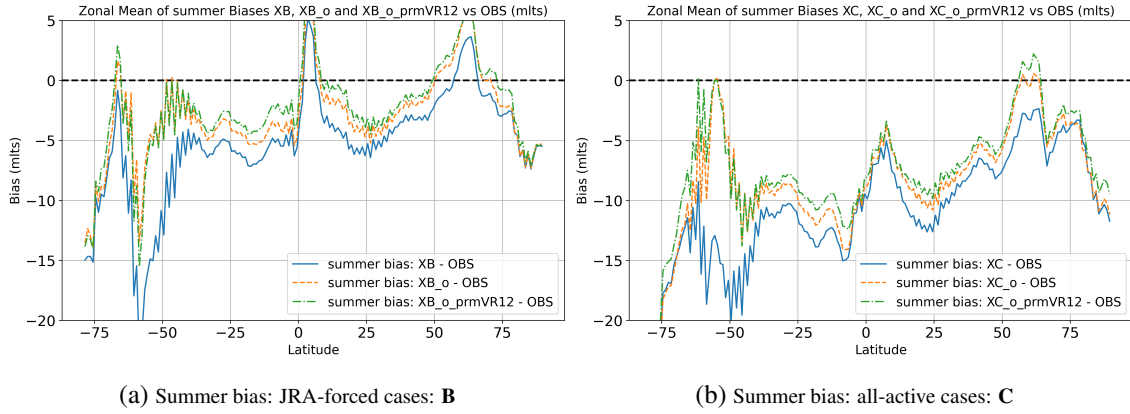


Figure 9: Zonal-averaged bias of the summer-mean MLD against [de Boyer Montégut et al. \[2004\]](#) observations, for Left) **B** experiments, and right) **C** experiments. The summer-mean is over JAS for the NH and JFM for the SH.

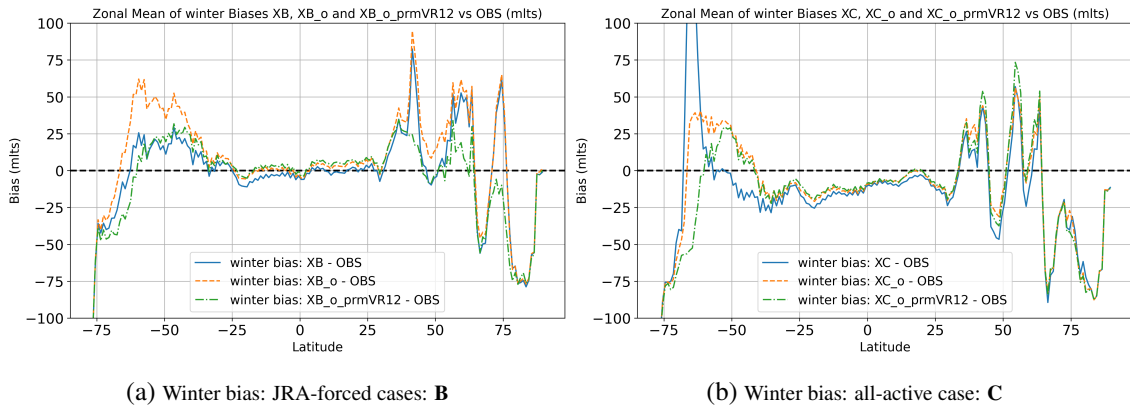


Figure 10: Same as in Fig. 9 but for the zonal-mean bias of the winter-mean MLD.

Corresponding to the MLD deepening, the enhanced LT mixing triggers SST cooling during summer, and SST warming during winter, which is more prominent in the **B** experiments as shown in timeseries Fig. 11 and summer/winter-mean Fig. 12. In contrast, in the **C** experiments where there is a presence of the atmospheric feedback, the SST is overall reduced specially in the southern hemisphere after the first 20 year of integration (see Fig. 11(b)).

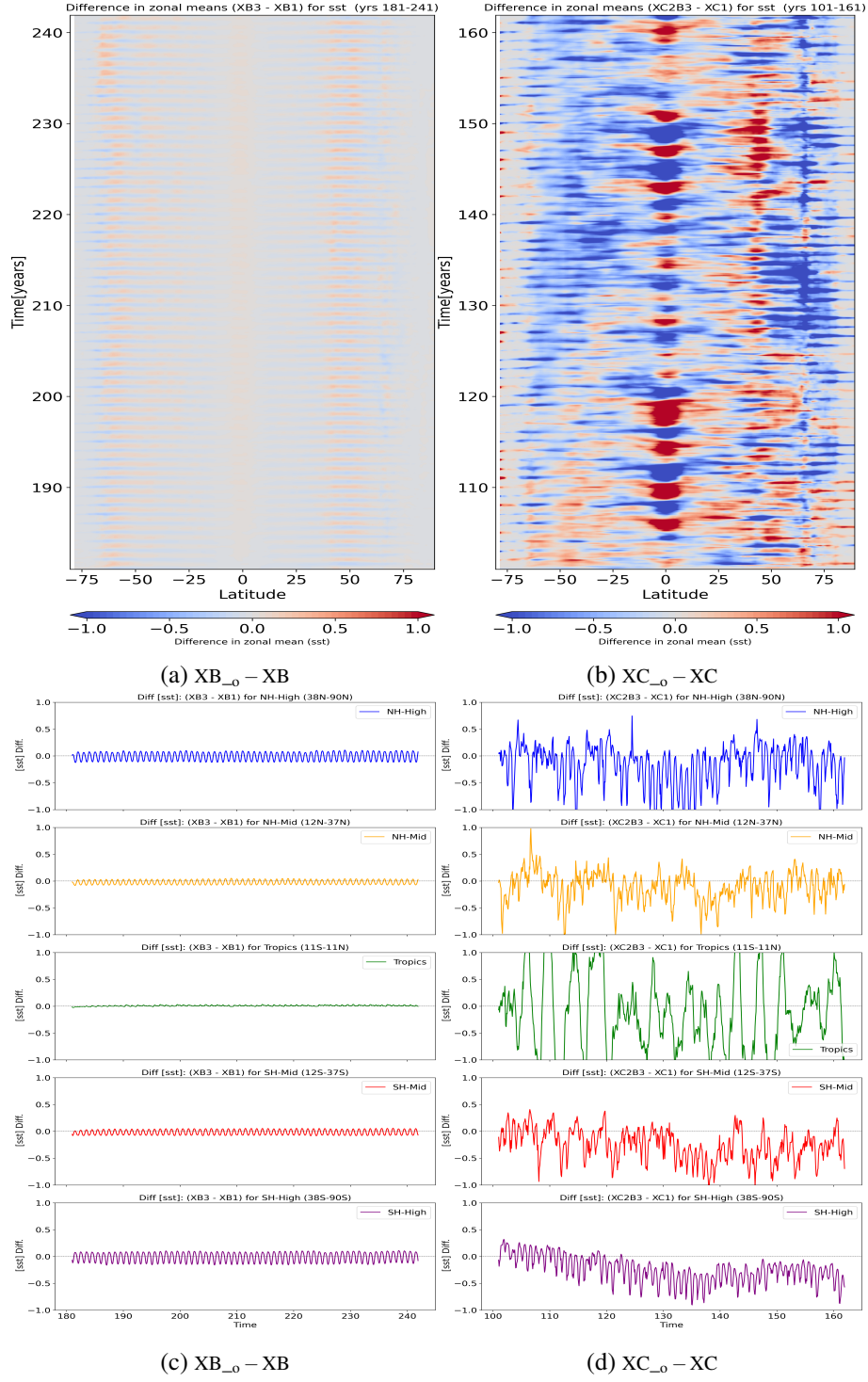


Figure 11: Same as in Fig. 5 but for the impact on the SST.

Results about wave-induced SST cooling during summer and warming during winter are also reported by [Li et al., 2025, see their Figs. 1&2]. The enhanced LT mixing trig-

gers downwards heat transport that cool SST and warm subsurface water during summer, while during winter it triggers upwards heat transport that warm SST. This appears to be prominent in the **B** experiments. In contrast, in the **C** experiment, during winter, the SST does not seems to get sufficient increase to compensate the summer SST cooling, see maps of the SST difference in Fig. 12, the corresponding zonal-averaged in Fig. 13 and the cross-section plot in Fig. 16.

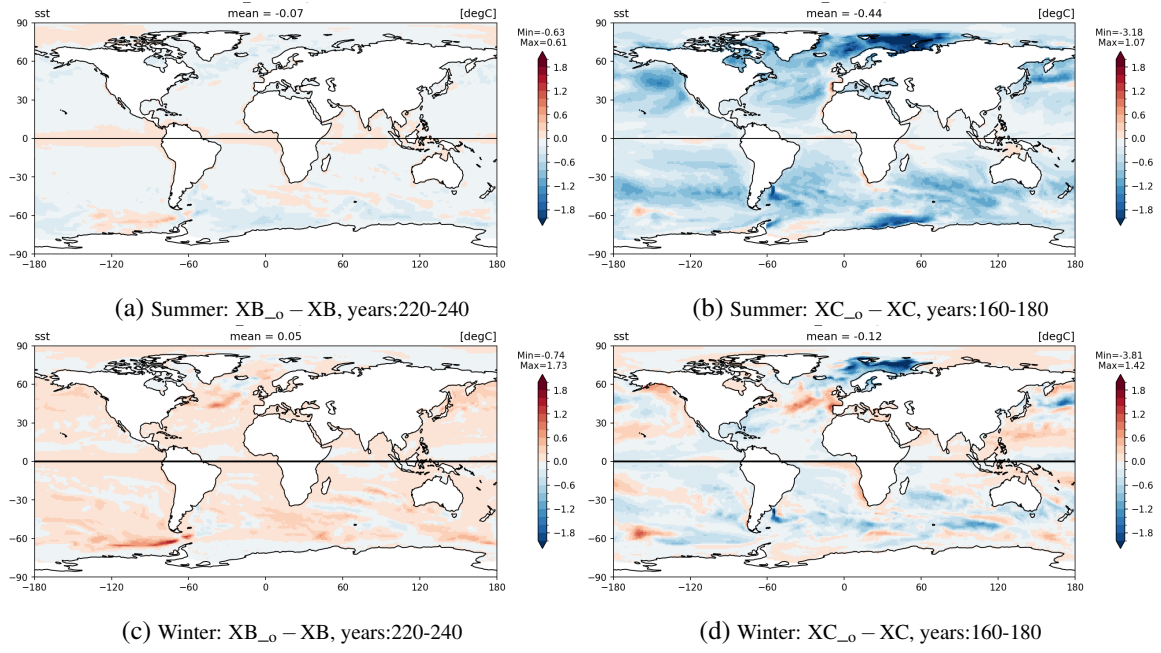


Figure 12: The LT mixing impact on the top) summer-mean and bottom) winter-mean sea surface temperature (SST) for both hemispheres. (Left) JRA-forced runs and (right) fully coupled runs .

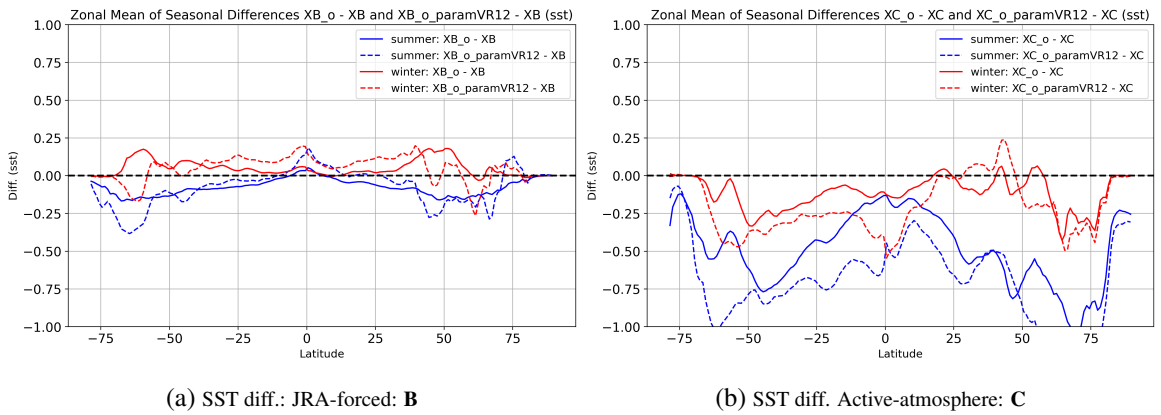


Figure 13: Zonal mean of the differences in the seasonal SST computed from Fig. 12.

Due to the LT mixing enhancement, the amplitude of the SST seasonal cycle has been reduced by approximately 0.11°C on the SH-High zones in the **B** experiments as illustrated in Fig. 15. On the other hand, in the **C** experiments shown in Fig. 14, the amplitude has been reduced by approximately 0.28°C on both the NH-High and SH-High regions. The green curve in Fig. 14 represents the SST seasonal cycle from WOA2018 observations. As shown, the reduction in the SST seasonal cycle amplitude has substantially improved its representation by decreasing the bias, particularly in the all active case: **B**, presented in Fig. 14, where the bias is reduced by about 41% and 55% over the NH-High and SH-High regions, respectively. The reduction in the amplitude bias of SST seasonal cycle, induced by the LT mixing enhancement, is further supported in a fully coupled simulation with LT mixing effect by [Li et al., 2017, see their Fig. 11]. The simulation with the parametrized wave from WTM (prmVR12 type) also shows consistent reduction in the amplitude of the SST seasonal cycle, compared to the case when using the ww3LF17 type where coupling with WW3 is used.

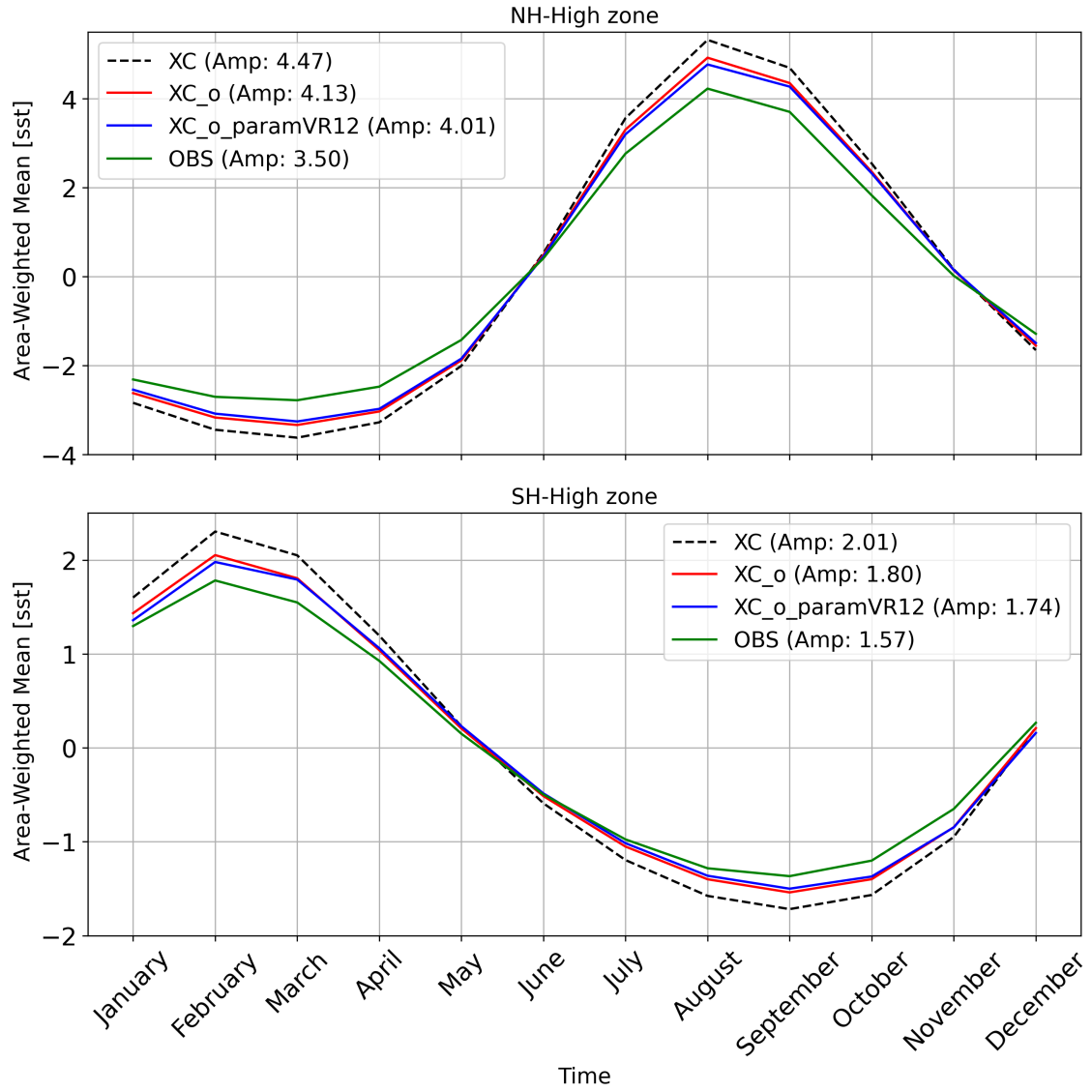


Figure 14: The seasonal cycle of SST averaged over both NH-High and SH-High regions (computed by subtracting the annual mean of the corresponding case), for the active-atmosphere **C** case, years:160-180. SST observations plotted in the green line are taken from the World Ocean Atlas 2018 [Locarnini et al., 2019].

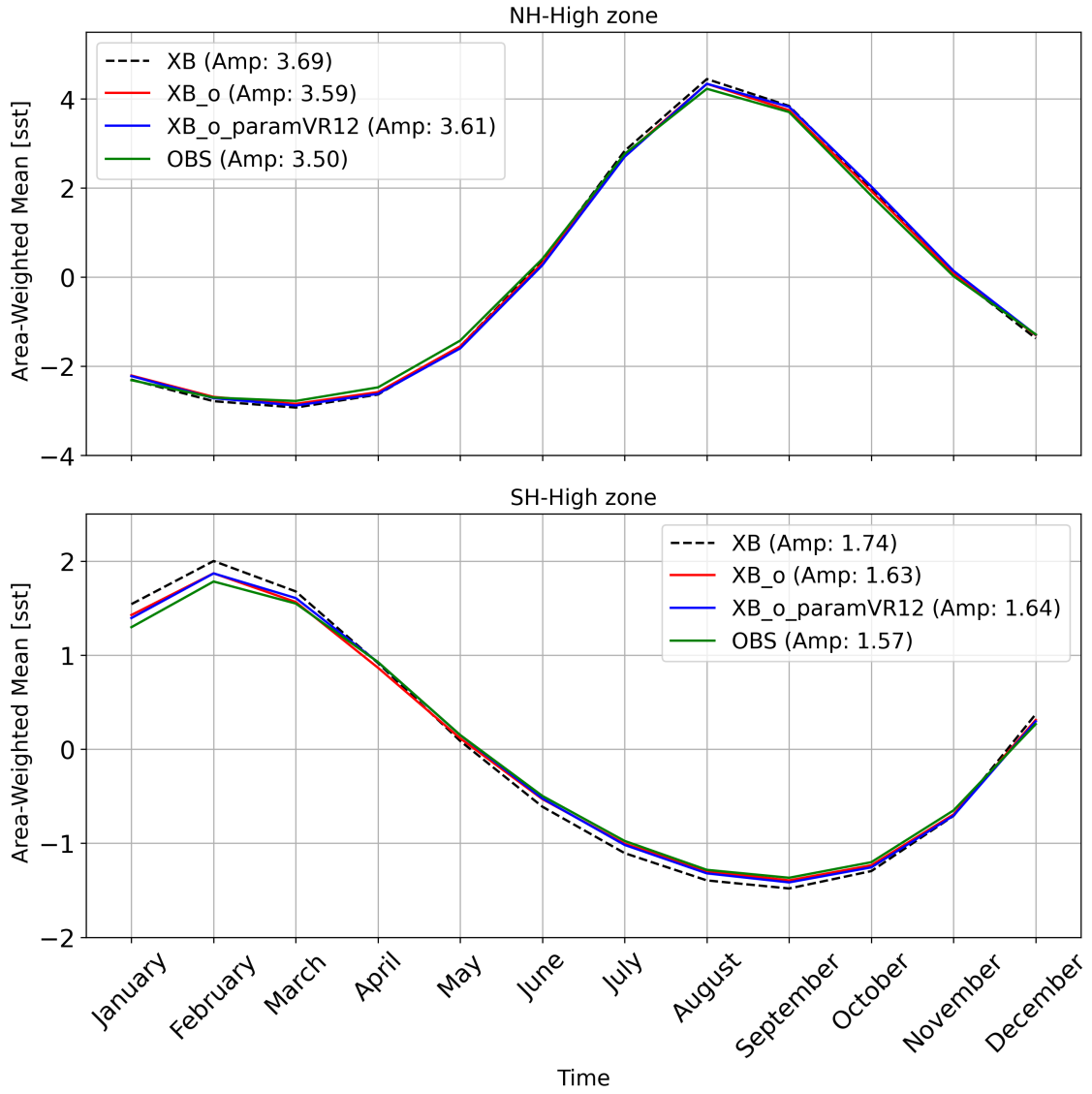


Figure 15: Same as in Fig. 14 but for the SST seasonal cycle in the JRA-forced **B** case, years:220-240

The enhanced vertical mixing transports heat downward, warming the relatively cold subsurface water, as clearly demonstrated in the cross-section (depth,latitude) of the seasonal mean temperature in Fig. 16 and the Hovmöller plot (depth,time) in Fig. 17.

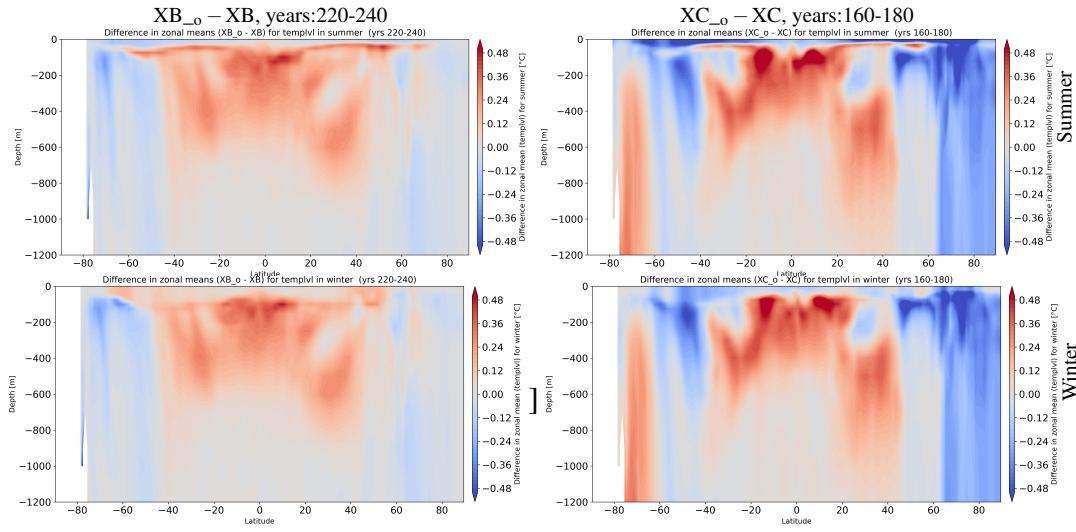


Figure 16: Cross-sections illustrating the differences in the zonal-mean global ocean temperature for top) summer-mean and bottom) winter-mean conditions. The left panels show $XB_o - XB$, while the right panels show $XC_o - XC$.

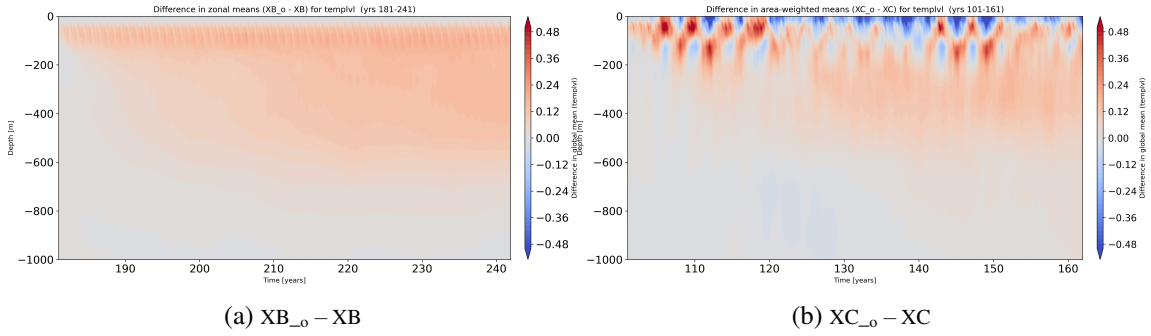


Figure 17: Hovmöller plot illustrating the temporal evolution of the differences in the area-averaged global ocean temperature. Left: $XB_o - XB$ and right $XC_o - XC$.

To examine how the enhanced vertical mixing of temperature manifests locally, the evolution of the differences in temperature are computed for the regions NH-High, NH-Mid, Tropics, SH-Mid and SH-High and presented in the Hovmöller plot shown in Fig 18.

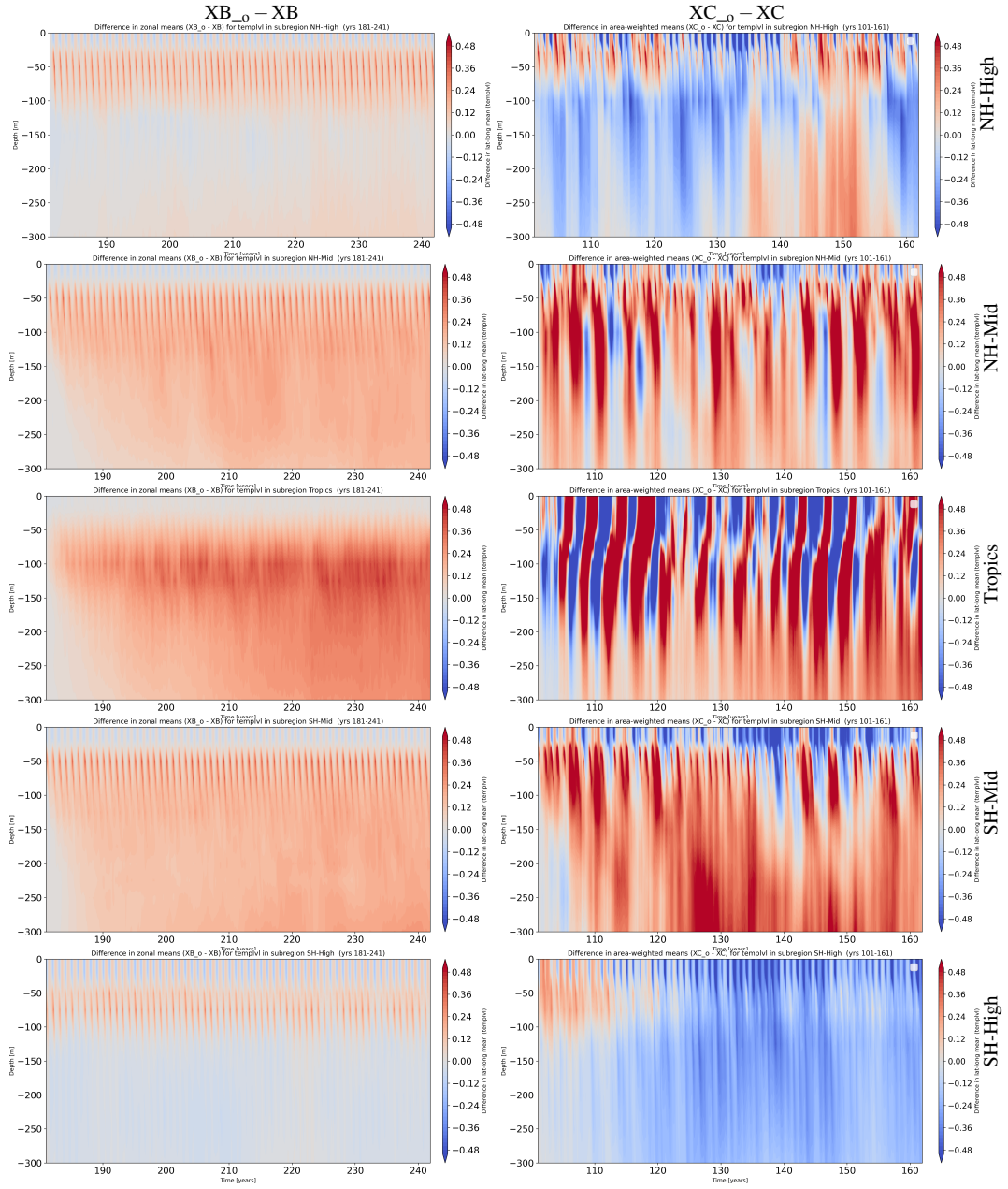


Figure 18: Hovmöller plots illustrating the temporal evolution of the area-averaged differences in the ocean temperature over subregions R1 to R5, where the left panels show $XB_o - XB$ and the right panels show $XC_o - XC$. The black and green lines represent the area-averaged MLD.

Extra simulations have been conducted in order to perform a pure comparison between coupling with WW3 vs using the WTM. Figure 19 shows the impact of the LT mixing on the zonal-mean global ocean temperature, within depth of 1200m, when coupling to WW3 (ww3VR12) compared to using parametrized waves from WTM (prmVR12), for both summer-mean and winter-mean. Here, almost the same LT mixing parametrization

is used and the only difference is in the consideration of wave-wind misalignment and the source of wave: WW3 or WTM. As can be seen, it clearly demonstrates the using WTM is able to reasonably capture the influence of LT mixing enhancement during both summer and winter, and can therefore replace coupling with WW3.

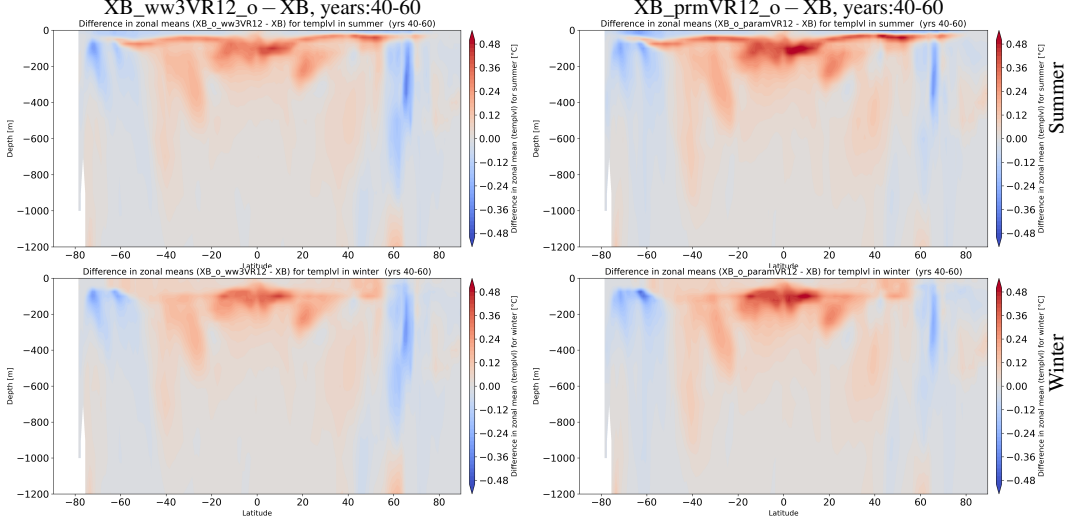


Figure 19: Cross-sections illustrating the differences in the zonal-mean global ocean temperature for (top) summer-mean and (bottom) winter-mean conditions from 20-year-average climatology. The used wave sources are: (left) WW3 and (right) WTM.

Following the analysis of [Li et al. \[2025\]](#), see their Fig. 3(a)], the seasonal upper ocean temperature difference in zone 80°S - 40°S between XB_{o} and XB is shown in Fig. (20). In [Li et al. \[2025\]](#), wave-induced mixing is parameterized by adding a turbulent kinetic energy production term, based on [Ghantous and Babanin \[2014\]](#), to the energetics based Planetary Boundary Layer (ePBL) scheme [[Reichl and Hallberg, 2018](#)]. Furthermore, they show the temperature difference in year 2 after wave-ocean coupling, while Fig. (20) is an average for years 31-60 after the onset of wave-ocean coupling. Despite the differences in vertical mixing scheme (KPP vs ePBL), wave-induced mixing parameterizations and averaging period after wave-ocean coupling, the temperature responses are very similar. Due to the wave-induced mixing, the summer heating and winter cooling are distributed over a thicker ocean surface mixed layer, leading to a colder and warmer SSTs during summer and winter, respectively.

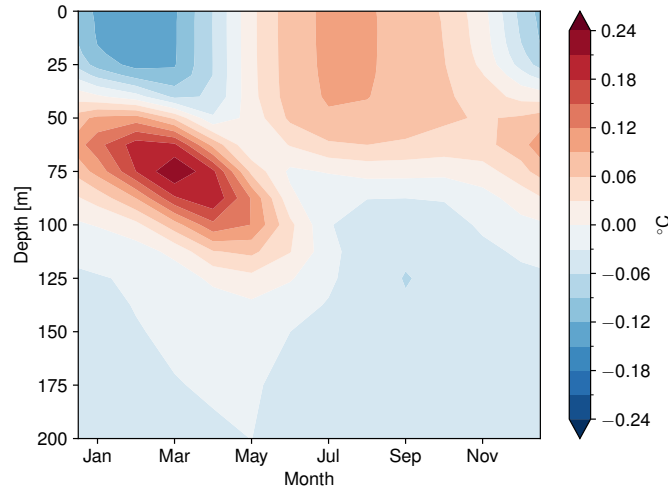


Figure 20: Annual cycle of upper ocean temperature differences between XB_o and XB averaging years 31-60 after wave-ocean coupling in the zonal band 80°S - 40°S .

Figure 21 shows the annual cycle of upper ocean temperature response due to wave-ocean coupling in the subregions of Fig. 2. Responses 1-10 and 31-60 years after the onset of wave-ocean coupling are shown for both XB and XC experiments. Structurally similar temperature response in XB and XC experiments during years 1-10, but generally stronger temperature differences in fully coupled simulations. Notable differences are cooling below the mixed layer in $\text{XC}_o - \text{XC}$, region NH-High and Tropics , which is not seen in $\text{XB}_o - \text{XB}$. For years 31-60, the overall cooler SSTs in $\text{XC}_o - \text{XC}$ become evident, consistent with Fig. 12. Particularly for regions NH-High and SH-High , the warm summer and cold winter SSTs relative to control simulations prevail for both XB and XC , although SH-High for years 31-60 of $\text{XC}_o - \text{XC}$ is offset by an overall cooling in the depth range shown.

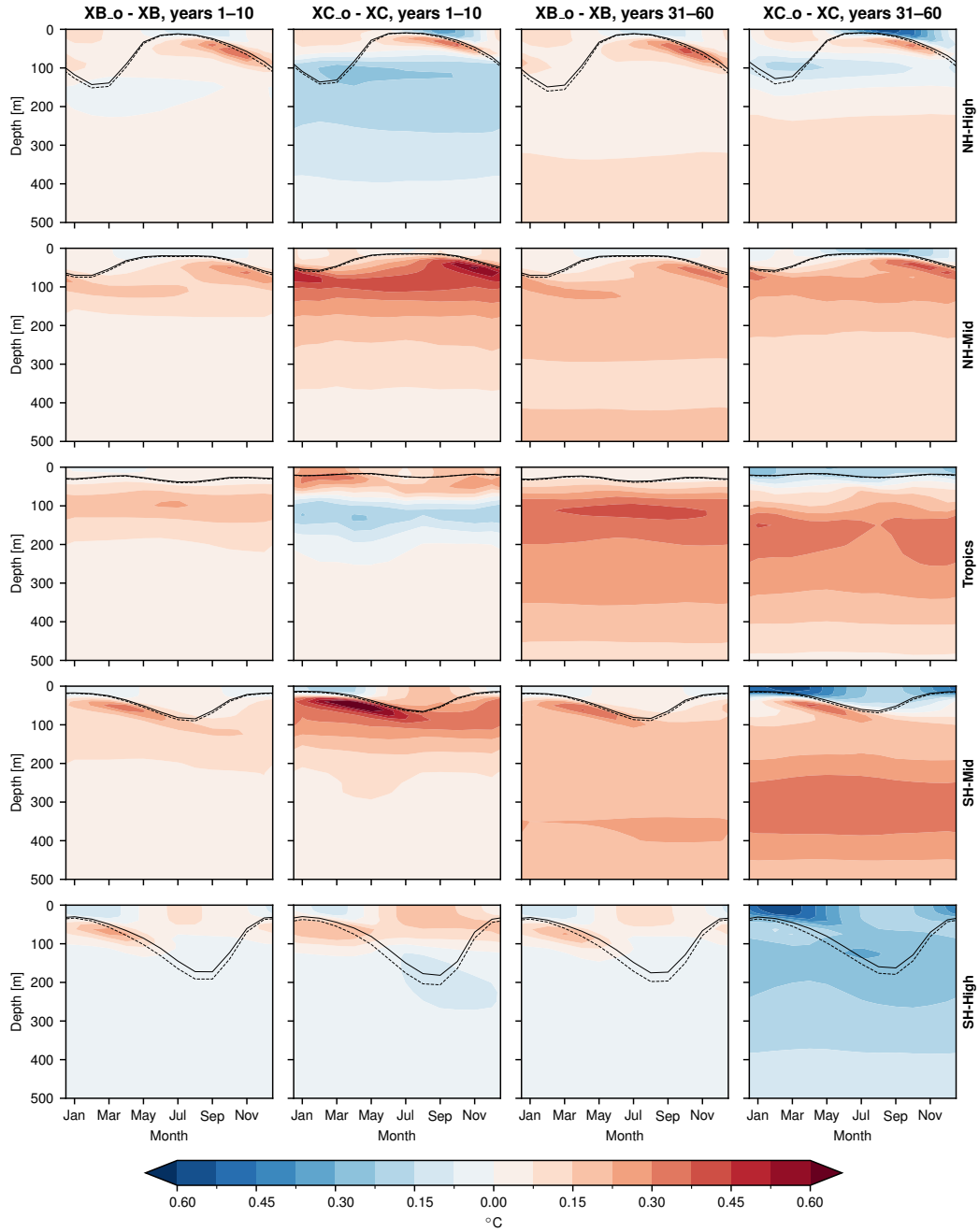


Figure 21: Annual cycle of upper ocean temperature differences averaged over subregions defined in Fig. 2, where regions are shown row-wise. Columns 1 and 3 compare XB_o to XB and columns 2 and 4 compare XC_o to XC . Columns 1 and 2 are the temperature difference for years 1–10 after wave-ocean coupling, while columns 3 and 4 are for years 31–60 after wave-ocean coupling. Solid and dashed lines show mixed layer depth for control and wave-coupled experiments, respectively.

Figure 22 shows XC_o minus XC total, solar and non-solar heat fluxes received by the ocean, separated by southern extratropical, tropical and northern extratropical contribu-

tions. Beyond internal variability, there is no clear systematic response of the wave-ocean coupling on the total heat flux in neither of the regions. However, less solar and more non-solar heat flux is received by the ocean south of 20°S in XC_{o} compared to XC . This southern extratropical response to wave-ocean coupling will be further discussed in Sect. 4.4.

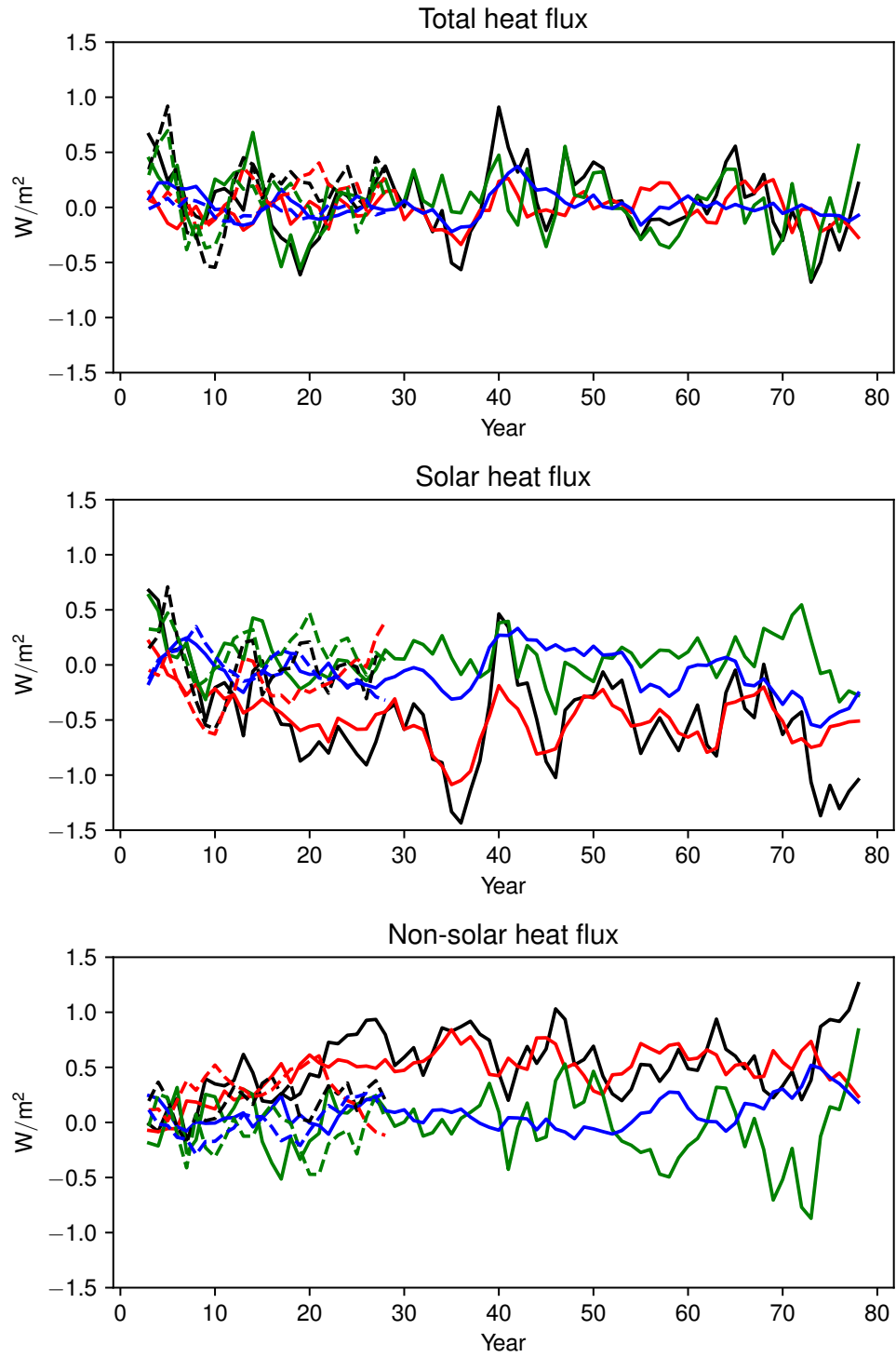


Figure 22: The upper panel shows the time-series of XC_{o} minus XC total heat flux received by the ocean (black line). The lower two panels show the solar and non-solar fractions of the total heat flux. The red, green and blue lines shows the heat fluxes south of 20°S , between 20°S and 20°N , and north of 20°N , respectively. Dashed lines are for the 30-year XC_{o} simulation with different initial conditions. For all time-series, a 5-year running mean filter has been applied.

Figures 23 and 24 show the differences in hemispheric sea ice area and sea ice volume between the $\text{XB}_{\text{o}} - \text{XB}$ and $\text{XC}_{\text{o}} - \text{XC}$, respectively. The results from the **B** experiments show a small, but systematic response to the Langmuir mixing with increased area during summer, and decreased during winter in both hemispheres. The sea ice volume show variations on multidecadal time scale with an increase in the start of the simulation. For the Southern Hemisphere, there is a consistent increase in the volume, especially during spring. Because the sea ice is strongly constrained by the imposed atmosphere which is the same in both **B** experiments, the differences between the experiments are muted, with anomalies in area less than 0.5 mill km^2 .

The **C** experiments with an interactive atmosphere (Fig. 24) show much larger differences between the experiments. The amplitudes of the anomalies are up to 4 mill km^2 in the Southern Hemisphere during late winter (note the difference in scale between Figs. 23 and 24). The response is in overall agreement with the SST-differences shown in Fig. 11. A warm Southern Ocean SST during the first years of the experiments is associated with reduced sea ice, while the colder SSTs later associated with more sea ice later in the simulation. In the NH, the it is warmer with less sea ice during the first 10 years of the simulation. Later, the situation is more variable, but with a general colder state with more sea ice.

$$XB_{_o} - XB$$

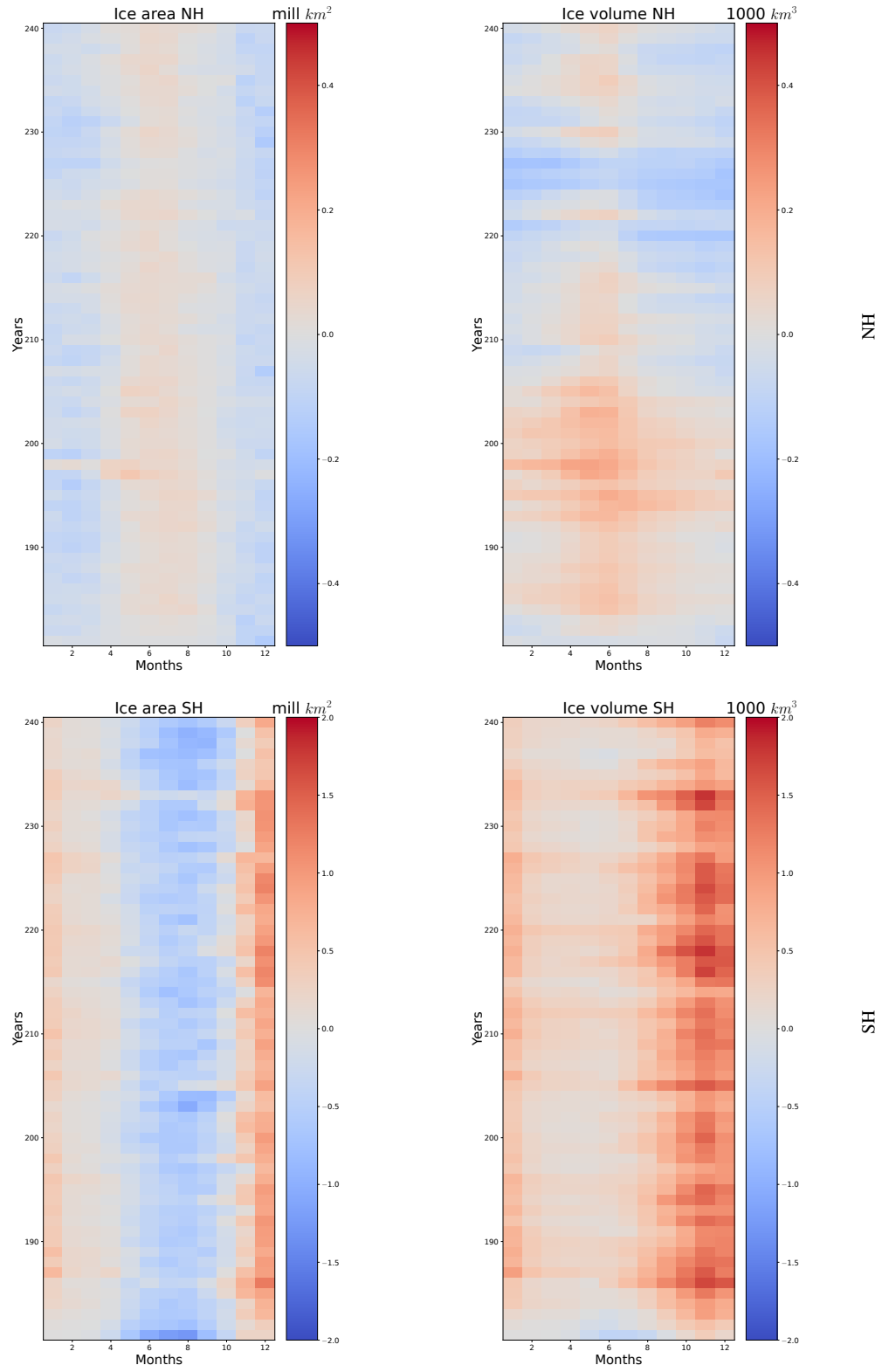


Figure 23: Month — year plots illustrating the temporal evolution of the differences in NH (upper) and SH (lower) sea ice area (left) and sea ice volume (right) between the $XB_{_o}$ and XB .

$$XC_{o} - XC$$

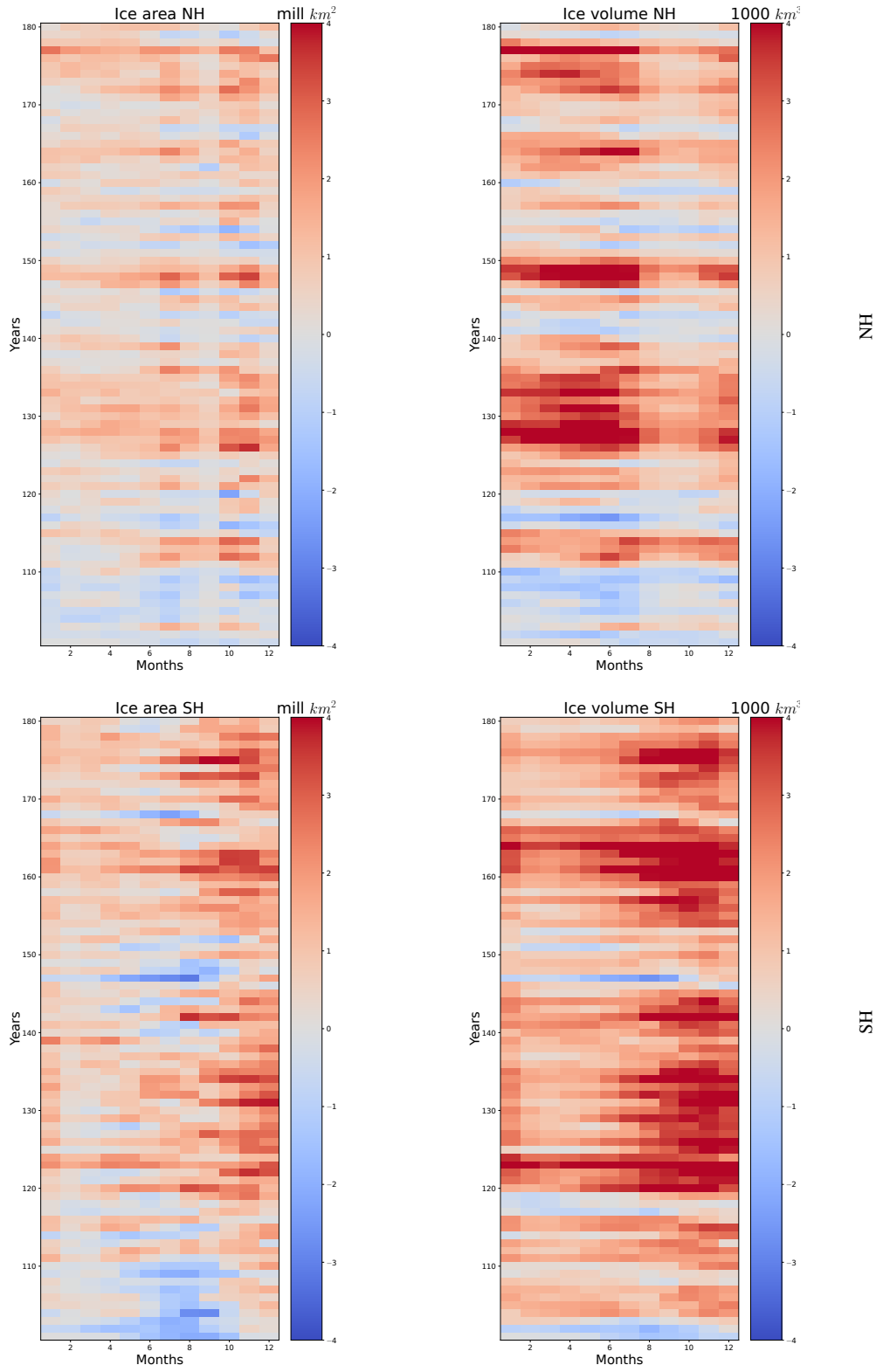


Figure 24: Same as Fig. 23 but for the XC_o and XC experiments.

4.3 Wave-Sea-ice coupling

As expected, the wave – sea ice coupling show small impacts due to the minor change in total stress experienced in the Marginal Ice Zone (MIZ). However, in the **B** experiments (Fig. 25) there is a consistent reduction of the NH ice area during summer, and a reduction in the NH sea ice volume. This indicates increase shortwave absorption in the MIZ, and further reduced winter growth, likely due to more absorbed heat.

$$XB_{-i} - XB$$

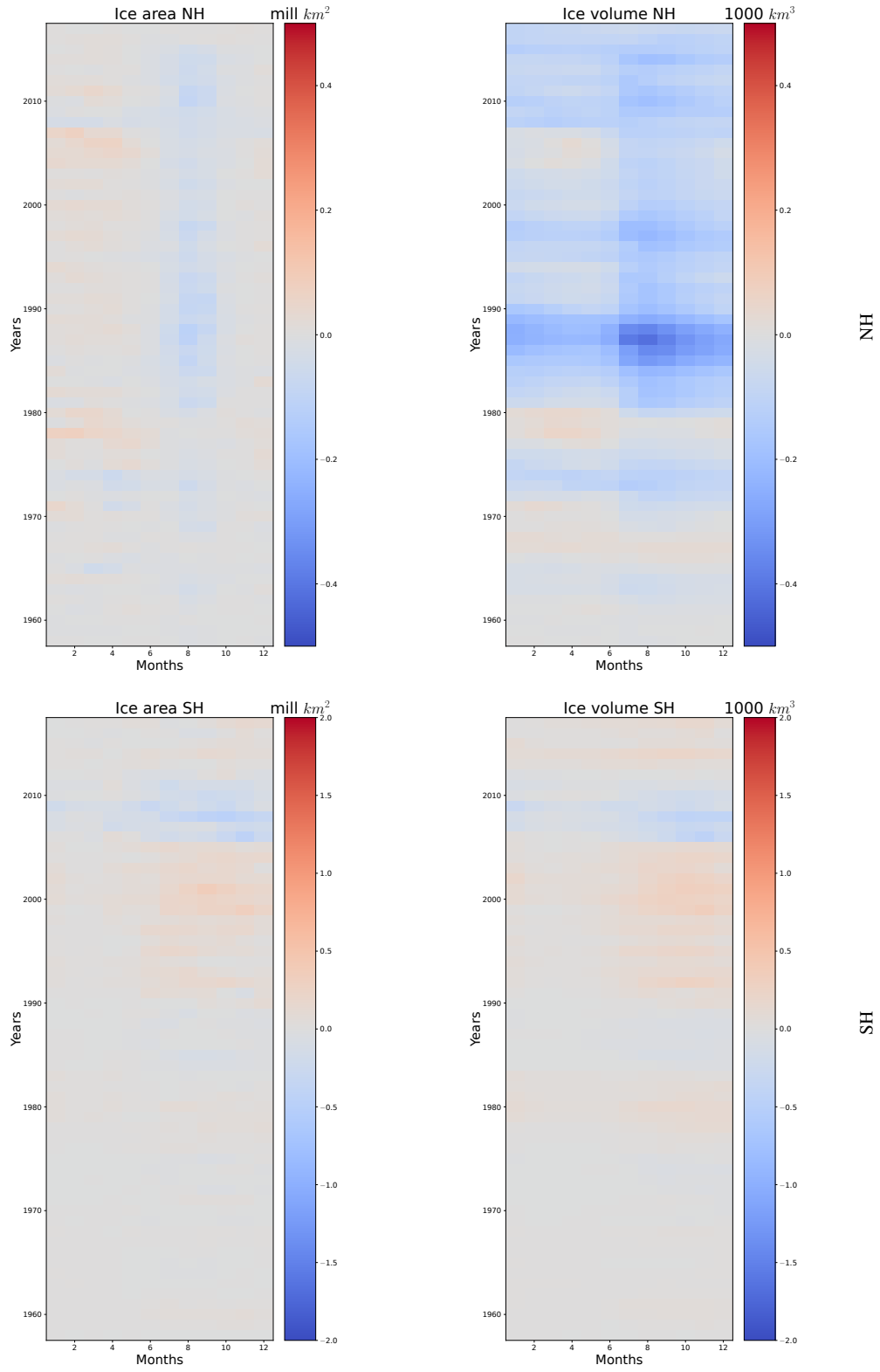


Figure 25: Same as Fig. 23 but for the XB_{-i} and XB experiments.

$$XC_{oi} - XC_o$$

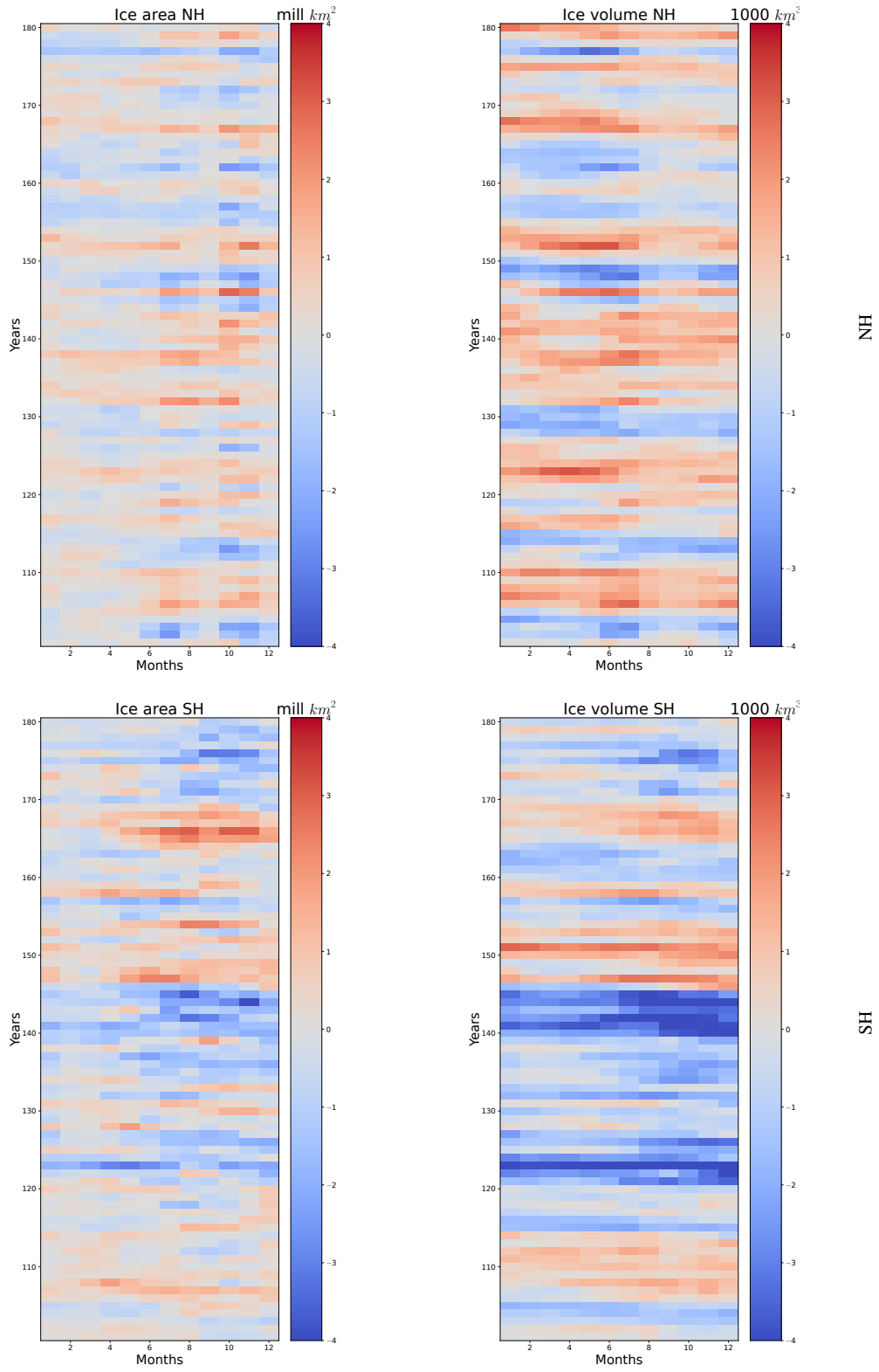


Figure 26: Same as Fig. 23 but for the XC_{oi} and XC experiments.

4.4 Wave-Langmuir coupling and model climatology

As discussed above, the wave-dependent Langmuir mixing causes a noticeable global cooling in the all active simulations: **C** configurations. This effect is very consistent across all all-active **C** configurations, and is not seen in the **B** configurations, indicating the presence of an atmospheric feedback that causes such cooling.

Analysis shows that the dominant feedback is through sea-ice albedo. Figure 27 illustrates this point. We take advantage of the consistency of the Langmuir effect (or, in other words, of the comparative smallness of the ice-wave coupling and of the wave-Charnock coupling effects) in the all-active experiments by forming a mini-ensemble of five independent differences, namely $XC_{-o} - XC$ (2 cases at different branching dates from the control), $XC_{-oi} - XC$, $XC_{-ao} - XC_{-o}$ (first case, at the same branching date), and $XC_{-aoi} - XC_{-oi}$, and trace the evolution of surface temperature, sea-ice fraction, and surface heat fluxes during the first 20 years of the Langmuir-mixing experiments. The result of averaging over these is shown in the bottom row of the figure. Focusing on the global averages (solid lines) first, it may be seen that sea-ice fraction trends upwards even while surface temperatures do not change. At the same time, net downward short-wave surface radiation trends downwards, along with the total downward surface heat flux (including sensible and latent heats fluxes). It may also be noted that the annual averages (black lines) generally follow the OND averages (green lines). The sign of the sea-ice concentration differences is definite positive in SH spring after year 2 in all experiments, with no such consistency seen in either surface temperatures or surface radiative cloud forcing. Limited to the SH extratropics, Spring season differences clearly indicate a radiation-driven surface heating deficit not related to surface temperature or cloud forcing. The year-to-year variation of SH extratropics, SH Spring season average (green dashed lines) appear to be matched in all other means, including the global annual mean.

We interpret this behaviour as indicative of a sea-ice albedo feedback. It originates in the delayed seasonal cycle of SH sea-ice cover extend, exemplified in Fig. 28, which in turn is probably linked to slower warming of the (deeper) ocean mixed layer during spring. This effect does not depend on coupling with the atmosphere, and is also evident in the **B** experiments (cf. Fig. 29).

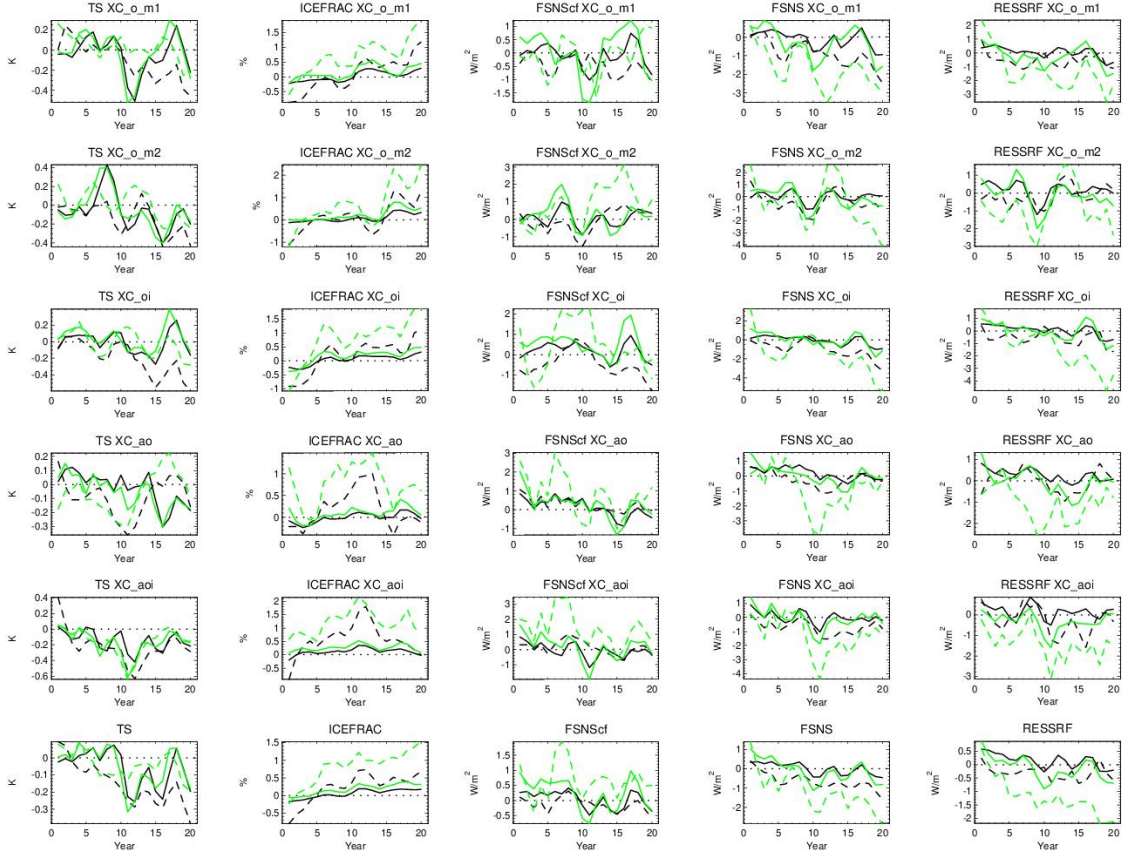


Figure 27: Evolution of surface temperature (TS, K), sea-ice fraction (ICEFRAC, %), net downward surface short-wave cloud forcing (FSNScf, W/m^2), net downward surface SW (FSNS, W/m^2), and total downward surface heat flux (RESSRF, W/m^2) for the first 20 years of all-active experiments, as differences from the respective control experiments (XC for experiments without wave-Charnock parametrization, XC_{ao} for experiments with wave-Charnock parametrization) that do not include Langmuir mixing. The experiments are ordered by row, with their name indicated at the top of each panel. The last row of plots is the average among the five experiments in the first five rows. Solid lines are used to indicate global averages, and dashed lines to indicate spatial averages over the SH extra-tropics ($38.8^\circ\text{S} - 90^\circ\text{S}$) only; black lines indicate annual means, and green lines OND means.

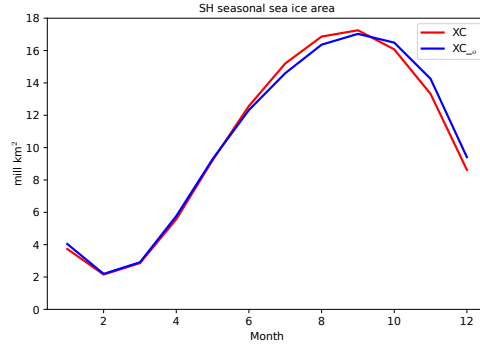


Figure 28: Monthly climatology of SH sea-ice area in XC_0 and XC based on the first 20 years of the simulations.

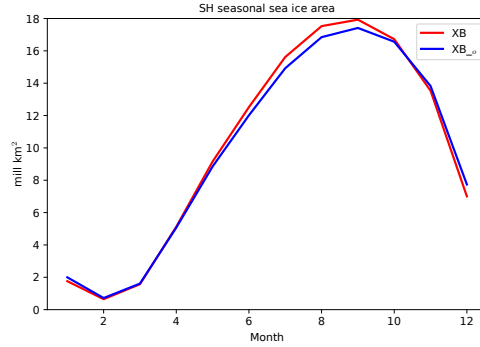


Figure 29: Same as Fig. 28 but for the XB_0 and XB experiments.

A delay in spring-time sea-ice melt, even if it were compensated by similar delay in its formation during early winter, without any change in annual-mean albedo in the annual average, inevitably leads to a net radiative forcing, because insolation is higher in Spring than in Winter. Therefore, the spring-time sea-ice albedo feedback is uncompensated in the annual average. This effect should be present also in the **B** experiments, but in the JRA configuration non-radiative surface fluxes always compensate for temperature differences between the surface and the prescribed 2-meter air temperature, compensating any radiative flux deficits.

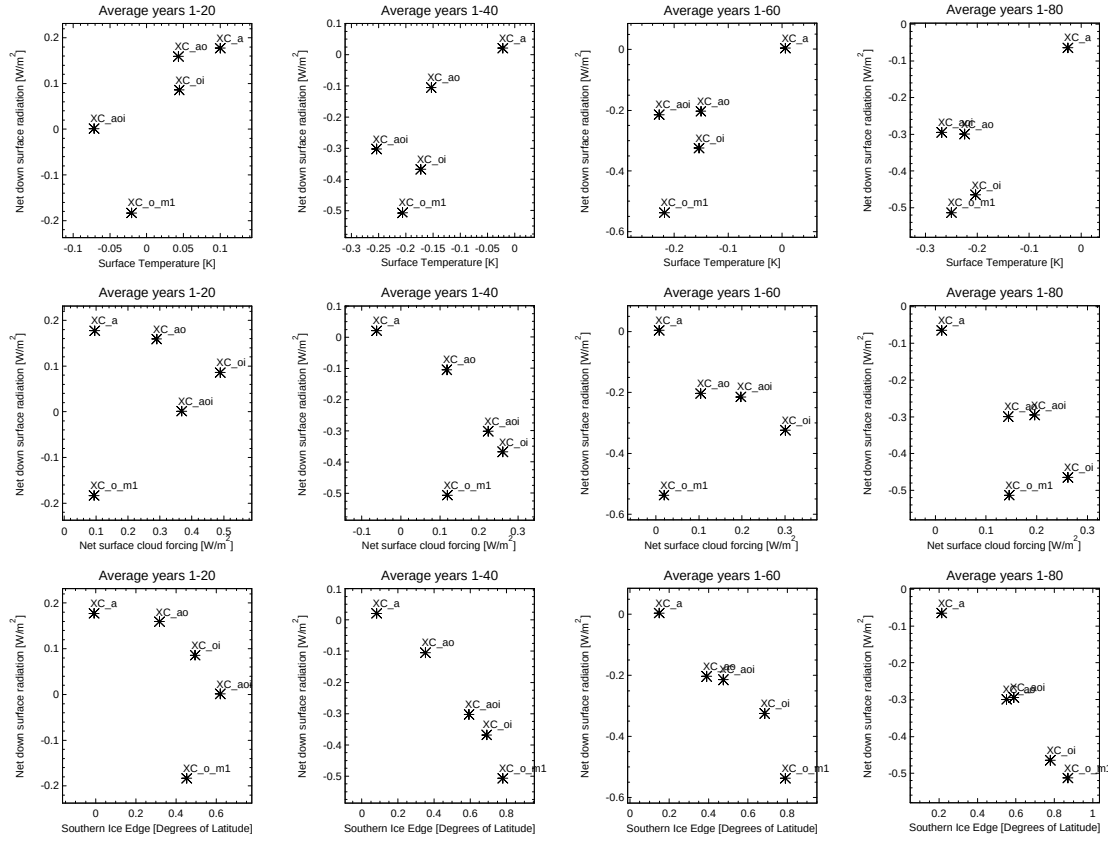


Figure 30: Scatter plots of global average net downward surface short-wave radiation (W/m^2) against surface temperature (K; top row), cloud radiative forcing (W/m^2 , middle row), and mean northern latitude of SH sea-ice edge (degrees of latitude, bottom row). OND means averaged over the first 20 (left), 40 (left of centre), 60 (right of centre), and 80 (right) years.

In the all active experiments, by contrast, the SH spring-time sea-ice albedo effect is left uncompensated throughout our simulations, with no signs of intervention by other potential restoring feedbacks. Figure 30 demonstrates this by showing the transition from initial adjustment to model climatologies. The total net downward short-wave radiation is permanently reduced and correlates linearly with the SH spring-time northern sea-ice edge in the climatologies across our five all-active experiments.

We note that this result is slightly unusual, as frequently cloud-radiative effects are seen to matter in terms of the simulated climatology when the model physics is modified. Moreover, the seasonal variation in surface temperature differences seen in the **B** experiments would justify the hypothesis that a cloud response should have an influence, as greater summer-time surface stability may promote low-cloud formation and cause negative (cooling) cloud radiative effects, while reduced winter-time stability may have a much smaller net cloud radiative effect. In order to test this possibility, which the analy-

sis presented in this section so far cannot completely exclude, we have performed a slab experiments in an XA-like model configuration, i.e. with slab Q-fluxes obtained from an integration with relaxation towards the XA SST climatology; and subsequently increased the slab thickness, without changes in Q-fluxes or prescribed sea-ice concentrations, in a follow-on experiment. The result was negative, i.e. in the absence of a sea-ice albedo feedback, the annual average global mean temperature of the slab was found to be insensitive to the summer SST cooling and winter SST warming caused by its increased thermal inertia.

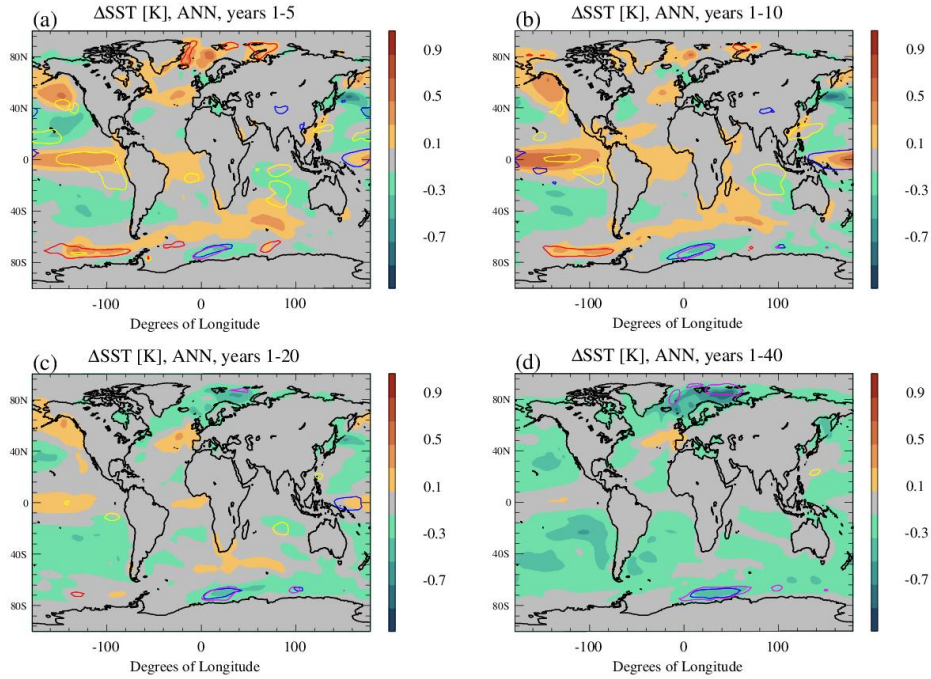


Figure 31: Annual-mean, ensemble-average differences of SST, sea-ice concentration, and net downward short-wave flux climatologies over different averaging period. The "ensemble" is represented by all available XC_xo-XC_x differences where may be nil, 'a', or 'ai'. Colour-filled contours represent SST differences, in K, with the scale shown by the colour-bars. Yellow and blue contours represent plus and minus 4 W/m^2 net downward surface short-wave radiation, respectively. Red and magenta contours represent plus and minus 4% sea-ice concentrations, respectively.

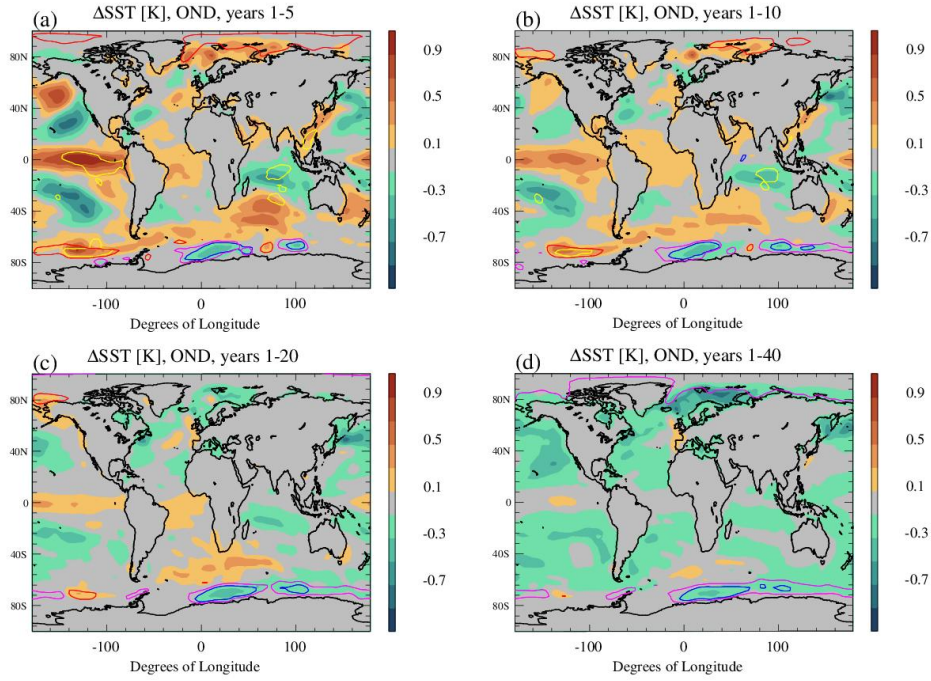


Figure 32: Like Figure 31, but for averages over southern-hemisphere Spring (OND) only. The contour levels for net downward surface short-wave radiation however are 10 W/m^2 here.

It is of some interest to note the regional pattern of increased sea-ice and reduced surface net SW radiation. Figure 31 illustrates this. The model consistently places co-located, positive sea-ice fraction and negative net downward short-wave surface radiation anomalies north of East Antarctica, just east of the Weddel Sea. Excess sea-ice in the NH is not associated with a matching short-wave forcing. Indeed, with Fig. 32 we reiterate the point that the annual means depend almost entirely on such effect during the SH spring season.

Figure 33 shows the Southern Ocean spring (OND) response of MLD, SST and sea-ice fraction due to wave-ocean coupling for years 1-20 in C experiments. The change in SST and sea-ice fraction corresponds well with what already shown in Figs. 31 and 32. For MLD, large isolated differences are found west of Weddell Sea, indicating differences in polynya-like episodes. Strongly negative net downward short-wave surface radiation anomalies of Fig. 32 are indeed co-located with these large MLD difference regions, although the anomalies extend further east where there also is increased sea-ice fraction in XC₀ compared to XC. Realism of the occurrence of polynyas in these simulations has not been fully assessed, and ongoing work with the development version of BLOM has

revealed issues with physics parameterizations that might trigger polynya-like episodes in an unrealistic manner. The sensitivity of the simulated global climatology to this particular, limited region of the Southern Ocean requires further investigation, as we cannot completely exclude that it may be a model artifact.

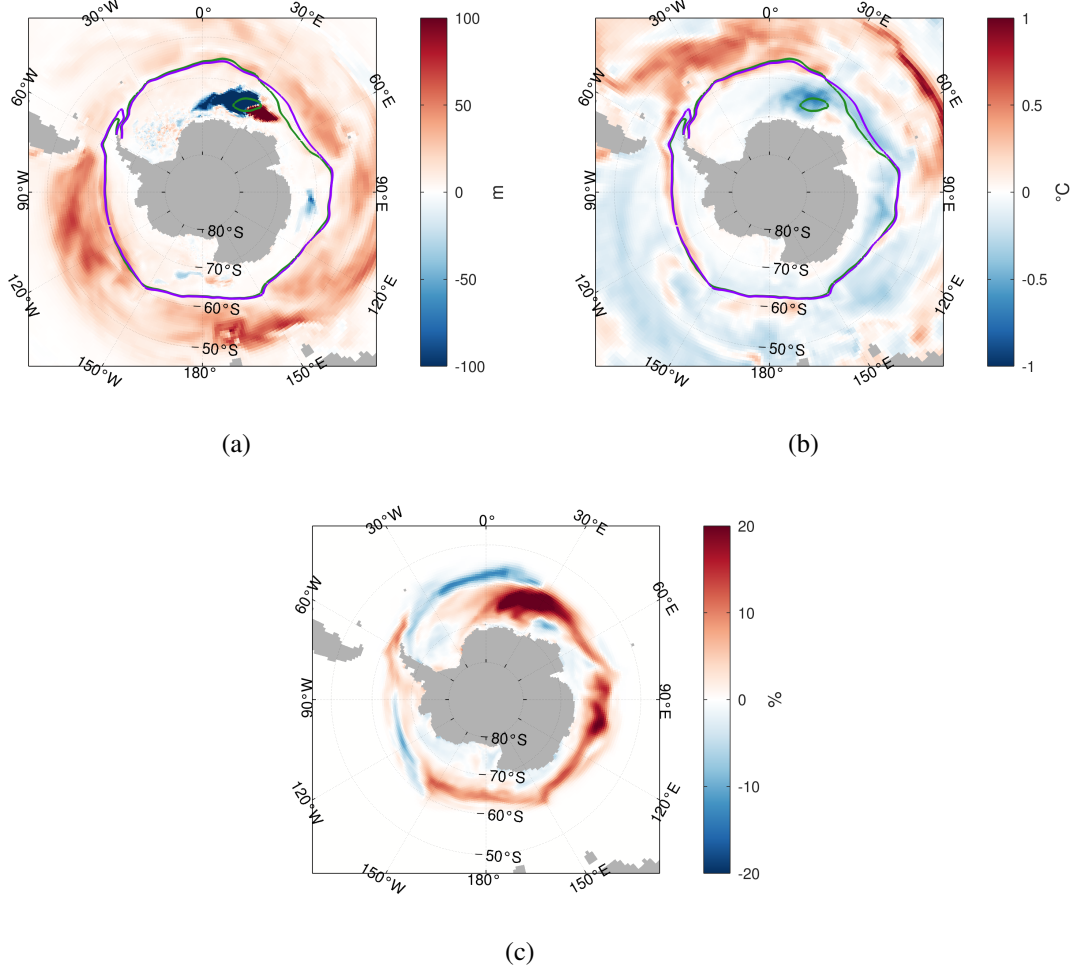


Figure 33: Southern Ocean spring (OND) difference of (a) MLD, (b) SST and (c) sea-ice fraction between XC_o and XC for years 1–20. In (a) and (b) the contours of 15% sea-ice fraction of XC (green lines) and XC_o (violet lines) are also shown.

4.5 Wave-Charnock coupling and model climatology

Contrary to the consistent effects of wave-dependent Langmuir mixing in the ocean, the effect of the wave-Charnock parametrization in all-active simulations is small and, for the most part, entirely negligible.

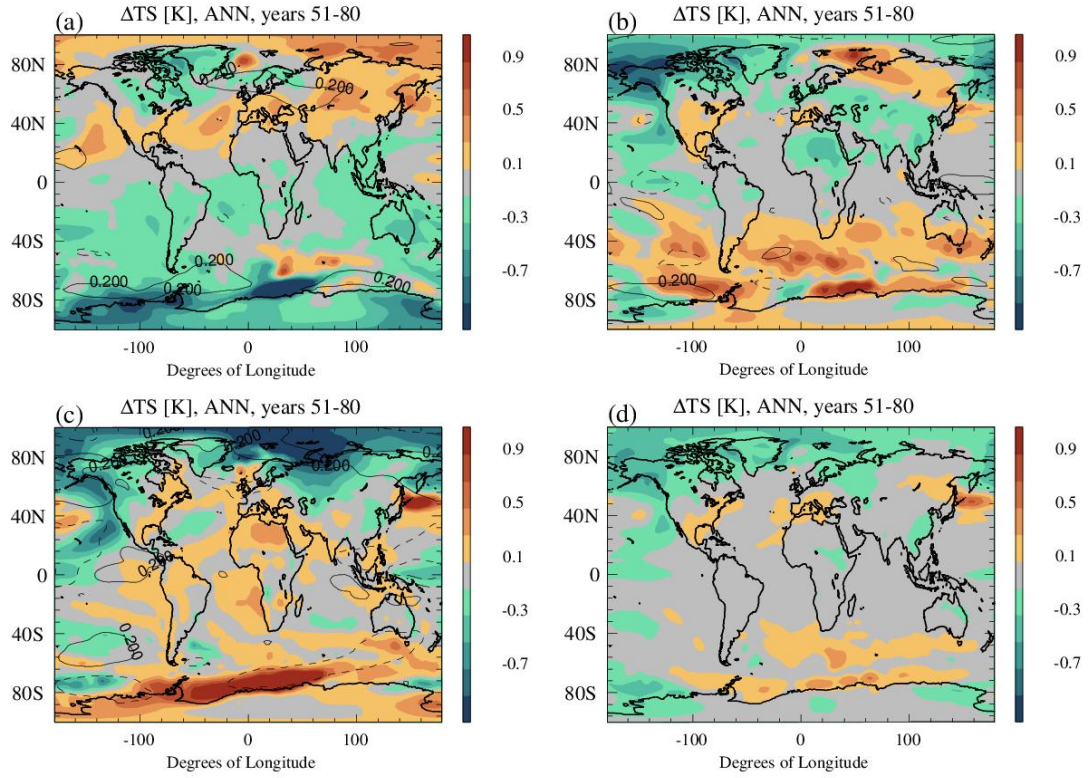


Figure 34: Difference in mean annual climatology of surface temperature (colour-filled contours, interval 0.2 K; orange-red for positive values, cyan-blue for negative values) and 850 hPa zonal wind (line contours, interval 0.4 m/s; solid for positive values, stippled for negative values) for (a) $XC_a - XC$, (b) $XC_{ao} - XC_o$, (c) $XC_{aoi} - XC_{oi}$, and (d) the average of (a), (b) and (c).

This is exemplified in Fig. 34, where we are again showing a mini-ensemble of simulations by making use of XC_o and XC_{oi} experiments. It is clear that the differences are small and dominated by internal variability; the "ensemble" average suggests a general Arctic cooling in the annual mean, linked with the boreal summer season (not shown), but this result is not stable against different averaging periods. Away from the surface, even at low elevation, any climatological differences disappear entirely, as exemplified in the Figure with the differences in zonal wind at 850 hPa, which remain below 0.2 m/s in absolute value. We note that this outcome is somewhat contrasting with the wave-Charnock effects noted in Sect. 4.1 when SSTs are prescribed. There is however no contradiction, for we have discussed there how such effects are indeed dependent on the prescribed SST climatology. In all-active simulations, SSTs tend to respond to changes in non-radiative surface heat flux anomalies in such a way as to damp them. In this scenario, changes

in the Charnock surface roughness parameter will not be effective in altering the amount of heat extracted from the ocean at a given point, negating the effects under high-wind conditions mentioned in Sect. 4.1.

5 Summary

To explore and understand the effect of modulation by ocean surface waves of air-sea interactions, upper ocean mixing, and the distribution and properties of sea-ice in numerical climate simulations, the prognostic wave model WAVEWATCH III (WW3) has been included in NorESM as an additional model component. Published model representation of the coupling between waves, the atmospheric boundary layer, the ocean mixed layer, and sea ice were employed, and the required state variables between WW3 and the relevant model components were exchanged via the NUOPC coupler. These variables included the significant wave height, the wave-supported stress, the surface Stokes drift velocity, the Langmuir enhancement factor, the Charnock parameter and the wave-ice radiation stress. A corresponding suite of simulations experiments has been conducted to investigate wave-atmosphere coupling using wave-modified Charnock parameter, wave-ocean coupling via Langmuir Turbulence (LT) effects, and wave-sea-ice coupling through wave-ice radiation stress. To distinguish between the direct effects of wave-component interactions and subsequent coupled feed-backs, two separate model configuration were used in addition to the "all-active" configuration, namely a data-ocean atmosphere-wave-land configuration and a data-atmosphere ocean-sea-ice configuration.

In all model configurations, we see significant and physically meaningful effects of wave coupling on the resulting model climatology. However, such effects differ by magnitude and robustness. In particular, the effects of wave-Charnock wave-atmosphere coupling were found to be highly sensitive to the model climatology – that is, the changes in model climatology are primarily dependent on the climatology itself, rather than the modelled additional interaction with prognostic waves. This makes the use of the wave-Charnock parametrization in coupled configuration inadvisable, at least in isolation, because it depends on underlying model systematic biases and risks compounding them.

We could not explore the full waves-sea-ice interactions, as originally intended, due to problems with the CICE code related to the immaturity of the development version we had to use within the time-frame of the project. We only explored the effect of wave-ice radiation stress on the sea-ice margin, but not the probably greater effect of ice-flow

break-up due to wave activity. Thus, even though we can report a consistent "damming" effects of sea-ice due to wave-ice radiation stress, this effects is very small on the scales of the grid spacing of the coupled model configuration, and it cannot meaningfully feed back on the overall ocean and sea-ice state to modify it further.

We do, find a significant and very consistent impact of Langmuir turbulence on the state of the upper ocean, or ocean mixed layer. Moreover, the ocean changes are seen to feed back to the sea-ice, broadly by causing a time delay in the seasonal cycle of sea-ice cover; and this in turn has a small but persistent and measurable impact on the net solar radiation absorbed by the simulated climate system, and therefore on the simulated SST climatology, with a global cooling of several tenths of degree C, and a pattern concentrated around the sea-ice margins.

To understand the added value of a prognostic wave model, we also carried out parallel integration with Langmuir turbulence effects using a wind-wave parameterization (see Sect. 3.1.6), namely the WTM, as a potentially efficient and useful alternative to the WW3 model. The parameterized LT mixing effects when using WTM were found to be very similar to those when coupling to WW3, with significantly less computational cost.

In agreement with previous studies, the impacts due to the inclusion of LT mixing enhancement are mainly deepening of MLD and warming of subsurface water in both the data-atmosphere forced and the all active cases. As a result, the MLD shallow biases are considerably reduced. The data-atmosphere forced cases exhibit SST warming during winter and cooling during summer, while in the all active cases the SST is generally reduced specifically at high latitudes regions. As the MLD deepens, and with increased ocean capacity, the amplitude of the seasonal cycle of the mean SST over the high latitude regions is reduced. As a result, the bias of SST seasonal cycle has substantially been reduced as shown in Fig. 14. The indirect influence of including LT mixing on sea-ice is mainly exhibited via changes in the SST where summer SST cooling increases the sea ice thickness and extent, while winter warming decrease them as in the data-atmosphere forced case. The overall SST reduction in the all-active case increases the sea ice extend and thickness as shown in Fig. 24. The wave-induced effects on the MLD and SST through the LT-modified KPP, which is a first order scheme, resemble those shown by the second order ePBL scheme with a parameterized non-breaking wave mixing in Li et al. [2025].

Acknowledgments

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