



## FORTE 3.0 and its simulation of equatorial Pacific warming

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**Abstract.** FORTE 3 is an intermediate-resolution coupled atmosphere–ocean general circulation model (AOGCM) consisting of the Intermediate General Circulation Model 4 (IGCM4.1), a T42 spectral atmosphere with 35  $\sigma$ -layers, coupled to Modular Ocean Model – Array (MOMA), a  $2^\circ \times 2^\circ$  ocean with 15 z-layer depth levels. It is a significant upgrade from its predecessor, the fast flexible AOGCM FORTE2, in terms of its mean climate and climate response. The model has been compiled using the Intel ifx compiler. Here we describe the changes between FORTE2 and FORTE3 and demonstrate its use by simulating global and Pacific ocean change under both idealised CO<sub>2</sub> forcing and stratospheric O<sub>3</sub> depletion. In the atmosphere a new formulation for convective entrainment in IGCM4.1 alleviates a cold upper tropospheric bias in IGCM4. The formulation for precipitating convection has been changed to reduce an existing strong negative cloud forcing bias in the model, and a new formulation for stratocumulus cloud has been introduced. Effects of changes are especially evidenced in climate simulations: the transient climate response in FORTE3 is 1.8 K, and the equilibrium climate sensitivity in a slab version of IGCM4.1 is 4.2 K, compared to FORTE2/slab IGCM4 values of 1.0 K and 2.1 K respectively, and are more consistent with CMIP6 GCMs and observations. In the ocean, a new parameterisation of equatorial mixing based on the meridional sea surface temperature gradient alleviates the equatorial cold bias. The Atlantic Meridional Overturning Circulation at 30° N is 17 Sv, while the transport of the Antarctic Circumpolar Current is 130 Sv, slightly stronger, and more in agreement in observations than FORTE2. We demonstrate the use of FORTE3 by conducting sensitivity experiments where CO<sub>2</sub> is increased by 1 % per year. We find that the introduction of a "high-volume" Andes mountain range does not change the pattern of global warming over the Pacific Ocean; conversely, the effect of stratospheric O<sub>3</sub> depletion does appear to change the equatorial Pacific gradient slightly.

### 1 Introduction

Climate models are an essential tool to aid in addressing fundamental challenges in contemporary climate science, including understanding the Earth's climate system and generating reliable future projections (Randall et al., 2007). Within the hierarchy of climate models, intermediate-complexity coupled climate models offer a valuable balance between computational efficiency



and the representation of key large-scale climatic components and major modes of climate variability, allowing for long-term simulations and sensitivity studies.

Building upon the established framework of FORTE 2.0 (Blaker et al., 2021), FORTE 3.0 introduces significant advancements in its physical reparametrisations. These updates address persistent model biases in its predecessor, and enhance its representation of crucial climate system components. This paper introduces FORTE 3.0, detailing these updated model physics and evaluation against observational datasets.

Beyond these core model improvements and validation, FORTE 3.0's flexibility offers the ability to easily explore optional model configurations to investigate specific scientific questions. We leverage these capabilities to explore the mechanisms governing the equatorial Pacific sea surface temperature (SST) gradient response to warming, focusing on the role of stratospheric ozone depletion (Arblaster and Meehl, 2006; Dong et al., 2025), and the influence of the Andes orography (Takahashi and Battisti, 2007; Xu and Lee, 2021). These experiments aim to provide a deeper understanding of the complex interactions shaping the equatorial Pacific SST gradient under future warming scenarios.

The remainder of this manuscript is structured as follows: Section 2 provides a description of the FORTE 3.0 model, briefly outlining the core architecture of FORTE and highlighting the specific updates implemented in version 3.0. Section 3 presents the results of the 2000-year deep ocean spin-up, and details the model's pre-industrial state. Section 4 describes the effect of the optional model configurations employed in this study, and the application of FORTE 3.0 to investigate the equatorial Pacific SST gradient response under climate change. Finally, Section 5 offers a discussion of the key findings and conclusions.

## 2 Model Description

FORTE 3.0 is built upon the Intermediate General Circulation Model (IGCM) framework and the Modular Ocean Model (MOM). MOM is a widely used code for ocean circulation, originated at NOAA's GFDL (Griffies et al., 2015). The IGCM has history rooted in the Reading Spectral Model (Hoskins and Simmons, 1975) where its spectral atmospheric approach offers computational efficiency and accuracy for large-scale dynamics. The IGCM3 (Forster et al., 2000) was first coupled with the Modular Ocean Model – Array version (MOMA) (Webb, 1996) to create FORTE 1.0 (Sinha and Smith, 2002; Smith et al., 2004), a computationally practical system for long atmosphere-ocean simulations. A significant evolution of the IGCM code to IGCM4 (Joshi et al., 2015) provided the atmospheric component for FORTE 2.0 (Blaker et al., 2021), which coupled IGCM4 with MOMA, achieving stable simulations without flux adjustments and offering flexible atmospheric vertical resolution. FORTE 3.0 maintains this fundamental coupled structure of FORTE 2.0 while incorporating key updates to its atmospheric physics schemes.

### 2.1 A Brief Overview of FORTE Architecture

The FORTE model family, including FORTE 3.0, is characterised by its intermediate resolution, balancing computational cost with the representation of essential climate processes. The spectral atmosphere uses a jagged-triangular truncation at wave-number 42 (T42) which corresponds to a horizontal grid of 64 latitudinal and 128 longitudinal points with a horizontal



resolution of approximately  $2.8^\circ$ . It employs a semi-implicit, semi-Lagrangian time-stepping scheme (Hoskins and Simmons, 1975; Blackburn, 1985) for stability and efficiency and utilises a relatively high-top with 35 vertical sigma-levels (L35) reaching around 65 km in altitude. FORTE 3.0 now employs 120 time steps per day (12 minute time steps), increased from 96 in FORTE 2.0, to enhance running stability by better resolving high-frequency atmospheric dynamics. Several key atmospheric routines remain unchanged between FORTE 2.0 and FORTE 3.0. Notably these include: radiatively active species inclusion, aerosol treatment, gravity wave drag parametrisation, run-off and river outflow, surface fluxes and turbulence over land. For further details on these unchanged schemes see Joshi et al. (2015) and Forster et al. (2000).

MOMA, the ocean component, is a z-level model with 15 vertical levels, ranging from a 30 m mixed layer at the surface (matching the IGCM4 slab setup) to thicknesses of up to 800 m in its abyssal layers (O'Callaghan, 2016). The atmospheric and oceanic components are coupled using the OASIS coupler (Valcke et al., 2000). OASIS is a standard tool in the climate modelling community for exchanging fluxes and state variables between different model components. The majority of updates incorporated into FORTE 3.0 are fed in via updates to the atmospheric component of the model. The model has the ability to run with prescribed aerosols, simulating both their direct and indirect effects (Stjern et al., 2024).

FORTE3 and IGCM4.1 run on the UEA High Performance Compute System, and have been compiled under the ifx FORTRAN compiler.

## 2.2 IGCM4.1 Changes

A new surface climatology for vegetation type has been implemented, which controls snow-free albedo, snow-covered albedo, and roughness length, splitting the "desert" type into "Sahara" and "non-Sahara", with the former type having a higher surface albedo, consistent with observations (Rechid et al., 2009). A new "Sahel" semi-arid grasslands type has also been created, enabling exploration of the effects of changing surface properties in this area. The albedo of the Antarctic and Greenland ice sheets has been increased from 0.8 to 0.85 to provide a better match to observations of temperature.

The albedo of the ocean has been changed from its FORTE2 value to include a small "bump" to include the effect of sea-spray aerosol direct and indirect cloud effects in the Southern Ocean. Such an approach ensures that biases in the mean state of the model winds do not bias the location of what is an important regional component of the Earth's energy balance (Hyder et al., 2018). The albedo of the ocean is set to be

$$A = 0.35 - 0.29 \exp\left(-\left(\frac{\phi}{65}\right)^4\right) + 0.04 \exp\left(-\left(\frac{\phi + 60}{11}\right)^2\right), \quad (1)$$

where  $\phi$  is latitude in degrees, to very simply take account of the effect of zenith angle on ocean albedo (Jin et al., 2004). The newer tropical albedo is lower and more realistic than in IGCM4/FORTE2, and balances the larger amounts of marine stratocumulus cloud (MScC) in FORTE3 (see later).

### 2.2.1 Convection Scheme

The IGCM has a deep precipitating convection scheme, which relaxes  $\theta$  and  $q$  to stable profiles over a timescale  $\tau_c$  of 3 hours (2.5 hours in FORTE2). The scheme has been modified so that in both IGCM4.1 and FORTE3,  $\tau_c$  is 2.5 hours for  $q$ , but is



2.08 hours for  $\theta$  over the ocean only. The change is meant to represent convection over the ocean entraining relatively large amounts of drier air (Takahashi et al., 2023), which implies a lower timescale for mixing  $\theta$  compared to  $q$ . The result is that the Hadley circulation strengthens and widens slightly, alleviating a persistent mid-tropospheric cold bias in IGCM4 (see Joshi et al. (2015)), and an equatorial bias in the southern midlatitude jet.

## 90 2.2.2 Convective Cloud

The cloud fraction associated with deep precipitating convection in IGCM3 and IGCM4 has been parameterised by a function that varies with  $\log(P)$ , where  $P$  is precipitation. Such a function causes a large negative cloud radiative forcing under climate change scenarios, since  $P$  increases with surface temperature. Given that it is unclear whether such a relationship should hold under such different forcings, the relative simplicity of the diagnostic cloud scheme which, for instance, has a prescribed effective cloud water particle size, and the unrealistically large negative forcing, the parameterisation has been replaced with a simplified deep cloud fraction of 0.2.

## 2.2.3 Marine Stratocumulus Cloud

Deficiencies in simulating Marine Stratocumulus Cloud (MScC), which significantly impact solar radiation, are a common source of persistent biases in coupled atmosphere-ocean models (Clement et al., 2009). Accurately representing sub-tropical Southeast Pacific stratocumulus clouds and their associated cloud feedbacks requires improved model physics and realistic large-scale temperature structures Lin et al. (2014). IGCM4/FORTE2 had a MScC parameterisation based on static stability in the lowest two layers of the model. This has now been replaced by a scheme dependent on surface fluxes and atmospheric humidity, such that

$$\text{MScCF} = 0.25 \left( 1 + \tanh \left( \frac{q_r - 1.55}{0.1} \right) \right) \left( 1 - \tanh \left( \frac{F_L + F_S - 190}{10} \right) \right), \quad (2)$$

where MScCF is MScC fraction,  $q_r$  is the ratio of  $q$  between the lowest and second lowest layers of the atmosphere, and  $F_L$  and  $F_S$  are the latent and sensible heat fluxes from ocean to atmosphere, respectively. The formulation allows a smooth transition from zero MScCF over most of the oceans to high values when high vertical gradients in  $q$  occur, typically in MSc zones, and when  $F_L$  and  $F_S$  fall below  $190 \text{ W m}^{-2}$ , typical of colder tropical regions or midlatitude regions. The dependence on  $F_L$  and  $F_S$  follows arguments by e.g. Bretherton (1997), who suggests that MScC clouds break up when  $F_L$  and  $F_S$  become large enough to allow convection and cumulus clouds to form. The parameterisation is tuned in order to increase MScCF in subtropical eastern ocean basins compared to IGCM4/FORTE2, while ensuring that MScCF does not unduly increase in the equatorial Pacific ocean, which encounters similar conditions because of the cold equatorial ocean bias in the model, typical of most GCMs.

It is well-known that convective schemes can significantly impact a model's sensitivity to radiative forcing (Tomassini et al., 2015). The net sum of the above changes is that the equilibrium climate sensitivity (ECS) of IGCM4.1 coupled to a simplified slab ocean is 4.2 K, which is more consistent with observations compared to earlier versions of the model, such as IGCM3 (1.6 K, see Forster et al. (2000)) or IGCM4 (2.1 K, see Joshi et al. (2015)).



#### 2.2.4 Equatorial Surface Stress

Sea surface temperature biases over the equatorial Pacific can impact the simulation of crucial climate phenomena such as the El Niño-Southern Oscillation (ENSO) (Wu et al., 2022) and, through atmospheric teleconnections, can therefore influence global weather patterns and remote climate variability (Bayr et al., 2019). An update to the treatment of equatorial oceans has aimed for a more realistic simulation of the observed SST structure in this critical region and its far-reaching impacts. A parameterisation of mixing has been implemented which reduces the surface stress felt by the ocean in a manner that is dependent on the meridional surface temperature gradient  $\partial T / \partial \phi$  at the equator. The surface stress  $\tau$  is replaced with a new adjusted value such that  $\tau_a = A\tau$ , where

$$A = 0.6 + 0.4 \tanh \left( 2.5 + 2.8 \operatorname{sgn}(\phi) \frac{\partial T}{\partial \phi} \right). \quad (3)$$

The new formulation reduces  $\tau$  when  $T$  decreases towards the equator, implying the presence of baroclinic instability, and hence tropical instability waves. Observations have suggested that the presence of such waves can reduce mean stresses (Chelton et al., 2001). The targeted reduction of equatorial wind stress in this manner helps to alleviate cold biases in the models caused by feedbacks involving overly strong easterly stresses and upwelling of cold water.

#### 2.2.5 Other Ocean updates

The wind-dependent drag formulation in FORTE2 has been removed, because coupled biases in the mean state meant that this formulation was making the mean state of the model worse in the tropics. The net sum of the above changes is that the transient climate response of FORTE3, i.e. the increase in global temperature after 70 years of increasing CO<sub>2</sub> at a rate of 1% per year, is 1.8 K, which is more consistent with observations and CMIP6 GCMs compared to FORTE2, which was 1 K (see e.g. Stjern et al. (2024)). The ECS and TCR values of 4.2 K and 1.8 K place FORTE3 in the middle of the pack of CMIP6 GCMs (see Nijssse et al. (2020) Fig. 2).

### 3 Spin-up and pre-industrial state

Attaining a stable and well-equilibrated deep ocean state is a crucial prerequisite for meaningful climate change and long-term variability simulations with atmosphere-ocean models (Bryan, 1984). FORTE3 was accordingly spun up for 2000 model years from rest with initial salinity and temperature oceanic conditions taken from Levitus (Levitus and Boyer, 1998; Levitus et al., 1998), and with pre-industrial conditions (285 ppmv CO<sub>2</sub> and default orography); this is denoted the ‘Pre-industrial Control’ (PiC) run. The model was then integrated forward for another 50 years and data from years 2000–2050 extracted for analysis.

#### 3.1 Large-Scale Climate

Figure 1 shows the difference in the zonally-averaged state between IGCM4 and IGCM4.1, which demonstrates the effect of the change in convection outlined above without the complicating factor of differing SSTs between FORTE2 and FORTE3.



In DJF (Fig. 1a) a strengthened Hadley circulation has shifted the southern jet stream polewards, and is more consistent with observations. The tropical mid-tropospheric cold bias in IGCM4, which reached 5–10 K in amplitude (Joshi et al., 2015) has been alleviated significantly, with IGCM4.1 displaying tropical mid-tropospheric temperatures 4 K warmer than IGCM4. The actual zonally-averaged state in FORTE3 is shown in Fig. 2. The model replicates the pattern of weak stratospheric westerlies in the northern hemisphere in DJF and stronger stratospheric westerlies in JJA in the southern hemisphere.

Figure 3 shows the differences in surface temperatures between EN4 observations (Good et al., 2013; Gouretski and Reseghetti, 2010) and both FORTE2 and FORTE3. Both models display similar biases, with a western tropical Pacific that is slightly too cold and an eastern tropical Pacific that is too warm; the warm bias in FORTE3 has been alleviated by the improved representation of MScC. A notable difference is that in both DJF and JJA, FORTE3 exhibits much warmer North-East Atlantic SSTs, associated with its stronger Atlantic Meridional Overturning Circulation (AMOC) (see below).

The changes in low-level clouds, especially MScC, that cause the above tropical changes are shown in Fig. 4. We show MScC in both the atmosphere-only and coupled versions of the model to demonstrate the effects of coupled feedbacks. Compared to ISCCP observations (Fig. 4e) (Rossow et al., 2016; Young et al., 2018), both IGCM4.1 and FORTE3 (Fig. 4b and d respectively) have too little low cloud, which is partially a result of tuning the new MScC parameterisation in order to avoid too much low level cloud over the equatorial Pacific region. Even then, some low cloud does persist over the equatorial Pacific region. However, when comparing IGCM4.1 to IGCM4 (Fig. 4b and a), and FORTE3 to FORTE2 (Fig. 4d and c), it can be seen that these models have more low cloud in general, and more MScC cloud in the tropical oceans, which is more consistent with observations.

Figure 5 shows DJF and JJA precipitation and differences between 20CRv3 (Compo et al., 2011; Slivinski et al., 2019) reanalysis of the period 1871–1896: a proxy for the pre-industrial period. Figure 5a and 5b show that DJF precipitation is represented quite well in FORTE3, albeit with a South Pacific Convergence Zone (SPCZ) that is slightly weak. Figure 5c and 5d show that JJA precipitation is well-represented. Over Asia there are biases, to be expected of a model that does not resolve the complex high topography in this region. Figure 5e and 5f show that the atmosphere-only IGCM4.1 represents DJF precipitation very well. Figure 5g and 5h show similar for JJA. A wet bias exists in the Northwest tropical Pacific, but this bias is smaller in magnitude than in IGCM4.

### 3.2 Ocean state and circulation

The global mean sea surface temperature in FORTE3 equilibrates after about 750 years to a stable value of 18.8 °C. Key circulation indicators in FORTE3 are shown in Fig. 6, which shows the Antarctic Circumpolar Current (ACC) (Fig. 6a) and AMOC (Fig. 6b). Both currents are stronger than in FORTE2, with the transport of the ACC at 140 Sv compared to 115 Sv in FORTE2, and the AMOC at 16–18 Sv in strength, compared to 13–15 Sv in FORTE2; both are closer to observations in FORTE3 (Meredith et al., 2011; Koenig et al., 2014; Donohue et al., 2016; Johnson et al., 2025). The stronger AMOC drives a significantly warmer Northeast Atlantic SST compared to FORTE2, shown in Fig. 3. Figure 7 shows the overturning streamfunction globally and in the North Atlantic Ocean in more detail. The North Atlantic streamfunction in particular extends to 70° N (Fig. 7b), consistent with the warmer Northeast Atlantic shown in Fig. 3c and f.



The evolution of area-averaged global ocean temperature and salinity in FORTE3 is shown in Fig. 8. Below 4000 m, the temperature warms compared to the initial conditions, opposite to what was seen in FORTE2. The differences are likely due to the stronger AMOC moving more heat to abyssal depths. The behaviour of salinity is very similar to FORTE2, with an anomaly of +0.2 in the top 2000 m and -0.1 in the lowest 2000 m. Surface salinity is shown in Fig. 9. Overall, the pattern of a saltier Atlantic and a fresher Pacific is reproduced. Compared to WOA23 observations (Reagan et al., 2023, 2024), the tropics are slightly fresh, while the extratropics are slightly more saline. A bias in North Pacific salinity in FORTE2 of +2 has been significantly reduced.

#### 4 Sensitivity experiments

Over the equatorial Pacific there is a notable discrepancy between observational evidence and general circulation model behaviour. Since the mid-20th century observations reveal a strengthening of the tropical Pacific zonal SST (Seager et al., 2019; Collins, 2005), which GCMs typically fail to simulate. Sensitivity tests were performed with FORTE3 to assess the sensitivity of the change in zonal equatorial Pacific gradient to certain model parameters, such as Andes topography, which is known to affect the eastern tropical Pacific climate (Takahashi and Battisti, 2007; Xu and Lee, 2021), and stratospheric  $O_3$  depletion, which has been hypothesised to cause a cooling signal in the Southern Ocean (Arblaster and Meehl, 2006), which can then propagate northwards into the Southeast Pacific region (Kang et al., 2026).

Following the 2000-year deep ocean spin-up (see Section 3) two FORTE3 runs were carried out by restarting from rest, but with ocean temperature and salinity conditions obtained from the end of the 2000-year long spinup. Two configurations of the model were run, one with the default Andes topography (denoted run I1), and one with a higher maximum height (run I2), these configurations are shown in Fig. 10. The aim was to spin up two model configurations which could be compared to each other after enough spin up time needed for any effects of the Andes topography to equilibrate. These configurations were run for 550 years, and the last 50 years of the integrations compared to investigate the effect of changing Andes topography on the mean state. Note that the volume as well as the height of the Andes was changed: the gradient of the Andes is limited by numerical considerations in any numerical model, and so simply increasing the height, and therefore gradient, of the Andes would cause numerical problems. Run I2 does have side effects on the simulation of South American climate, but is meant to be an idealised demonstration of the effect of the Andes on climate and climate change.

After 550 years,  $CO_2$  was increased in each run by 1% per year for 80 years. An additional experiment, denoted I3, was carried out using the default Andes topography, but with stratospheric  $O_3$  depleted using the following idealised formulation at every model layer:

$$\text{Adjusted } O_3 = \text{Standard } O_3 \times \left[ 1 - (1 - L_f) \exp \left( - \left( \frac{\cos(D_y + 60) - 1}{1.5} \right)^2 \right) \right], \quad (4)$$

where  $D_y$  is the day of year from 0 to 360,

$$L_f = Y_f + (1 - Y_f) \tanh \left( \frac{\phi + 50}{5} \right), \quad (5)$$



and

$$Y_f = 1 - 0.35 \exp \left( - \left( \frac{Y - 40}{Y_s} \right)^2 \right), \quad (6)$$

where  $Y$  is the time in years since  $\text{CO}_2$  started to increase from pre-industrial levels. If  $Y < 40$ ,  $Y_s = 20$  and if  $Y > 40$ ,  $Y_s = 40$ .

215 The function in Eq. (4) represents a seasonal variation of the idealised ozone hole that maximises when  $D_y = 300$ , the function in Eq. (5) limits the ozone hole to over Antarctica, and the function in Eq. (6) denotes the time evolution of the ozone hole. The aim of this experiment was to see if an idealised, tunable, yet realistic, stratospheric  $\text{O}_3$  change, with an atmospheric response similar to previous work (Rosier and Shine, 2000) would affect the Southern Ocean, and potentially the Pacific Ocean, and is denoted run I3. The difference in surface temperature in years 500–550 between I2 and I1 is shown in Fig. 11. As seen in Xu  
 220 and Lee (2021), a cooling does occur on the eastern boundary of the Pacific Ocean. Other changes to mean climate are less than 1 K.

When  $\text{CO}_2$  is then increased by 1% per year, the global mean surface temperature change is shown in Fig. 12. I2 and I1 indicate a TCR of 1.8 K, showing that the presence of the high-volume Andes topography has not affected climate feedbacks globally. I3 warms slightly more slowly, consistent with the small negative radiative forcing from stratospheric  $\text{O}_3$  depletion in  
 225 this configuration (Forster and Shine, 1997).

The effects of stratospheric  $\text{O}_3$  depletion are shown in Fig. 13. In the 20 years after the maximum in  $\text{O}_3$  depletion, the stratosphere does display a stronger jet and a colder polar stratosphere, consistent with earlier simulations (Rosier and Shine, 2000). The dynamical signal of the cooling does extend to the surface, giving a characteristic dipole in zonally averaged zonal wind, consistent with a positive forcing of the Southern Annular Mode (SAM), which has been seen in observations, e.g.  
 230 Arblaster and Meehl (2006). The SAM change can drive a cooling of the Southern Ocean: the difference between runs I3 and I1 is shown in Fig. 14. The Southern Ocean in I3 warms more slowly than I1 in the first 50 years after  $\text{CO}_2$  increase, and the Indian Ocean sector cools slightly (Fig. 14b).

Figure 15 shows the evolution of time-averaged equatorial ( $10^\circ \text{S}$ – $10^\circ \text{N}$ ) Pacific SST in the three runs. Run I1 displays a small sign of an increase in the E-W SST gradient seen in observations at the start of the run, although this signal does  
 235 disappear 50 years after  $\text{CO}_2$  starts to increase. FORTE3 does not display the cooling trend seen in observations. Neither runs I2 or I1 give any indication of any significant and sustained Eastern Pacific cooling trend compared to I1, suggesting that in this model, neither changed topography or significant  $\text{O}_3$  depletion drive an intensification of the equatorial Pacific SST gradient as seen in observations.

## 5 Conclusions

240 We present an updated version of the FORTE AOGCM, “FORTE3.0”. This version of the GCM has a mean state, cloud feedbacks, ECS and TCR which lie in the range of CMIP GCMs and observations, improving its utility as a tool for studying climate change and circulation. Its speed, flexibility and resolution of the stratosphere make it uniquely suited to investigating problems such understanding the effect of stratospheric  $\text{O}_3$  depletion on the coupled climate system under a warming climate.



245 Future work will involve conducting more sensitivity studies with the O<sub>3</sub> distribution in space and time, with more ensemble members to assess signal/noise.

*Code availability.* The code, compilation instructions and example run scripts, together with all necessary ancillary files, are accessible at <https://doi.org/10.5281/zenodo.22307102> (Robbins et al., 2026). The configuration committed to the Zenodo archive (v3.0.0) is the one used to produce the simulations presented in this paper.

250 Processing of the IGCM4 output requires the BGFLUX programme, a copy of which is accessible from the FORTE 3.0.0 GitHub repository linked to the Zenodo archive.

*Author contributions.* CR, AG, MJ and DS performed model experiments and analysis; MJ, DS, MW and AB contributed to model development; All authors contributed to writing.

*Competing interests.* We declare no competing interests.

255 *Acknowledgements.* The research presented in this paper was carried out on the High Performance Computing Cluster supported by the Research and Specialist Computing Support service at the University of East Anglia. We acknowledge the support of UKRI-NERC grant ImPOse (NE/W005239/1) for this research. Mark Webb was supported by the Met Office Hadley Centre Climate Programme funded by DSIT. Adam Blaker was supported by the NERC AtlantiS programme (NE/Y005589/1) EN.4.2.2 (g10) data were obtained from <https://www.metoffice.gov.uk/hadobs/en4/> and are © Crown Copyright, Met Office, provided under a Non-Commercial Government Licence <http://www.nationalarchives.gov.uk/doc/non-commercial-government-licence/version/2/>, these data were accessed in April 2026. Support  
260 for the Twentieth Century Reanalysis Project dataset is provided by the U.S. Department of Energy, Office of Science Innovative and Novel Computational Impact on Theory and Experiment (DOE INCITE) program, and Office of Biological and Environmental Research (BER), and by the National Oceanic and Atmospheric Administration Climate Program Office.



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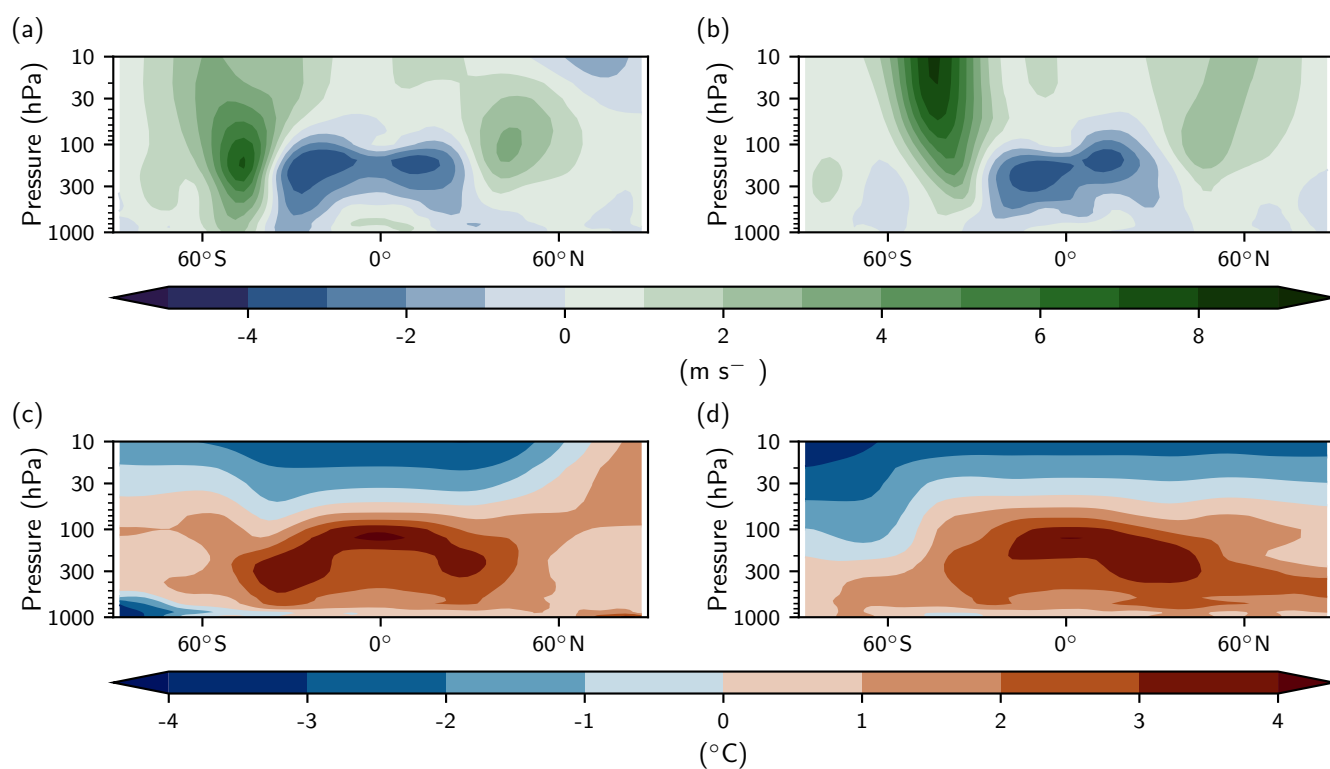
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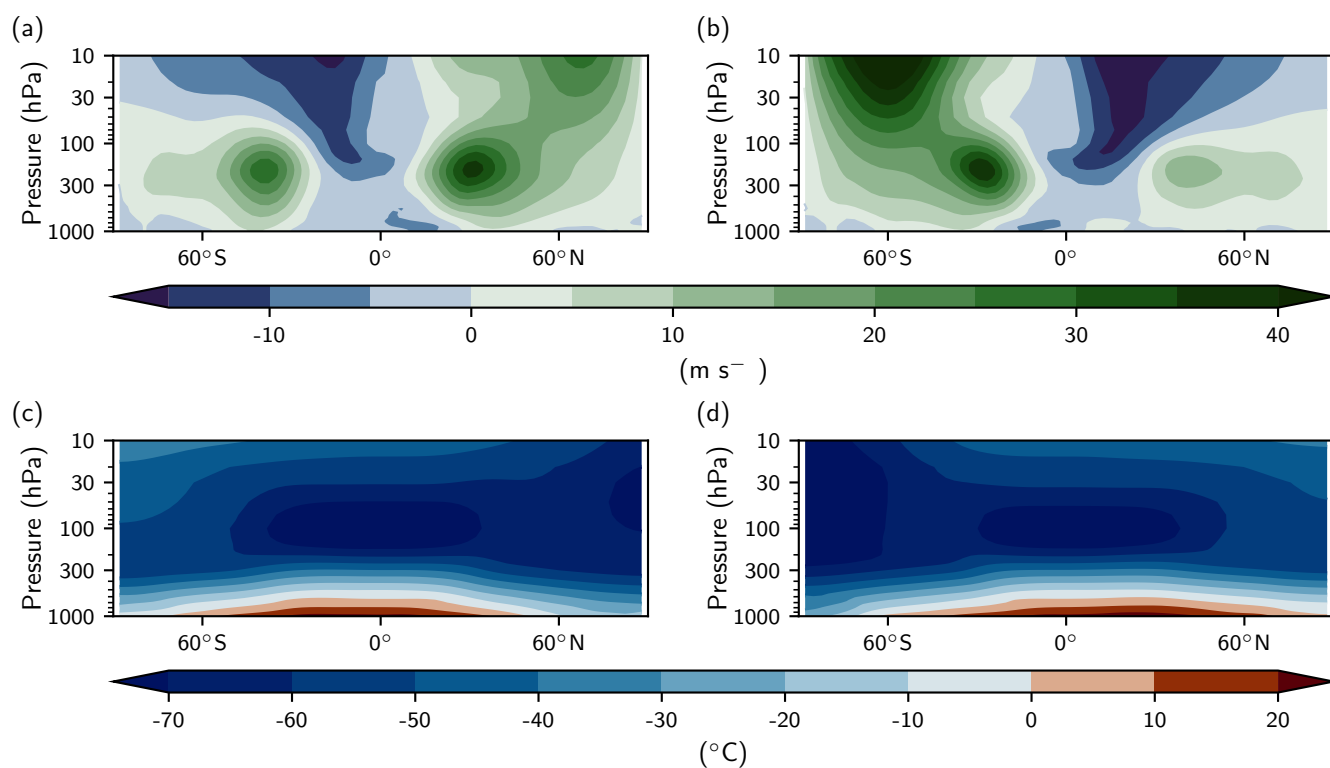
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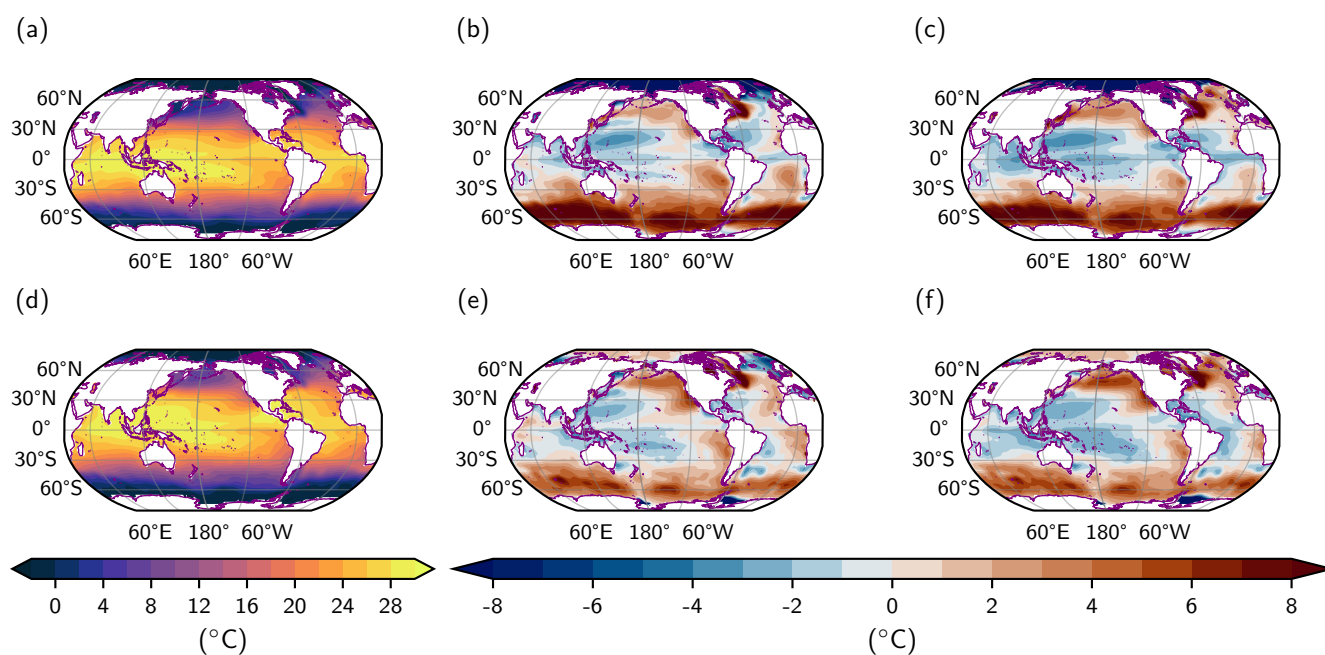
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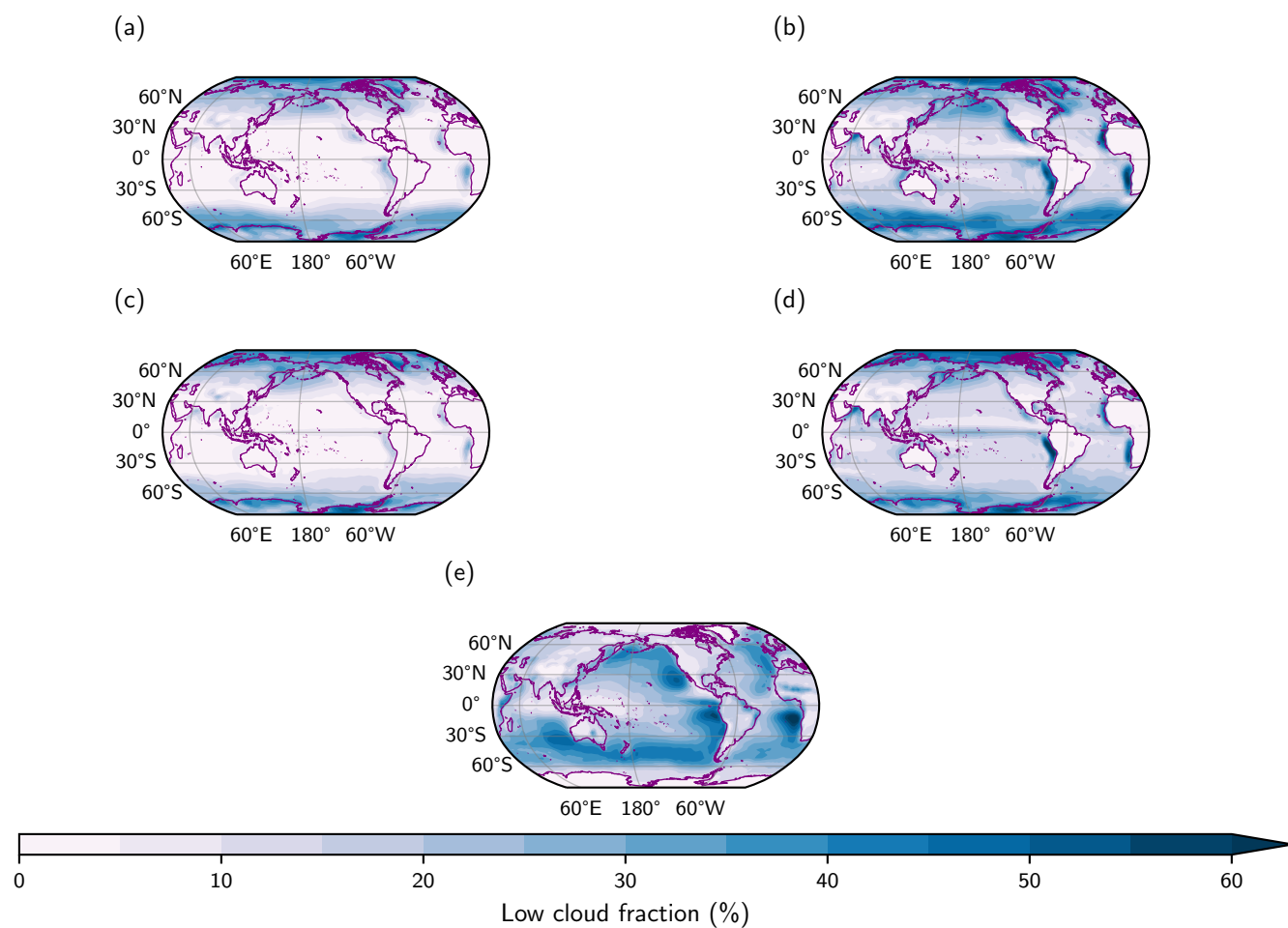
**Figure 1.** Difference (IGCM4.1 minus IGCM4) in the vertical structure of the seasonal mean zonally averaged zonal wind field for (a) DJF and (b) JJA, and zonally averaged temperature field for (c) DJF and (d) JJA.



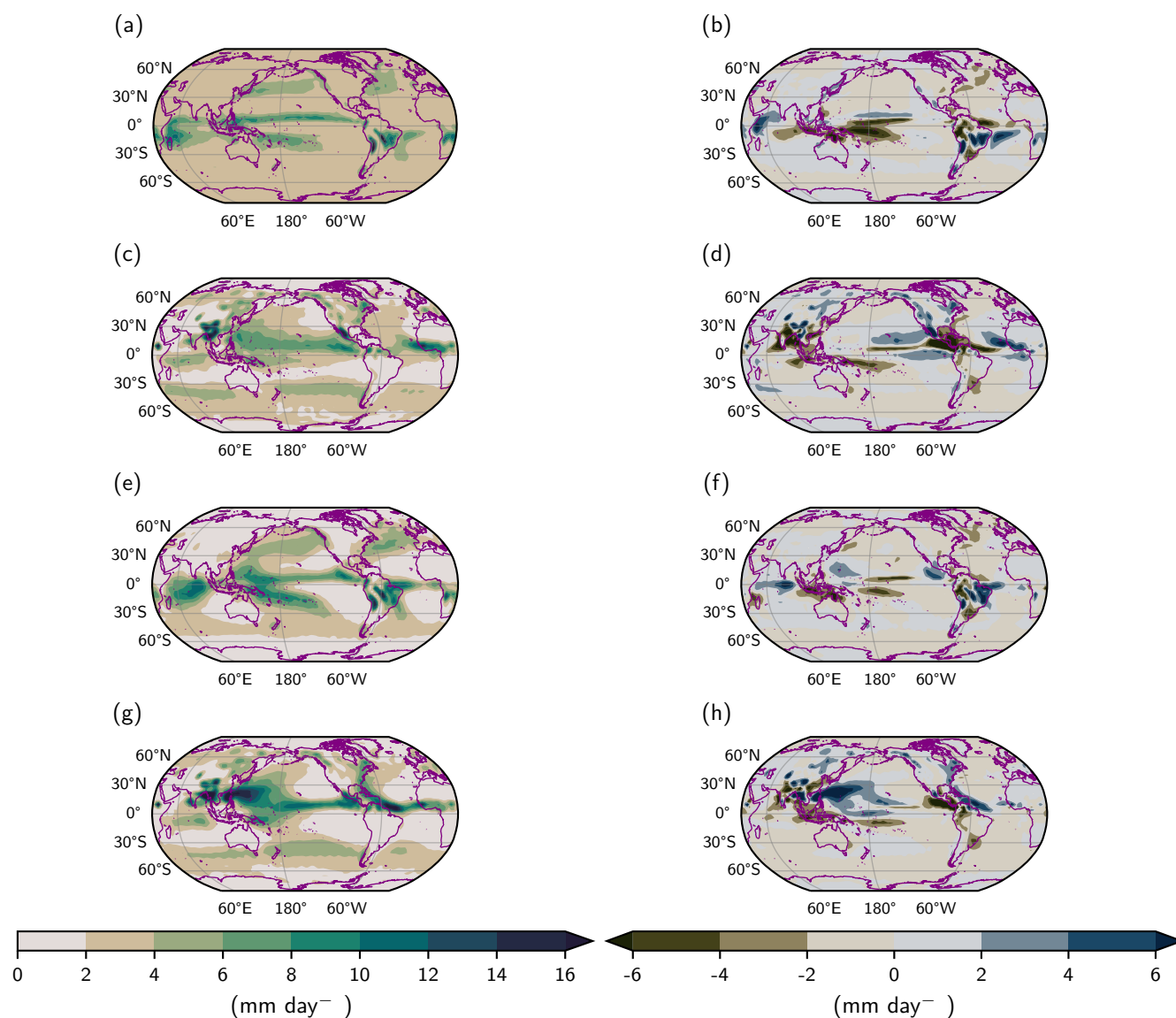
**Figure 2.** Vertical structure of the seasonal mean zonally averaged zonal wind field in FORTE3 for (a) DJF and (b) JJA, and zonally averaged temperature field for (c) DJF and (d) JJA.



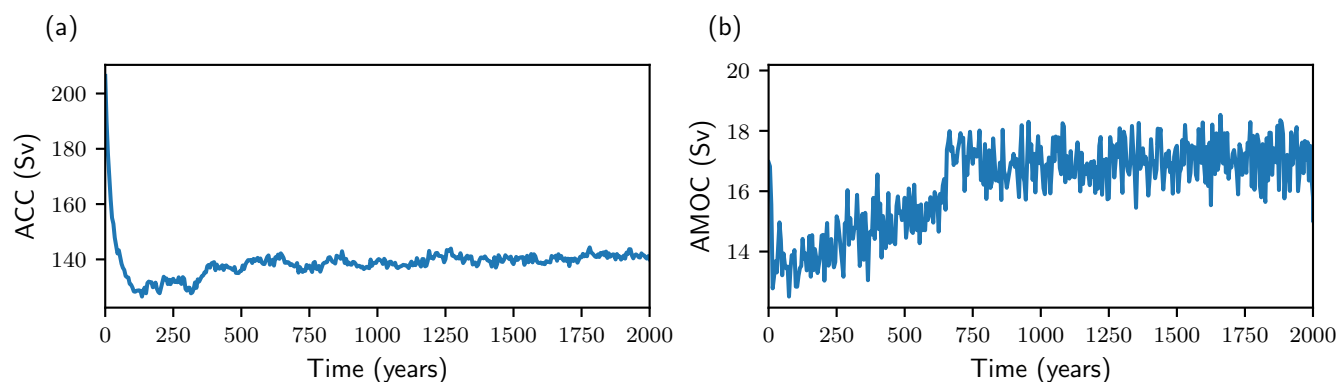
**Figure 3.** Mean sea surface temperatures and their differences from observations (EN4 1900–1920). Panels are: (a) DJF observations; (b) DJF FORTE2 - observations; (c) DJF FORTE3 - observations; (d) JJA observations; (e) JJA FORTE2 - observations; (f) JJA FORTE3 - observations.



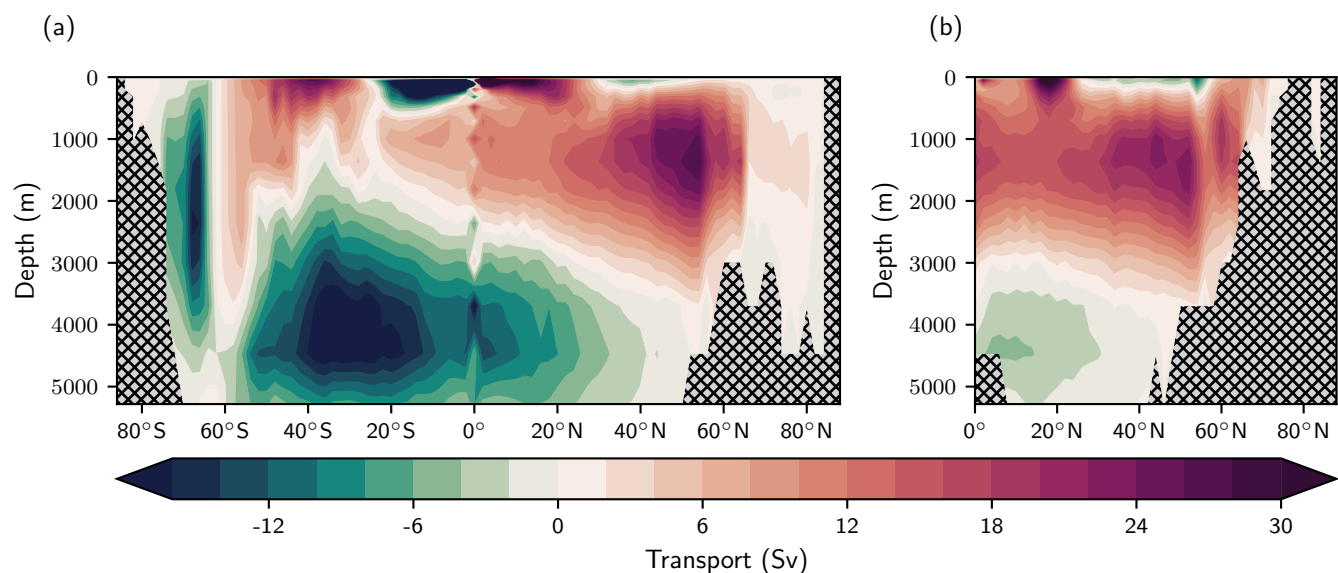
**Figure 4.** Annual mean low level cloud cover in (a) IGC4, (b) IGC4.1, (c) FORTE2, (d) FORTE3, and (e) ISCCP 1983–2017.



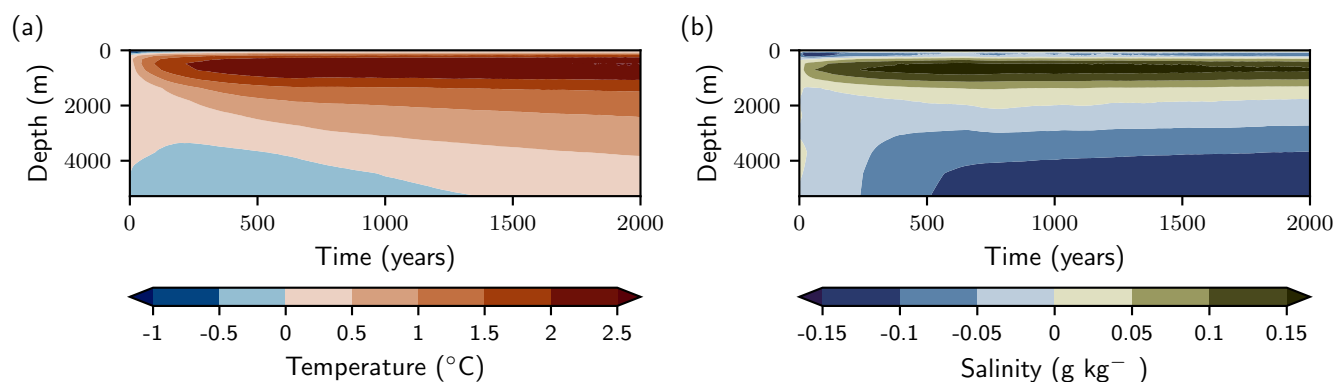
**Figure 5.** Mean precipitation for (a) FORTE3 DJF; (c) FORTE3 JJA; (e) IGCM4.1 DJF; (g) IGCM4.1 JJA. Panels (b, d, f, h) show the differences of the model output on that row from reanalysis (20CR 1871–1896) for the corresponding seasons.



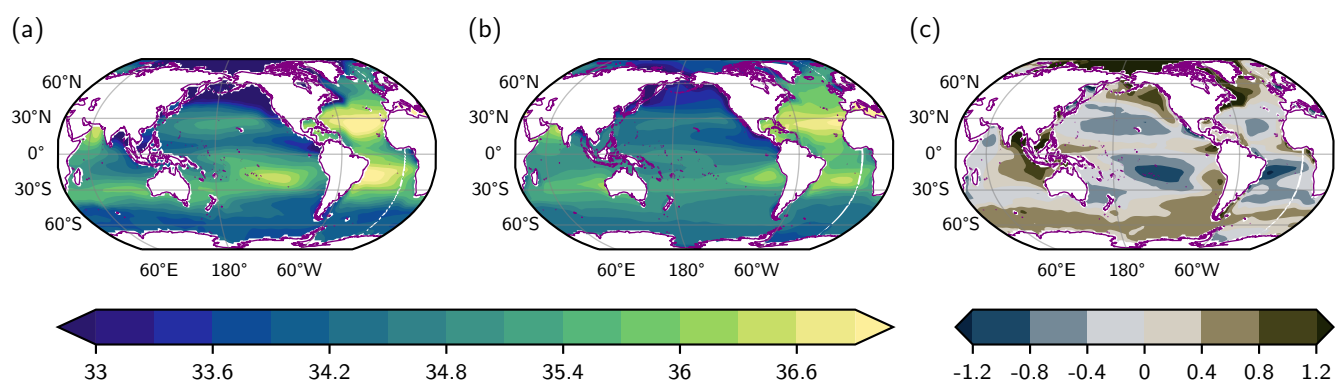
**Figure 6.** Transport of the (a) ACC and (b) AMOC throughout the PiC model run.



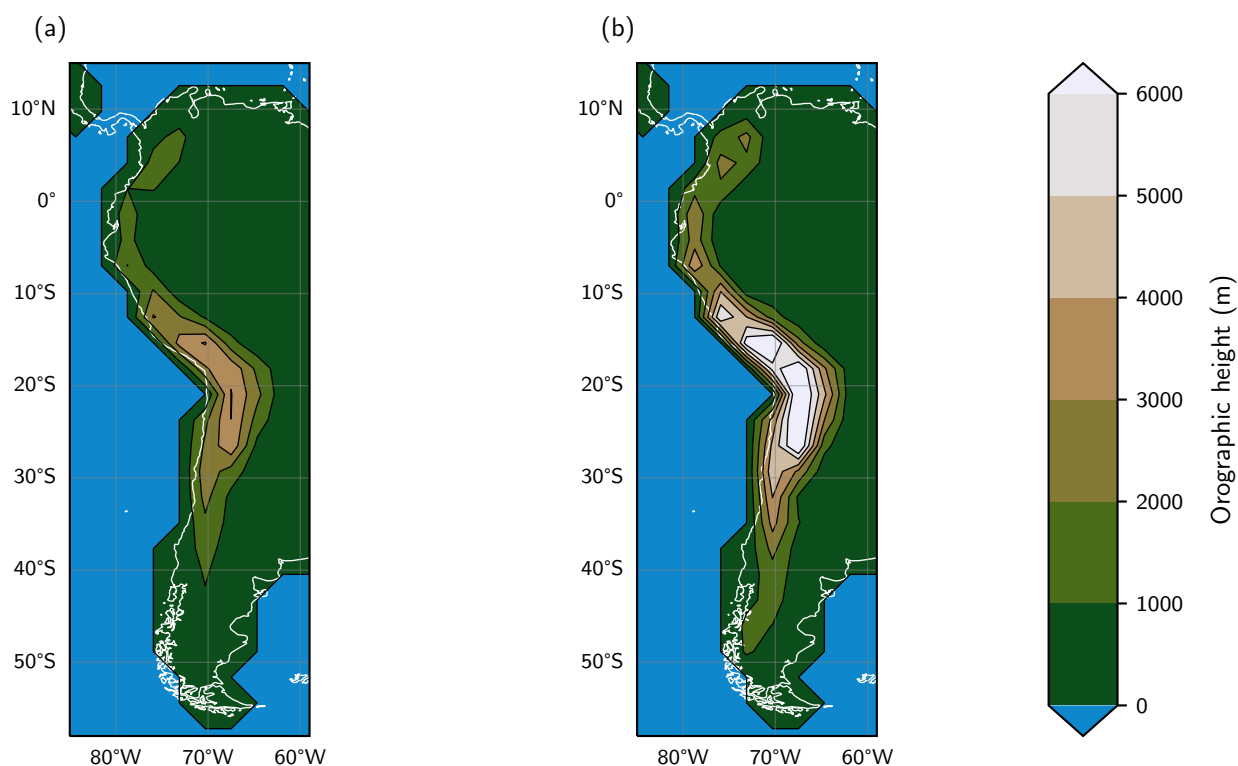
**Figure 7.** Ocean meridional overturning plots FORTE3 mean 2000–2050 (a) global, (b) Atlantic.



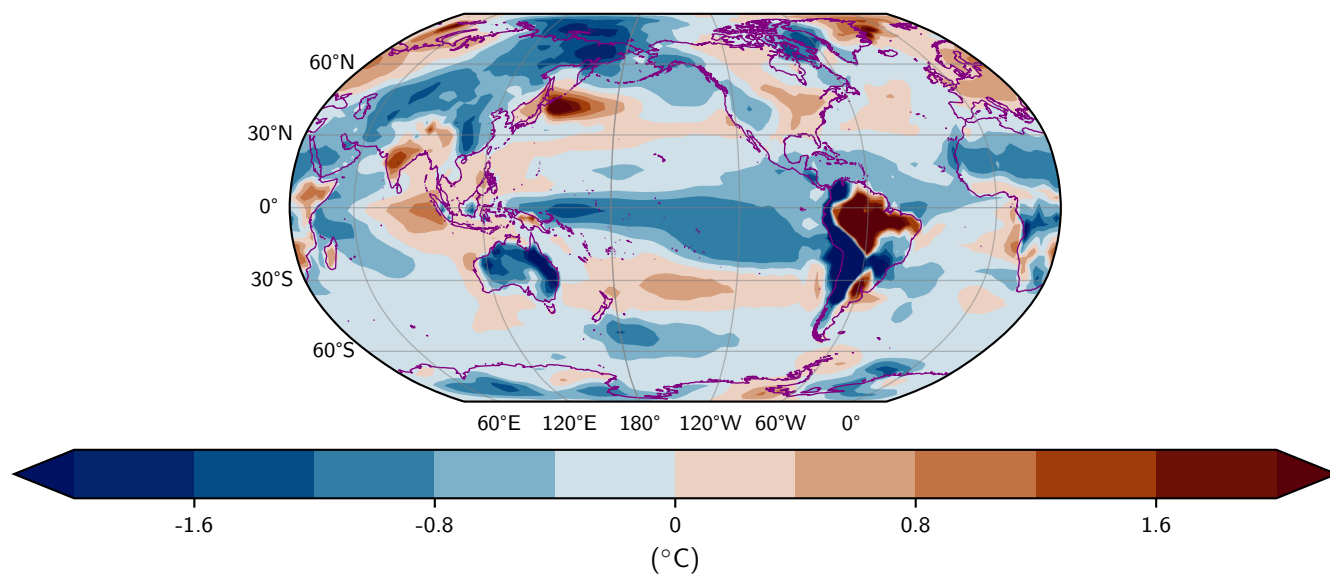
**Figure 8.** Evolution of the mean vertical structure of temperature (a) and salinity (b) fields during the PiC model run.



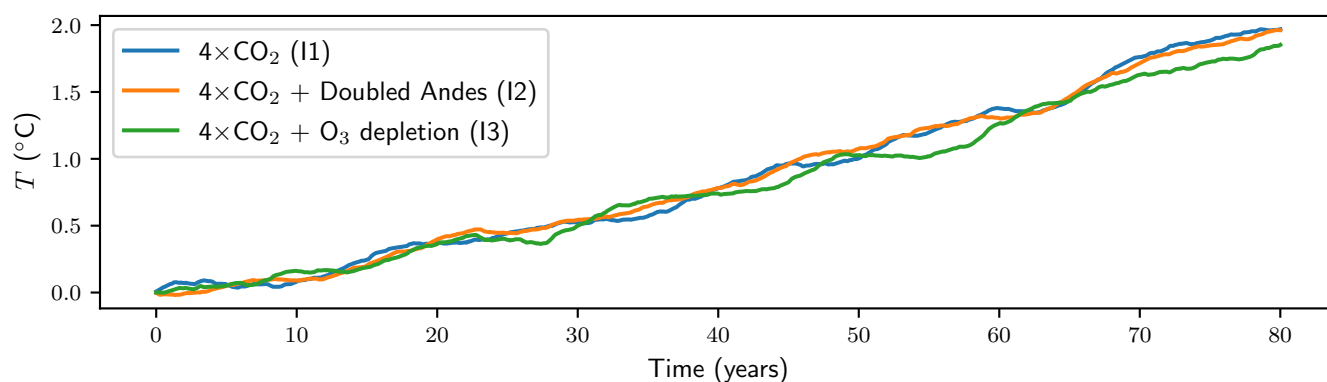
**Figure 9.** Mean sea surface salinity for (a) WOA23 (1971–2000) and (b) FORTE3. The difference FORTE3 minus observations is shown in (c).



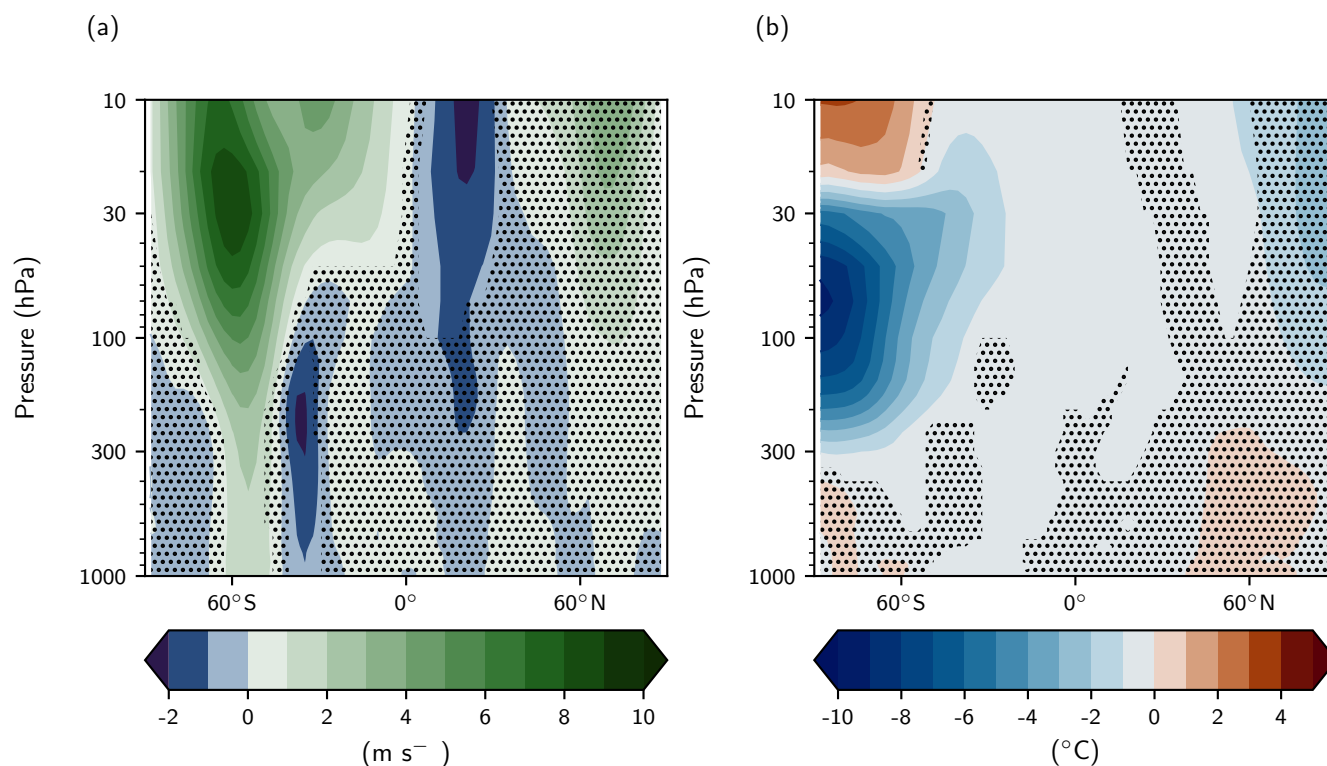
**Figure 10.** Comparison of the Andes topography configurations: (a) default Andes; (b) higher volume Andes (used in run I2).



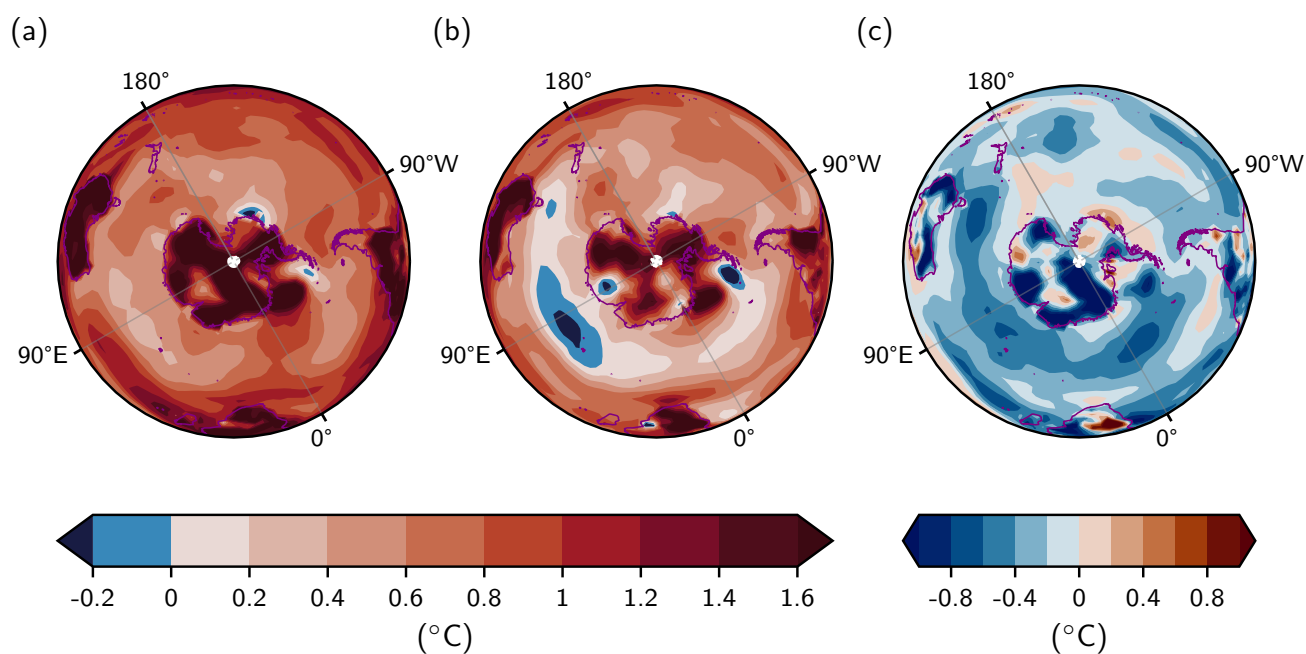
**Figure 11.** Difference in DJF surface temperature, I2 minus I1 (years 500–550).



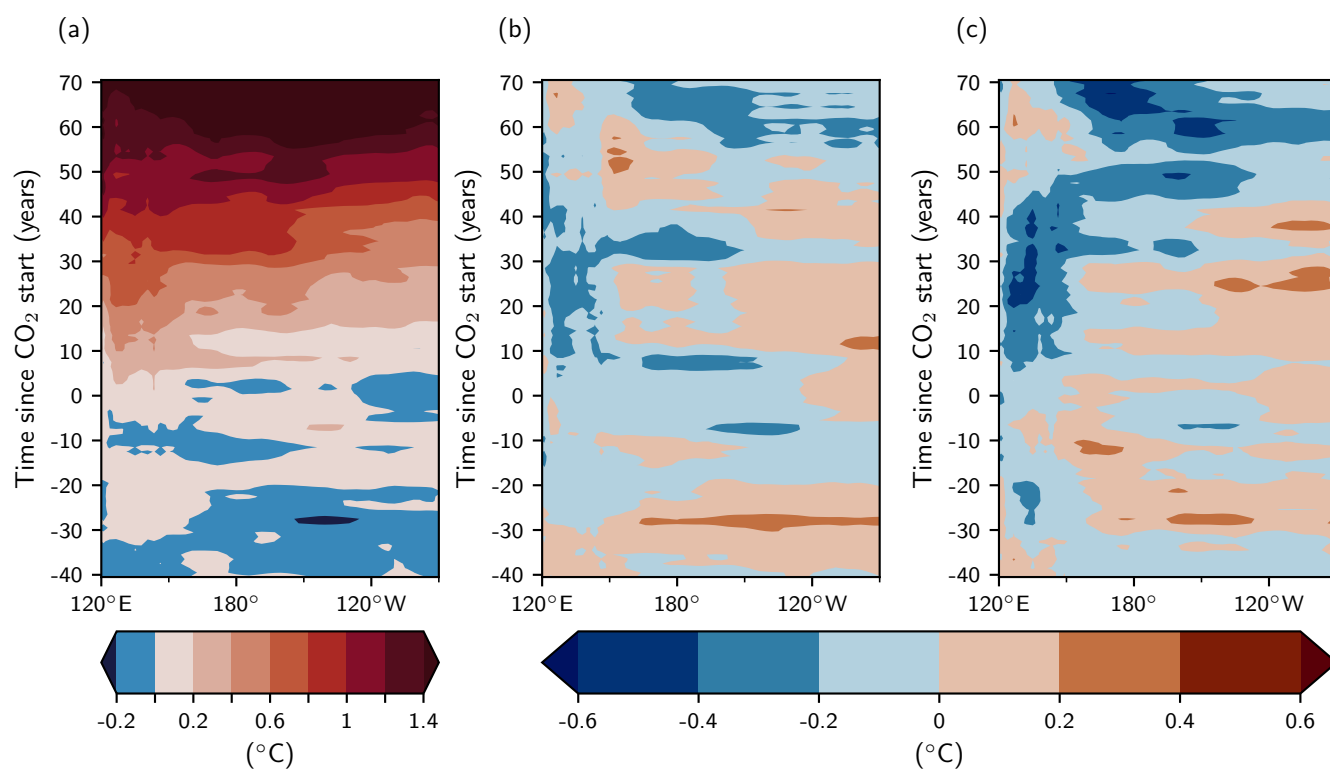
**Figure 12.** Difference in the global previous 5-year rolling mean surface temperature from its value at the initiation of  $\text{CO}_2$  increase.



**Figure 13.** 20 year average from 590–609, December zonally averaged zonal wind and temperature, I3 minus I1, stippling where Welch's t-test gives  $p > 0.05$  (i.e., where insignificant).



**Figure 14.** Mean DJF surface temperatures from years 590–609 minus that of years 500–549 in (a) run I1, (b) run I3, (c) run I3 minus run I1.



**Figure 15.** DJF surface temperature, first averaged 10° S–10° N, before subtracting the average of years 500–550, and finally taking a 20 year rolling average. (a) shows I1, then the difference from I1 is shown for: (b) I2; (c) I3.