



Increased horizontal resolution provides limited benefits for simulated Arctic and Antarctic sea ice climatology and recent extremes in a NEMO4-SI3 hierarchy

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Abstract. We assess the sensitivity of Arctic and Antarctic sea ice simulations to horizontal resolution using an ocean-sea ice NEMO4-SI3 hierarchy at 1°, 1/4° and 1/12° nominal resolutions, but with identical vertical resolution, forced by the ERA5 atmospheric reanalysis. Increased resolution improves selected regional and seasonal features, particularly where ocean circulation strongly controls the sea ice edge. In the Arctic, improvements are modest and mainly occur in the Greenland Sea, where the higher-resolution configurations (1/4° and 1/12°) better represent Atlantic inflow west of Svalbard and reduce the excessive winter ice extent simulated at 1°. In the Antarctic, the response is more widespread. Higher horizontal resolution improves the winter ice edge downstream of major bathymetric features, while the 1/4° configuration also improves austral summer sea ice extent and thickness in the Weddell and Ross Seas. Additionally, the 1/12° configuration enhances vertical mixing with warm subsurface waters, promoting thinner ice and a more poleward ice margin. Overall, across both poles, resolution does not systematically improve sea ice extremes or hemispheric biases and benefits depend strongly on the diagnostic considered.

1 Introduction

The simulation of climate and weather phenomena is moving toward kilometre-scale “digital twins” of the Earth system that fuse high-resolution models, observations and machine learning on exascale computers (Doblas-Reyes et al., 2026; Hoffmann et al., 2023; Schneider et al., 2023; Bonavita et al., 2023). The rise of kilometre-scale Earth system modelling has in part been motivated by the promise that regional-scale biases would be partially eliminated by resolving small-scale processes explicitly rather than by parametrizing them (Shaw et al., 2024). A standard assumption of climate science is that large- and small-scale climatic processes are well separated. Yet, mean state and variance biases seen in climate models – including atmospheric and oceanic circulation patterns, extremes and associated risks – do not necessarily fit under this assumption, and their understanding requires new paradigms in modelling (Shaw and Stevens, 2025). High-resolution modelling offers a straightforward solution to tackling these challenges, but these frameworks are usually associated with computationally-demanding constraints (e.g., large storage space, high CPU time), which calls for carefully designed experiments to test their added value but also to avoid drawing too hasty conclusions about their impact.



The polar regions are recognized as key components in regulating the response of the global climate system to anthropogenic forcing (Williams et al., 2023; Screen and Simmonds, 2010), therefore improving the reliability of polar climate simulations is crucial. Quantifying the benefits of higher ocean-sea ice spatial resolution remains central both for the development of “regular-scale” (1° to $1/4^\circ$, typical of the resolutions of models participating to the Coupled Model Intercomparison Projects) Earth system models and for emerging kilometre-scale “digital twin” frameworks. A large body of work has examined the impact of horizontal resolution in both coupled climate models and atmosphere-forced (i.e., standalone) ocean-sea ice models. The Coupled Model Intercomparison Phase 6 (CMIP6) High Resolution Model Intercomparison Project (HighResMIP; Haarsma et al., 2016) provided the first coordinated multi-model protocol to compare standard (1° ocean) and enhanced ($1/4^\circ$) resolutions in both atmosphere and ocean, explicitly asking whether higher resolution yields a more realistic climate mean state and variability. Subsequent analyses within HighResMIP showed that increased resolution can improve the structure of western boundary ocean currents, the global ocean circulation pathways as well as atmospheric storm tracks, but that improvements in sea ice climatology are often modest and model-dependent (Docquier et al., 2019; Koenigk et al., 2021; Selivanova et al., 2024a, b). Hewitt et al. (2020) synthesized the broader literature, concluding that whether ocean mesoscale eddies are explicitly resolved or parametrised affects not only the ocean mean state and variability but also the coupled response, and that resolution choices will continue to influence projections for the foreseeable future.

In parallel with coupled HighResMIP simulations, a set of traceable, atmosphere-forced ocean-sea ice hierarchies has emerged. The Ocean Model Intercomparison Project phase 2 (OMIP-2) protocol, driven by the JRA55-do atmospheric re-analysis, has been used to compare low- (1°) and high-resolution ($1/4^\circ$ - $1/10^\circ$) ocean-sea ice models across multiple centres (e.g., Chassignet et al., 2020; Tsujino et al., 2018). Chassignet et al. (2020) analysed four pairs of non-eddy and eddy-rich global models and showed that higher resolution generally reduces long-standing sea surface temperature (SST) and sea surface salinity (SSS) biases and yields more realistic western boundary currents and Antarctic Circumpolar Current (ACC) structure, but does not uniformly fix all large-scale errors, including those in sea ice extent and thickness. The ACCESS-OM2 hierarchy at 1° , $1/4^\circ$ and $1/10^\circ$ (Kiss et al. 2020; Bi et al. 2013), the GFDL-OM4 models at $1/2^\circ$ and $1/4^\circ$ (Adcroft et al., 2019) and a recent work with AWI-CM3 (Zapponini et al., 2026) likewise demonstrate substantial improvements in straits, boundary currents, mixed-layer depths and some aspects of sea ice at medium ($1/4^\circ$) to high ($1/10^\circ$) resolution, while also revealing sensitivities to Gent-McWilliams-based parametrizations and vertical mixing parametrizations.

Within the Nucleus for European Modelling of the Ocean (NEMO) family, Megann et al. (2014) introduced the eddy-permitting GO5.0 configuration. Storkey et al. (2018) later developed the GO6/GO7 hierarchy at 1° , $1/4^\circ$ and $1/12^\circ$, emphasizing traceability across resolutions and documenting reduced Southern Ocean biases and improved sea ice when moving to eddy-permitting resolutions. Building on GO6/GO7, Guiavarc’h et al. (2025) developed the recent GOSI9 suite using NEMO 4.0.4 coupled with the Sea Ice modelling Integrated Initiative (SI3) sea ice model at 1° , 0.25° and $1/12^\circ$ under common CORE2 forcing to evaluate how updated physics and numerics affect ocean and sea ice climatology. Iovino et al. (2023) similarly compared ORCA1, ORCA025 and global $1/16^\circ$ NEMO-SI3 configurations forced by JRA55-do, showing reduced temperature and salinity biases and improved circulation at higher resolution, though challenges persist in simulating the Arctic halocline.



OMIP-style studies indicate that enhanced ocean resolution primarily benefits regions characterized by strong mesoscale activity – e.g. Gulf Stream, Kuroshio, ACC and Arctic through-flows. Treguier et al. (2023, 2025) showed that higher resolution reduces some winter mixed-layer depth (MLD) biases in deep-water formation regions and modifies multi-decadal trends. Regional analyses similarly highlight improvements in ocean boundary currents and dense-water processes at higher resolution (e.g., Aksenov et al., 2016; Dufour et al., 2017; Wekerle et al., 2017; Danek et al., 2019; Schmidt et al., 2025).

Sea ice, however, is more tightly constrained by atmospheric fluxes in forced setups (Goosse et al., 2023). Biases in sea ice extent and thickness are often linked to forcing uncertainties and parameter choices (Hunke, 2010; Ilıcak et al., 2016; Jackson et al., 2023; MacDermid et al., 2025). Similarly, Lin et al. (2021) showed that higher ocean resolution sometimes improves ice-edge location and drift, but that atmospheric forcing dominates skill. High-resolution ice-centric studies also show a “saturation scale”: refining the grid mainly affects deformation statistics rather than large-scale climatology (Hutter et al., 2018; Williams and Tremblay, 2018; Xu et al., 2021). Previous downscaling studies emphasize that marginal ice zone structure improves with resolution, but large-scale biases remain (e.g., Rinke et al., 1999; Maslowski and Lipscomb, 2003; Dorn et al., 2019). Meanwhile, ocean heat transport through key gateways strongly shapes sea ice, including Fram Strait, Nordic Seas, and the Barents Sea (Årthun et al., 2012; Wekerle et al., 2017; Shu et al., 2019), therefore improvements in ocean circulation due to refined resolution remains a key aspect to focus on.

Nonetheless, systematic multi-resolution forced model hierarchies that focus on both poles remain rare, despite ACCESS-OM2, OM4 and GLOB16 providing important precedents (Adcroft et al., 2019; Kiss et al., 2020; Iovino et al., 2023). Here we evaluate a traceable hierarchy of global NEMO–SI3 configurations at 1° , $1/4^\circ$ and $1/12^\circ$, forced identically and keeping tuneable parameters as identical as possible (more details in Section 2). Building upon GO6/GO7 and GOSI9 design principles, we examine climatological sea ice thickness, extent, drift, the ice edge structure and ice melt/freeze onset using standard observational datasets, building on a recently developed Python diagnostic tool (SITool: Lin et al., 2021). We briefly also assess ocean hydrography and oceanic heat transport diagnostics. Our analysis addresses the following questions:

- Can increased horizontal oceanic resolution improve sea ice climatology in standalone configurations (forced by atmospheric reanalysis)?
- Where and when do we see benefits of increased horizontal resolution for ocean-sea ice climatology, and which regions appear less sensitive to the resolution?
- Is there evidence of a “saturation scale” beyond which refinement (e.g., from $1/4^\circ$ to $1/12^\circ$) yields diminishing returns?

In this work, we quantify what is gained – specifically for sea ice and high-latitude ocean climatology – by increasing horizontal resolution in NEMO4–SI3 atmosphere-forced configurations. The remaining of the paper is organized as follows. In Section 2 we describe the model configuration details, the observational datasets used for evaluation and the analysis methods. In Sections 3 and 4 – for the Arctic and Antarctic, respectively – we briefly assess the three configurations against observational reference datasets before investigating the physical mechanisms underlying the most notable differences in the sea ice state across the model configurations. We also analyse the impact of the horizontal resolution on the simulation of selected observed sea ice extremes. In Section 5 we discuss our findings and conclude the paper.



2 Data and Methods

2.1 General description and configuration

In this study, we use the Nucleus for European Modelling of the Ocean (NEMO) version 4.2.2 (Madec et al., 2023), which includes the Sea Ice modelling Integrated Initiative (SI3) model (Vancoppenolle et al., 2023). The model is configured on the global eORCA tripolar grid, with three nominal horizontal resolutions increasing from 1° (111 km at the equator, 49 km at the North Pole) through $1/4^\circ$ (28 km at the equator, 12 km at the North Pole) to $1/12^\circ$ (9 km at the equator, 4 km at the North Pole). The vertical grid is set up on 75 z-coordinate levels, identically for all three configurations, with 8 levels in the first 10 m (24 in the first 100 m), allowing to properly resolve the shallow summer mixed layer in the Arctic Ocean. NEMO is a primitive equation model using a three-dimensional, free surface, hydrostatic, Boussinesq-approximation approach. SI3 is a dynamic-thermodynamic continuum sea ice model, using a two-dimensional elastic-viscous plastic rheology for dynamics and a one-dimensional energy- and salt-conserving method for thermodynamics. In the configuration used here, a subgrid-scale distribution of ice thickness is discretised into five categories and the sea ice-snow medium in each category is further divided into two ice layers and two snow layers. Sea ice dynamic processes simulated by SI3 include horizontal advection, rheology, and ridging/rafting. Regarding the thermodynamic processes, the model includes energy conservation, an explicit representation of sea ice salinity, solar radiation penetration and transmission, a surface albedo function of ice temperature, thickness, snow depth, cloud fraction, and melt ponds, lateral melting through ice floe size parametrisation (dependent on sea ice concentration), snow-to-ice conversion, and level-ice melt ponds including frozen lids. The $1/4^\circ$ and $1/12^\circ$ configurations include open ice shelf cavities, and the model explicitly simulates ice-shelves-ocean interactions; the 1° cannot resolve those cavities due to coarse resolution.

All parameters for the sea ice model are rigorously identical between the three configurations. Ocean model parameters are as similar as possible albeit with some differences in the $1/12^\circ$ configuration to compensate for the lack of representation of mesoscale processes in the 1° and $1/4^\circ$ resolutions. The most important differences include the eddy-induced velocity parametrization (Gent and McWilliams, 1990) and the lateral diffusion of momentum, that are turned off in the eddy-resolving $1/12^\circ$ configuration. Moreover, the minimum background vertical mixing coefficient, eddy viscosity and diffusivity are higher in the 1° configuration, although all three configurations compute the actual coefficients using a turbulent eddy kinetic energy (TKE) closure scheme.

2.2 External forcing

All configurations are forced identically at the atmospheric boundaries using the ECMWF ReAnalysis v5 (ERA5, Hersbach et al., 2020). ERA5 provides atmospheric forcing fields (winds at 10 meter: u,v; air temperature at 2 meter: T2M; surface pressure and dew point temperature at 2 meter, from which we derive the specific humidity at 2 meter that we provide to the model: q; downwelling short- and longwave radiations: DSR and DLR; total and solid precipitation; mean sea-level pressure) on a $1/4^\circ$ horizontal resolution regular grid, which are regridded to the 1° , $1/4^\circ$ and $1/12^\circ$ eORCA grids through bi-linear (for scalar fields) and bi-cubic (for wind vectors) interpolation. Fluxes are then computed by NEMO using the NCAR bulk formulae



(Large and Yeager, 2009). ERA5 is known to have a winter warm bias in the polar regions; the associated reduction in winter thermodynamic ice growth arises from the lack of an explicit snow layer over sea ice in the ERA5 surface scheme (Batak and Müller, 2019), as well as from misrepresented cloud properties (Wang et al., 2024b; Zampieri et al., 2023). To account for the impact of this warm bias, we increase the thermal conductivity of snow to $0.5 \text{ W m}^{-1} \text{ K}^{-1}$ in all configurations.

2.3 Initialization and restoring

For the 1° and $1/4^\circ$ configurations, the ocean model starts from rest from 3D temperature and salinity fields obtained from the World Ocean Atlas 2018 (WOA18, references to take from <https://www.ncei.noaa.gov/access/world-ocean-atlas-2018/>) and is run over the period 1960–2024. Because of the higher computational cost, the $1/12^\circ$ configuration is initialized from interpolated temperature, salinity and sea ice (concentration, thickness and snow thickness) fields of the $1/4^\circ$ configuration state in year 1995 and was run until 2024. To allow for spin-up of the $1/12^\circ$ simulations, we perform the comparison analysis between the configurations focusing on the 25-year period covering 2000–2024. A sea surface salinity (SSS) restoring towards the WOA18 data is applied in the form of a corrective surface freshwater flux, with a strength corresponding to a time scale of 227 days for a MLD of 50 m. Such a restoring is commonly applied in forced ocean models, which do not have a closed freshwater budget, and is aimed to prevent unrealistic drifts of the SSS. This flux is scaled by a factor of $1 - c$, with c denoting the sea ice concentration, to progressively disable the restoring in ice-covered areas and avoid interfering with the impacts of sea ice formation or melting on salinity.

2.4 Computational and storage aspects

All simulations were performed on the European High-Performance Computing (EuroHPC) supercomputer LUMI, located in Finland. The 1° configuration was run on 128 cores (120 for computation, 8 for input-output management), with an average resource use of 128 core hour per year of simulation. For the $1/4^\circ$ resolution, the simulation necessitated 1280 cores (1257 for computation, 23 for input-output management) with an average of 3851 core hour per year of simulation. Finally, the $1/12^\circ$ simulation, particularly expensive, requested 4480 cores (4406 for computation, 74 for input-output management), with resource usage of 159 704 core hours per year of simulation. Overall, considering queue time, the 65 years of 1° and $1/4^\circ$ simulations took around 3 days and 15 days to complete, respectively. The 30 years of the $1/12^\circ$ simulation took 86 days to complete. In summary, and accounting for all aspects involved in running these simulations (run time and queue time), a four-fold and a twelve-fold increase in horizontal spatial resolutions (i.e., from 1° to $1/4^\circ$ and from 1° to $1/12^\circ$) correspond to a five-fold and sixty-fold increase in physical time needed to achieve one year of simulation. Regarding storage resources, storing one two-dimensional, uncompressed variable at the daily temporal resolution over 1 year of simulation requires 238M for the 1° configuration, 2.5G (i.e., ten times as much) for the $1/4^\circ$ configuration and 24G (i.e., hundred times as much as the 1°) for the $1/12^\circ$ configuration. Sea ice variables used in this study include sea ice concentration (SIC: fraction of grid area covered by sea ice), sea ice thickness (SIT: sea ice volume per grid cell area), sea ice velocities, and the dynamical and thermodynamical flux terms in the sea ice mass balance (ice convergence/divergence, basal-, surface- and lateral melt, frazil- and basal growth, snow-ice conversion, ice formation due to freezing of seawater trapped within porous ridges and sublimation). All sea ice



variables are written out as daily outputs, while ocean-related 3D properties such as salinity and temperature are written out at monthly time steps.

2.5 Observational and reanalysis products used for model evaluation

160 To compare the output of numerical simulations to observational references but also to better highlight the impact of horizontal resolution on the model results, we use daily and monthly SIC products from the National Snow and Ice Data Center (NSIDC) version 6 dataset (Meier et al., 2026). We make use of all three sea ice retrieval algorithms (NASA-Team, Bootstrap and Climate Data Record) to estimate observational uncertainties when comparing our model experiments with the observed sea ice record. For sea ice drift, we use the low-resolution sea ice drift product (OSI-455) from the European Organization for the Exploitation
165 of Meteorological Satellites Ocean and Sea Ice Satellite Application Facility (EUMETSAT) release 1 dataset processed from passive microwave satellite data at a spatial resolution of 75 km on an EASE2 grid at daily temporal resolution (OSI-SAF). Due to constraints on the availability of the OSI SAF product, we focus on the 2000-2020 period when comparing the modelled and observed sea ice drifts for both polar regions. We also use the Polar Pathfinder Daily 25 km Equal Area Scalable Earth (EASE) Grid Sea Ice Motion Vectors, Version 4 from the NSIDC for the period 2000-2024 (Tschudi et al., 2019). This dataset
170 is a compilation of velocity vectors from the International Arctic Buoy Program (IABP) buoy data, satellite observations, as well as NCEP-NCAR reanalysis. This merged product suffers from known artifacts of the method used to incorporate buoy data, especially in terms of ice divergence and convergence (Szanyi et al., 2016). Nonetheless, we keep both products to sample uncertainties in observing sea ice drift in the Arctic (Sumata et al., 2014).

We use monthly mean gridded Arctic SIT (sea ice volume per unit grid cell area) reanalysis for the period 2000-2024
175 derived from the Pan-Arctic Ice Ocean Modeling and Assimilation System (PIOMAS; Zhang and Rothrock, 2003). The model assimilates observed sea ice concentrations and sea surface temperatures, and is driven by daily NCEP-NCAR reanalysis surface forcing fields at a mean ~ 22 km horizontal resolution. PIOMAS has been shown to simulate Arctic SIT within the observational uncertainties of satellite measurements with CryoSat (Landy et al., 2022) and agree comparably well with other observational products (Schweiger et al., 2011). As for Antarctic SIT, we use the Global Ice-Ocean Modelling and Assimilation
180 System (GIOMAS Zhang and Rothrock, 2003) reanalysis, which also shows relatively good agreement with satellite-based products (Shi et al., 2021a; Liao et al., 2022). However, taken the intrinsic difficulties of observing SIT (Kwok and Sulsky, 2010) and the associated uncertainties in both observed and reanalysis SIT products for both polar regions (Zygmuntowska et al., 2014; Kern and Spreen, 2015), the evaluation of the three NEMO configurations' SIT against PIOMAS/GIOMAS is more intended as a way to highlight any differences between the three configurations than a strict model-bias assessment.

185 Temperature and salinity profiles in the Southern Ocean are compared to the Met Office Hadley Centre EN version 4 data set (EN4), a monthly objective analyses incorporating temperature and salinity in-situ profiles (Good et al., 2013). We use an updated version that incorporates XBT and MBT corrections. We compute a seasonal climatology over the 2000-2024 to match the model data.



2.6 Analysis methods

190 When comparing the model configurations' SIC and SIT to the observational reference products, we first remap all – both the model and PIOMAS/GIOMAS – products to the polar stereographic grid provided by the NSIDC at 25 km horizontal resolution (NSIDC-0771; Stewart et al., 2022) using conservative remapping. For the analysis, we used some of the “SITool” Python functions developed by Lin et al. (2021) and the associated code and functions are shared on GitHub (see Code Availability). We consider a regional decomposition of the Arctic Ocean into seven regions following Richaud et al. (2026) (Central Arctic; 195 East Canadian Archipelago; Greenland Sea; Barents-Kara Seas; Siberian Sea; Chukchi Sea and Beaufort Sea) and a modified sectoral decomposition of the Southern Ocean based on Raphael and Hobbs (2014) (Weddell Sea; King-Håkon VII Sea; East Antarctica; Ross Sea; Amundsen Sea and Bellingshausen Sea, see Fig. A1).

We calculate sea ice extent (SIE) as the sum of grid element areas where SIC > 15 %. We primarily assess SIE rather than sea ice area, to acknowledge inherent limitations of the satellite observations (Serreze and Meier, 2019). The sea ice edge is 200 defined as the 15 % ice concentration contour. Arctic and Antarctic sea ice speeds (the magnitude of the sea ice velocity) are computed as the square root of the sum of the squares of the daily horizontal sea ice velocity components for both observations (NSIDC and OSI-SAF) and the model output. Monthly mean sea ice speeds are then derived from the daily mean velocity fields rather than from the monthly outputs, because monthly output can underestimate sea ice drift by a factor of two due to higher temporal variability on daily timescales (Docquier et al., 2017). When plotting the ocean surface and sea ice velocities, 205 we account for the rotation needed to transform vectors from the eORCA Arakawa C-grid to actual geographical (east/north) directions.

Climatologies are computed over the period 2000-2024 and the anomalies are defined relative to that 25-yr climatology. When showing a monthly time-step climatological seasonal cycle, it is either computed as monthly averages of daily data or directly from the monthly mean outputs without any smoothing of the daily climatologies. However, when computing the daily 210 sea ice extent anomalies (relative to the 2000-2024 climatology) to study SIE extremes (such as 2012 in the Arctic: see Section 3.1.5), we apply a smoothing of the daily climatologies; a 365-day day-of-year (DOY) climatology was constructed and then double-smoothed with centred 11-day and 31-day running means. For leap years, no integration is performed for the last day of the year.

We conduct a sea ice mass budget decomposition using the individual flux terms contributing to the sea ice mass balance 215 (following Richaud et al., 2026) to investigate the processes underlying the differences in sea ice seasonal cycles. Ice mass gain or loss can be decomposed into melting and growing fluxes and an ice transport term. The melting flux term can itself be further decomposed into basal, surface and lateral flux terms, and the growth term can be decomposed into frazil and basal growth fluxes, as well as ice formation due to freezing of seawater trapped within porous ridges. Other terms (e.g. sublimation) can be neglected (Keen and Blockley, 2018). The sea ice mass balance sign convention is from the perspective of sea ice: a 220 positive flux means a mass gain, following previous work (Richaud et al., 2026).

Using daily SIC, we also conduct a comparison between the configurations' and the observed (NSIDC) melt and freeze onset climatologies. We apply slightly different methods – but both following previous work (Bliss et al. 2019; Smith and Jahn 2019)



– for identifying melt and freeze onset dates in the Arctic and Antarctic, respectively, to account for hemispheric differences in SIC bias characteristics. As such, the observed climatological melt and freeze onset dates may differ from those computed directly by the NSIDC or by earlier works. However, the comparison between the observations and the model output required a common computation method and our goal is primarily to compare the three resolutions using some observed baseline rather than quantifying the biases.

Arctic melt onset was defined from the first sustained spring SIC decrease, taken as the first day of at least 5 consecutive days with $SIC < 0.95$ and reported in a day of year (DOY) unit. If no such early season decrease was detected, a marginal ice zone fallback was applied: grid cells with fewer than 5 days of $SIC \geq 0.8$ were treated as effectively ice-free and assigned no onset date; otherwise, onset was assigned as the last day with $SIC < 0.8$ before the first ice-free day. This follows the logic of Arctic sea ice phenology products, where continuous melt in peripheral ice zones is inferred from SIC thresholds when a clear passive-microwave melt signal is unavailable (Bliss et al., 2019; Smith and Jahn, 2019). Arctic freeze onset was diagnosed from autumn SIC recovery: after the annual post-summer SIC minimum, freeze onset was defined as the first day with monotonically increasing SIC for at least 5 consecutive days, giving a persistence-based estimate rather than a simple threshold crossing. Freeze onset is also reported in a day of year (DOY) unit.

Antarctic melt onset was defined from the first sustained seasonal SIC decrease within an August-September seasonal year. Continuous melt onset was the first day of the first sequence of at least 5 consecutive days with $SIC < 0.85$, reported in DOY units, with 1 August as day 1. Antarctic freeze onset was defined as the first day after the last sequence of at least 5 consecutive melt days, where melt days had $SIC < 0.85$. The lower Southern Ocean SIC threshold, compared with the Arctic threshold of 0.95, was used to avoid spurious early freeze onsets caused by underestimated austral summer-autumn SIC in the configurations.

Ocean heat advection was computed using ocean velocity and potential temperature fields from the ocean model at monthly time steps on the native model grid. Terms necessary to compute the contribution from eddies were not saved, and we can therefore only provide an estimate for ocean heat advection, and not total ocean heat transport. The computation is done over selected transects by integrating the advective heat flux across the transect surface, vertically and along the transect. The heat advection Q_T is formally computed as:

$$Q_T = \rho_0 c_p \int_0^L \int_{-H}^0 \theta(s, z) \mathbf{v}(s, z) \cdot \mathbf{n}(s) dz ds \quad (1)$$

where ρ_0 is reference seawater density, c_p is specific heat capacity of seawater, θ is potential temperature, $\mathbf{v}$ is the ocean velocity vector, H is the water column depth, s is the distance along the transect of total length L and $\mathbf{n}$ is the unit vector normal to the transect. We note that in practice, model computations detect all grid cells intersected by the transect, generating a discrete staircase-like transect aligned with the native model grid, and integrate the heat advection over each grid-cell faces that constitute the discrete transect. In the conventions used here, a net northward heat flow is positive.



3 Results

255 The primary goal of this paper is to assess the impact of the horizontal resolution on the simulation of Arctic and Antarctic sea ice climatologies and recent extremes. In what follows, we focus on the sea ice climatological state (i.e., the long-term average over 2000-2024), except when assessing the impact of the horizontal resolution on the simulation of sea ice extremes – involving the Arctic 2007 and 2012 September sea ice lows, and the Antarctic summer 2022 and winter 2023 low anomalies –, which involves the analysis of daily sea ice extent anomalies (departures from the 2000-2024 climatology).

260 3.1 Arctic

3.1.1 Boreal summer Arctic sea ice monthly climatological characteristics

All three model configurations substantially underestimate summer SIE, amounting to an almost 2 million km² low bias in August and September compared with the mean of the three NSIDC products (Fig. 1a). For context, the range of the observational products is largest in August and September, where they span less than 1 million km². The main differences
265 in the climatological seasonal cycle between the model and the NSIDC reflect known characteristics of atmosphere-forced NEMO-SI3 configurations showing a persistent low summer bias (e.g., Rousset et al., 2015). The summer low bias in NEMO-SI3 was shown to be associated with multiple causes: uncertainties in the forcing datasets (Lin et al., 2023), and in particular to the lingering effects of a warm winter surface temperature bias in ERA5 (for a sensitivity analysis, see Topal et al., 2026),
270 or to the representation of vertical turbulent mixing under sea ice and its impact on the Arctic Ocean MLD (Allende et al., 2024). The climatological seasonal cycle of sea ice volume (SIV) similarly shows a low summer bias in all three configurations (Fig. 1b), which is a combination of the summer SIC and SIT differences between the model and the reference datasets (NSIDC and PIOMAS) (Fig. 2a-h).

Compared with the NSIDC, the strongest September SIC differences are located around the Central Arctic, with up to 70 % lower SIC north of the Laptev Sea (Fig. 2a-d;i-k). Compared with PIOMAS, modelled sea ice is also ~15-20 centimetres
275 thicker over the East Canadian Archipelago and the north Greenland Sea, and thinner in all other regions of the Arctic (Fig. 2e-h;l-n). Strongest SIT differences, up to 1.5 m thinner, are located in the Central Arctic, just north of Greenland. SIT in the Beaufort Sea and Canada Basin exhibit a relatively good agreement between the model and PIOMAS (Fig. 2l-n).

Regarding the differences between the configurations, the pan-Arctic summer sea ice characteristics are almost identical between the three configurations for both the SIE and SIV, with no clear sign of a systematic benefit of increased horizontal
280 resolution (Fig. 1a-b). Nonetheless, some marginal differences appear in the corresponding spatial climatology maps shown for September (Fig. 2a-h). Those differences become clearer when applying a decomposition of the Arctic Ocean into regional sectors (Fig. A2-A3). The few regions where the horizontal resolution shows an impact on the simulated late summer (August-September) SIC and SIE are located in the Central Arctic and in the Greenland and Chukchi Seas, although those differences are small and not consistent between resolutions and therefore not discussed further (Fig. 2i-k; Fig. A2-A3).

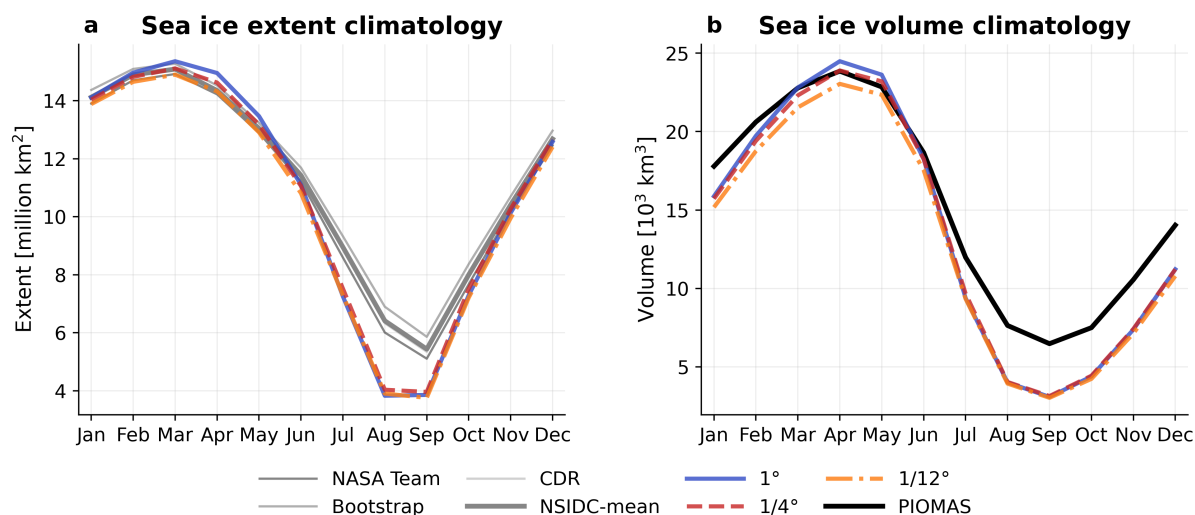


Figure 1. Pan-Arctic climatological seasonal cycle of (a) sea ice extent and (b) sea ice volume in the three configurations (1°; 1/4°; 1/12°). In panel (a) satellite observations (from the NSIDC) are shown with thin grey lines and their mean with thick black line. In panel (b) the black line shows the PIOMAS reanalysis-derived thickness.

285 3.1.2 Boreal winter Arctic sea ice monthly climatological properties

In boreal winter (defined here as the December-March period), the pan-Arctic SIE and SIV climatological seasonal cycles are in better agreement with the observational references than in summer (Fig. 1a-b). Prominent SIE biases appear only in the Greenland Sea, where the configurations show a large overestimation of SIE (Fig. A2). There, the simulated March sea ice edge is located too southward and too offshore compared with the NSIDC observational reference, especially in the 1° configuration (Fig. 3a-d;i-k). The March sea ice edge bias in the Greenland Sea is well reflected in the SIT spatial climatology maps (Fig. 3e-h). In addition, the SIT climatological differences between the model and PIOMAS indicate 0.8 to 1 m thinner sea ice in the Central Arctic for all configurations, with a spatial pattern similar to that seen in September (Fig. 3l-n vs. Fig. 2l-n).

The differences among the three configurations in the simulated sea ice state are more pronounced in winter than in summer, although they remain modest overall. Two main resolution-dependent features stand out. First, in March, the south-eastward extended ice cover in the Greenland Sea in the 1° configuration contrasts sharply with the more realistic ice edge simulated in the 1/4° and 1/12° configurations (Fig. 3a-d;i-k). Second, higher-resolution configurations systematically show thinner sea ice north of the Chukchi Sea, most pronounced in March, but also visible in annual mean (Fig. 3l-n). We now investigate the associated mechanisms for each of these features separately.

The change in winter ice edge location with increased resolution is closely linked to resolution-dependent changes in both surface ocean currents (Fig. 4) and sea ice velocity fields (not shown). Specifically, the 1° configuration fails to reproduce the Fram Strait branch of the North Atlantic Current (NAC), which carries relatively warm Atlantic water along the western

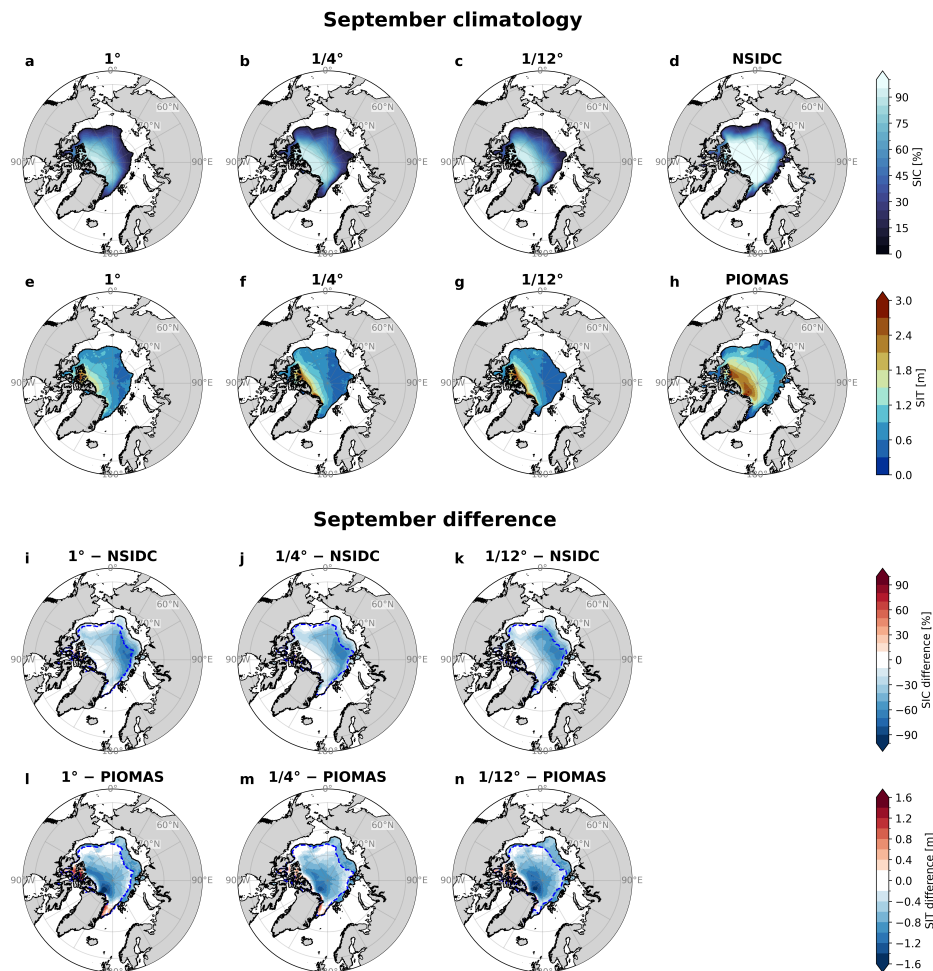


Figure 2. September Arctic climatological sea ice concentrations (SIC: %) simulated by the (a) 1° , (b) $1/4^\circ$ and (c) $1/12^\circ$ configurations and as (d) observed (from the NSIDC Climate Data Record: CDR). September climatological sea ice thickness (SIT: m) simulated by the (e) 1° , (f) $1/4^\circ$ and (g) $1/12^\circ$ configurations and (h) from the PIOMAS reanalysis. The model simulated and observed (NSIDC CDR) climatological sea ice edges (15 % sea ice concentration contour), respectively, are also shown with black line on panels (a)-(h)). Also shown: September Arctic climatological SIC differences (relative to the NSIDC CDR) in the (i) 1° , (j) $1/4^\circ$ and (k) $1/12^\circ$ configurations. September climatological SIT differences (relative to PIOMAS) in the (l) 1° , (m) $1/4^\circ$ and (n) $1/12^\circ$ configurations. In panels (i)-(n) the model simulated ice edge is shown with dashed blue lines and the NSIDC CDR ice edge is shown with solid black line.

Svalbard margin (Fig. 4a;d). By contrast, this branch of currents is resolved properly in the $1/4^\circ$ and $1/12^\circ$ configurations, allowing warmer Atlantic-origin waters to penetrate poleward across Fram Strait, and maintaining a realistic winter ice edge (Fig. 3a-d and 4b-c;e-f). Associated resolution dependent differences are also apparent in poleward ocean heat advection (Fig. 5). Although the total poleward ocean heat advection across the Denmark Strait is comparable among all three configu-

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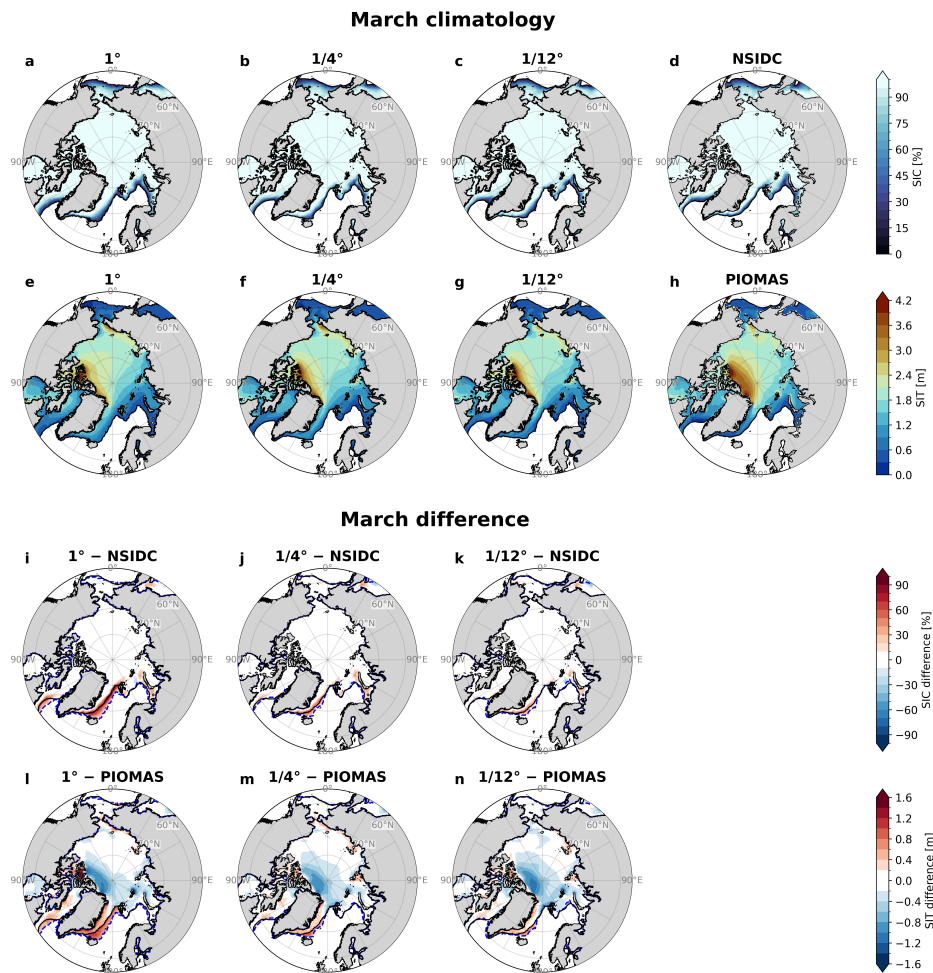


Figure 3. March Arctic climatological sea ice concentrations (SIC: %) simulated by the (a) 1°, (b) 1/4° and (c) 1/12° configurations and as (d) observed (from the NSIDC Climate Data Record: CDR). March climatological sea ice thickness (SIT: m; ice volume per cell area) simulated by the (e) 1°, (f) 1/4° and (g) 1/12° configurations and (h) from the PIOMAS reanalysis. The model simulated and observed (NSIDC CDR) climatological sea ice edges (15 % sea ice concentration contour), respectively, are also shown with black line on panels (a)-(h)). Also shown: March Arctic climatological SIC differences (relative to the NSIDC CDR) in the (i) 1°, (j) 1/4° and (k) 1/12° configurations. March climatological SIT differences (relative to PIOMAS) in the (l) 1°, (m) 1/4° and (n) 1/12° configurations. In panels (i)-(n) the model simulated ice edge is shown with dashed blue lines and the NSIDC CDR ice edge is shown with solid black line.

rations (Fig. 5c), marked differences emerge farther downstream: the 1° configuration exhibits significantly reduced poleward heat advection along the Iceland–Svalbard transect and consequently through Fram Strait (Fig. 5a-c). This is especially true in boreal winter, when simulated poleward heat advection in the 1° is almost half of the advection in the two higher-resolution configurations (Fig. 5c vs. a-b). The relatively weaker Atlantic inflow limits oceanic heat available for sea ice melt in the



310 Greenland Sea, permitting the ice edge to extend unrealistically far south. Ocean heat advection in both Fram Strait and the
Iceland-to-Svalbard transect simulated in the $1/12^\circ$ configuration agrees well with another study involving a similar model in
an atmosphere-sea ice-ocean coupled configuration albeit with 20 to 30 TW lower simulated heat transport in the Denmark
Strait (Treguier et al., 2021). Over half of this discrepancy in Denmark Strait is related to a slightly different period of interest
(constant forcing values for 2000 in the coupled model), while the rest of the discrepancy is related to a weaker poleward
315 flow close to Iceland in our model configurations (not shown). In addition, clear resolution-dependent features appear in the
East Greenland Current system (Fig. 4). In the $1/12^\circ$ configuration, localized recirculation cells controlled by bathymetry are
present adjacent to the Greenland coast, influencing both ocean surface and sea ice velocities. These features are absent or
poorly represented at coarser resolutions. Although these small-scale velocity structures do not dominate the large-scale bias
in the ice edge, they further illustrate the added realism enabled by resolving mesoscale and coastal oceanic processes. Overall,
320 our results suggest that the primary cause of the winter ice edge differences across the three configurations is tied to ocean
dynamics rather than sea ice-specific processes.

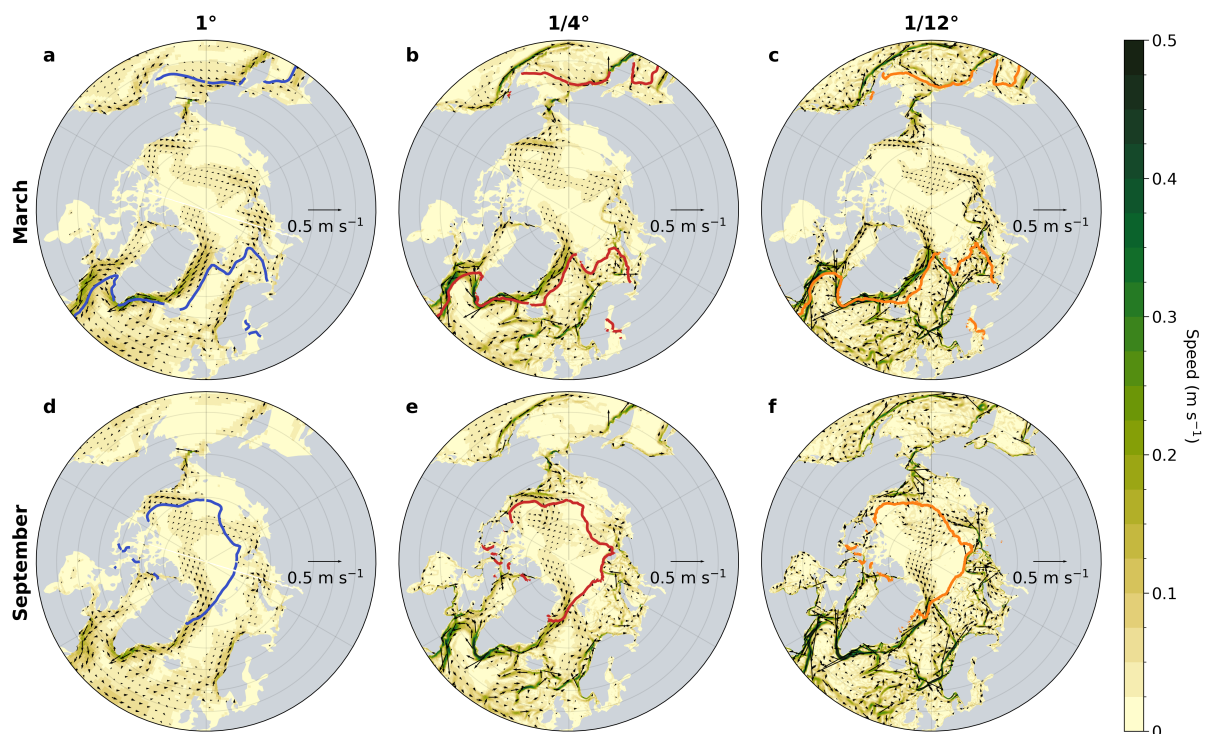


Figure 4. Arctic Ocean surface currents for the 1° (a), $1/4^\circ$ (b) and $1/12^\circ$ (c) configurations in March, with arrows representing the direction and background shading for speed. Contour lines indicate the simulated sea ice edge (15 % sea ice concentration contour). Panels (d)-(f) is the same as (a)-(c) but for September.

To better understand the processes responsible for the second resolution-dependent winter Arctic sea ice feature, we diagnose the sea ice mass budget integrated over a region north of the Chukchi Sea, defined between longitudes of 160°E and 230°E

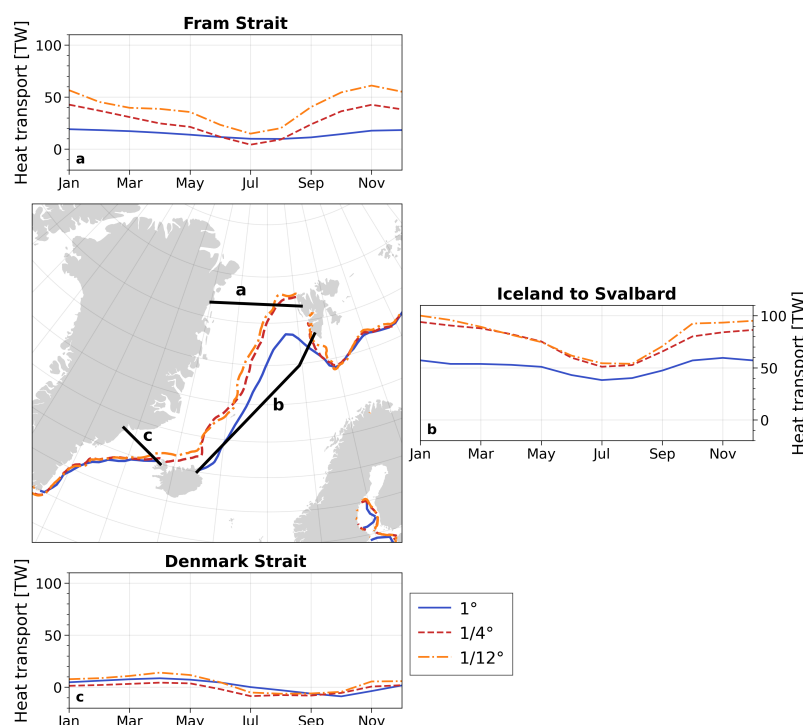


Figure 5. Poleward oceanic heat transport across three transects bounding the East Greenland Sea, where the largest differences in the simulated Arctic sea ice edge are observed. Map: climatological winter sea ice edge for the three configurations (contour lines) and location of the three transects bounding the area (black lines) across which the poleward ocean heat transport is calculated. Side panels: ocean heat transport across (a) Fram Strait, (b) Iceland-to-Svalbard and (c) Denmark Strait in the three configurations (1° ; $1/4^\circ$ and $1/12^\circ$) as shown in the legend. Ocean heat transport is defined positive northward.

and latitudes of 72°N and 90°N (see map inset, Fig. 6) and track the climatological annual cycle of the ice growth and melt terms. The decomposition reveals that distinct mechanisms drive the SIT differences between 1° and $1/4^\circ$ resolutions, and between $1/4^\circ$ and $1/12^\circ$. The contrast between the 1° and $1/4^\circ$ configurations is primarily dynamical. The 1° configuration features substantially greater ice transport into the region of interest between January and May, whereas transport is nearly negligible in the $1/4^\circ$ configuration (Fig. 6). This enhanced import in the 1° simulation feeds increased ice ridging, which in turn promotes additional mechanical thickening and freezing of seawater in the porous ridges (Fig. 6). The resulting increase in mean SIT modifies thermodynamic feedbacks: thicker ice reduces the conductive heat flux, thereby decreasing basal growth and altering both basal and surface melt later in the melt season (Fig. 6). Thus, the initial dynamical differences set into motion a cascade of thermodynamic adjustments that maintain the thickness contrast through spring and summer. The differences between the $1/4^\circ$ and $1/12^\circ$ simulations on the other hand are smaller in magnitude and mostly localized to the Bering Strait and East Siberian Sea regions (not shown). As with the 1° versus $1/4^\circ$ comparison, dynamic import again plays a key role: the $1/4^\circ$ model transports slightly more ice into the region than the $1/12^\circ$ one. This enhances frazil ice production in leads



and ice ridging, but these processes are partially offset by reduced basal ice growth across the entire domain from November to March. In April, the balance shifts: frazil growth and ridging contributions exceed those from transport. Later in the melt season, transport becomes increasingly influential in the East Siberian Sea, with the $1/4^\circ$ model exhibiting both greater import onto the shelves and reduced export farther offshore (not shown). These dynamic tendencies outweigh the increased basal melt that co-occurs with slightly reduced surface melt through July. By late summer, however, basal melt strengthens, eventually eroding the small thickness surplus in the $1/4^\circ$ configuration.

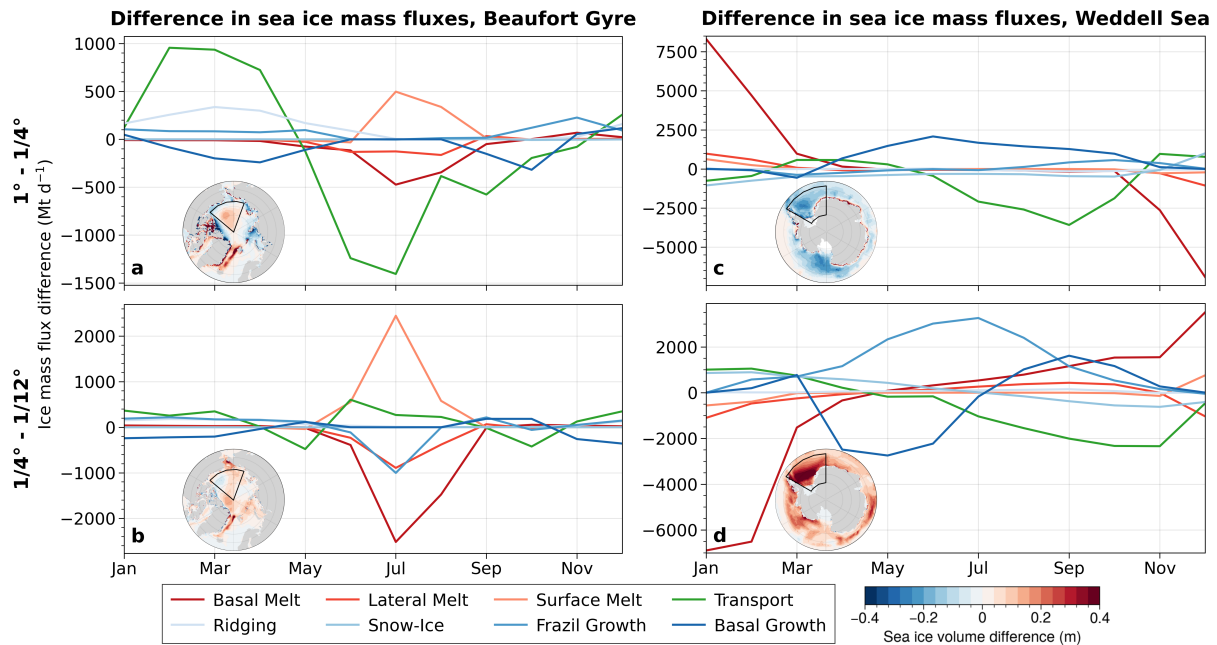


Figure 6. Differences in the climatological annual cycle of the sea ice mass balance terms between (a) the 1° and $1/4^\circ$ configurations ($1^\circ - 1/4^\circ$) and (b) the $1/4^\circ$ and $1/12^\circ$ configurations integrated over the Beaufort Gyre ($1/4^\circ - 1/12^\circ$). Panels (c) and (d) are the same as (a) and (b) but for the Weddell Sea. In our conventions, a positive flux indicates ice gain. Inset maps indicate the annual mean sea ice volume difference between the $1^\circ - 1/4^\circ$ for panels (a) and (c), and $1/4^\circ - 1/12^\circ$ for panels (b) and (d), as well as the location of the regions over which sea ice mass fluxes are integrated (Beaufort Gyre in panels (a) and (b), Weddell Sea in panels (c) and (d)).

3.1.3 Pan-Arctic sea ice drift

Next, we examine climatological differences in pan-Arctic sea ice drift among the configurations. We compute the monthly mean climatological seasonal cycle of pan-Arctic ($>70^\circ$) sea ice speeds across the three resolutions derived from daily frequency sea ice velocity components, for 2000-2020 (see Section 2.6). Relative to the observational estimates (OSI SAF and NSIDC), the model broadly reproduces the climatological seasonal cycle, although the annual minimum occurs 1 month earlier (Fig. 7a). In addition, the model exhibits comparably large sea ice speeds year-round, with the largest differences seen in early summer (June-July) and smallest differences between October to May.



Regarding the resolution dependent features, the two higher resolution configurations (1/4° and 1/12°) show notably lower
 350 sea ice speeds (i.e., closer to the observational references) from October to April compared with the 1° configuration. This
 difference is also reflected by the distribution of the daily sea ice speeds across all years between 2000 and 2020 (Fig. 7b). In
 winter, the higher resolutions lead to faster drifts in narrower, more constrained locations, while the 1° resolution simulates a
 wide outflow of ice, e.g. in the East Greenland Current (Fig. 4). As a result, this skews the distribution of the 1° sea ice velocity
 towards higher values (Fig. 7b). In summer, the loosened ice pack is more sensitive to wind forcing and the ice drift is almost
 355 directly proportional to wind stress. Because wind stress depends quadratically on wind speed in the bulk formulas employed
 in NEMO4-SI3, the greater wind speed variability resolved at higher resolutions likely enhances the effective wind forcing,
 contributing to higher sea ice speeds in the 1/4° and 1/12° configurations.

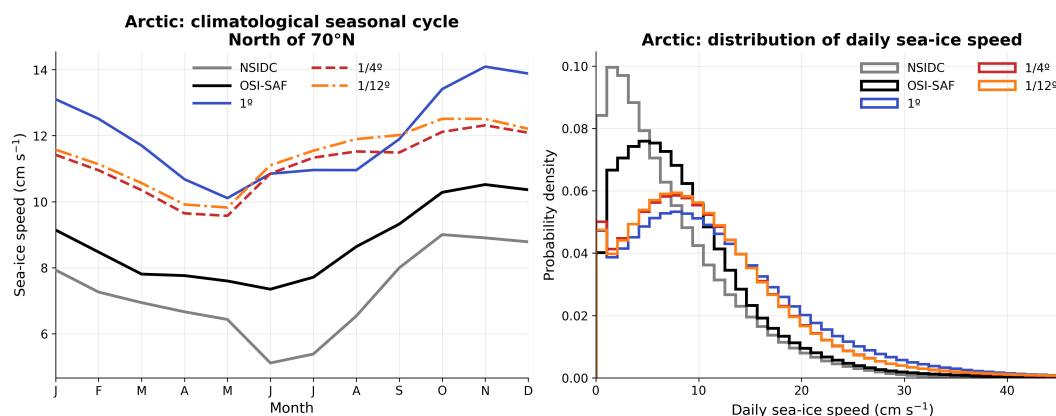


Figure 7. Climatological seasonal cycle of Arctic sea ice speed for the period 2000-2020 as (left) simulated by the three configurations and as observed from the OSI SAF and NSIDC products as shown in the legend. Also shown: (right) probability density functions of the simulated and observed (as shown in the legend) distributions of daily sea ice speed across the 2000-2020 period.

3.1.4 Pan-Arctic melt/freeze onset climatologies

We also compute the melt/freeze onset dates from daily SIC fields (Fig. 5; see also Section 2.6 for a definition). The relatively
 360 earlier (2 weeks) melt onset in the model – compared with the NSIDC CDR – likely reflects the climatological differences in
 the marginal ice zone SIC fields and a relatively faster transition to the melting season (Fig. 8; Fig. 1a). The modelled shape
 of the melt onset distribution shows a relatively sharper peak of melt early in the season compared with the NSIDC, which
 shows a more gradually distributed melt onset DOY during summer (Fig. 8). Regarding freeze onset, the model captures the
 two-step shape of the distribution seen in the NSIDC record, albeit with an approximately 1 month earlier onset of freeze up
 365 in the Central Arctic (first step) – likely due to the low SIC biases in the model (Fig. 5). Overall, no clear differences in either
 melt or freeze onset exist across the three configurations, suggesting a minor role of horizontal resolution in constraining the
 vertical thermodynamics of the sea ice melt onset.

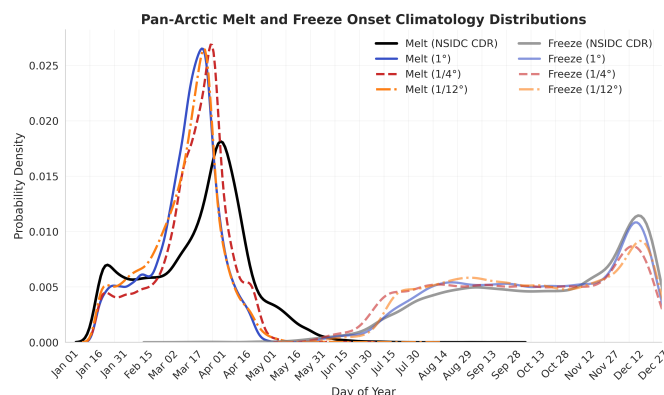


Figure 8. Probability density functions of the Arctic climatological timing of melt and freeze onset shown as “day of year” for the period 2000-2024 in the three configurations (1°; 1/4° and 1/12°) and from the NSIDC Climate Data Record (CDR) as shown in the legend.

3.1.5 Arctic sea ice lows: 2007 and 2012

The foregoing analysis has solely focused on the climatological sea ice state. However, we also assess whether resolution-dependent differences emerge during extreme sea ice loss events. In doing so, we briefly investigate any potential differences in simulating the two historical sea ice low events observed in September 2007 and 2012 (Richaud et al., 2026). We compute the daily SIE anomalies relative to the 2000-2024 climatology of each day (see Section 2.6) and compare the model simulated pan-Arctic SIE anomalies with that obtained from the NSIDC between 1 August and 30 September in 2007 and 2012, respectively. Regarding the 2007 extreme low SIE anomaly, all three configurations simulate the late summer evolution of daily SIE anomalies relatively well compared with the NSIDC (Fig. 9a). However, they consistently underestimate the SIE loss, by up to 0.5 million km² (Fig. 9a), in particular in the Central Arctic and the east Kara and Laptev Seas in September (Fig. A4). Yet, the biased, low climatological state of the sea ice extent still leads to the models overestimating the area of ice-free regions (Fig. A5). As for the 2012 summer minimum, the configurations relatively better match the magnitudes of the observed daily sea ice anomalies during August and September 2012 compared with 2007, except during the first week of August (Fig. 9b). Overall, no pronounced difference in the simulated daily SIE anomalies appears among the configurations in any of the regional sectors (Fig. A6). In summary, without a prominent impact on the mean sea ice state, increased horizontal resolution does not show sufficient benefits for a better simulation of the 2007 and 2012 extreme low summer SIE anomalies in the Arctic.

3.2 Antarctic

Sea ice processes in the Southern Ocean are more tightly coupled with oceanic processes than in the Arctic (e.g., Goosse et al., 2025, 2026; Silvano et al., 2025). We can therefore expect more diverse impacts of increased horizontal resolution on the Antarctic sea ice state than seen for the Arctic. This is confirmed by the more widespread resolution dependent features over the Southern Ocean than in the Arctic, exhibited year-round (Fig. 10).

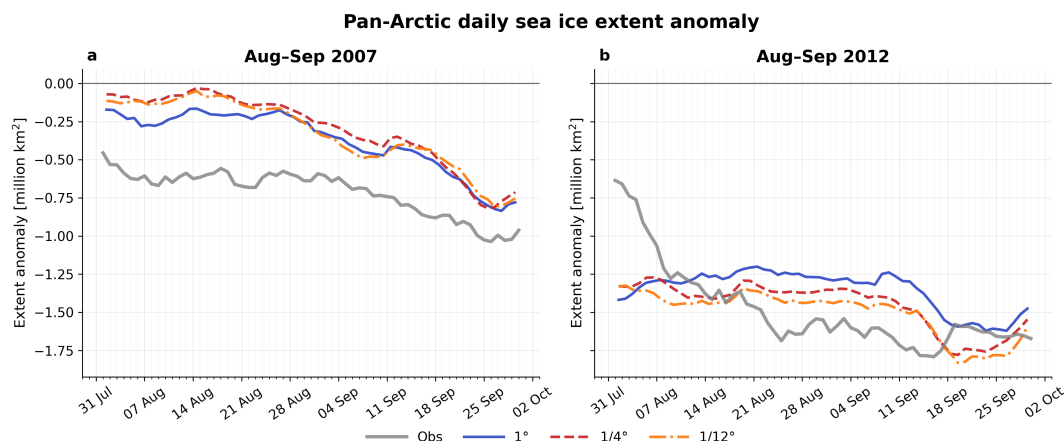


Figure 9. Daily sea ice extent anomalies (relative to the 2000–2024 climatological mean) between 1 August and 30 September in (a) 2007 and (b) 2012 in the three configurations (1° ; $1/4^\circ$ and $1/12^\circ$) and from the NSIDC Climate Data Record (CDR) as shown in the legend.

3.2.1 Austral summer Antarctic sea ice monthly climatological properties

Around the Antarctic SIE seasonal minimum from January to March, all three configurations underestimate SIE relative to the NSIDC record (Fig. 10a). The negative bias in February is largest in the 1° and $1/12^\circ$ configurations, reaching approximately 2 million km^2 , whereas it is smaller in the $1/4^\circ$ configuration, at approximately 0.9 million km^2 . A broadly similar picture emerges for SIV, with the $1/4^\circ$ configuration showing the closest agreement with GIOMAS during January–March (Fig. 10b). Spatially, all three configurations underestimate SIC and SIT in the western Amundsen Sea, in the Bellingshausen Seas and in the eastern part of the Southern Ocean, suggesting that these regional biases are relatively insensitive to horizontal resolution.

In contrast, the Weddell and Ross Seas show clearer configuration dependent differences: the $1/4^\circ$ configuration maintains more extensive and thicker sea ice in these sectors than either the 1° or $1/12^\circ$ configurations (Fig. 11; Figs. A7–A8), explaining much of its larger pan-Antarctic SIE during January–March (Fig. 10a). As a result, the $1/4^\circ$ configuration reduces the climatological SIC bias in the Weddell and Ross Seas by up to approximately 40 % relative to the other two configurations, while the corresponding SIT biases are reduced by up to approximately 50 % (Fig. 11i–n).

3.2.2 Austral winter Antarctic sea ice monthly climatological properties

The relative agreement between the $1/4^\circ$ configuration and the observed climatological summer SIE starts to diminish from June onwards. By September, SIE is overestimated by 1.4 million km^2 in the $1/4^\circ$ configuration and by 1.6 million km^2 in the 1° one, while it is underestimated by 0.65 million km^2 in the $1/12^\circ$ one (Fig. 10a). Notably, both the SIE and SIV climatological seasonal cycles peak one month later in the 1° and $1/4^\circ$ configurations than in NSIDC and GIOMAS, while this offset is less apparent in the $1/12^\circ$ configuration, at least for SIE (Fig. 10a–b). The overestimation of the winter SIE and the timing offset is seen in the regional SIE and SIV climatological seasonal cycles, reflecting biases primarily in SIE (Figs. A7–A8; Barthélemy

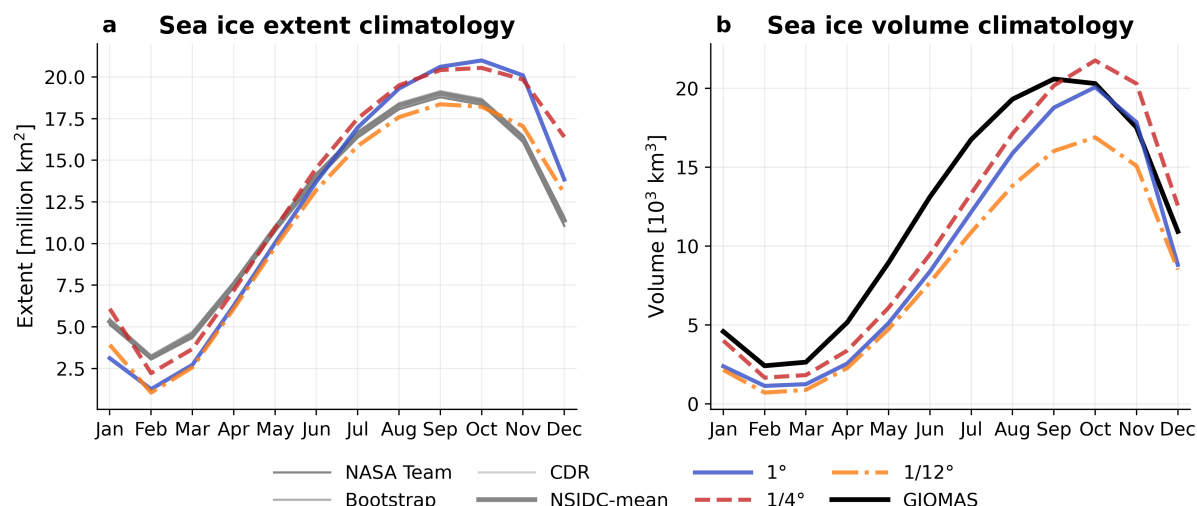


Figure 10. Pan-Antarctic climatological seasonal cycle of (a) sea ice extent and (b) sea ice volume in the three configurations (1°; 1/4°; 1/12°). In panel (a) satellite observations (from the NSIDC) are shown with thin grey lines and their mean with thick black line. In panel (b) the thin grey lines show GIOMAS reanalysis-derived thickness multiplied by the NSIDC sea ice concentrations from the three products, respectively, and their mean (thick black line) as shown in the legend.

et al., 2018). Spatially, all three configurations show similar winter SIC biases in the Ross Sea and the western part of the Amundsen Sea (Fig. 12i-k). Similar biases also prevail in both the 1° and 1/4° configurations in the King Håkon VII sector (Fig. 12i-k).

410 We next discuss two of major divergences between configurations and explain the physical mechanisms, between them. First, the winter sea ice edge exhibits notable regional differences among the configurations downstream from the Kerguelen Plateau (approximately 100°E), the Pacific Antarctic Ridge off the Amundsen Sea (around 230°E) and Bouvet Island in the King Håkon VII Sea sector (near 20°E) (Fig. 12a-c). Namely, the 1/4° and 1/12° configurations generally place the sea ice edge southward from the 1° and with an elbow-shaped edge, more in line with observations, although the 1/12° seems to “over-

415 correct” downstream of those bathymetric obstacles (Fig. 12i-k). Given the strong control of the Antarctic Circumpolar Current (ACC) and its bathymetrically steered fronts on the winter Antarctic ice edge (Goosse et al., 2025), these differences point to ocean circulation and ocean-sea ice coupled processes as key mechanisms behind the resolution dependent sea ice response. In the two higher-resolution configurations – and especially the 1/12° – boundary currents near the ice edge are narrower, faster and more tightly aligned with the ACC frontal structure than in the 1° configuration, where currents are broader, weaker and

420 spatially smoother (Fig. 13). Sharper ocean currents can also transport relatively warm and salty waters closer to the winter ice edge, strengthening the horizontal temperature and salinity gradients and sharpening the oceanic fronts that separate the open Southern Ocean from the sea ice zone. This behaviour is closely linked to the role of mesoscale eddies in setting the ACC dynamics and cross frontal exchanges, since eddy rectification and eddy-induced transports are central to the structure of

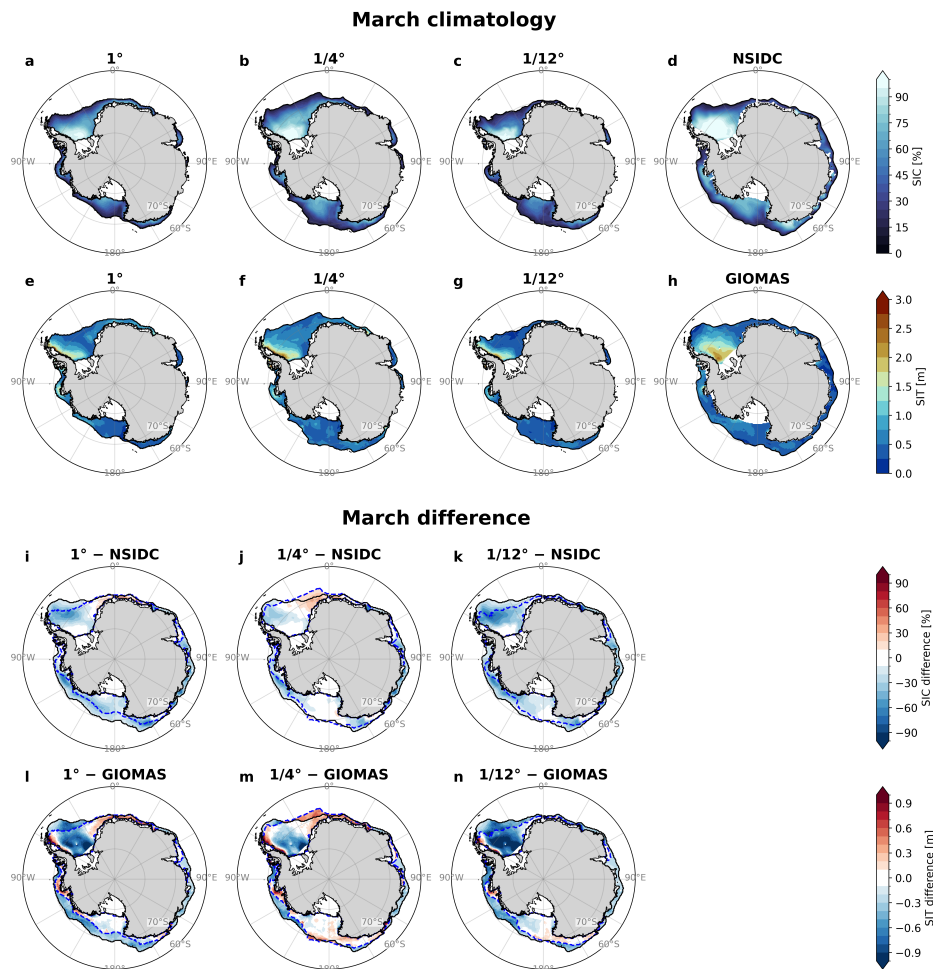


Figure 11. March Antarctic climatological sea ice concentrations (SIC: %) simulated by the (a) 1°, (b) 1/4° and (c) 1/12° configurations and as (d) observed (from the NSIDC Climate Data Record: CDR). September climatological sea ice thickness (SIT: m) simulated by the (e) 1°, (f) 1/4° and (g) 1/12° configurations and (h) from the GIOMAS reanalysis. The model simulated and observed (NSIDC CDR) climatological sea ice edge (15 % sea ice concentration contour), respectively, are also shown with black line on panels (a)-(h)). Also shown: March Arctic climatological SIC differences (relative to the NSIDC CDR) in the (i) 1°, (j) 1/4° and (k) 1/12° configurations. March climatological SIT differences (relative to GIOMAS) in the (l) 1°, (m) 1/4° and (n) 1/12° configurations. In panels (i)-(n) the model simulated ice edge is shown with dashed blue lines, and the NSIDC CDR ice edge is shown with solid black line.

the ACC and to the redistribution of heat and salt across its fronts (Farneti et al., 2010; Naveira Garabato et al., 2011; Dufour et al., 2015). The regions where the ice edge improves most in our higher resolution configurations – downstream of the Kerguelen Plateau, along the Pacific Abyssal Plains, and downstream from Bouvet Island – are also regions where mesoscale eddy activity is enhanced through bathymetric steering. More vigorous horizontal mixing and eddy-induced heat transport can

increase the delivery of oceanic heat from the ACC frontal region toward the winter sea ice edge, thereby limiting northward sea ice expansion and contributing to a more poleward ice margin (see also Goose et al., 2025).

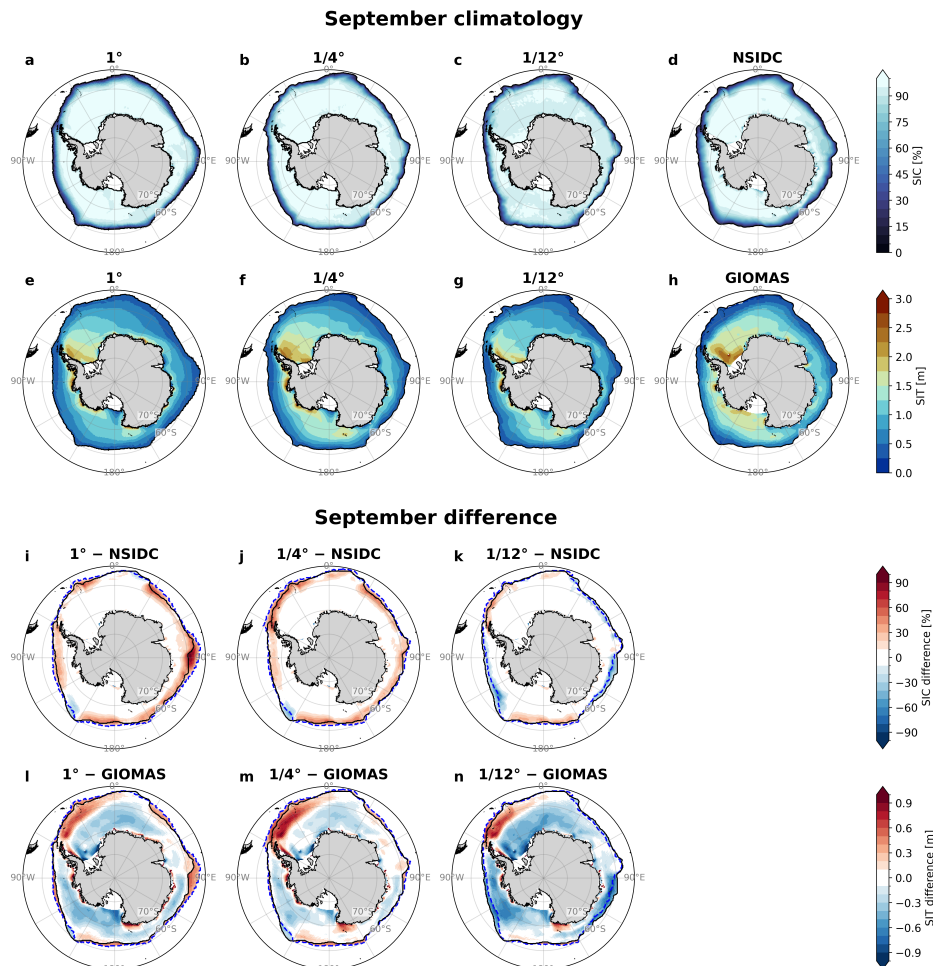


Figure 12. September Antarctic climatological sea ice concentrations (SIC: %) simulated by the (a) 1°, (b) 1/4° and (c) 1/12° configurations and as (d) observed (from the NSIDC Climate Data Record: CDR). September climatological sea ice thickness (SIT: m) simulated by the (e) 1°, (f) 1/4° and (g) 1/12° configurations and (h) from the GIOMAS reanalysis. The model simulated and observed (NSIDC CDR) climatological sea ice edge (15 % sea ice concentration contour), respectively, are also shown with black line on panels (a)-(h)). Also shown: September Arctic climatological SIC differences (relative to the NSIDC CDR) in the (i) 1°, (j) 1/4° and (k) 1/12° configurations. September climatological SIT differences (relative to GIOMAS) in the (l) 1°, (m) 1/4° and (n) 1/12° configurations. In panels (i)-(n) the model simulated ice edge is shown with dashed blue lines and the NSIDC CDR ice edge is shown with solid black line.

430 The simulation of mesoscale eddies in the 1/12° configuration induces a better simulation of oceanic fronts and to significant differences in the vertical structure of the water column (Fig. 14). At the sea ice edge in the Kerguelen Plateau and in the Pacific Abyssal Plains, the halocline is deeper and weaker in the 1/12° configuration, visible below 200 m depth while it sits

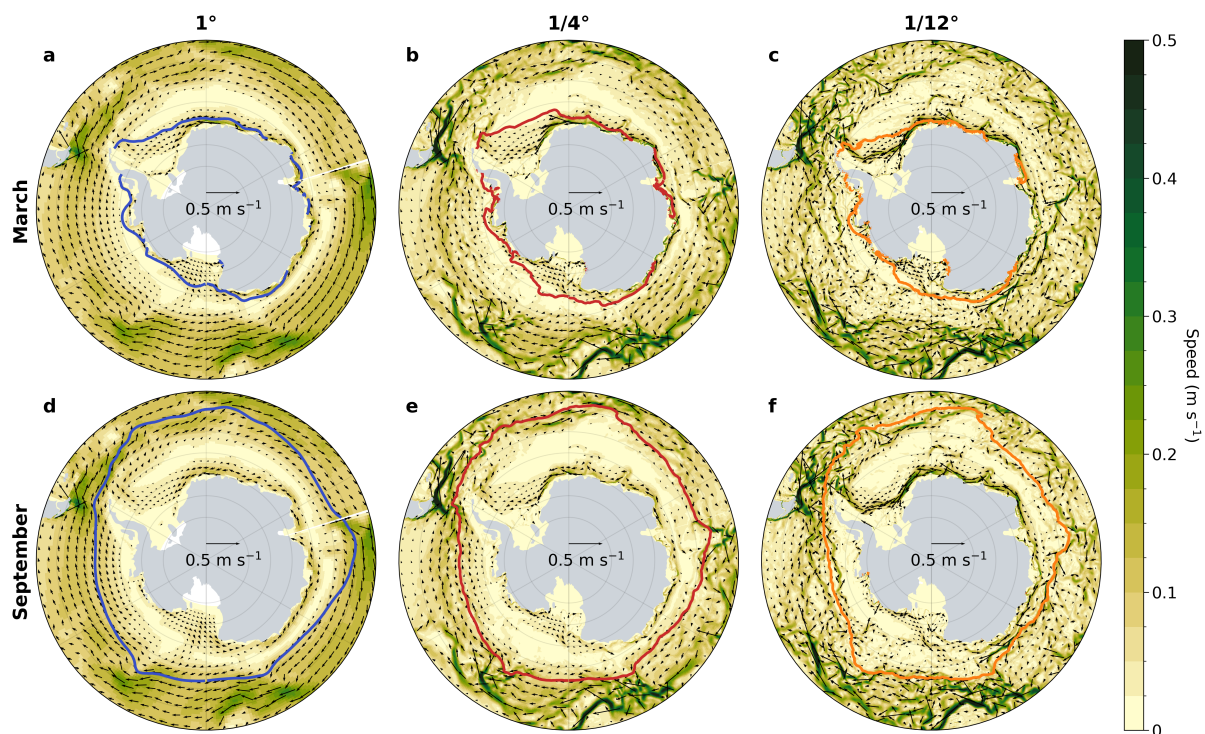


Figure 13. Southern Ocean surface currents for the 1° (a), $1/4^\circ$ (b) and $1/12^\circ$ (c) configurations in March, with arrows representing the direction and background shading for speed. Contour lines indicate the simulated sea ice edge (15 % sea ice concentration contour). Panels (d)-(f) is the same as (a)-(c) but for September.

at 100 m depth in the other configurations. This leads not only to a deeper mixed layer, but also to a weaker stratification and enhanced vertical mixing with warm, subsurface waters. This feeds and maintain the higher surface temperatures in the $1/12^\circ$ configuration, reaching positive values, compared to the 1° and $1/4^\circ$ where surface temperatures remain at the freezing point (Fig. 14a-b). A similar feature is also visible close to Bouvet Island, though less exacerbated (Fig. 14c). Those higher surface temperatures prevent the subsistence of sea ice and could be maintained by a positive feedback loop where less sea ice leads to less freshwater input by meltwater, further decreasing the salinity-driven stratification (Haumann et al., 2016). The observation-based EN4 data indicate a weaker stratification at those locations, with the halocline sitting around 150 m over the Kerguelen Plateau. The $1/4^\circ$ resolution seems closest to those observations, though the weaker stratification is better simulated by the $1/12^\circ$ configuration.

A second major resolution-dependent feature is that the $1/4^\circ$ simulation produces systematically thicker sea ice than both the 1° and the $1/12^\circ$ configurations, a pattern evident throughout most of the circumpolar sea ice zone (Fig. 12). This contrast is particularly pronounced in the Weddell Sea, which we therefore use as a case study to examine the processes underlying the resolution dependent sea ice thickness differences. The Weddell Sea T-S profiles provide useful context for this analysis, as they illustrate differences in near-surface stratification and in the temperature-salinity properties of waters beneath the cold surface

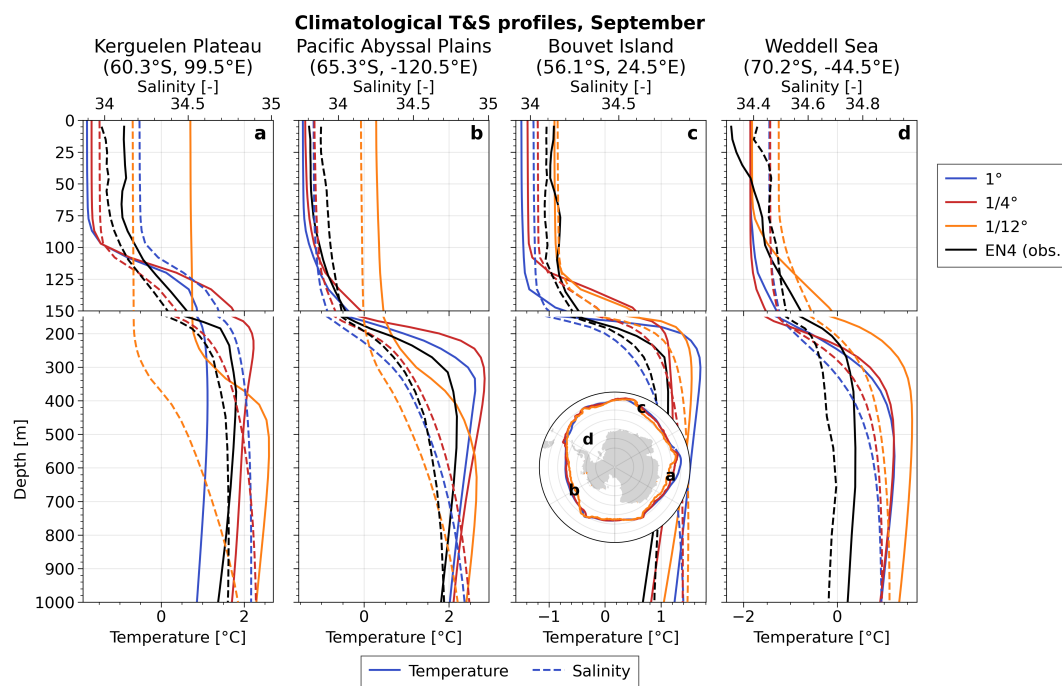


Figure 14. Temperature (solid lines, lower x-axis) and salinity (dashed lines, upper x-axis) profiles for the three configurations (coloured lines) and observation-based analyses EN4 (black lines) around the Southern Ocean. Hydrographic conditions are shown close to the winter ice edge, where differences in the sea ice edge are most obvious, in (a) downstream from Kerguelen Plateau, (b) over the Pacific Abyssal Plains and (c) downstream from Bouvet Island. Also shown: (d) hydrographic conditions illustrated under pack ice in the Weddell Sea, where differences in sea ice thickness are most noticeable. All locations are indicated in the inset map in panel (c), along with the climatological September sea ice edge for all three configurations. Note the change of scale in the y-axis at 150 m depth, zooming on the upper part of the water column.

layers across the three configurations (Fig. 14d). The $1/4^\circ$ configuration maintains a colder and/or less ventilated upper ocean relative to the $1/12^\circ$ configuration, indicating limited upward ocean heat supply and stronger basal ice growth (Martinson, 1990). By contrast, the $1/12^\circ$ configuration shows a weaker stratification and warmer waters within the halocline, hinting towards stronger exchange between the relatively warmer and saltier subsurface waters and the upper ocean, consistent with earlier basal melt and reduced winter ice growth. To further reveal the mechanisms underlying the large-scale SIT differences, we apply a similar decomposition of the sea ice mass budget as for the Beaufort Gyre in the Arctic but integrated over the Weddell Sea and focusing on the seasonal buildup of thickness anomalies. The main conclusions from this analysis are similar when looking at other regions exhibiting large differences in sea ice thickness, such as the Ross Sea (not shown).

The comparison between the $1/4^\circ$ and 1° configurations shows that reduced export in the $1/4^\circ$ configuration is the dominant contributor to its positive ice mass anomaly in the Weddell Sea (Fig. 6c). From winter to early summer, the 1° configuration loses more ice through transport, while the $1/4^\circ$ retains more ice. The relatively enhanced transport in the 1° is only partially



compensated by stronger frazil ice production in leads and basal ice growth, and the overall dynamic mass loss is persistent. Basal ice melt also begins earlier in the 1° configuration, from November onward, consistent with its thinner ice cover and a more equatorward winter ice edge (Fig. 12; Fig. 14). This earlier melt may be facilitated by earlier breakup of the marginal ice zone and by an associated albedo feedback once thinner ice and open water areas become more prevalent. By contrast, the $1/4^\circ$ configuration shows a comparatively delayed onset of basal ice melt and therefore maintains thicker ice through winter and early summer, even though melt rates become more comparable later in the season (Fig. 6c).

The comparison between the $1/4^\circ$ and $1/12^\circ$ configurations reveals a different balance of processes (Fig. 6d). In the $1/4^\circ$ configuration, more ice is exported than in the $1/12^\circ$ one, which would, by itself, act to reduce SIT in the $1/4^\circ$ configuration. However, this dynamic ice loss is more than compensated by stronger thermodynamic ice production, particularly through enhanced frazil ice formation in leads and basal ice growth. As a result, the $1/4^\circ$ configuration maintains a positive ice mass and thickness anomaly compared with the $1/12^\circ$ one. The $1/12^\circ$ configuration, in contrast, exhibits earlier basal ice melt, similar in timing to the 1° configuration, although the $1/4^\circ$ configuration eventually reaches comparable melt rates later in summer. These differences suggest that the thinner ice in the $1/12^\circ$ configuration is not primarily caused by stronger export, but by a thermodynamic balance in which reduced winter ice growth and earlier basal ice melt limit the seasonal buildup of ice. This is consistent with the T-S profiles (Fig. 14).

3.2.3 Pan-Antarctic sea ice drift

Next, we investigate sea ice drift focusing on sea ice speed south of 50°S (see Sections 2.5;2.6). The broad features of the climatological seasonal cycle are reproduced in the three configurations compared with both the OSI SAF and NSDIC products (Fig. 15a). Nevertheless, clear resolution dependent differences emerge during the season of relatively fast ice motion from April to October, when the $1/4^\circ$ and $1/12^\circ$ configurations simulate slower sea ice speeds than the 1° configuration and are closer to OSI SAF, with the $1/4^\circ$ case matching particularly well from June to September (Fig. 15a). The distribution of the daily sea ice speeds across 2000-2020 also reflect that the two higher resolution configurations better capture the high-speed tail, especially for speeds between 20 cm s^{-1} and 45 cm s^{-1} , while the 1° configuration overestimates the occurrence of fast-moving ice (Fig. 15b). Yet, the $1/4^\circ$ and $1/12^\circ$ configurations fail to reproduce the magnitude of the observed decrease in sea ice speed from August to January, whereas the 1° configuration shows a more realistic seasonal decline.

Ocean circulation related mechanisms also help interpret the resolution dependence of Antarctic sea ice drift. As discussed in the previous section (Section 3.2.1), the two higher resolution configurations –especially the $1/12^\circ$ simulation – produce narrower, faster boundary currents that are more closely aligned with the ACC frontal structure than in the 1° one (Fig. 13). These stronger currents near the winter ice edge likely contribute to the improved representation of fast Antarctic sea ice drift in the higher-resolution configurations (Fig. 13b), particularly during the growth season when the marginal ice zone is most expanded and dynamically exposed to ocean forcing. However, the same intensified ocean circulation may also maintain excessive ice mobility into spring, thereby weakening the observed August-January slowdown in sea ice motion. Moreover, the stronger sea ice drift in austral winter in the 1° configuration is well in line with the above-mentioned increased sea ice export



in Weddell Sea, which leads to reduced ice mass in the 1° simulation compared to the $1/4^\circ$ one (Fig. 6c). The slightly higher drift in the $1/12^\circ$ simulation compared to the $1/4^\circ$ one is also coherent with previous analyses (Section. 3.2.2).

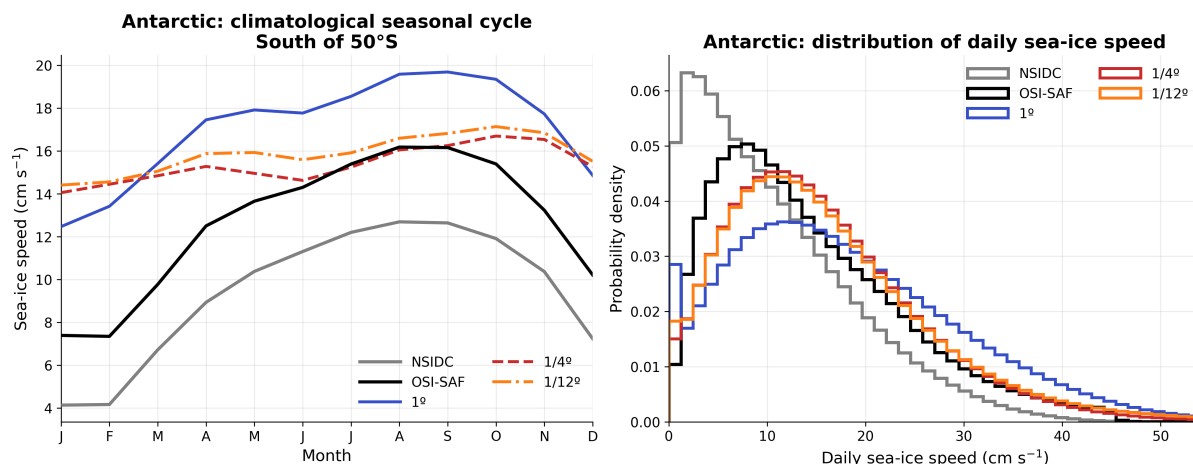


Figure 15. Climatological seasonal cycle of Antarctic sea ice speed for the period 2000–2020 as (left) simulated by the three configurations and as observed from the OSI SAF and NSIDC products as shown in the legend. Also shown: (right) probability density functions of the simulated and observed (as shown in the legend) distributions of daily sea ice speed across the 2000–2020 period.

3.3 Pan-Antarctic melt/freeze onset

We also investigate potential differences in the seasonal onset of Antarctic sea ice melt and freeze (Fig. 14; see also Section 2.6). The configurations do not fully reproduce the observed two-peak structure of the Antarctic melt-onset distribution (Fig. 16). All configurations show excessive early melt onset in November and underestimate the occurrence of later melt onset in January. The modelled SIE biases in spring, with overestimated SIE in November for all 3 configurations and insufficient SIE in January for the 1° and $1/12^\circ$ configurations, can be linked to this early sea ice melt onset in all three configurations. The early bias is smaller in the $1/4^\circ$ configuration (<1 week) relative to the 1° and $1/12^\circ$ ones, which both exhibit approximately 2 weeks earlier peak of melt onset. Whether there is a clear causal direction between SIE and melt onset biases is unclear. This behaviour is similar to the Arctic, where the model simulations also favour an earlier, sharper transition into the melting season compared with observations. The simulations also broadly capture the skewed distribution of the observed freeze onset, including the late winter peak in August. However, the simulated freezing season begins too early, especially in the $1/4^\circ$ configuration. Among the three configurations, the 1° simulation shows the closest agreement with the observed peak of freeze onset distribution in August. The lack of a monotonic dependence of the melt and freeze onset distributions on the horizontal resolution indicates that the thermodynamic transition between the melting and freezing seasons are not primarily controlled by horizontal resolution alone. Rather, the seasonal advance and retreat of the ice cover is strongly constrained by the simulated mean state and the atmospheric boundary forcing.

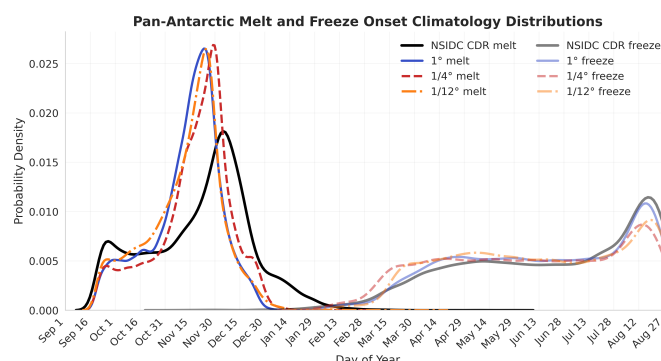


Figure 16. Probability density functions of the Antarctic climatological timing of melt and freeze onset shown as “day of year” for the period 2000–2024 in the three configurations (1°; 1/4° and 1/12°) and from the NSIDC Climate Data Record (CDR) as shown in the legend.

3.3.1 Antarctic sea ice lows: summer 2022 and winter 2023

510 Finally, we briefly investigate potential differences among the three configurations in simulating the two historical Antarctic sea ice low events observed in February–March 2022 (Turner et al., 2022) and during early-to-mid winter 2023 (Espinosa et al., 2024; Wang et al., 2024a). During the 2022 late summer event, the observed pan-Antarctic SIE anomalies remained negative from January to March, reaching approximately -1 million km^2 with the largest contribution coming from the Ross and Weddell Seas (Fig. 17a and A10). Overall, all configurations reproduce the magnitude – but less so the temporal evolution

515 – of the observed pan-Antarctic SIE anomalies relatively well. The closest overall alignment between all three configurations and the observations (during the whole January–March 2022 period) occurs in the Weddell Sea (Fig. A10). Among them, the 1/4° configuration shows the closest agreement with the observations during February and early March 2022, which reflects closer-to-observed SIE anomalies in the Ross Sea relative to the other two configurations. Furthermore, the 1/12° configuration exhibits a much closer match with the observed SIE anomalies relative to the 1° and 1/4° ones in the Bellingshausen Sea.

520 However, this region does not contribute much to the pan-Antarctic minimum (Fig. A10). The spatial structure of the simulated anomaly biases broadly resembles the climatological SIC mean state biases of the configurations (Fig. A11a–f; Fig. A9i–k). In particular, the relatively better representation of summertime daily SIE anomalies in the Ross Sea by the 1/4° configuration is consistent with its improved February–March mean sea ice state in this region compared with the 1° and 1/12° configurations (Fig. A11i–k; Fig. A9i–k). Nevertheless, increased horizontal resolution does not lead to a robust improvement in simulating

525 the 2022 Antarctic extreme low sea ice event.

On the other hand, the early winter (May–June–July) Antarctic mean sea ice state is comparably well simulated in all three configurations, which is also reflected in a relatively close match between the simulated and observed daily pan-Antarctic SIE anomalies during May–July 2023 (Fig. 17b). The 2023 pan-Antarctic deficit is associated with strong negative anomalies in all sectors except the Amundsen Sea, where anomalies are positive (Fig. A12). The largest model deficiency occurs in the King

530 Håkon VII and East Antarctic sectors, where all configurations simulate too weak a deficit (Fig. A12e–f). The configurations



also underestimate the negative anomalies in the Bellingshausen Sea through much of the period, with the 1° configuration showing the closest alignment with the observations, especially from the second half of June 2023 (Fig. A12c). In the Ross Sea, all configurations show comparably good skill in reproducing both the magnitude and timing of the observed anomalies (Fig. A12a). However, we find no clear reduction in SIC or SIT anomaly regional biases with increasing resolution for the period May-July 2023 relative to the reference datasets (NSIDC and GIOMAS) (Fig. A11g-l).

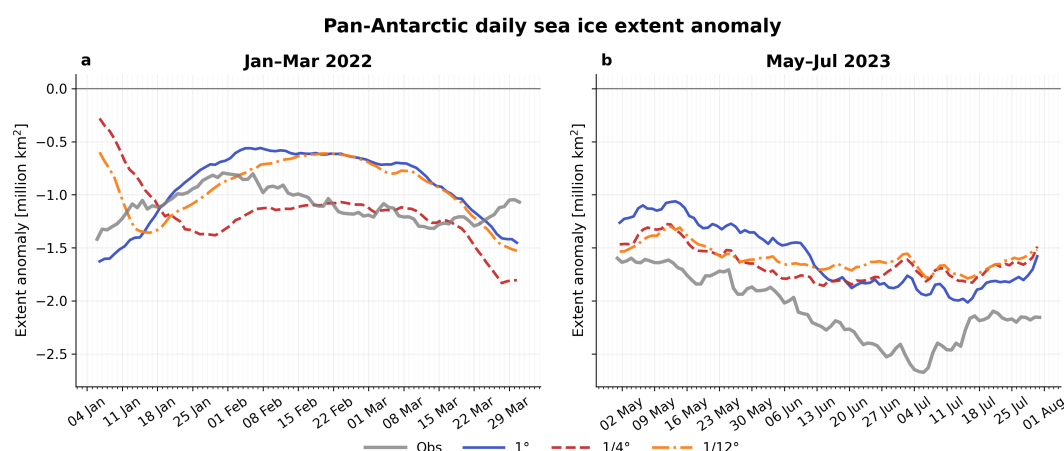


Figure 17. Daily sea ice extent anomalies (relative to the 2000-2024 climatological mean) between 1 January and 31 March in (a) 2022 and (b) between 1 May and 31 July 2023 in the three configurations (1° ; $1/4^\circ$ and $1/12^\circ$) and from the NSIDC Climate Data Record (CDR) as shown in the legend.

4 Summary and Discussion

In this paper, we present a hierarchy of NEMO4-SI3 coupled ocean-sea ice simulations, forced by the ERA5 atmospheric reanalysis, at three horizontal resolutions: 1° ; $1/4^\circ$ and $1/12^\circ$. The simulations are used to assess the sensitivity of Arctic and Antarctic sea ice climatologies and low sea ice events to horizontal resolution, and to evaluate the cost-benefits of dedicating more computational resources to improve physical realism. Overall, increased ocean-sea ice resolution improves some seasonal and regional sea ice properties, particularly where ocean dynamics strongly controls the ice edge, but it does not systematically reduce the dominant hemispheric biases. Across both hemispheres, much of the benefit is already achieved in the transition from 1° to $1/4^\circ$, with the $1/12^\circ$ configuration providing limited additional improvement relative to its computational cost.

In the Arctic, benefits of increased resolution appear clearest in the Greenland Sea, where the $1/4^\circ$ and $1/12^\circ$ configurations better represent the Atlantic inflow pathway west of Svalbard and reduce the excessive winter sea ice extension simulated at 1° . These results show that resolution affects Arctic sea ice partly through improved ocean circulation and heat transport, rather than through sea ice thermodynamics alone. However, the dominant Arctic summer biases in SIE and SIV remain similar across all configurations, indicating that model improvements are required beyond horizontal resolution refinement.



The Antarctic response is more widespread, consistent with stronger ocean-sea ice coupling in the Southern Ocean than in the Arctic (Goosse et al., 2025, 2026; Silvano et al., 2025). Higher resolution can improve the winter ice edge downstream of major bathymetric features by better representing ACC fronts, bathymetric steering and mesoscale activity, but the simulation of mesoscale eddies by the $1/12^\circ$ can also enhance vertical mixing with warm subsurface waters, promoting thinner ice and a more poleward ice margin. Thus, the impact of increasing resolution has a dual effect: it can improve the dynamical representation of oceanic fronts, currents and bathymetric controls, but it can also expose or amplify biases in vertical mixing, stratification and ocean heat delivery to the ice.

Thus, the impact of increasing resolution has a dual effect: it can improve the dynamical representation of fronts, currents, and bathymetric controls, but it can also expose or amplify biases in vertical mixing, stratification, and ocean heat delivery to the ice. This is particularly true for the Antarctic region, consistent with stronger ocean-sea ice coupling in the Southern Ocean than in the Arctic (Goosse et al., 2025, 2026; Silvano et al., 2025). However, the dominant summer biases in SIE and SIV in both poles remain similar across all configurations, indicating that improvements are required beyond horizontal resolution refinement.

Other important configuration differences are related to SIT. In the Arctic Ocean, thicker ice is associated with coarser resolution, consistent with stronger ridging and rafting at lower resolution. In the Antarctic, by contrast, the thickest ice occurs in the $1/4^\circ$ configuration, while the 1° and $1/12^\circ$ simulations are more similar. Although coarser resolutions tend to produce a more dynamic sea ice pack and greater equatorward export, which can thin the ice, this does not explain why Antarctic ice remains thicker in the $1/4^\circ$ configuration than in the $1/12^\circ$. We hypothesize that this contrast reflects differences in ocean stratification and subsurface heat access, with weaker stratification above warmer subsurface waters in the Weddell and Ross Gyres in the $1/12^\circ$ configuration. This is consistent with the documented excessive strengthen of the subpolar gyres in eddy-permitting ($1/4^\circ$) configurations, along with reduced import of Circumpolar Deep Water onto the shelf (Storkey et al., 2025)(Storkey et al., 2025). Additional drivers could also be related to the differences in vertical mixing parametrizations among the configurations.

The benefits of increased resolution also depend strongly on the diagnostic considered. Higher resolution improves some aspects of ice motion, including fast-drift regimes in the Antarctic marginal ice zone and pan-Arctic drift magnitudes, but it does not systematically improve seasonal phasing or transition timing. Similarly, increased resolution does not robustly improve Arctic or Antarctic low sea ice events. The 2022 Antarctic summer event is poorly captured because the configurations simulate too little ice in key sectors to reproduce the observed anomaly evolution, whereas the 2023 winter anomaly is better represented but not consistently improved with resolution.

These results suggest that several factors beyond horizontal resolution may contribute to the remaining model deficiencies. Potential candidates include atmospheric forcing, sea ice parameter choices and missing or incomplete physical processes (e.g., Barthélemy et al., 2018; Notz and Community, 2020; Nie et al., 2023). Atmospheric reanalyses, including ERA5, have known polar warm biases (Batak and Müller, 2019), but sensitivity tests with corrected ERA5 forcing show only limited improvement in the summer sea ice bias (Topal et al., 2026), suggesting that forcing errors are unlikely to be the sole cause. Parameter choices are also important: because of the high computational cost of the $1/12^\circ$ configuration, no dedicated tuning of sea ice or ocean parameters was performed, and values optimized for the $1/4^\circ$ configuration were retained. These parameters



may not be optimal at higher resolution and, in any case, reflect trade-offs that do not fully remove the negative summer sea ice
585 bias. Machine learning approaches may provide a practical route for optimizing such parameters at high resolution and reduced
computational cost.

Further progress will require both human and observational resources to better represent complex polar physical processes.
Several developments already point toward possible improvements in sea ice simulations, including coupling with biogeo-
chemistry leading to a better representation of absorbed solar radiation (Lim et al., 2019) and adding an additional degree of
590 freedom for the ice-ocean interface temperature (Shi et al., 2021b). Another potential mechanism affecting thickness differ-
ences between the 1° and $1/4^\circ$ configurations is the reduced variance of the atmospheric forcing at 1° , caused by interpolation of
 $1/4^\circ$ ERA5 fields onto the coarser grid. Such smoothing can lead to rectification effects and thicker ice, as shown for the central
Arctic (Richaud et al., 2025). This mechanism is less likely to explain differences between the $1/4^\circ$ and $1/12^\circ$ configurations,
where interpolation to a finer grid preserves monotonicity.

595 Future progress in polar ocean-sea ice modelling will therefore require treating horizontal resolution as one component of a
broader model-development strategy. Priority should be given to targeted process diagnostics in regions of strong sensitivity,
improved representation of ocean vertical mixing and ocean-ice heat exchanges, and sensitivity experiments addressing at-
mospheric forcing uncertainty. Fully coupled atmosphere-ocean-sea ice simulations are also needed, because many remaining
biases involve feedbacks through clouds, radiation, turbulent heat fluxes, near-surface stability and atmospheric circulation.
600 However, coupling should not be viewed as a guaranteed solution, as recent coupled-model studies also show modest and
regionally dependent improvements (e.g., Zapponini et al., 2026). Overall, the $1/4^\circ$ configuration appears to offer the strongest
cost-benefit compromise, while further gains will likely depend on coordinated improvements in forcing, vertical physics, sea
ice thermodynamics and coupled feedbacks.



Appendix A: Additional Figures

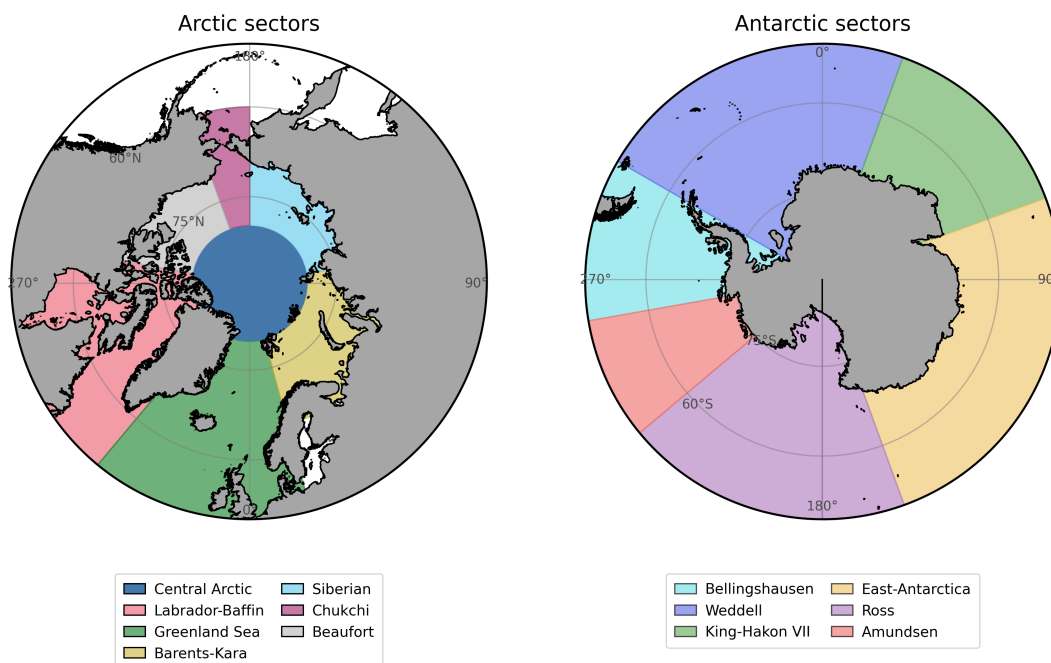


Figure A1. Sectoral decomposition of (a) the Arctic Ocean and (b) the Southern Ocean used in the study as shown in the legend.

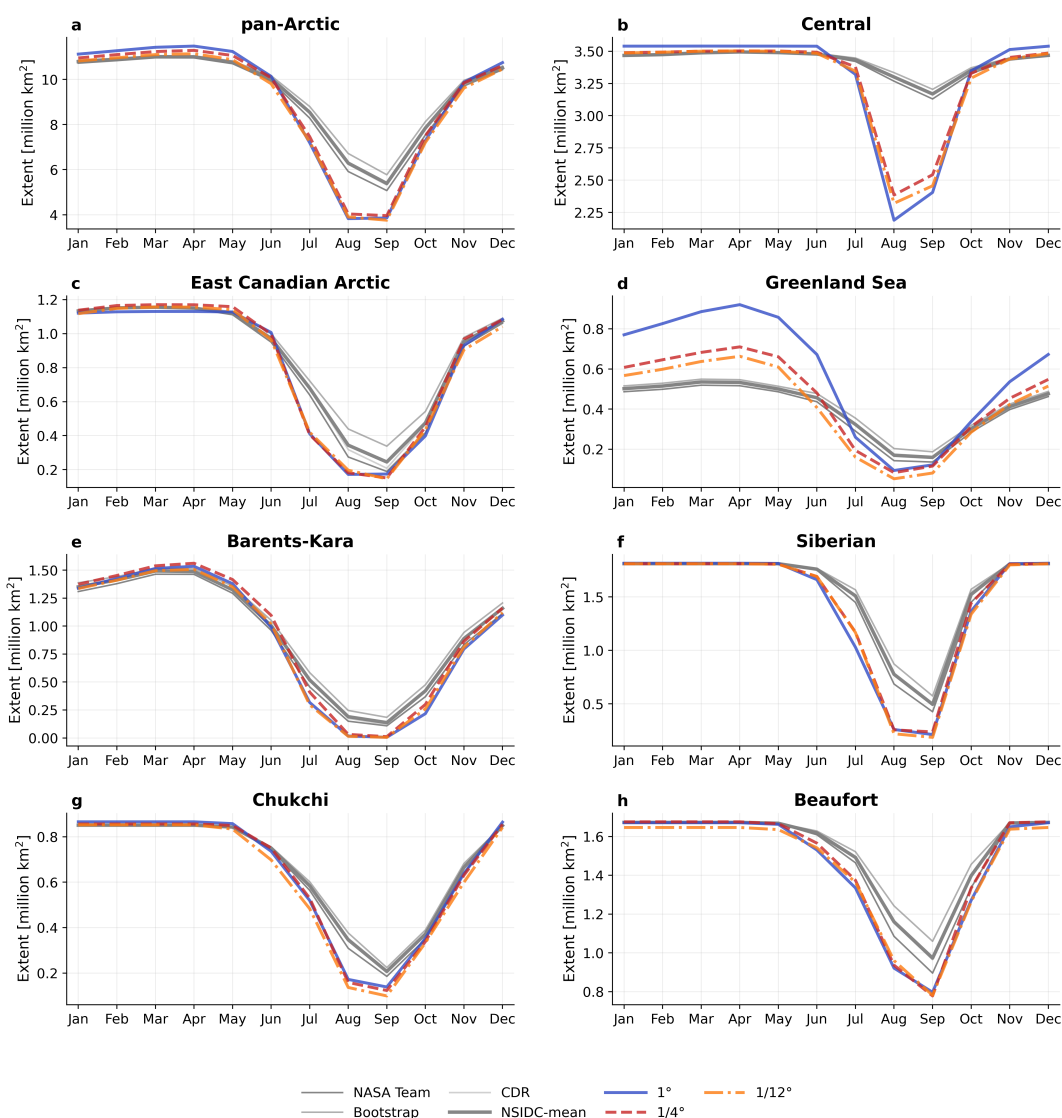


Figure A2. Climatological seasonal cycle of Arctic sea ice extent (SIE) computed for regional sectors of the Arctic Ocean as indicated in the panel titles (Central Arctic; East Canadian Arctic; Greenland Sea; Barents-Kara Sea; Siberian Sea; Chukchi Sea – including the Barents Sea and the Beaufort Sea), where pan-Arctic refers to SIE computed over the entire Arctic Ocean, in the three configurations (coloured lines) and in the NSIDC products (grey lines) as well as their mean (thick grey line) – as shown in the legend.

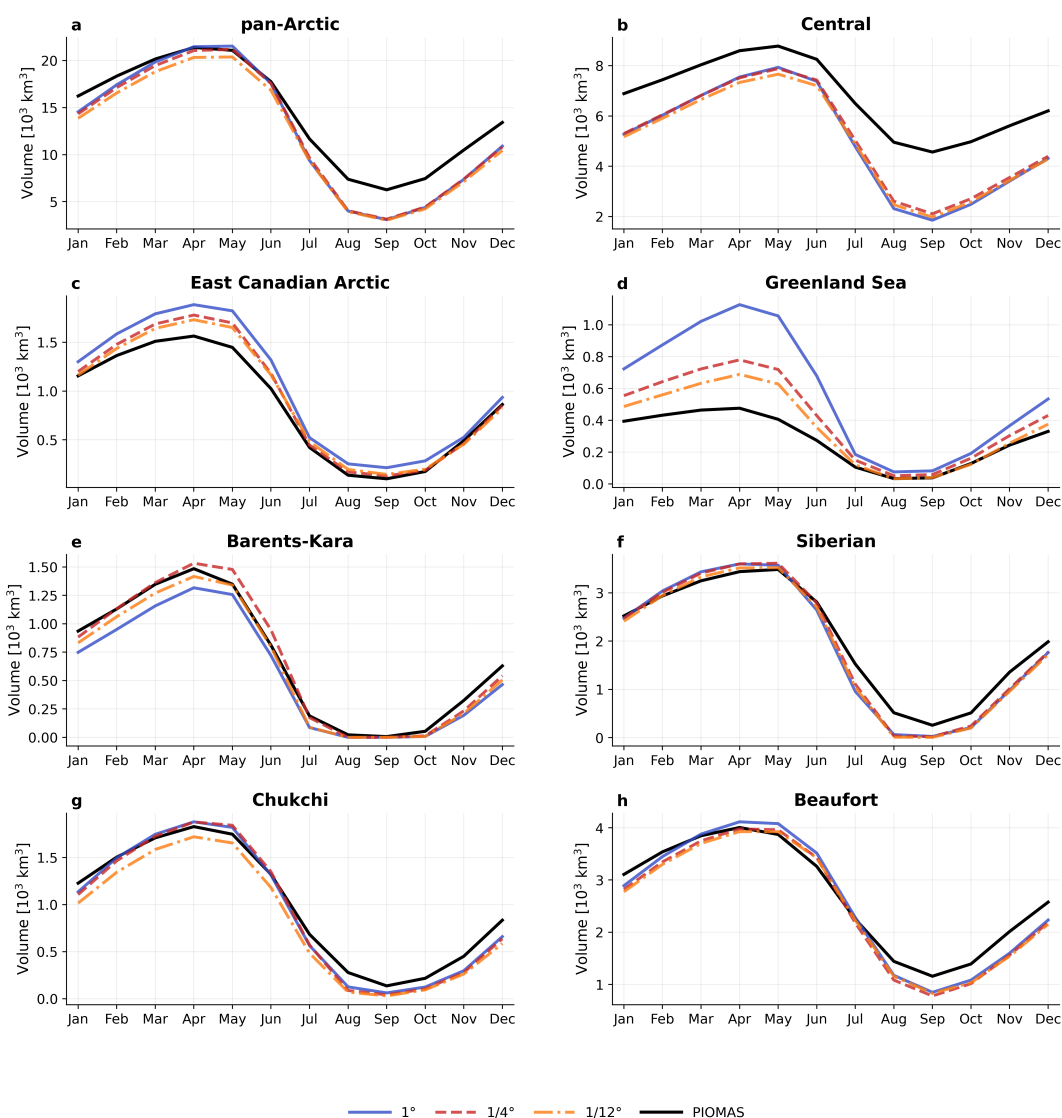
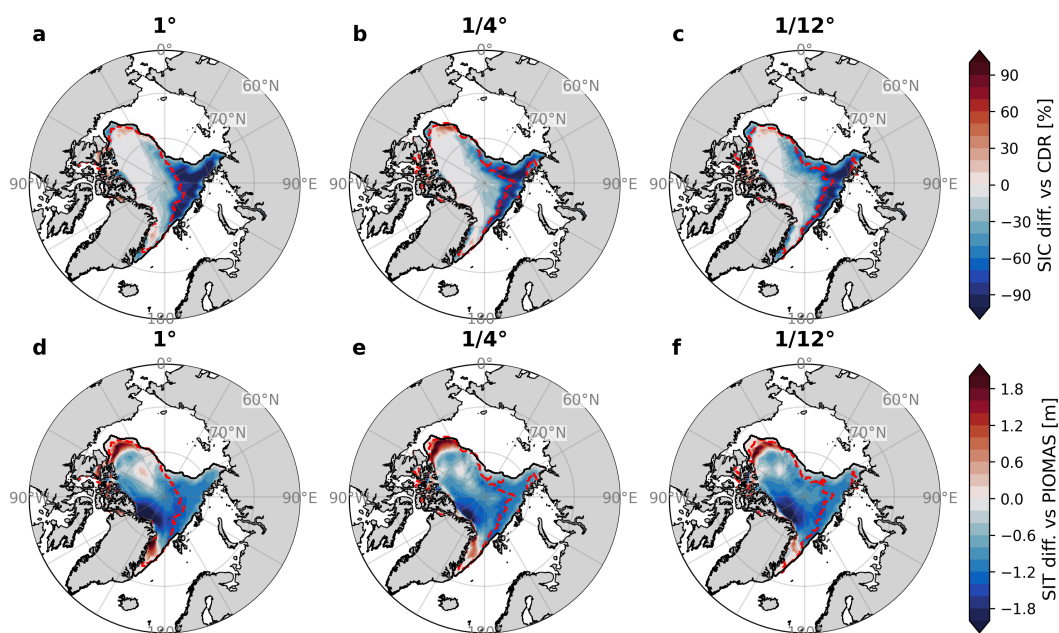


Figure A3. Climatological seasonal cycle of Arctic sea ice volume (SIV) computed for regional sectors of the Arctic Ocean as indicated in the panel titles and in Figure A2 (Central Arctic; East Canadian Arctic; Greenland Sea; Barents-Kara Sea; Siberian Sea; Chukchi Sea – including the Barents Sea and the Beaufort Sea), where pan-Arctic refers to SIV integrated over the whole Arctic Ocean, in the three configurations (coloured lines) and in PIOMAS reanalysis product (black line) as shown in the legend.



September 2007 difference



September 2012 difference

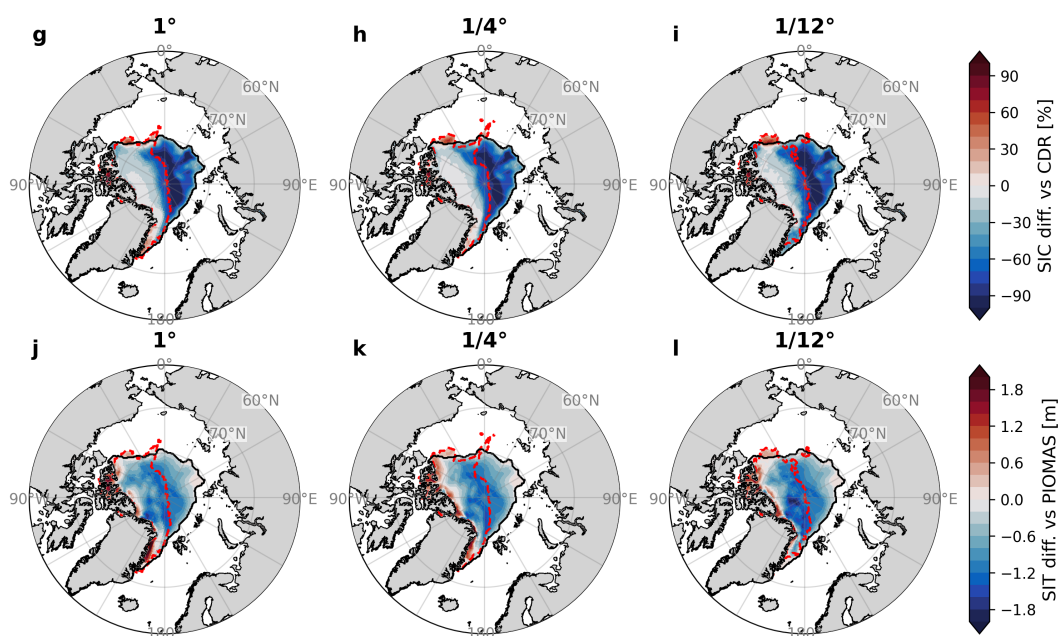


Figure A4. The spatial distribution of the monthly mean anomaly differences in sea ice concentration (SIC) (relative to the NSIDC Climate Data Record: CDR) in the three configurations for September (a)-(c) 2007 and (g)-(i) 2012. Panels (d)-(f) show the anomaly differences in sea ice thickness relative to the PIOMAS reanalysis in September 2007 and (j)-(l) in 2012. The model simulated ice edge (15 % ice concentration) contours are shown with red dashed lines, the NSIDC CDR ice edge is shown with black solid lines.

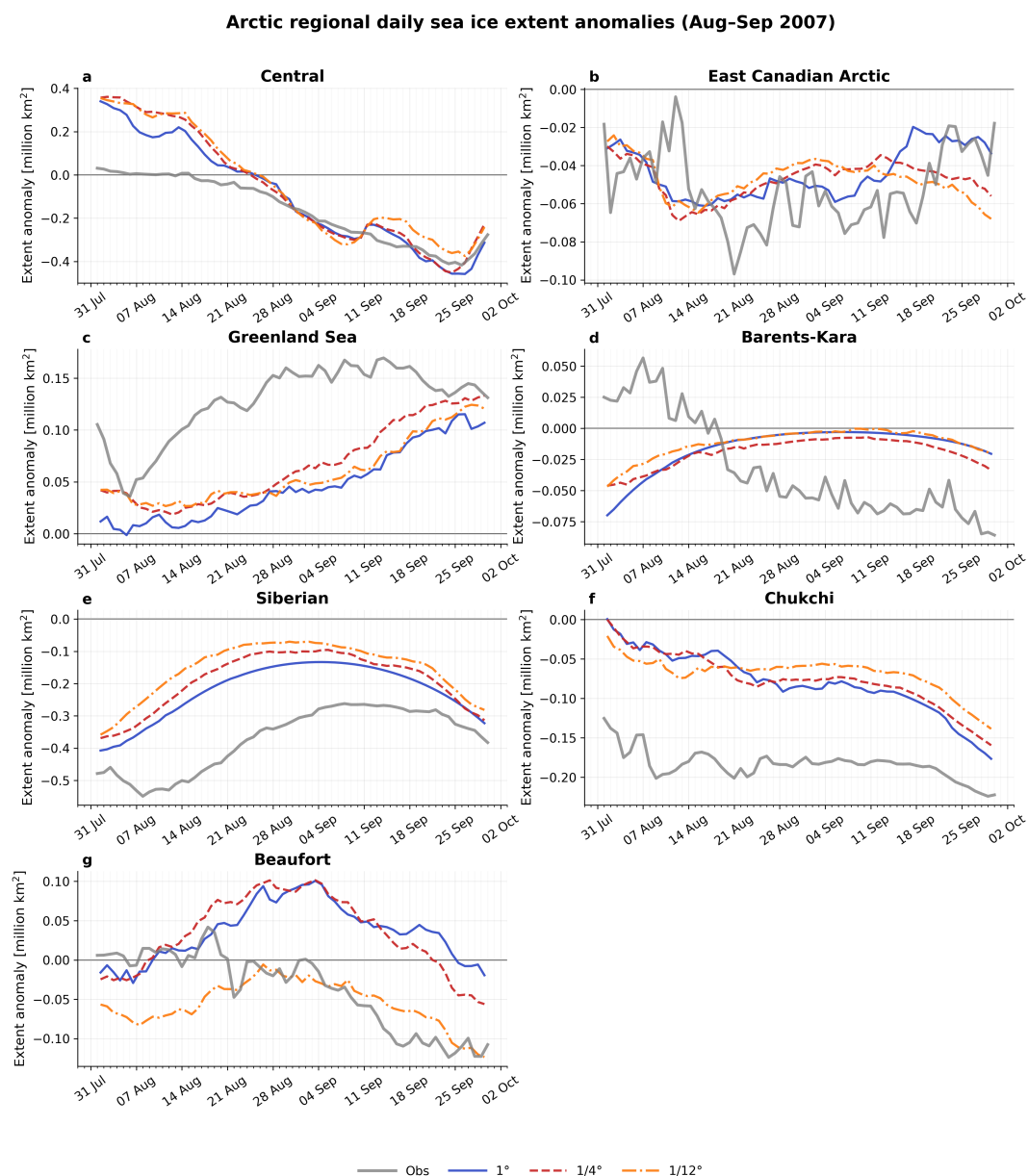


Figure A5. Arctic regional daily sea ice extent (SIE) anomalies in the three configurations and in the NSIDC Climate Data Record (CDR) – as shown in the legend – during August-September 2007.

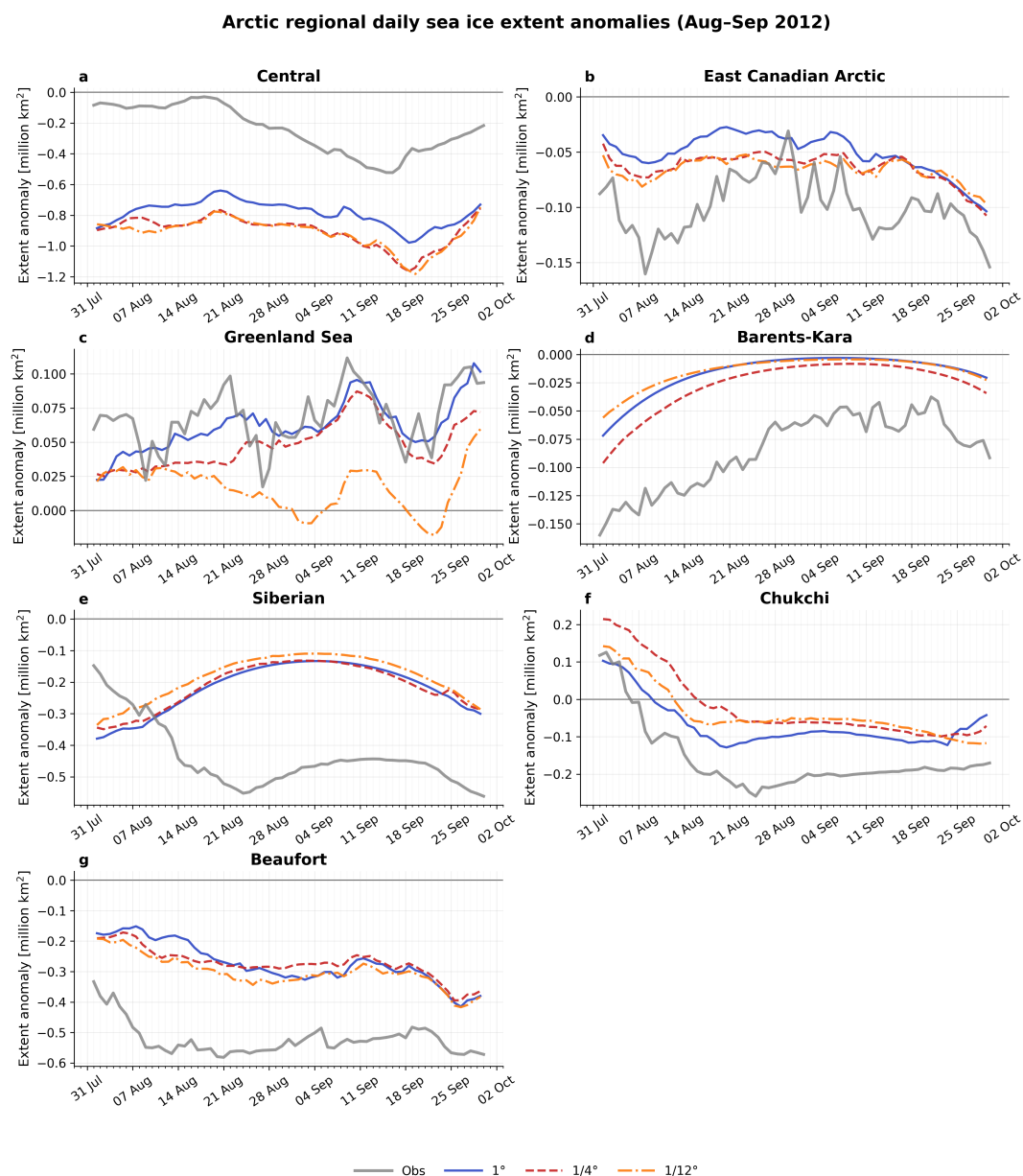


Figure A6. Arctic regional daily sea ice extent (SIE) anomalies in the three configurations and in the NSIDC Climate Data Record (CDR) – as shown in the legend – during August-September 2012.

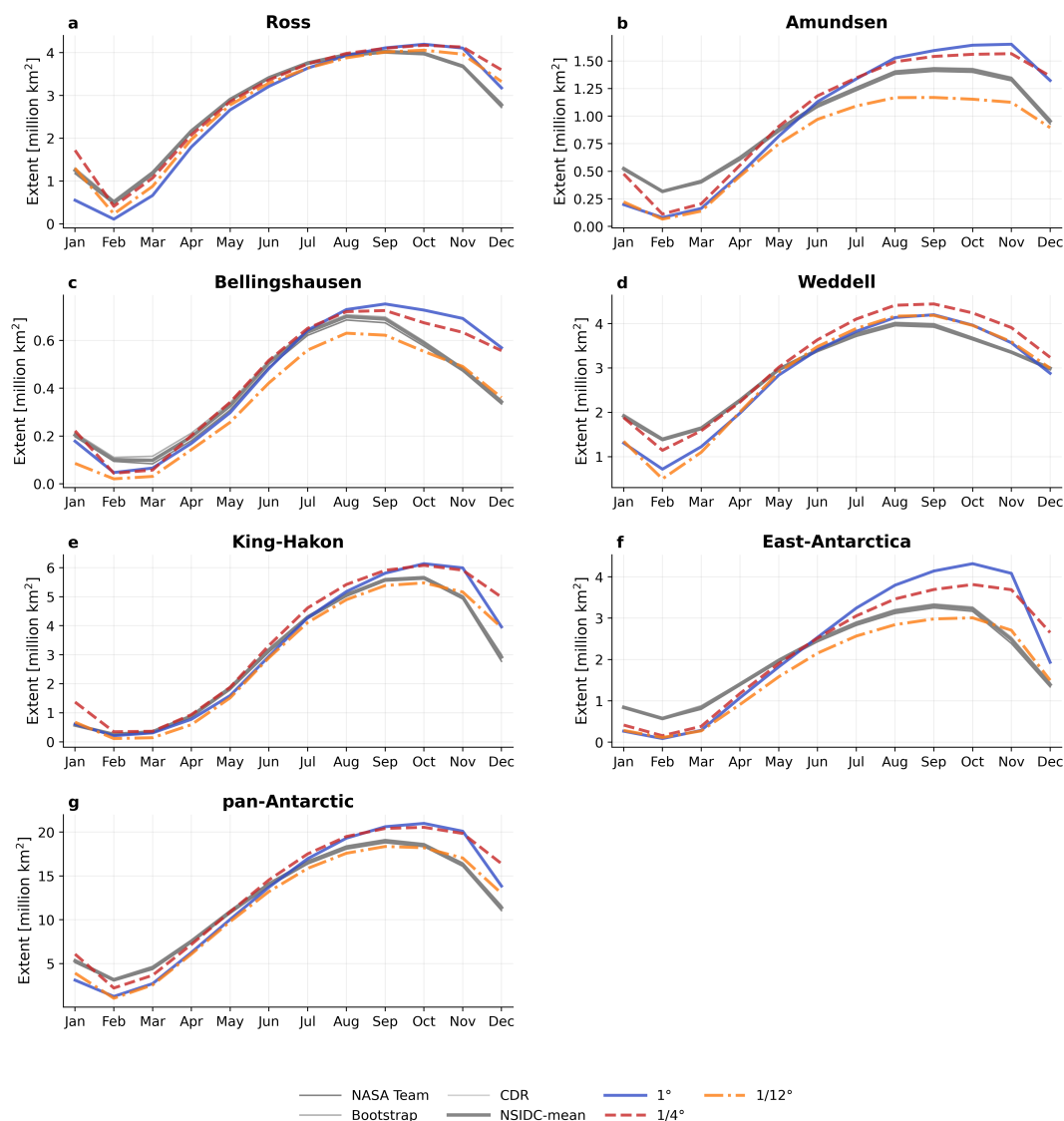


Figure A7. Climatological seasonal cycle of Antarctic sea ice extent (SIE) computed for regional sectors of the Southern Ocean as indicated in the panel titles (Ross Sea; Amundsen Sea; Bellingshausen Sea; Weddell Sea; King Håkon VII Sea; East Antarctica) where pan-Antarctic refers to SIE computed over the entire Southern Ocean, in the three configurations (coloured lines) and in the NSIDC products (grey lines) as well as their mean (thick grey line) – as shown in the legend.

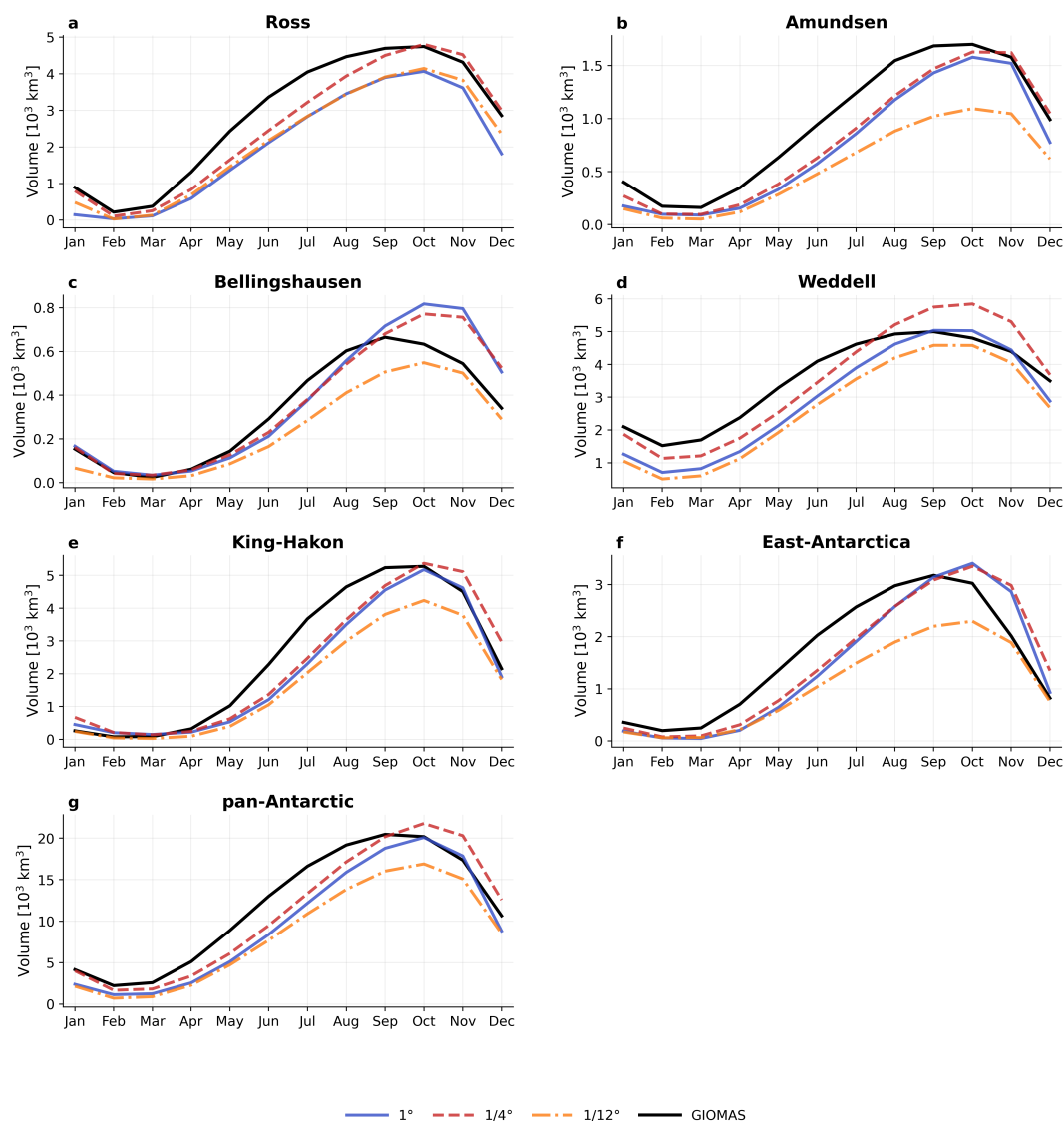


Figure A8. Climatological seasonal cycle of Antarctic sea ice volume (SIV) computed for regional sectors of the Southern Ocean as indicated in the panel titles (Ross Sea; Amundsen Sea; Bellingshausen Sea; Weddell Sea; King Håkon VII Sea; East Antarctica) where pan-Antarctic refers to SIT computed over the entire Southern Ocean, in the three configurations (coloured lines) and in GIOMAS reanalysis product (black line) as shown in the legend.

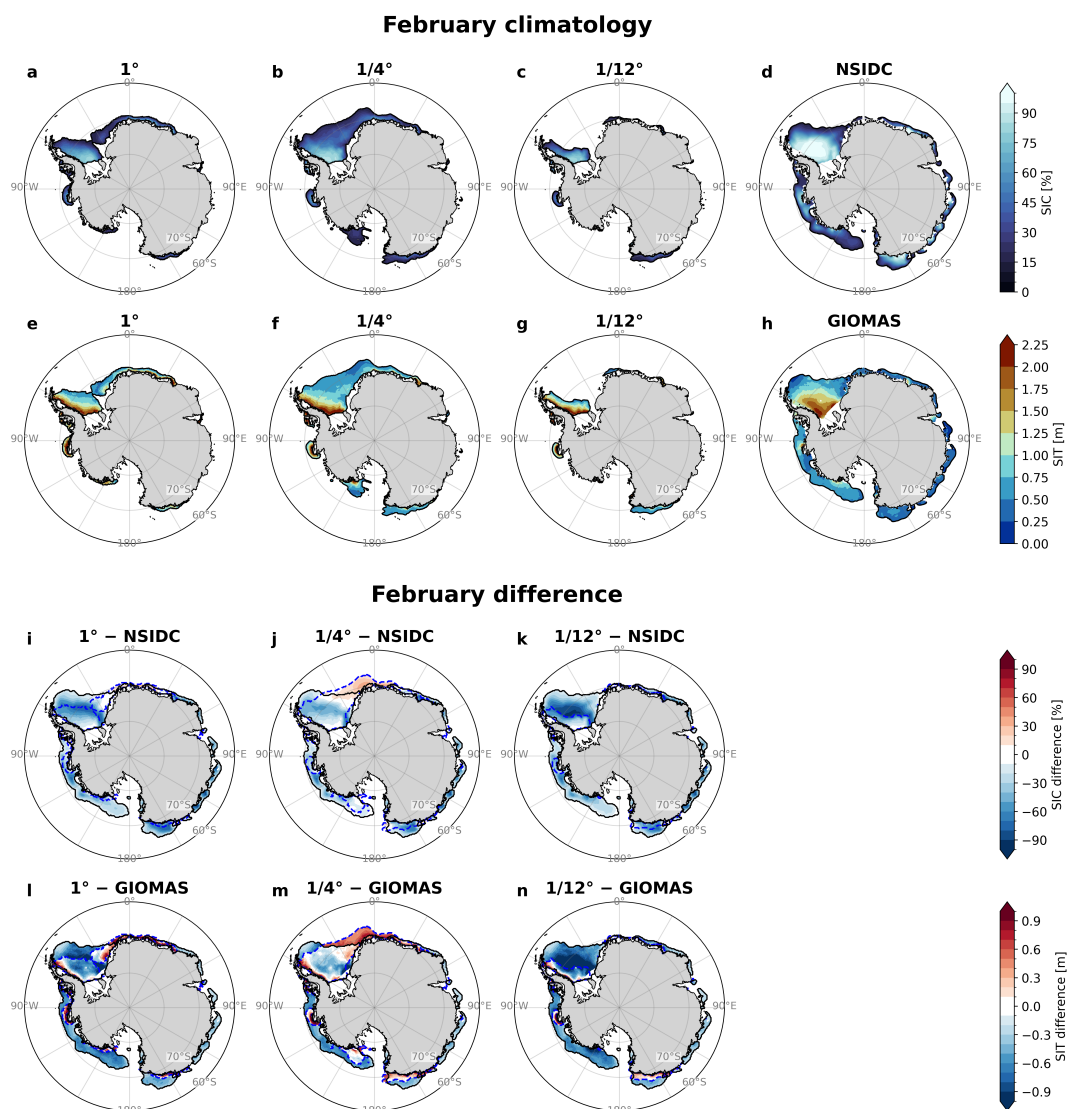


Figure A9. February Antarctic climatological sea ice concentrations (SIC: %) simulated by the (a) 1°, (b) 1/4° and (c) 1/12° configurations and as (d) observed (from the NSIDC Climate Data Record: CDR). September climatological sea ice thickness (SIT: m) simulated by the (e) 1°, (f) 1/4° and (g) 1/12° configurations and (h) from the GIOMAS reanalysis. The model simulated and observed (NSIDC CDR) climatological sea ice edge (15 % sea ice concentration contour), respectively, are also shown with black line on panels (a)-(h)). Also shown: February Arctic climatological SIC differences (relative to the NSIDC CDR) in the (i) 1°, (j) 1/4° and (k) 1/12° configurations. February climatological SIT differences (relative to GIOMAS) in the (l) 1°, (m) 1/4° and (n) 1/12° configurations. In panels (i)-(n) the model simulated ice edge is shown with dashed blue line and the NSIDC CDR ice edge is shown with solid black line.

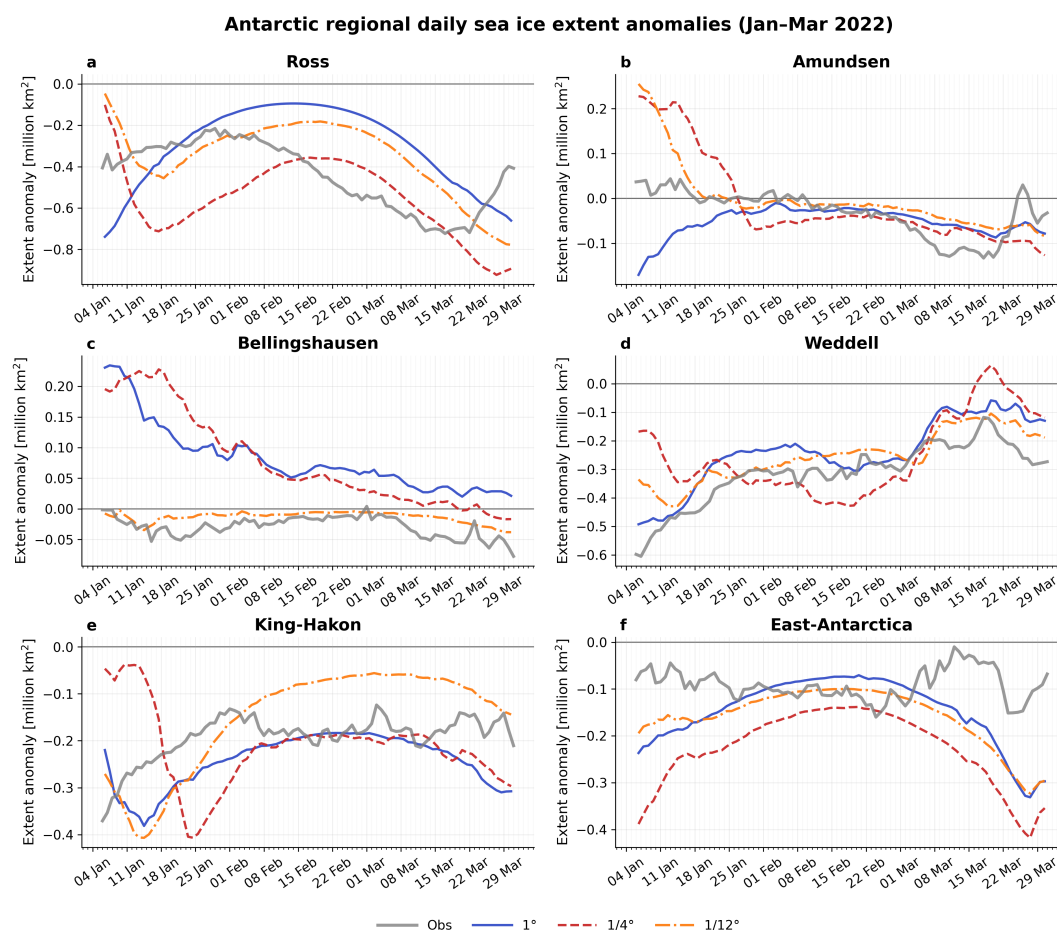
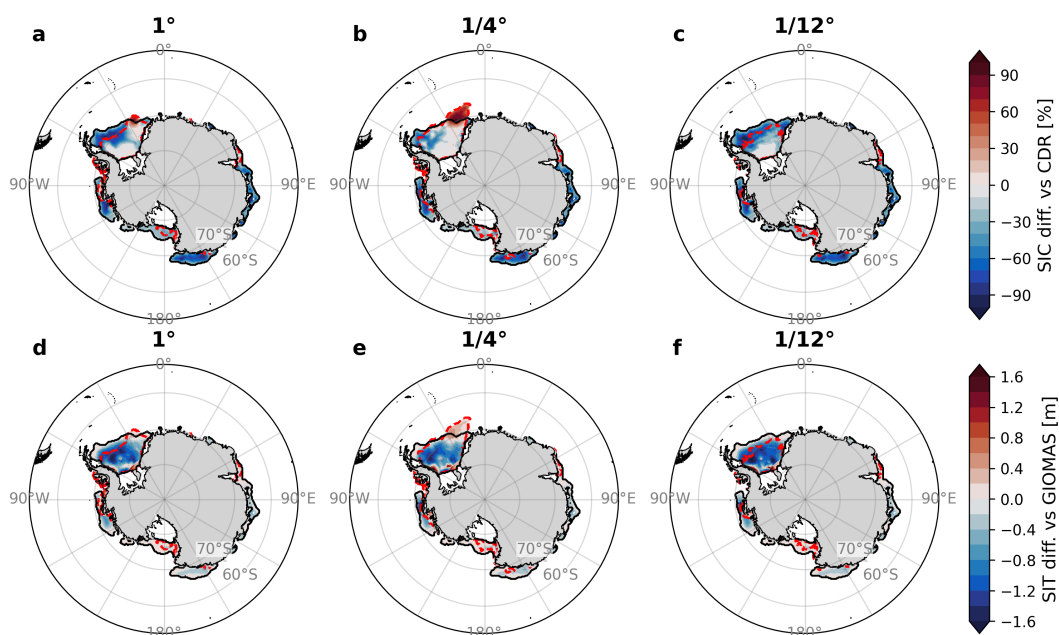


Figure A10. Antarctic regional daily sea ice extent (SIE) anomalies in the three configurations and in the NSIDC Climate Data Record (CDR) – as shown in the legend – during January-March 2022.



March 2022 difference



May-July mean 2023 difference

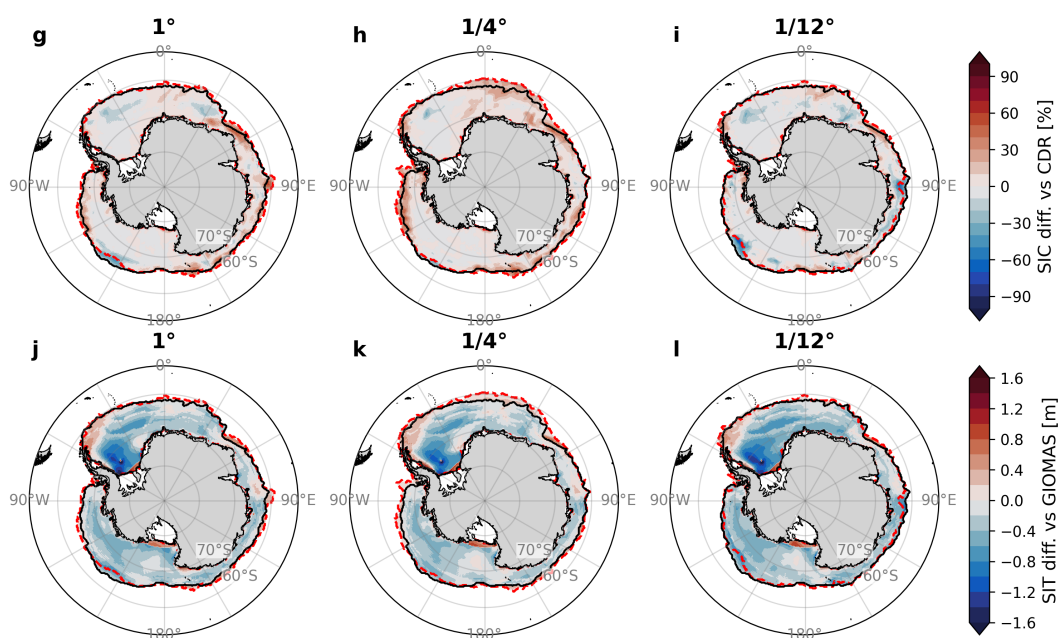


Figure A11. The spatial distribution of the monthly mean anomaly differences in sea ice concentration (SIC) (relative to the NSIDC Climate Data Record: CDR) in the three configurations for (a)-(c) March 2022 and (g)-(i) May-July 2023 (seasonal mean). Panels (d)-(f) show the anomaly differences in sea ice thickness relative to the PIOMAS reanalysis in March 2022 and (j)-(l) in May-July 2023 (seasonal mean). The model simulated ice edge (15 % ice concentration) contours are shown with red dashed lines, the NSIDC CDR ice edge is shown with black solid lines.

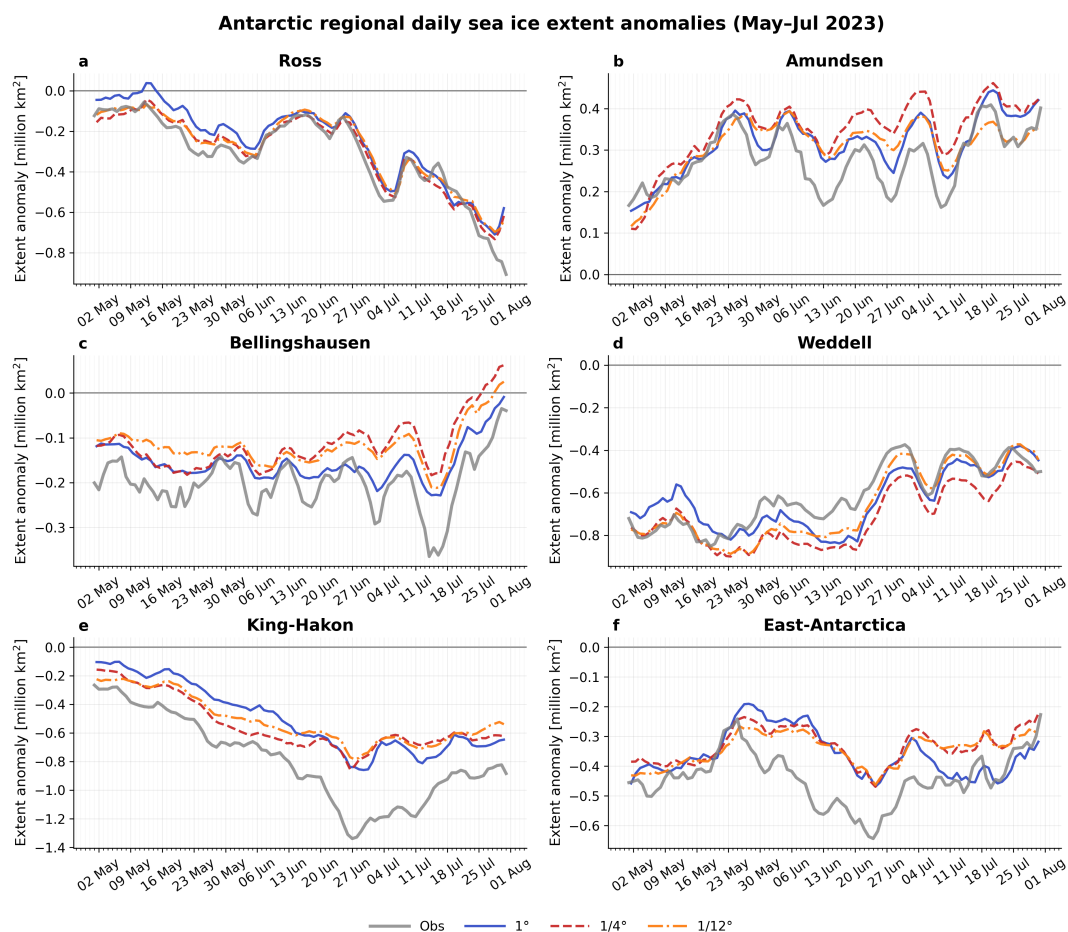


Figure A12. Antarctic regional daily sea ice extent (SIE) anomalies in the three configurations and in the NSIDC Climate Data Record (CDR) – as shown in the legend – during May-July 2023.



605 *Code and data availability.* Data are publicly available. The NSIDC sea ice concentration product can be found here: <https://nsidc.org/data/g02202/versions/6> (Meier et al., 2026). The sea ice reanalysis product PIOMAS can be found at <https://psc.apl.uw.edu/research/projects/arctic-seaice-volume-anomaly/>, while the GIOMAS product can be found at <https://psc.apl.uw.edu/data/global-seaice-giomas-data-sets/> (Zhang and Rothrock, 2003). EN4.2.2 data were obtained from <https://www.metoffice.gov.uk/hadobs/en4/> and are ©British Crown Copyright, Met Office, [2024], provided under a Non-Commercial Government Licence (<http://www.nationalarchives.gov.uk/doc/non-commercial-government-licence/version/2/>).

610 Model code for NEMO and SI3 is available through the official website at <https://www.nemo-ocean.eu> (last access: 23 July 2026). Configuration files for all three configurations used in this study are available at <https://forge.uclouvain.be/nemo-forks/nemo-4.2.2>. A table containing all namelist parameters differences across the three configurations is available from the corresponding authors on request. The ERA5 reanalysis product used to force the model is publicly available at <https://cds.climate.copernicus.eu/datasets/reanalysis-era5-single-levels?tab=download> (Hersbach et al., 2020; Copernicus Climate Change Service, Climate Data Store, 2023). Finally, the model outputs used for this study are publicly available at <https://doi.org/10.14428/DVN/JSDC8Q> (Barthélemy et al., 2026).

The code used to process and analyse the model outputs can also be accessed publicly on Zenodo at <https://doi.org/10.5281/zenodo.21508270> (Topal and Richaud, 2026).

Author contributions. B.R. and D.T. contributed equally to the analysis of the model results, the co-first authors' are listed in alphabetical order. B.R. and D.T. together led writing of the manuscript with contributions from all authors. D.T. created Figs. 1-5;9-14;17 and Figs. A1-A12; B.R. created Figs. 6-8;15-16. T.F., H.G., and F.M. conceived the idea for the creation of the simulations with discussions involving all co-authors on the experimental design. A.B. performed the simulations. D.T. has contributed with initial sensitivity experiments for parameter tuning.

Competing interests. The authors declare no conflict of interest.

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