

Reviewer 1

We thank the reviewer for their comments. We agree with the reviewer that the evidence to support some of the conclusions regarding the mechanism is not as strong as we would like. However, as the reviewer commented, this is unavoidable given the limitation of a multi-model analysis. We have softened some of our conclusions and included more discussion/analysis of the robustness of our results. Responses to the reviewer's explicit comments are listed below (in red).

1. Most conclusions of this paper is based on correlation analysis of climatologies among models. As the authors already pointed out, these correlation are sensitive to the choices of models included into the analysis. Thus, even a seemly high and statistically significant correlation may not be robust after all. For example, the correlation between $\phi_{v=0}$ and X_CO50 in JJA is 0.65 (Fig. 10b). But if excluding CAM-C1, WACCM-C1 and GEOS-C1, the correlation becomes much weaker and may even changes sign.

We agree with the reviewer that the correlations are somewhat sensitive to the choice of the models. We had acknowledged this in the original version, but have made this clearer in the revised version. However, we think the correlations are actually fairly robust. We have calculated the correlation coefficients for different subsets of models, e.g., excluding SD, excluding models with similar heritage, and there are generally only small changes in correlations and often a higher correlation than using all models (e.g. using just the free-running models results in higher correlations between $\phi_{v=0}$ and CO-50, see Table R1). Excluding CAM-C1, WACCM-C1 and GEOS-C1 does have a big impact but we do not think this is a reasonable thing to do as these simulations are not physical outliers and their exclusion greatly reduces the range of $\phi_{v=0}$ and leaves only 2 independent FR models, so that resulting correlation is determined only by SD models. As we now make clear in the manuscript, the relationships are much less robust if only SD models are considered.

2. The authors argued the mechanisms leading to the correlation between $\phi_{v=0}$ and X_CO50 is that if the divergence zone locates north of the emission, then the northward transport from the source region is limited. This mechanism should work on the interannual time scales as well as the climatologies. Yet, the interannual variation within each model does not show similar correlations to the climatologies among different models. This is evident in Fig. 10e, which shows that the interannual variations of $\phi_{v=0}$ within each model is comparable to the inter-model spread of climatology, but the corresponding Arctic CO concentration does not show any negative correlation. In models such as CMAM-C1SD, the correlation from inter-annual variations even seems to be positive.

We agree with the reviewer that the interannual variation between different metrics (ϕ_{jet} , mean flux, and $\phi_{v=0}$) and CO50 for individual models generally doesn't show the same relationship as between models. This is now discussed in detail in the revised manuscript. There are two main reasons for this disagreement: (i) multiple processes influence the transport of CO50 into the Arctic, and although the mid-latitude mean meridional flux may dominate the variations between the models, other processes can have a larger influence on the interannual variation. (ii) The metrics used to define jet location and Hadley Cell edge can have large uncertainties for monthly or even seasonal data (e.g., jet can be weak with no distinct peak or v can be close to 0 over a broad latitude range, for given year) and differences in the metrics between two years may not be representative of the differences in the jet and HC width. Averaging over multiple years is required to get robust values of these metrics.

An example of another process that influences the transport of CO50 into the Arctic is the meridional transport at high latitudes. The climatological meridional velocity and mean flux poleward of 70N are similar among models as compared to the intermodal variations in the midlatitudes (see Figures 9 and 10) and hence the transport difference due to advection by this flow does not contribute to the spread among models (see Table R2). However, there can be large interannual variations in the high latitude meridional flow/flux within individual models (see Figure R1). This plays an important role in the interannual variations in Arctic CO50 in most of the models with high or moderate interannual correlations between

high-latitude meridional flux and CO50 (see Table R2). This is one example, and there are likely other processes that make a larger contribution to interannual variations than to the spread among models.

As indicated above, another reason for the differences between interannual correlations and correlations among model climatologies is the large uncertainty / non-representativeness of metrics for individual seasons. A clear example of this is the $\phi_{v=0}$ during JJA. In many models, v is close to 0 for a wide range of latitudes in this season, and although there are only small differences in the mean meridional flow (and mean meridional flux) over this region between the years there can be large variations in $\phi_{v=0}$, see Figure R1. In particular, in the SD models the $\phi_{v=0}$ can vary between 30°N and 40°N for individual years, even though for all years there is a very weak meridional flow over this range of latitudes. As a result, the interannual spread in $\phi_{v=0}$ is not representative of interannual differences in meridional transport of CO50 between the years, and the interannual variations in $\phi_{v=0}$ (shown in Figure 10b) are generally larger than those in mean fluxes (Figure 9b).

3. The authors argue the importance of the zonal mean circulation for the tracer transport by comparing the zonal mean and eddy component in two models: GEOS-C1 and GEOS-C1SD. But as shown in Fig. 9a, GEOS-C1 is clearly an outlier in terms of the zonal mean contribution to the tracer flux. If comparing GEOS-C1SD versus WACCM- C1 instead of GEOS-C1, the difference in the zonal mean flux would be much smaller. But the Arctic CO concentration is similar in WACCM-C1 and GEOS-C1, this would implies that the eddy component may be more important to explains the difference between the two models.

We have followed the reviewer's suggestion below (comment 4) and now show in Figure 9 the vertically integrated mean flux calculates over the lowest levels (700-1000 hPa). GEOS-C1 is no longer an outlier, and the zonal-mean transport in GEOS-C1 is not structurally different from other models (e.g., GEOS-C1SD). Fig. 8 is also adjusted accordingly.

4. Since the argument is about the low level divergence zone (800-950 hPa), why the tracer flux is integrated over the whole troposphere (200-1000 hPa) rather than just the low level? The mean circulation pointing to opposite direction in the lower and upper troposphere, and hence there may be canceling effects when integrating over the whole troposphere.

This is a good point, and we have modified the figure as the reviewer suggested. There are no major changes to the conclusions, but the new figure should help ease the reviewer's concern on GEOS-C1 being as an outlier.

5. Other Editorial comments

We have revised the manuscript following reviewer's suggestions.

Reviewer 2

We thank the reviewer for their comments. We agree with the reviewer that the evidence to support some of the conclusions regarding the mechanism is not as strong as we would like, due to the limited number of models, and we have softened some of our conclusions and included more discussion/analysis of the robustness of our results. Responses to the reviewer's explicit comments are below (in red).

1. While the physical mechanisms proposed by the authors are plausible, the supporting evidence is based on correlations of decadal averages within a very limited number of models rather than a clear causal indication of the process. It is clear that the number of models that can be examined conclusions can be reached using these methods. In particular, the number of independent pieces of information is unclear. WACCM and CAM-1 seem nearly identical in the scatter plots. The EMAC-47 level model and EMAC-90 level model are definitely not independent, and considering the frequent grouping of these with their SD runs in the scatter plots, it would be useful to know how many independent data points these four represent. Since, as the authors note, the correlations can be easily influenced easily by which models are included the choice of weighting the single ACCESS run as equivalent to one of the EMAC runs, for example, could change the results significantly.

We agree that it is not possible to make strong conclusions given the limited number of models, and we have softened our language so we are saying the analysis is suggestive rather than proves or shows. We also agree that the models are not all independent, and we have included discussion of the relations between models in the section 4.1. Specifically, we have now noted in the text the similarity between WACCM and CAM, between ACCESS and NIWA, and different EMAC versions. We have also calculated the correlation coefficients only using one of each of these pairs and the values are essentially the same to those using all the models (see Table R1).

2. I would expect that the mechanisms the authors point to would work on shorter timescales than decadal, so it is strange that the small dots of the annual averages do not fall closer to the relationships that are calculated. Perhaps an annual average is too short for the process to be valid, but why should a decadal average be necessary? Some filtered time series showing that this process is valid within a model and not just between models would be helpful, as it would help convince the reader that this physical mechanism is correct. Maybe GEOS has a longer record that could be used? And if the relationship only holds between models (i.e. the explanation of the jet and the tracer concentration only explains the difference between models, not the physical processes within the models), then it just implies that biases in models between transport of tracers and large scale circulation near the tracer sources are correlated.

We agree with the reviewer that the interannual variation between different metrics (ϕ_{jet} , mean flux, $\phi_{v=0}$) and CO50 for individual models generally doesn't show the same relationship as between models. Please see our response to comment #2 from reviewer 1, who raised the same issue.

3. The statistical methodology is problematic. The correlations are done using least squares, which necessarily holds one variable as dependent and one as independent. This is why all of the "fits" where there is no correlation are horizontal lines. Since there is no fundamental reason to expect that either quantity should be an independent variable, this method is not sufficient. Either the fit needs to be done with the x and y reversed (in $y=ax+b$) and that slope should be plotted as well, or a reduced major axis regression, with each quantity scaled by its variance, should be employed. With small r values (e.g., Figure 3), no fit line should be shown at all.

We followed the reviewer's suggestion and used a total least square method (Petráš and Bednářová, 2010) to calculate the linear fit. Note that the Pearson's correlation coefficient is not affected by the method of

fitting. We have also followed the reviewer's suggestion and do not show the linear fit when the correlation is not significant.

4. The analysis of the differences between GEOS and GEOS SD is fine but since GEOS is shown to have quite different behavior from the other models in Figure 9, the conclusions drawn from this comparison are not obviously going to apply to other models.

We have followed comment #4 by reviewer 1 and now show in Figure 9 the vertically integrating the mean flux only over low levels (700-1000 hPa), which is the most relevant for transport away from the source. In the new Figure 9 GEOS-C1 is no longer an outlier, and therefore the proposed mechanism of location shift in mean meridional circulations and associated changes in low-level meridional flow is still applicable to the difference between GEOS-C1 and GEOS-C1SD.

5. My final concern is that analyzing transport in SD runs seems inherently problematic, since it is unclear whether they are actually conserving tracers. I assume that, since the authors are analyzing these, they have reason to believe that it is not a problem. A discussion of the extent to which the SD runs do or do not conserve mass and tracer concentrations would be helpful. I'm not sure if such a study already exists or if the authors could do some analysis of their own with the GEOS-SD model.

Mass conservation is not inherently problematic for the SD runs. These simulations used the same numerical framework as the free-running (FR) counterpart, so have the same transport algorithm and same implementation of emissions and chemistry as the FR. The only difference between a pair of SD and FR simulations is the winds that are used in the transport. While there may be conservation issues with the winds directly from the reanalyses, in all but one of the SD runs the horizontal winds are only nudged towards the reanalyses (and not set directly equal to the reanalyses) and winds are constrained together with the vertical velocity to conserve mass. The one exception is the GEOS-CTM, the only CTM analyzed in this study, which uses horizontal and vertical winds directly from the reanalyses, and adjusts surface pressure to conserve mass. To reassure the reviewer, we calculated the global CO₅₀ burden in each CCMI simulation (see Table R3), and only small differences can be found among models (except GEOS-CTM), with these small differences likely due to interpolation of the emissions onto different resolutions of model grids. In particular, when considering the C1-C1SD pairs within the same base model, the differences are very small (less than 1%).

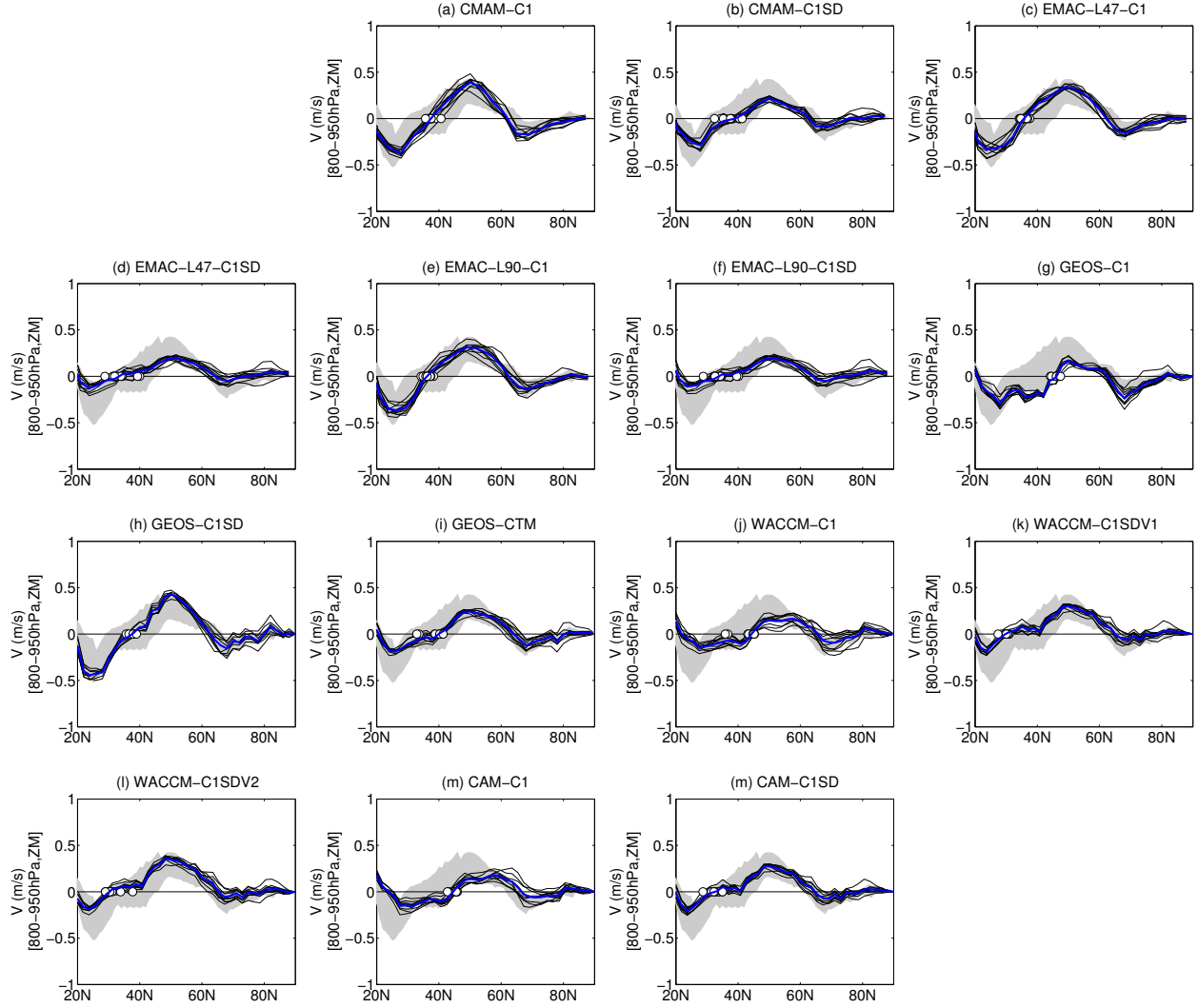


Figure R1 Latitudinal profile of JJA low-level zonal mean meridional wind v (800 – 950 hPa) in each simulation. The black lines and circles denote the interannual variations of v and $\phi_{v=0}$ within the model, while gray shades give the multi-model spread among model climatology, as shown in Figure 10(a). The blue thick line denotes the climatological v of the simulation.

Table R1 Correlation coefficients between Arctic CO50 concentration and physical process related metrics (such as ϕ_{jet} , mean flux, and $\phi_{v=0}$ as listed in the 1st column) calculated using all available models versus only the free-running (C1) models versus clustered models (i.e., exclude NIWA-C1 for similarity to ACCESS-C1, exclude EMAC-L90-C1/C1SD for similarity to EMAC-L47-C1/C1SD, exclude WACCM-C1SDV2 for similarity to WACCM-C1SDV1, and exclude CAM-C1/C1SD for similarity to WACCM-C1/C1SDV1). Calculation of correlation is based on climatology, and those are statistically significant (95%) are marked in **bold**.

	DJF			JJA		
	All	C1	Cluster	All	C1	Cluster
ϕ_{jet}	-0.63	-0.92	-0.64	-0.84	-0.79	-0.83
mean flux	0.69	0.68	0.69	0.78	0.96	0.79
$\phi_{v=0}$	-0.76	-0.95	-0.76	-0.58	-0.70	-0.51

Table R2 Correlation coefficients between mean meridional flux over high latitudes (60°N-80°N) and Arctic CO50 concentrations (500-800 hPa, 70°N-90°N, zonal mean). For mean meridional flux over high latitudes, it is vertically integrated in the low levels (700-1000 hPa) during DJF but in the upper levels (300-500 hPa) during JJA, considering differences in the CO50 vertical maximum between seasons (Fig. 1 and Fig. 2(c,d)). The correlation among model climatologies are shown in the 2nd row (similar to results shown in Fig. 9 (b,d)), while interannual correlations in individual simulations are shown in the 3rd – 16th rows. Coefficients that are statistically significant (95%) are marked in **bold**.

	DJF	JJA
Multi-model Climatology	0.28	0.37
CMAM-C1	0.67	-0.26
CMAM-C1SD	0.35	0.74
EMAC-L47-C1	0.50	0.22
EMAC-L47-C1SD	0.43	0.60
EMAC-L90-C1	0.51	-0.25
EMAc-L90-C1SD	0.43	0.55
GEOS-C1	0.88	0.70
GEOS-C1SD	0.12	0.75
GEOS-CTM	0.46	0.84
WACCM-C1	-0.02	0.52
WACCM-C1SDV1	0.09	0.78
WACCM-C1SDV2	0.53	0.83
CAM-C1	0.50	0.51
CAM-C1SD	0.15	0.75

Table R3 Climatological global CO50 burden (units: $\times 10^{10}$ kg) in each CCM simulation. C1-C1SD pairs in the same base model are highlighted in the same box. Note that molecular weight of CO50 is assumed to be the same as CO (28g/mol).

Model	Global CO50 burden ($\times 10^{10}$ kg)
ACCESS-C1	6.18
CMAM-C1	6.39
CMAM-C1SD	6.40
EMAC-L47-C1	6.35
EMAC-L47-C1SD	6.37
EMAC-L90-C1	6.38
EMAC-L90-C1SD	6.40
GEOS-C1	6.42
GEOS-C1SD	6.46
GEOS-CTM	5.95
WACCM-C1	6.22
WACCM-C1SDV1	6.21
WACCM-C1SDV2	6.20
CAM-C1	6.24
CAM-C1SD	6.23
NIWA-C1	6.18

Large-scale transport into the Arctic: the roles of the midlatitude jet and the Hadley Cell

Huang Yang¹, Darryn W. Waugh^{1,2}, Clara Orbe³, Guang Zeng⁴, Olaf Morgenstern⁴, Douglas E. Kinnison⁵, Jean-Francois Lamarque⁵, Simone Tilmes⁵, David A. Plummer⁶, Patrick Jöckel⁷, Susan E. Strahan^{8,9}, Kane A. Stone^{10,11,a}, and Robyn Schofield^{10,11}

¹Department of Earth and Planetary Sciences, Johns Hopkins University, Baltimore, Maryland, USA

²School of Mathematics, University of New South Wales, Sydney, Australia

³NASA Goddard Institute for Space Studies, New York, New York, USA

⁴National Institute of Water and Atmospheric Research, Wellington, New Zealand

⁵National Center for Atmospheric Research (NCAR), Atmospheric Chemistry Observations and Modeling (ACOM) Laboratory, Boulder, Colorado, USA

⁶Climate Research Branch, Environment and Climate Change Canada, Montreal, QC, Canada

⁷Deutsches Zentrum für Luft- und Raumfahrt (DLR), Institut für Physik der Atmosphäre, Oberpfaffenhofen, Germany

⁸Atmospheric Chemistry and Dynamics Laboratory, NASA Goddard Space Flight Center, Greenbelt, Maryland, USA

⁹Universities Space Research Association, Columbia, Maryland, USA

¹⁰School of Earth Sciences, University of Melbourne, Melbourne, Victoria 3010, Australia

¹¹ARC Centre of Excellence for Climate System Science, University of New South Wales, Sydney, New South Wales 2052, Australia

^anow at: Department of Earth, Atmospheric and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, Massachusetts 02139-4307, USA

Correspondence: Huang Yang (hyang61@jhu.edu)

Abstract. Transport from the Northern Hemisphere (NH) midlatitudes to the Arctic plays a crucial role in determining the abundance of trace gases and aerosols that are important to Arctic climate via impacts on radiation and chemistry. Here we examine this transport using an idealized tracer with a fixed lifetime and predominantly midlatitude land-based sources in models participating in the Chemistry Climate Model Initiative (CCMI). We show that there is a 20% - 40% difference in the Arctic concentrations of this tracer among the models. **This spread is correlated with the spread in the location of the Pacific jet, as well as the spread in the location of the Hadley Cell (HC) edge which varies consistently with jet latitude. Our results suggest that it is likely that the HC-related zonal-mean meridional transport rather than the jet-related eddy mixing is the major contributor to the inter-model spread in the transport of land-based tracer into the Arctic.** Specifically, in models with a more northern jet, the HC generally extends further north and the tracer source region is mostly covered by surface southward flow associated with the lower branch of the HC, resulting in less efficient transport poleward to the Arctic. During boreal summer, there are poleward biases in jet location in free-running models, and these models likely underestimate the rate of transport into the Arctic. Models using specified dynamics do not have biases in the jet location, but do have biases in the surface meridional flow, which **may result** in differences in transport into the Arctic. In addition to the land-based tracer, the midlatitude-to-Arctic transport is further examined by another idealized tracer with zonally uniform sources. With equal sources **from both land and ocean**, the intermodel spread of this zonally uniform tracer is more related

to variations of parameterized convection over oceans **rather** than variations of HC extent particularly during boreal **winter**. This suggests that transport of land-based and oceanic tracers or aerosols towards the Arctic differ in pathways and therefore their corresponding intermodel variabilities result from different physical processes.

1 Introduction

5 The Arctic is characterized by the largest climate sensitivity with surface temperatures increasing much more rapidly than the global average in recent decades (IPCC, 2013). Trace gases and aerosols have been shown to be important for Arctic climate via their direct radiative influences and indirect effects on cloud properties (Garrett and Zhao, 2006; Lubin and Vogelmann, 2006; Coopman et al., 2018). Since the majority of these trace gases and aerosols originate over **the** Northern-Hemisphere (NH) midlatitudes, where anthropogenic emissions are **the** largest (Bottenheim et al., 2004; Fisher et al., 2010; Kupiszewski et al., 10 2013), long-range transport from NH midlatitude source regions plays a crucial role in determining their Arctic distributions. Transport therefore has a remote impact on the Arctic climate as important as local forcings (Shindell, 2007). Shindell et al. (2008) further showed that the multi-model spread of simulated Arctic carbon monoxide (CO) and ozone (O₃) concentrations is as large as the corresponding multi-model mean and this large multi-model spread may be related to large differences in long-range transport. It is therefore important that models correctly represent this transport.

15 Orbe et al. (2018) recently analyzed the transport in models participating in the Chemistry Climate Model Initiative (CCMI), and showed a large spread among the models. A 30%-40% difference of the multi-model mean is found in the Arctic concentrations of idealized tracers originating in the NH midlatitudes. Orbe et al. (2018) attributed much of these differences in transport into the Arctic to differences in midlatitude convective transport (**primarily over the oceans**) among the models, particularly during boreal winter. Specifically, they showed that, for tracers with zonally uniform sources over the NH midlatitudes, stronger 20 convection over the oceans tends to enhance tracer concentrations in the upper troposphere and dilute tracer concentrations at the surface. While this enhances transport into the upper troposphere, it also weakens along-isentropic transport from **the mid-latitude surface** into the Arctic middle troposphere, manifesting as negative correlations between midlatitude convection and Arctic tracer concentrations.

A limitation of the study of Orbe et al. (2018) is that the authors focused on idealized tracers with zonally uniform sources, 25 and hence it is unclear whether their conclusions apply to more realistic tracers. Most chemical tracers of interest have strong zonal asymmetries in their source regions, with emissions primarily over midlatitude continents (e.g., tracers with anthropogenic emissions). These tracers may be less sensitive to differences in the simulated convection (which occurs predominantly over the oceans), and there may be less spread among the models in the transport of these tracers into the Arctic.

Here we revisit the issue of transport into the Arctic within the CCMI models, considering an idealized “CO50” tracer. 30 This tracer has realistic, zonally-varying, emissions corresponding to anthropogenic carbon monoxide (CO) emissions but with **an** idealized, fixed decay time of 50 days. We examine the transport of CO50, and also the “NH50” tracer (the same 50-day lifetime but with a zonally uniform boundary condition) considered by Orbe et al. (2018), into the Arctic within the CCMI models, and show **that** there is a spread in Arctic concentrations of both tracers among the models. This spread of CO50 is,

however, not closely linked to differences in convective mass fluxes but **is more likely due to** differences in the midlatitude jet over the Pacific Ocean and **in** the mean meridional circulation.

Section 2 introduces the models, tracers, and dynamical metrics examined. Section 3 shows the multi-model mean tracer distribution followed by highlights on the multi-model spread of tracer concentrations over the Arctic. Section 4 focuses on examining the influences of **the midlatitude jet** on the poleward transport of tracers, in which we further explore the mechanisms. To further examine the role of tracer boundary condition and lifetime, Section 5 compares transport of CO50 (zonally asymmetric emissions) to the Arctic with another idealized tracer NH50 with zonally uniform sources as well as a realistic tracer carbon monoxide (CO) with a temporally and spatially varying chemical lifetime. Conclusions and discussions are given in Section 6.

2 Methods

2.1 Models and experiments

This study analyzes simulation results from models participating in the Chemistry Climate Model Initiative (CCMI) phase 1 (Morgenstern et al., 2017). CCMI is a joint activity of the international Global Atmospheric Chemistry (IGAC) and Stratosphere-troposphere Processes And their Role in Climate (SPARC) projects that aims to better quantify stratospheric and tropospheric ozone and other important chemical species using state-of-the-art chemistry-climate models (Eyring et al., 2013). Here, we examine distributions of the idealized CO50 and NH50 tracers (see Section 2.2) from 15 CCMs **and 1 Chemistry-Transport Model (CTM)** (Table 1). These models mostly overlap with those considered by Orbe et al. (2018), and we use the same model names. Several simulations analyzed by Orbe et al. (2018) are not used here because the CO50 tracer is **either not included or incorrectly implemented** in these simulations. As in Orbe et al. (2018), we focus on two types of hindcast reference simulations, namely the C1 simulation (i.e., referred to as REF-C1 in CCMI) and the C1SD simulation (i.e., REF-C1SD in CCMI). Both C1 and C1SD simulations were forced by observed sea surface temperatures (SSTs) and sea ice concentrations (SICs) from the UK Met Office HadISST1 data set (Rayner, 2003), but they differ in the source of meteorological fields. C1 simulations calculate the meteorological fields within the model, whereas C1SD simulations use (or relax towards) meteorological fields from meteorological reanalyses. The models used in this study differ widely in many respects, including the model resolution, **dynamical** core, physical parameterizations, and **chemical schemes** (Table 1 and Morgenstern et al. (2017)).

We analyze monthly output from 2000 to 2009 for all CCMI simulations (except for GEOS-C1 and GEOS-C1SD, see Table 1), and calculate 10-year climatologies for northern winter and summer by averaging the months of December-January-February (DJF) and June-July-August (JJA), respectively. In addition, interpolation is applied from each simulation's native vertical levels (isobaric, hybrid pressure or hybrid altitude) to standard isobaric vertical coordinates consisting of 19 tropospheric levels (from 1000 hPa to 100 hPa with a uniform spacing of 50 hPa) and 4 stratospheric levels (at 80, 50, 30, and 10 hPa). Analysis of individual models is done using each model's native horizontal grid, but when forming multi-model mean fields, the model output is interpolated onto a standard $1^\circ \times 1^\circ$ grid at every isobaric level after the interpolation to common levels noted above.

2.2 Tracers

To quantify the large-scale transport from the NH midlatitude land sources to the middle-troposphere Arctic, we examine the idealized CO50 tracer. The CO50 tracer has a flux boundary condition, corresponding to the annual mean value of anthropogenic emissions of CO for 2000, from the Hemispheric Transport of Air Pollution (HTAP) REanalysis of the TROpospheric chemical composition (RETRO) (Eyring et al., 2013), and a spatially uniform loss with a 50 day e-folding decay time. **One exception are the EMAC models that use the annual cycle rather than the annual mean value of anthropogenic CO emissions for 2000. This results in higher CO50 concentrations during winter and lower concentrations during summer compared to the other models, which influences some results in this study but is unlikely to change any major conclusion (as we show in the remainder of the paper).** There are strong zonal asymmetries in the CO50 emissions (**white-dotted regions** in Fig. 1), with the largest contributions from East and South Asia. Note that a similar 50 day CO-like tracer has also been used in many previous studies for diagnosing long-range transport, but in these previous studies the emissions included those from biomass burning (Shindell et al., 2008; Fang et al., 2011; Doherty et al., 2017).

We also compare the simulated CO50 to the NH50 idealized tracer. The NH50 also has a spatially uniform 50-day loss, but with a different boundary condition. The concentration of NH50 (χ_{50}) in the bottom model level is specified as a fixed mixing ratio (i.e., 10 ppmv) over the NH midlatitude region (30°N - 50°N, 180°E-180°W). Wu et al. (2018) have shown the spatial distribution of NH50, particularly its inter-hemispheric gradient, is strongly associated with the seasonal shift of the Intertropical Convergence Zone (ITCZ) and also likely the Southern Pacific Convergence Zone (SPCZ) over the oceans. As noted in Section 1, Orbe et al. (2018) further documented a wide spread of NH50 concentrations amongst CCMI models both over the Arctic and in the SH, and attributed this spread to the intermodel variation of low-level parameterized convection primarily over the oceans.

Last, to examine how **well** CO50 can represent real tracers with land sources, we compare it with carbon monoxide (CO) that undergoes the full chemistry (spatially and temporally varying) in the models. CO is removed from the troposphere primarily by reacting with the hydroxyl radical (OH) that yields a global mean annual mean lifetime of ~ 2 months. However, as OH concentrations are much higher as well as for the temperature during summer, CO lifetime is much shorter in summer than that in winter. The emissions of CO generally resembles that of CO50, but it has additional sources from biomass burning, which features large emissions from forests in West Africa, South America all year round, as well as Siberia during summer. The latter is particularly important for the Arctic abundance of CO, and complicates comparisons with CO50.

2.3 Dynamical fields

As previous studies have indicated the importance of **the midlatitude jet streams and associated storm tracks** for tracer transport into the Arctic (e.g., Eckhardt et al., 2003), we also examine the relationship of the distributions of the above tracers with dynamical (meteorological) fields. In particular, we decompose the tracer transport into a zonally asymmetric component and a zonally symmetric component. The zonally asymmetric transport is associated with eddy mixing and we examine the

relationship with the NH midlatitude jet. The zonally symmetric transport is associated with the zonal-mean flow advection and we examine the relationship with the Hadley Cell (HC) circulation.

For the midlatitude jet, we focus on the zonal wind u over the Pacific Ocean, as this plays an important role in the transport of CO₅₀ away from the major source region over East Asia, and examine the variation of the latitude of the Pacific jet (ϕ_{jet}).

5 This latitude is where the zonally (135°E-125°W) and vertically (500-800 hPa) averaged u maximizes **in the midlatitudes (25°N-65°N)**. To account for differences in model resolution, ϕ_{jet} is calculated as the location of the maximum of a quadratic function fitted to the zonally and vertically averaged u at its maximum grid point and the two points either side (Barnes and Polvani, 2013). ϕ_{jet} is calculated at every **season** of the integration and the wintertime and summertime climatologies of jet position are then derived for inter-model comparison.

10 For the HC we examine the 800 hPa - 950 hPa averaged zonal-mean meridional wind \bar{v} , and calculate the latitude ($\phi_{v=0}$) at which $\bar{v} = 0$ between 20°N and 50°N. This latitude corresponds to the surface divergence zone separating the NH Hadley and Ferrel Cells. Again, $\phi_{v=0}$ is calculated **seasonally**, and winter and summer climatologies are compared between the models.

3 Distributions of CO₅₀

We first examine the multi-model mean (i.e., C1 and C1SD simulations combined) distributions of CO₅₀, and then examine
15 the spread among the models with a focus on distributions in the Arctic. The CCMI multi-model mean horizontal and vertical distributions of CO₅₀ are shown in Fig. 1. In both the lower troposphere (850 hPa) and the middle troposphere (500 hPa), and for both seasons, there are higher concentrations of CO₅₀ ($\chi_{\text{CO}_{50}}$) over the midlatitudes than over the Arctic, with large zonal asymmetries over the midlatitudes but not in the Arctic. The maxima of $\chi_{\text{CO}_{50}}$ over the midlatitudes highlight the primary source regions of CO₅₀ in East Asia and South Asia, with $\chi_{\text{CO}_{50}}$ decreasing rapidly away from the source regions.

20 The meridional and vertical distribution of zonal-mean CO₅₀ varies with season. During boreal winter, CO₅₀ features a much stronger meridional transport near the surface in both the poleward and equatorward directions. The distribution of CO₅₀ also generally follows the slope of isentropic surfaces, **exhibiting stronger vertical transport north of the midlatitude CO₅₀ source region and suppressed vertical transport in the south**. During summer, $\chi_{\text{CO}_{50}}$ has a weak vertical tracer gradient and a secondary maximum at 200 - 300 hPa in the subtropics, indicating reduced meridional transport compared to winter. This
25 secondary maximum in CO₅₀ mixing ratio requires robust vertical transport with relatively slow chemical loss, which is likely due to **the** close proximity of the emissions to the strong continental convection underlying the summertime Asian monsoon anticyclone over the Tibetan Plateau (Park et al., 2007; Garny and Randel, 2013) (see Fig. S1). A similar maximum within the Asian monsoon anticyclone region near the tropopause was observed by balloon sondes for particle surface area density of aerosols (Yu et al., 2017), as well as for CO in the upper troposphere over East Asia by flight measurements (Holloway et al.,
30 2000; Palmer et al., 2003).

The spatial distribution of zonal-mean CO₅₀ for each model is similar to that for the multi-model mean distribution discussed above. This is illustrated in Fig. 2, which shows the latitudinal variation of lower troposphere (500 - 800 hPa) and vertical profiles of Arctic (70°N - 90°N) CO₅₀, respectively. Although the latitudinal and vertical structure of CO₅₀ concentrations

are similar among the models, there is a large spread in the magnitude of CO50 tracer concentrations. The multi-model spread is the largest over the midlatitude source region, decreasing rapidly in the tropics south of the source region while remaining relatively large north of the source towards the Arctic for both seasons.

We will focus here on model differences in CO50 concentrations over the Arctic, and the poleward transport from NH midlatitudes. The differences in Arctic CO50 concentrations among the models peak around 400 hPa during both winter and summer. In the middle and lower troposphere, the range of χ_{CO50} among the models decreases from ~ 11 ppbv in winter to ~ 5 ppbv in summer; **this** yields a 20% - 40% wintertime fractional spread (i.e., **the** multi-model spread relative to the corresponding multi-model mean) and a 25% - 30% summertime spread of Arctic χ_{CO50} . **Note that χ_{CO50} in the EMAC models are biased due to the use of seasonally varying CO50 emissions, which manifest as higher χ_{CO50} during winter and lower concentrations during summer. However, it is difficult to quantify how much the biases are remaining in the Arctic χ_{CO50} given the complexity of long-range transport. Nevertheless, if we assume no seasonality of CO50 emissions in the EMAC models (as the other CCMi models), we expect a lower Arctic χ_{CO50} during winter and a higher Arctic χ_{CO50} during summer; this would yield a smaller multi-model spread of Arctic χ_{CO50} during winter and likely a larger range during summer.**

The difference between pairs of simulations (and hence the ordering of simulations) is generally the same at all altitudes. For example, χ_{CO50} in winter is smaller in ACCESS-C1 and NIWA-C1 than that in the EMAC simulations throughout most of the tropospheric column. This suggests that the above model spread of Arctic CO50 is related to a vertically consistent difference in the poleward transport rather than a tracer redistribution between different levels.

The large spread in CO50 concentration among the models is consistent with the wide spread reported by Orbe et al. (2017, 2018) for idealized tracers with zonally-uniform sources. Also, Figs. 2 and S2 show that the spread in CO50 among the C1SD simulations (dashed lines) is comparable to or even larger than the spread among the C1 simulations (solid lines). This is again consistent with the results of Orbe et al. (2017, 2018), **and** provides further evidence that using specified dynamics simulations does not constrain climatological tropospheric transport any more than using free-running models.

4 Transport processes of CO50

Having shown a large model spread in the Arctic concentrations of CO50, we now examine possible causes for these differences. Shindell et al. (2008) suggested that the Arctic CO concentration in the middle troposphere is equally sensitive to changes of emissions over Europe, Asia, and North America. However, given the total amount of emissions from Asia (East Asia and South Asia, see Table 2 in Shindell et al. (2008)) is ~ 2 -3 times larger than those from Europe and North America, we first examine processes that are associated with the transport of Asian pollutants.

4.1 Relationship with midlatitude convection

Orbe et al. (2018) showed that differences in convection among **the** models contribute to differences in tracer distributions. They showed that models with stronger midlatitude convection tend to have lower Arctic concentrations of the idealized

NH5 tracer (this tracer has the same zonally symmetric boundary conditions as NH50 but **with** a shorter lifetime of 5 days), especially during northern winter. However, examination of CO50 shows a very weak relationship between the strength of the midlatitude convection and the Arctic χ_{CO50} in both winter and summer (Fig. 3). The strength of convection is measured using the convective mass flux (CMF) in the low-level midlatitudes, which is the average of 800 hPa - 950 hPa, 30°N-50°N, 130°E-170°E for boreal winter focusing on convection over West Pacific Ocean and 110°E-140°E for summer highlighting continental and maritime convection over East Asia, as in Orbe et al. (2018). The average zones overlap the strongest intensity and the largest intermodel variability of convection. Note that the χ_{CO50} - CMF relationship during winter is sensitive to which models are included in the correlation since simply excluding the ACCESS and NIWA models would produce a positive correlation whereas excluding the EMAC results would produce a negative correlation. The latter is more plausible given the fact that the Arctic χ_{CO50} are biased higher in the EMAC models during winter and a better negative χ_{CO50} - CMF correlation can be achieved if the Arctic χ_{CO50} in the EMAC models are lower. Despite the complications associated with using the EMAC model results, this large sensitivity to including or excluding a few models indicates that special care must be taken when interpreting correlations using only a limited number of CCMI models. Furthermore, there are a few models that have similar heritage (e.g., ACCESS and NIWA, WACCM and CAM) but such a similarity does not affect the robustness of the results presented in this study. In particular, we find that correlation coefficients calculated using only one of these similar pairs are essentially the same as those using all the models; furthermore, there is no change in statistical significance with reduced degree of freedom (see Table S1). Hence, variations in CMF do not seem to be the primary cause of variations in transporting CO50 into the Arctic.

The absence of a strong correlation between Arctic CO50 and midlatitude convection may be largely due to the zonally asymmetric boundary condition of CO50, particularly in winter. With primary sources over land, CO50 tends to be less impacted by the variability of convection that maximizes over the oceans during winter. In summer, despite midlatitude convection being the strongest and also having the largest model spread over the land-based emission regions, the poleward transport of CO50 along isentropic surfaces is much weaker than that in winter (comparing Fig. 1(f) with Fig. 1(e)). Therefore, the Arctic CO50 concentration during summer is less connected to CO50 concentration over the midlatitude surface source regions and consequently shows a weaker correlation with the midlatitude convection.

4.2 Relationship with the midlatitude jet

Figure 1(c,d) suggests that, in addition to convection, the zonal winds u , especially over the northern Pacific Ocean, also play an important role in the transport of CO50 from its source regions. We therefore start by examining the structure of the midlatitude jet over the Pacific Ocean in the models. Fig. 4 shows the latitudinal variation of lower-mid tropospheric (500 - 800 hPa) zonal wind u averaged over the Pacific Ocean (135°E-125°W) for each model. In winter there is a similar latitudinal variation in u among the models. There is a variation in the magnitude of the peak winds but the latitude of this peak ϕ_{jet} varies by only a few degrees ($\sim 35^\circ\text{N}$ - 40°N). The C1SD simulations, which use reanalyses winds, have very similar jet latitudes, with $\phi_{\text{jet}} \sim 36^\circ\text{N}$.

However, in summer, there is a much larger variation among the models, not only in the magnitude and location of peak winds but also the latitudinal structure. ϕ_{jet} varies from $\sim 45^\circ\text{N}$ to 57°N , with the C1SD models at the lower end ($\phi_{\text{jet}} \sim 45^\circ\text{N}$). This implies that the latitude of the Pacific jet in C1 simulations is generally biased poleward of the reanalyses. A similar bias **was** found for models participating in the phase 5 of the Coupled Model Intercomparison Project (CMIP5) (Barnes and Simpson, 2017). The variation in the summertime jet structure among the models is further illustrated in Fig. 5, which shows the 500 hPa - 800 hPa averaged u during summer in each individual simulation. The C1SD simulations show “strong” winds across the Pacific with **the jet** axis tilting SW-NE. The C1 simulations show a much more varied structure, with many showing a more northern and more east-west jet that does not extend across the whole Pacific Ocean.

The summertime distribution of the 500 - 800 hPa average CO50 concentration is also shown in Fig. 5 (colors), and there appears to be a relationship between the midlatitude jet and the Arctic CO50 concentration. In general, lower χ_{CO50} over the Arctic are found in simulations with a more northern jet over the Pacific Ocean and higher χ_{CO50} in simulations with a more southern jet (primarily the C1SD simulations). The correspondence between the latitude of the Pacific jet and the Arctic CO50 during summer is further quantified in Fig. 6(a), which shows a scatter plot of Arctic χ_{CO50} versus ϕ_{jet} of the Pacific jet for summer. This shows that lower χ_{CO50} is generally associated with a more northern ϕ_{jet} , with a clear negative correlation (-0.84) between the climatological-mean values for each model. This suggests that models with a more northern jet generally have weaker (slower) midlatitude-to-Arctic transport.

Repeating the above analysis for winter, we find the wintertime tracer transport from NH midlatitudes **into the** Arctic is also sensitive to **the** jet location, with negative correlations between χ_{CO50} and ϕ_{jet} , **as shown in** Fig. 6(b). This is somewhat surprising given the differences in jet structure among **the** models is much smaller in winter (see Fig. 4).

While there is a negative $\phi_{\text{jet}} \sim \chi_{\text{CO50}}$ correlation among the climatological means for each model, this does not hold for the interannual variations in individual models (small circles). One of the reasons may be that there are other aspects of the jet that can also impact the poleward transport of CO50, such as jet strength and jet structure (i.e., whether SW-NE tilted or zonal). These characteristics vary consistently with ϕ_{jet} for climatologies between models, but are less consistent at the interannual scale in individual models (not shown). Similar inconsistency occurs in $\phi_{v=0}$ that quantifies the location of the mean meridional circulation, as discussed next in Section 4.3.

4.3 Mechanisms

We now explore the underlying mechanisms for the above connection between the Pacific jet and transport into the Arctic. A strong jet with rapid zonal flow at its center can act as a barrier to meridional transport (e.g., Bowman and Carrie, 2002), but there can also be intensive transport on the flanks of the jet due to Rossby wave breaking (RWB) (e.g., Haynes and Shuckburgh, 2000). This RWB on the edge of the jets may explain the connection between jet location and transport into the Arctic. As shown by the schematics in Fig. 7, when the Pacific jet is in a more northern position (e.g., summertime jets in C1 simulations as shown in Fig. 5) the source region of CO50 is on the equatorward flank of the jet and the anti-cyclonic RWB occurring here transports CO50 equatorward and blocks transport to the Arctic. In contrast, when the Pacific jet has a more southern position and its western end tilts more southward (e.g., summertime jets in C1SD simulations), a fraction of the CO50 source region

overlaps the poleward flank of the jet and the cyclonic RWB occurring there transports CO₅₀ to higher latitudes and the Arctic. In other words, differences in the Arctic χ_{CO50} between models with different jet locations could be due to differences in the meridional eddy transport caused by RWB.

One approach to examine whether transport caused by RWB is the cause of differences in the transport into the Arctic is to decompose the tracer flux into a zonal-mean and an eddy components, i.e.,

$$\overline{v\chi_{\text{CO50}}} = \overline{v}\overline{\chi_{\text{CO50}}} + \overline{v'\chi'_{\text{CO50}}}, \quad (1)$$

where $\overline{(\)}$ denotes the zonal mean, $(\)'$ is the corresponding departure from the zonal mean, $\overline{v\chi_{\text{CO50}}}$ is the total flux, $\overline{v}\overline{\chi_{\text{CO50}}}$ is the zonal-mean component, and $\overline{v'\chi'_{\text{CO50}}}$ is the eddy component. The meridional fluxes are further vertically integrated to yield the corresponding **tracer mass flow rate** across each latitude, i.e. the vertically integrated flux is

$$\langle F \rangle = \int_{1000}^{p_0} dp \frac{2\pi a \cos \phi r_M}{g} F \quad (2)$$

where F is the total, mean, or eddy flux, r_M is the ratio of molecular mass weight between CO (28 g/mol) and dry air (28.97 g/mol). ϕ is latitude, p is pressure, a is the Earth's radius of 6370 km, $g = 9.8 \text{ kg ms}^{-2}$ **is the gravity of Earth, and p_0 is the upper limit for the vertical integral. An upper limit of 700 hPa is chosen as the tracer mass flux decays exponentially with altitude following variations of air density and hence the vertically integrated flux throughout the full tropospheric column is mostly captured by fluxes in the lower troposphere (Fig. S3(c,d)).** Positive flux is defined as northward transport while negative corresponds to transport to the south.

A substantial contribution of the eddy flux comes from synoptic eddies, and to calculate this flux requires v and χ_{CO50} at higher frequency than the monthly-mean output available from the CCMI archive. However, we have access to daily output from GEOS-C1 and GEOS-C1SD simulations **from 1990 to 1994**, which can be used to examine the relative roles of mean and eddy fluxes in the meridional transport. As the Arctic χ_{CO50} in GEOS-C1 is much lower than that in GEOS-C1SD (with the difference being almost the largest among CCMI simulations in summer; see Fig. 6), comparison of the fluxes between these simulations can test whether differences in eddy transport are the causes of differences in Arctic CO₅₀ concentrations.

The flux diagnostics for CO₅₀ meridional transport in the two GEOS simulations are shown in Fig. 8. During summer, there is an equatorward transport of CO₅₀ in the subtropics and a poleward transport in the extratropics in both simulations (bold curves in Figs. 8(a,b)). The latitude separating the equatorward transport from the poleward transport shifts from $\sim 40^\circ\text{N}$ in GEOS-C1 to $\sim 34^\circ\text{N}$ in GEOS-C1SD. Given that CO₅₀ is largely emitted from East Asia and South Asia over $20^\circ\text{N} - 40^\circ\text{N}$, most of the CO₅₀ source region is characterized by equatorward transport in GEOS-C1 but a significant fraction of CO₅₀ source stretches into the zone of poleward transport in GEOS-C1SD. This yields a much larger poleward total flux over the midlatitudes in GEOS-C1SD than that in GEOS-C1 (see Fig. 8(c)) consistent with a higher summertime Arctic χ_{CO50} in GEOS-C1SD than that in GEOS-C1. Examination of the zonal-mean and eddy fluxes shows that differences in the total fluxes are dominated by the zonal-mean and not the eddy component (Fig. 8(c)). During winter, the latitude that separates the equatorward transport from poleward transport in GEOS-C1SD ($\sim 36^\circ\text{N}$) is only slightly south of that in GEOS-C1 (\sim

38°N) (Fig. 8(d,e)). However, the total tracer flux of CO50 features a much larger poleward transport over the midlatitudes in GEOS-C1SD than that in GEOS-C1, and this large difference is again primarily due to difference in the zonal-mean fluxes.

The above analysis of tracer fluxes in GEOS-C1 and GEOS-C1SD contradicts our original speculation that the difference in Arctic χ_{CO50} is due to the jet-associated RWB (and eddy transport). Instead, it indicates that differences in the zonal-mean component dominates, which is linked to transport by the mean meridional circulation. We are unable to perform this tracer flux decomposition in all CCMI simulations due to a lack of daily data. However, we can approximate the zonal-mean components of tracer flux using monthly-mean fields as the zonal-mean flux is largely associated with the slowly varying mean meridional circulation. We have confirmed that the zonal-mean flux calculated using monthly output differs only slightly from the one using daily output in both GEOS-C1 and GOES-C1SD (see Fig. S3).

The results for the approximated zonal-mean flux **in each simulation** are shown in Fig. 9. Note that the ACCESS-C1 and NIWA-C1 simulations are excluded for the analysis because v in those simulations was output only at 850 hPa in the lower troposphere (800 hPa - 950 hPa) **which** cannot accurately represent the lower-tropospheric mean as compared to other CCMI simulations. **The latitudinal structure of the zonal-mean flux is generally similar among the models, with equatorward flux in the subtropics and poleward flux in the midlatitudes, but the magnitude of the flux as well as the location where the zonal-mean flux switches sign differ significantly among the models. More importantly, there is a high positive correlation between the approximated zonal-mean flux maximizing on the poleward flank of the midlatitude CO50 source region (700 hPa - 1000 hPa, and 40°N - 60°N during summer versus 30°N - 50°N during winter) and the Arctic χ_{CO50} during both seasons, which suggests that the dominant role of zonal-mean flux in separating the different poleward transport of CO50 between GEOS-C1 and GEOS-C1SD may also be one of the major causes of the spread of Arctic CO50 concentrations among the CCMI models. Note that such positive correlations generally hold in the EMAC models, despite both χ_{CO50} and mean flux in these simulations being biased during summer and winter.**

There are a few simulations deviating from this positive correlation between the zonal-mean flux and CO50 concentrations over the Arctic. **For example, the zonal-mean flux in CMAM-C1 is larger than that in CMAM-C1SD but the Arctic CO50 concentrations are similar between the two simulations.** Further analysis is required to determine what **are the other processes responsible for the variations in the Arctic CO50 concentrations among these simulations. Also, the positive mean flux - CO50 relationship does not hold for interannual variations in most simulations, which will be discussed in details below.**

The above results suggest an important role of mean meridional circulations in **separating the meridional transport of tracers among the CCMI models**, with larger poleward transport when the jet is located more equatorward. A possible reason for this connection between jet location and transport by the mean meridional circulation could be the well-known link between the jet latitude and **the** edge of the HC (e.g., Staten and Reichler, 2014) (also noted in Fig. 7). Specifically, it is shown that when the midlatitude jet is located more poleward, the HC extends further poleward. Thus, when the Pacific jet is in a more northern position, the HC likely also extends further north and the CO50 source region is mostly covered by the lower branch of the HC with southward surface flow, and this may result in less poleward transport. Fig. 10(a) shows the meridional profile of summertime low-level (800 hPa - 950 hPa averaged) zonal mean meridional wind \bar{v} . While there is an agreement in

the general shape of the latitudinal variation of \bar{v} , there is a large spread in the magnitude of the flow and, **equally important**, in the latitude where the flow changes from northerly to southerly.

To examine this possible relationship, we use $\phi_{v=0}$ (see details in Section 2.3) to identify the latitude where the surface meridional flow v changes from southward to northward flow. During summer, $\phi_{v=0}$ varies from 30°N - 46°N among the models, with even a spread of 30°N to 40°N for C1SD simulations. Furthermore, there **appears to be** a negative correlation between $\phi_{v=0}$ and the Arctic χ_{CO50} (Fig. 10(b)), that is, when $\phi_{v=0}$ is further north (south), there is a less (more) poleward transport of CO50. The spread in \bar{v} and $\phi_{v=0}$ among the models is smaller in winter, but there is again **roughly** a negative correlation between $\phi_{v=0}$ and the Arctic χ_{CO50} , see Fig. 10(d,e). **Biases of the Arctic χ_{CO50} in the EMAC models again can have an influence on the examined $\chi_{CO50} - \phi_{v=0}$ relationship noted above but it is difficult to disentangle cleanly.**

Putting those complications aside, the $\phi_{v=0} - CO50$ correlation is weaker than the mean flux - CO50 correlation especially during summer and among C1SD simulations (comparing Fig. 10(b) with Fig. 9(b)). This occurs because of a relatively weaker correlation between $\phi_{v=0}$ and low-level mean flux among C1SD simulations (not shown), which highlights the fact that $\phi_{v=0}$, which only represents the location of the mean meridional circulation cannot accurately represent variations in the surface mean meridional transport in some models. A large contributor to the weak $\phi_{v=0} - CO50$ correlation comes from the C1SD simulations, and there are much higher correlations for both seasons if only C1 simulations are included (see Table S1).

The $\phi_{v=0} - CO50$ relationship is also not well established at the interannual scale, again highlighting the non-representativeness of $\phi_{v=0}$ for interannual variations in the surface meridional flow. Specifically, in many C1SD simulations, v is close to zero for a wide range of latitudes in summer. Although there are only small differences in the mean meridional flow (and mean meridional flux) over this region between years, there can be large interannual variations in $\phi_{v=0}$ (see Fig. S4). Also, other processes can have a larger influence on the interannual variation of CO50 transport than the spread among the models. One example is the meridional transport at high latitudes (60°N - 80°N). The climatological meridional velocity and mean flux north of 70°N are similar among the models especially when comparing to the large intermodal variations in the midlatitudes (see Figs. 9 and 10). As such, transport difference by high-latitude meridional flow and mean flux does not seem to be important for difference of climatological Arctic CO50 concentrations between the models (see Table S2). In contrast, there can be large interannual variations in the high-latitude meridional flow/flux within individual models (see Fig. S4 as an example for summer), and this plays an important role in the interannual variations of poleward CO50 transport, with high or moderate interannual correlations between high-latitude meridional flux and Arctic CO50 in most of the models (see Table S2).

As noted above, previous studies have shown a connection between the latitudinal extent of the HC and the latitude of the midlatitude jet. **We verify this connection among the CCMi models by showing a positive correlation between $\phi_{v=0}$ and ϕ_{jet} , see Fig. 10(c,f). This explains why a negative correlation is also found between ϕ_{jet} and χ_{CO50} . Close inspection of Fig. 10(c,f) shows that there is a tighter $\phi_{jet} - \phi_{v=0}$ relationship for the C1 simulations, but a large spread for C1SD simulations. The C1SD simulations agree in latitude of the jet but there is a large spread (comparable to spread amongst C1 simulations) in $\phi_{v=0}$ (and a corresponding spread in χ_{CO50}) despite both u and v being constrained by reanalyses in C1SD simulations.**

In summary, we have proposed two mechanisms, illustrated in Fig. 7, for why there are generally larger Arctic CO50 concentrations in models with a more southern location of the Pacific jet: The first mechanism relates directly to a shift in jet location and associated changes in RWB, while the other mechanism does not involve the jet directly but instead relates to the surface meridional flow that varies consistently with the jet. Analysis of the zonal-mean and eddy tracer fluxes indicates that the second mechanism is **likely one of the dominant causes of the spread in Arctic CO50 concentrations among the models.** That is, **differences in the mean meridional circulations appear to be key drivers of the spread in poleward transport among the models.**

5 Comparison with other tracers

The above analysis **suggests** that variations in the near-surface extent of the HC (latitude where $v = 0$) among the models is **one of the major contributors** to the spread in transport of CO50 to the Arctic, and that variations in CMF play a minor role. This appears to contradict the studies of Orbe et al. (2017, 2018), which show that variations in CMF play a large role in the spread of the tracers they considered (i.e., NH5 as noted in Section 4.1). We also expect an important role of HC extent and associated zonal-mean transport for Arctic NH50, since NH50 has a zonally uniform boundary condition. We therefore revisit the spread in NH50 among the CCMI models, to compare the relative roles of CMF and HC extent. We also, briefly, examine CCMI model simulations of CO to see if there is also an impact on realistic tracers with full chemistry.

5.1 NH50

The multi-model mean distribution of NH50 features a stronger transport along isentropic surfaces so that Arctic χ_{NH50} peaks in the middle troposphere (~ 400 hPa, see gray lines in Fig. 11(c,d) and also in Fig. S9). Similar to CO50, the spread of NH50 concentrations among **the** models is the largest in midlatitudes and remains almost unchanged further north. The spread in the Arctic concentrations of NH50, in particular, is also comparable to that for CO50 (see Fig. 11; latitude and vertical profiles of each model), with a fractional spread of 20% - 25% in winter and 40% - 50% in summer. The overall similarity between CO50 and NH50 is further indicated in Fig. 12(a,e) with moderate correlations between the Arctic concentrations of the two tracers (0.58 in summer and 0.42 in winter). Again, the $\chi_{\text{CO50}} - \chi_{\text{NH50}}$ correlation during winter is sensitive to models of choice. The positive correlation presented in Fig. 12(e) is largely due to ACCESS and NIWA, and oppositely a negative correlation is rendered if these two models are excluded.

To explore the relative role of changes in CMF, latitude of the Pacific jet (ϕ_{jet}), and HC extent ($\phi_{v=0}$) in causing the spread in the Arctic NH50, we repeat the above analyses in Section 4, and examine the correlations of Arctic concentrations of NH50 with different quantities, see Fig. 12. In contrast to CO50, there is a stronger relationship of NH50 with CMF **during winter, see Fig. 12(f), but still a weak correlation during summer.** Consistent with the study of Orbe et al. (2018) for NH5, Arctic NH50 concentrations tend to be lower in simulations that feature larger low-level mid-latitude CMF and such a correlation is weaker in summer. The summertime CMF - NH50 correlation (-0.09) is much lower than the CMF - NH5 correlation (-0.45) reported by Orbe et al. (2018). This is not due to a difference between NH50 and NH5 but rather different models used in the

two studies. GEOS-CTM is included here but not by Orbe et al. (2018), and it has higher NH50 than models with similar CMF (and lowers the correlation). At the same time MRI simulations are included by Orbe et al. (2018), but not here (as they do not include the CO50 tracer). These MRI results have high CMF and low NH5 and thus increase the correlation.

Unlike CO50, NH50 exhibits only a moderate or weak correlation with ϕ_{jet} and $\phi_{v=0}$. Note that the ϕ_{jet} - NH50 seems to be stronger during winter (-0.5), but this correlation is largely due to the ACCESS and NIWA results. Without these two models, a moderate positive correlation is found instead, which is consistent with Fig. 12(g) showing a moderate $\phi_{v=0}$ - NH50 correlation during winter.

In summary, despite NH50 having a zonally uniform boundary condition, the multi-model spread of Arctic NH50 seems to be much less impacted by differences in the HC extent and associated zonal-mean transport among the models. Instead, NH50 shows a stronger correlation with low-level mid-latitude convection especially during boreal winter, as shown by Orbe et al. (2018). Therefore, in contrast to a minor role for transporting CO50 towards the Arctic, midlatitude convection predominantly contributes to the intermodel variations of Arctic NH50 concentrations in winter. In summer, convection may play a role as important as the HC extent (as for CO50), but the two processes may act oppositely so that correlations of summertime NH50 are weaker for both. The above results again suggest that transport of zonally uniform (or oceanic) tracers differ in pathways compared to land tracers, and low-level convection over the oceans seem to play a more significant role. **Another possible contributor to the CO50-NH50 differences could be the different latitudes of their sources, with NH50 further north than CO50. This possibility needs further analysis.**

5.2 CO

It is also of interest to examine whether the above conclusions based on idealized tracers apply to more realistic tracers with interactive chemistry. We therefore examine whether the spread, and relationship with the HC extent (i.e. $\phi_{v=0}$), found in CO50 can also be found in full chemistry simulations of CO from the CCMI models.

The comparison of CO50 and CO from CCMI results shows a positive relationship between the Arctic concentration of CO50 and CO in winter, but no relationship in summer; see Fig. 13(a,c). This suggests that differences in transport that cause differences in CO50 might explain a significant fraction of the multi-model spread of CO during winter when chemistry is relatively weak, but these transport differences are likely less important during summer when model differences in chemistry dominate. This is borne out in Fig. 13(b,d), which shows a weak-moderate negative relationship between $\phi_{v=0}$ and χ_{CO50} during winter but no relationship during summer. This indicates that chemistry may still determine the spatial distribution of real tracers, especially during summer when tracers are more chemically reactive. As to variations of chemistry among the models, a detailed examination on the spatiotemporal variability of tropospheric OH is needed.

In addition to chemistry, differences in emissions between CO and CO50 are also likely to result in their different sensitivities to variations of the HC extent among the models. In particular, CO features an additional summertime emission source from biomass burning over Siberia, which is in the close proximity to the Arctic and hence tends to have a strong influence on the Arctic CO concentration. However, this emission region is distant from the HC edge over the NH midlatitudes and tracer transport from this **higher-latitude** region is less likely to be impacted by variations of the HC extent.

6 Conclusions

In this study, we examine long-range transport into the Arctic using an idealized CO50 tracer with predominantly midlatitude Asian emissions in simulations from a suite of CCMs. There is a wide spread (20 - 40%) of the Arctic concentrations of CO50 among the simulations, indicating a large intermodel variability in the simulated NH midlatitude-to-Arctic transport. Further, this spread is found to **be correlated with the variation in** the location of the Pacific jet among the models, with lower Arctic tracer concentrations for a more northern Pacific jet. While the **inter-model** spread in transport to the Arctic is associated with the latitude of the jet, our analysis indicates that this **may be** an indirect relationship, with difference in the mean meridional flow (that is correlated with the jet latitude) being **the cause** of differences in the poleward transport of tracers. Specifically, in models with a more northern jet, the Hadley Cell (HC) generally extends further north and the tracer's source region is mostly covered by the lower branch of the HC with southward surface flow, resulting in less poleward transport. Differences in midlatitude convection among the models appear to play a secondary role.

While the **inter-model** spread in Arctic CO50 concentrations is **largely determined** by the **HC-related mean meridional transport**, this is not the case for the NH50 tracer that features zonally uniform midlatitude sources, which shows a larger correlation with midlatitude convection over the Pacific Ocean during winter, as shown for NH5 by Orbe et al. (2018). Thus, it is likely that variations in convection over the oceans are more efficient in influencing the transport of trace species towards the Arctic than variations in surface meridional flow during winter. Specifically, for NH50 that has similar sources from oceans and lands, the role of convection over the oceans overweights the influence of surface meridional flow. In contrast, for CO50 which has emissions primarily over land, variations in convection over the oceans are remote and less influential and therefore zonal-mean transport by surface meridional flow dominates. In summer, the relative importance of convection versus HC extent is more complex for NH50 suggesting comparable and offsetting effects from both convection and surface meridional flow.

The free-running model C1 simulations have a jet on average further poleward than observed during summer (a common bias in climate models (Barnes and Simpson, 2017)), with a corresponding bias in the latitudinal extent of the HC. The correlation between the transport into the Arctic and the latitude of the jet (or the HC edge) then implies that these models likely underestimate the transport into the Arctic. While we have focused on impacts on the CO50 and NH50 tracers, this bias likely exists for the transport of other tracers with predominantly land sources and relatively long lifetimes. Therefore, free-running climate models may underestimate the rate of transport into the Arctic for radiatively important land-based gases, especially during summer.

The specified dynamics simulations (C1SD) which use the same (or very similar) specified meteorological fields do not have bias in jet location, but, surprisingly, there is a spread in the latitude where $v = 0$ (i.e., $\phi_{v=0}$), which results in a spread in the rate of transport into the Arctic. Orbe et al. (2017, 2018) also noted a spread in the transport among C1SD simulations, which they related to the spread in transport due to differences in the parameterized convective mass fluxes. Here we **suggest** that variations in near-surface v is also a major contributor to differences in transport among C1SD models. Analysis of other metrics of the HC extent show agreement among C1SD simulations for a metric based on u (latitude where surface zonal wind vanishes) but a larger spread for a metric based on v (latitude where mean meridional stream function at 500 hPa switches the

sign) [person. comm. with Orbe]. It is an open question as to why the C1SD simulations agree on the latitude of the Pacific jet, but not on the latitude where $v = 0$, **or more fundamentally, why u is constrained while v is not. This needs more analysis in future studies.**

The results presented here suggest that differences in the HC extent and associated mean meridional transport is a major factor in causing the large spread in Arctic CO50 among the models. However, the rather small number of models available and the wide range of differences among these models limits how strong conclusions can be made about the relative importance of different processes. To be more definitive, studies are required where individual aspects of a model (or models) are varied to isolate the role of this process on the transport into the Arctic. Such experiments are planned for the future.

10 *Data availability.* Most of data from CCMI-1 used in this study can be obtained through the British Atmosphere Data Centre (BADC) archive (<ftp://ftp.ceda.ac.uk>, last access: 06 July 2018). Data from Community Earth System Model (CESM) can be obtained through the Climate Data Gateway at National Center for Atmospheric Research (NCAR) (<https://www.earthsystemgrid.org/search.html?Project=CCMI1>, last access: 6 July 2018). For instructions for access to both archives, see <http://blogs.reading.ac.uk/ccmi/badc-data-access/> (last access: 2 August 2018)

15 *Competing interests.* no competing interests

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Table 1. Simulations analyzed in this study^a and their corresponding selected period, horizontal and vertical configurations. Names of simulations follow the same fashion in Orbe et al. (2018). FD is finite difference; FV is finite volume; STL is spectral transform linear; STQ is spectral transform quadratic; TA is hybrid terrain-following altitude; TP is hybrid terrain-following pressure; P is pressure.

Simulations	Selected period	Hor. resolution (lat×long)	Hor. discretization	Vert. levels	Top level	Coord. sys.
ACCESS-C1	01/2000-12/2009	$2.5^{\circ} \times 3.75^{\circ}$	FD	60	84 km	TA/P ^c
CMAM-C1	01/2000-12/2009	$\sim 3.8^{\circ} \times 3.8^{\circ}$ (T47)	STL	71	0.08 Pa	TP
CMAM-C1SD	01/2000-12/2009	$\sim 3.8^{\circ} \times 3.8^{\circ}$ (T47)	STL	71	0.08 Pa	TP
EMAC-L47-C1	01/2000-12/2009	$\sim 2.8^{\circ} \times 2.8^{\circ}$ (T42)	STQ	47	1 Pa	TP
EMAC-L47-C1SD	01/2000-12/2009	$\sim 2.8^{\circ} \times 2.8^{\circ}$ (T42)	STQ	47	1 Pa	TP
EMAC-L90-C1	01/2000-12/2009	$\sim 2.8^{\circ} \times 2.8^{\circ}$ (T42)	STQ	90	1 Pa	TP
EMAC-L90-C1SD	01/2000-12/2009	$\sim 2.8^{\circ} \times 2.8^{\circ}$ (T42)	STQ	90	1 Pa	TP
GEOS-C1	01/1990-12/1998 ^b	$2^{\circ} \times 2^{\circ}$	FV	72	1.5 Pa	TP
GEOS-CTM	01/2000-12/2009	$2^{\circ} \times 2.5^{\circ}$	FV	72	1.5 Pa	TP
GEOS-C1SD	01/2000-12/2007 ^b	$2^{\circ} \times 2^{\circ}$	FV	72	1.5 Pa	TP
WACCM-C1	01/2000-12/2009	$\sim 1.9^{\circ} \times 2.5^{\circ}$	FV	66	140 km	TP
WACCM-C1SDV1	01/2000-12/2009	$\sim 1.9^{\circ} \times 2.5^{\circ}$	FV	88	140 km	TP
WACCM-C1SDV2	01/2000-12/2009	$\sim 1.9^{\circ} \times 2.5^{\circ}$	FV	88	140 km	TP
CAM-C1	01/2000-12/2009	$\sim 1.9^{\circ} \times 2.5^{\circ}$	FV	26	200 Pa	TP
CAM-C1SD	01/2000-12/2009	$\sim 1.9^{\circ} \times 2.5^{\circ}$	FV	56	200 Pa	TP
NIWA-C1	01/2000-12/2009	$2.5^{\circ} \times 3.75^{\circ}$	FD	60	84 km	TA/P ^c

^a **SOCOL-C1, MOCAGE-CTM, and ULAQ-C1 also output CO50, but are neglected from this study because CO50 is incorrectly implemented in these simulations.**

^b Negligible differences in climatology between these simulations and another two corresponding GEOS simulations averaged with the period of 01/2000-12/2009.

^c Simulations are based on the TA coordinates, but the meteorological fields have been particularly interpolated into the P coordinates with 31 levels.

Table 2. Tracers and dynamical/thermodynamic variables of models analyzed in the study. u and v are the zonal wind and meridional wind; CMF is the Convective Mass Flux by moist convection updraft. Available variables are marked by “x”. Variables of some simulations are scaled before the inter-model comparison and details are listed in table footnotes. (Re)analysis for specified dynamics in C1SD simulations are listed in the second last column, otherwise the meteorology is free-running (FR) in C1 simulations. Moreover, meteorological fields are specified by nudging in most of C1SD simulations, except GEOS-CTM uses CTM and GEOS-C1SD uses replay. The moist convection scheme is listed in the last column.

Simulations	CO50	NH50	CO	u	v	w	CMF	Meteor. fields	Moist conv. schemes
ACCESS-C1	x	x	x	x	x^a	x^a	x^b	FR	Hewitt et al. (2011)
CMAM-C1	x		x	x	x	x	x	FR	Zhang and McFarlane (1995)
CMAM-C1SD	x		x	x	x	x	x	ERA-interim	Zhang and McFarlane (1995)
EMAC-L47-C1	x	x^c	x	x	x	x	x	FR	Tiedtke (1989); Nordeng (1994)
EMAC-L47-C1SD	x	x^c	x	x	x	x	x	ERA-interim	Tiedtke (1989); Nordeng (1994)
EMAC-L90-C1SD	x	x^c	x	x	x	x	x	ERA-interim	Tiedtke (1989); Nordeng (1994)
GEOS-C1	x	x^d		x	x	x	x	FR	Moorthi and Suarez (1992); Bacmeister et al. (2006)
GEOS-CTM	x	x		x	x		x	MERRA	Moorthi and Suarez (1992); Bacmeister et al. (2006)
GEOS-C1SD	x	x^d		x	x	x	x	MERRA	Moorthi and Suarez (1992); Bacmeister et al. (2006)
WACCM-C1	x	x	x	x	x	x	x	FR	Hack (1994); Zhang and McFarlane (1995)
WACCM-C1SDV1	x	x	x	x	x	x	x	MERRA ^e	Hack (1994); Zhang and McFarlane (1995)
WACCM-C1SDV2	x	x	x	x	x	x	x	MERRA ^f	Hack (1994); Zhang and McFarlane (1995)
CAM-C1	x	x	x	x	x	x	x	FR	Hack (1994); Zhang and McFarlane (1995)
CAM-C1SD	x	x	x	x	x	x	x	MERRA ^e	Hack (1994); Zhang and McFarlane (1995)-
NIWA-C1	x	x	x	x	x^a	x^a	x^b	FR	Hewitt et al. (2011)

^a There are only two levels below 800 hPa (i.e., 850 hPa and 1000 hPa) for model output in P coordinates. Therefore, further vertical interpolation is problematic near the surface due to large impacts of topography at 1000 hPa, and analyses on lower-troposphere v and related diagnosis of the Hadley Cell exclude these simulations.

^b Scaled by 1/9.80665.

^c Scaled by 100.

^d Scaled by 0.001.

^e Relaxing timescale is 50 hours for nudging.

^f Relaxing timescale is 5 hours for nudging.

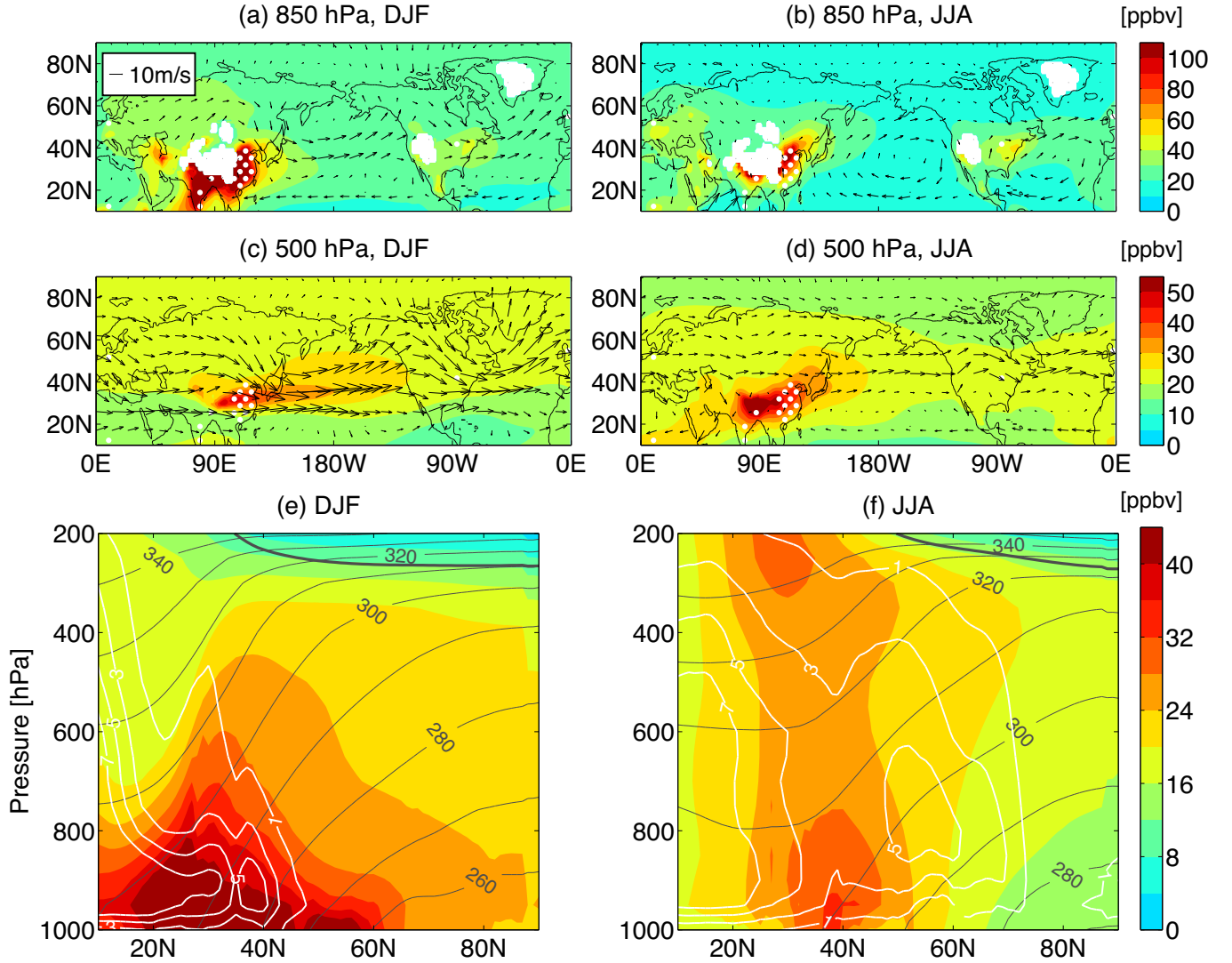


Figure 1. Ensemble mean of the horizontal distribution of CO50 concentration (shades, units: ppbv) at levels of (a,b) 850 hPa and (c,d) 500 hPa, and ensemble mean of the meridional and vertical transport of CO50 by showing the zonal-mean cross sections (e,f) during (a,c,e) DJF and (b,d,f) JJA. In (a,b,c,d), horizontal winds (u,v) are overlaid as vectors and regions of CO50 sources are highlighted by white stipples within which CO50 emission fluxes are larger than $0.4 \times 10^{-9} \text{ kg/m}^2/\text{s}$. In (e,f), isentropic surfaces are overlaid as dark gray isopleths (units: K), the tropopause is marked as the bold dark gray curve, and regional convective mass flux (CMF) over the East Asia (110°E - 140°E) are denoted by white contours (units: $10^{-3} \text{ kg/m}^2/\text{s}$).

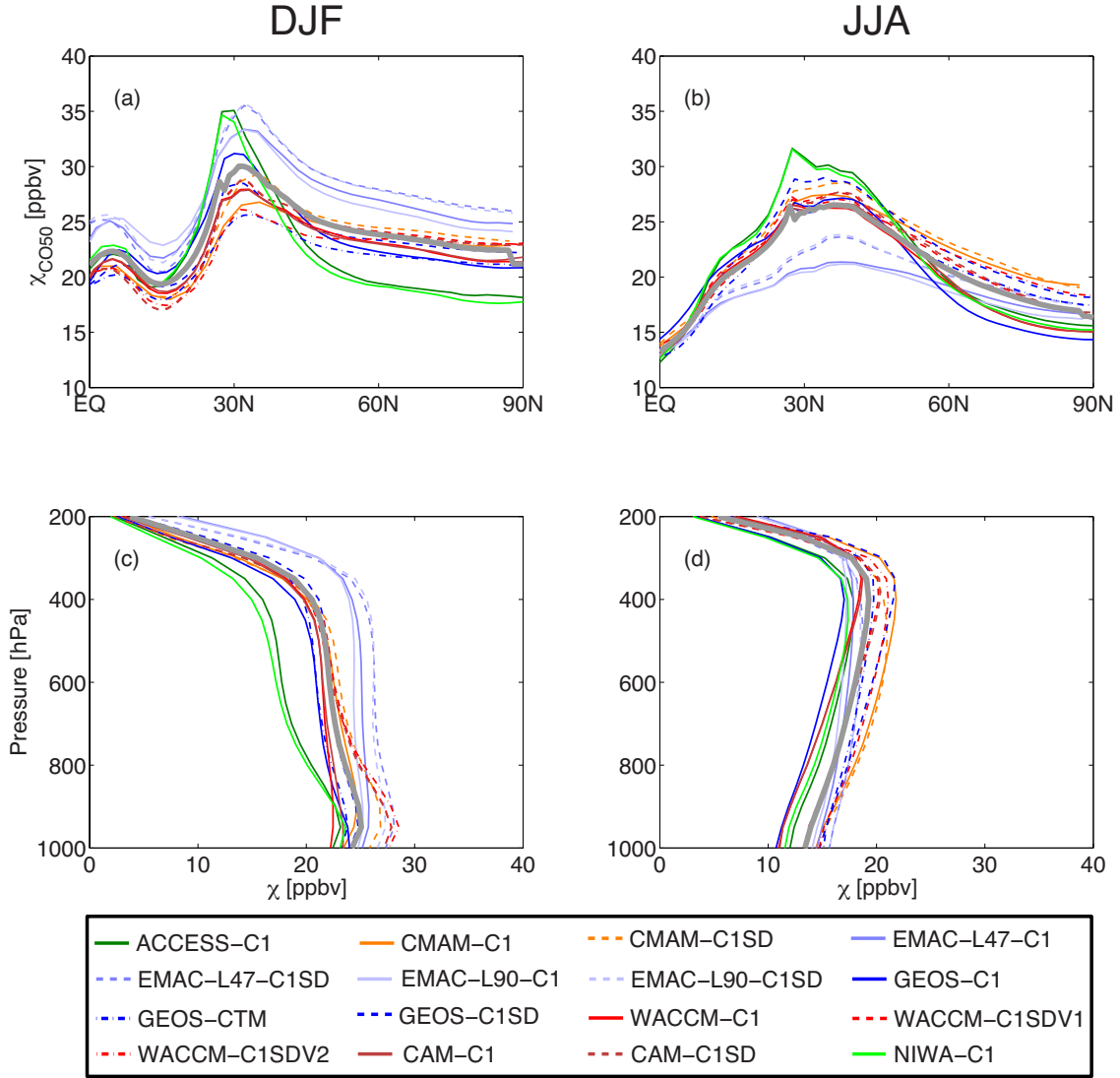


Figure 2. (a,b) Latitudinal and (c,d) vertical variations of zonal mean CO50 concentration χ_{CO50} (units: ppbv) for each model during (a,c) DJF and (b,d) JJA. The corresponding ensemble means are depicted as the heavy gray lines. In (a,b), χ_{CO50} are averaged in the middle and lower troposphere (500-800 hPa); whereas in (c,d), χ_{CO50} are averaged over the Arctic (70°N-90°N).

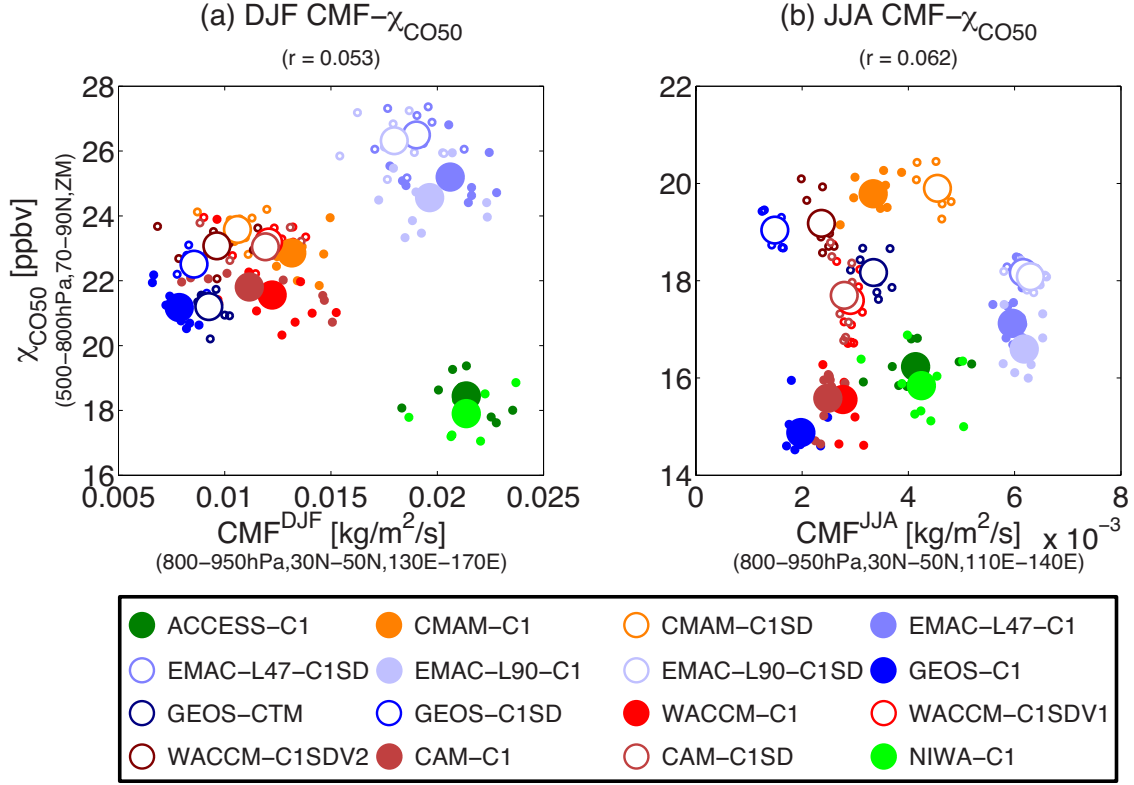


Figure 3. Correlation between Arctic CO50 concentration χ_{CO50} (units: ppbv) and low-level midlatitude CMF (units: $\text{kg}/\text{m}^2/\text{s}$) in (a) DJF and (b) JJA. Arctic χ_{CO50} is here the vertical average of 500 hPa - 800 hPa, latitudinal average of 70°N - 90°N, and zonal mean (ZM). CMF is here the vertical average of 800 hPa - 950 hPa, latitudinal average of 30°N - 50°N in both DJF and JJA, while longitudinal average window differs between seasons, as DJF CMF highlights the robust convection over the western Pacific Ocean (130°E-170°E) whereas JJA CMF focuses on the maritime convection over the East Asia (110°E-140°E) following Orbe et al. (2018). Large marks denote the 2000-2009 climatology (except GEOS-C1 and GEOS-C1SD) while small marks denote the corresponding interannual variations of each simulation. Results of least squares fitting based on climatological means are shown as the solid black lines, and the corresponding Pearson correlation coefficients are given in parentheses in the titles. **If this correlation coefficient is significant (95%), a corresponding linear regression is derived using the total least square method (Petráš and Bednářová, 2010).**

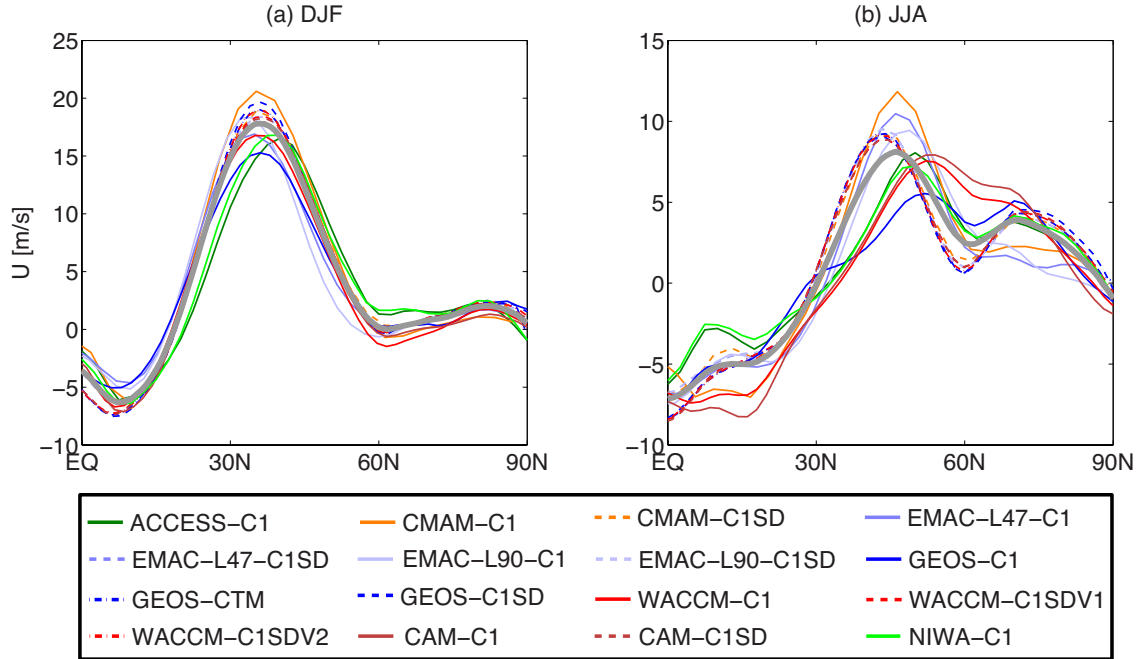


Figure 4. Multi-model spread of latitudinal profile of zonal wind u vertically averaged between 500 hPa and 800 hPa and longitudinally averaged over the Pacific Ocean (135°E-135°W) during (a) DJF and (b) JJA. C1 simulations are shown in solid lines while C1SD simulations are shown in either dashed or dotted-dashed lines. The corresponding ensemble means are depicted as the heavy gray lines.

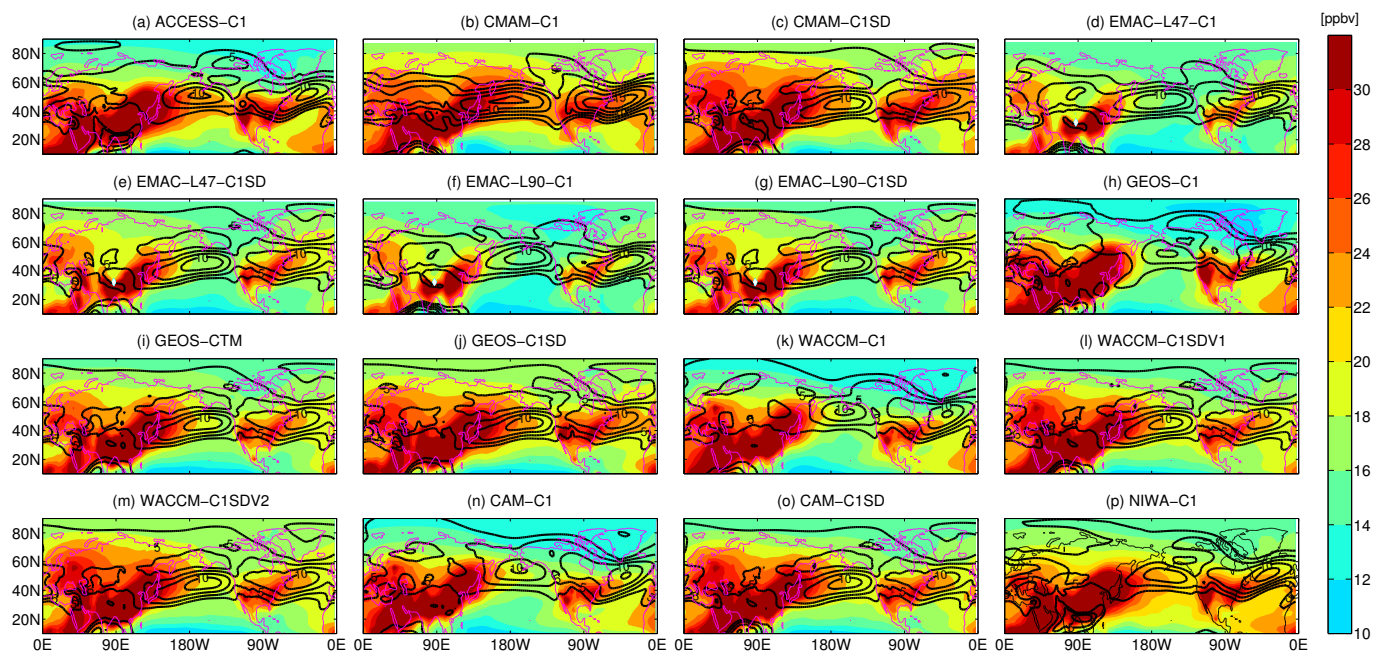


Figure 5. Maps of 500 hPa - 800 hPa averaged CO50 distribution (shades, units: ppbv) and the corresponding 500-800 hPa averaged u during JJA in each CCM simulation.

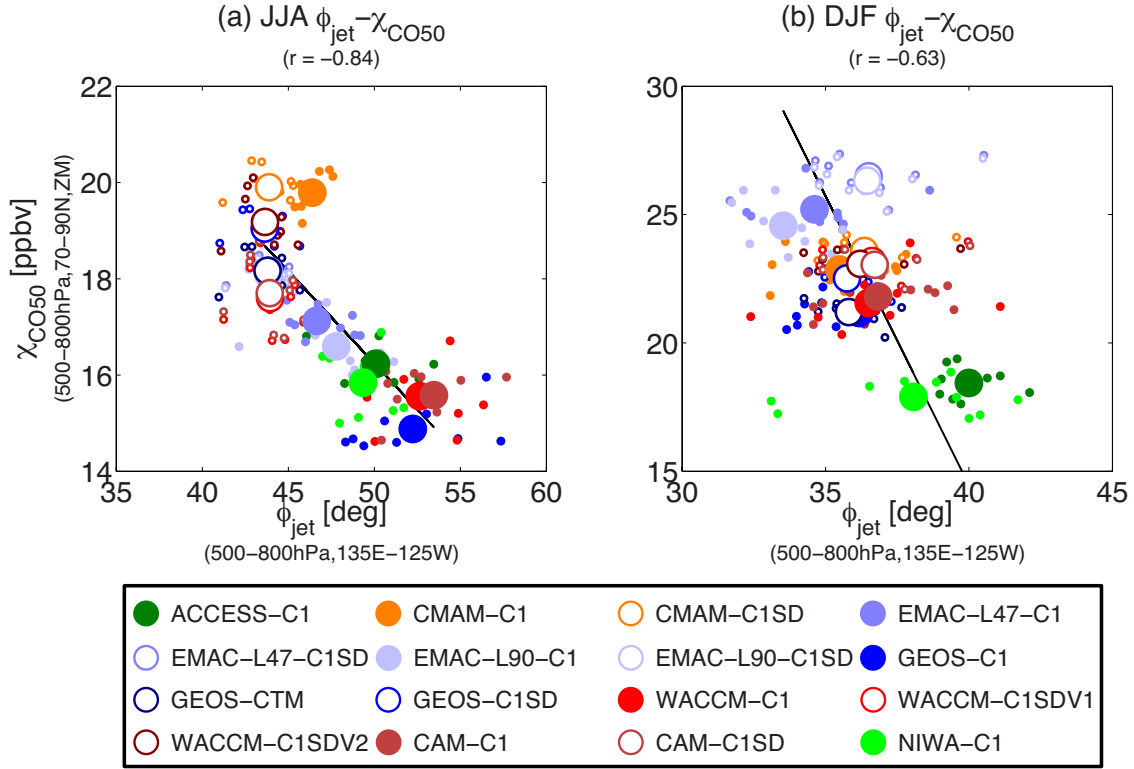


Figure 6. Similar to Fig. 3, but for the correlation between Arctic χ_{CO50} and latitudinal location of the NH midlatitude jet ϕ_{jet} over the Pacific Ocean (135°E–135°W) (Barnes and Polvani, 2013). Note that the sequence of displayed seasons switches with JJA in (a) and DJF in (b).

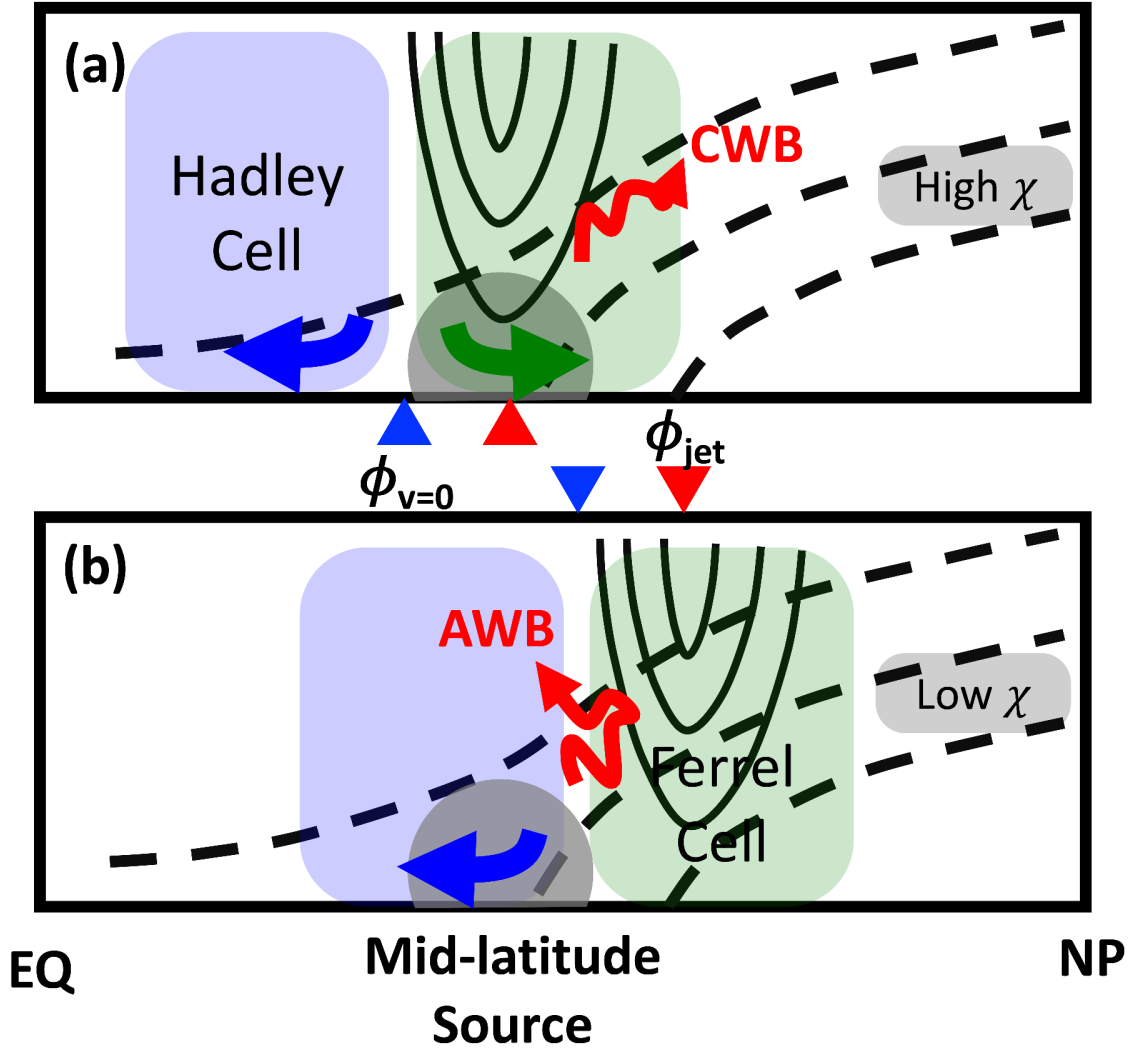


Figure 7. Schematics of mechanisms illustrating dynamic influences on the NH midlatitude-to-Arctic transport for the midlatitude jet situated more southern in (a) and more northern in (b). When jet location ϕ_{jet} and meridional flow switching point $\phi_{v=0}$ are more southern, cyclonic wave breaking (CWB) along the poleward flank of the jet and northward surface meridional flow (green arrow) result in high tracer concentrations in the high latitudes. In contrast, when ϕ_{jet} and $\phi_{v=0}$ are more northern, anticyclonic wave breaking (AWB) along the equatorward flank of the jet and southward surface meridional flow (blue arrow) result in low tracer concentrations over the Arctic. Tracer sources in the NH midlatitudes are denoted by the gray shades, isentropic surfaces are depicted as dashed lines, red and blue triangles mark the latitude of ϕ_{jet} and $\phi_{v=0}$ respectively.

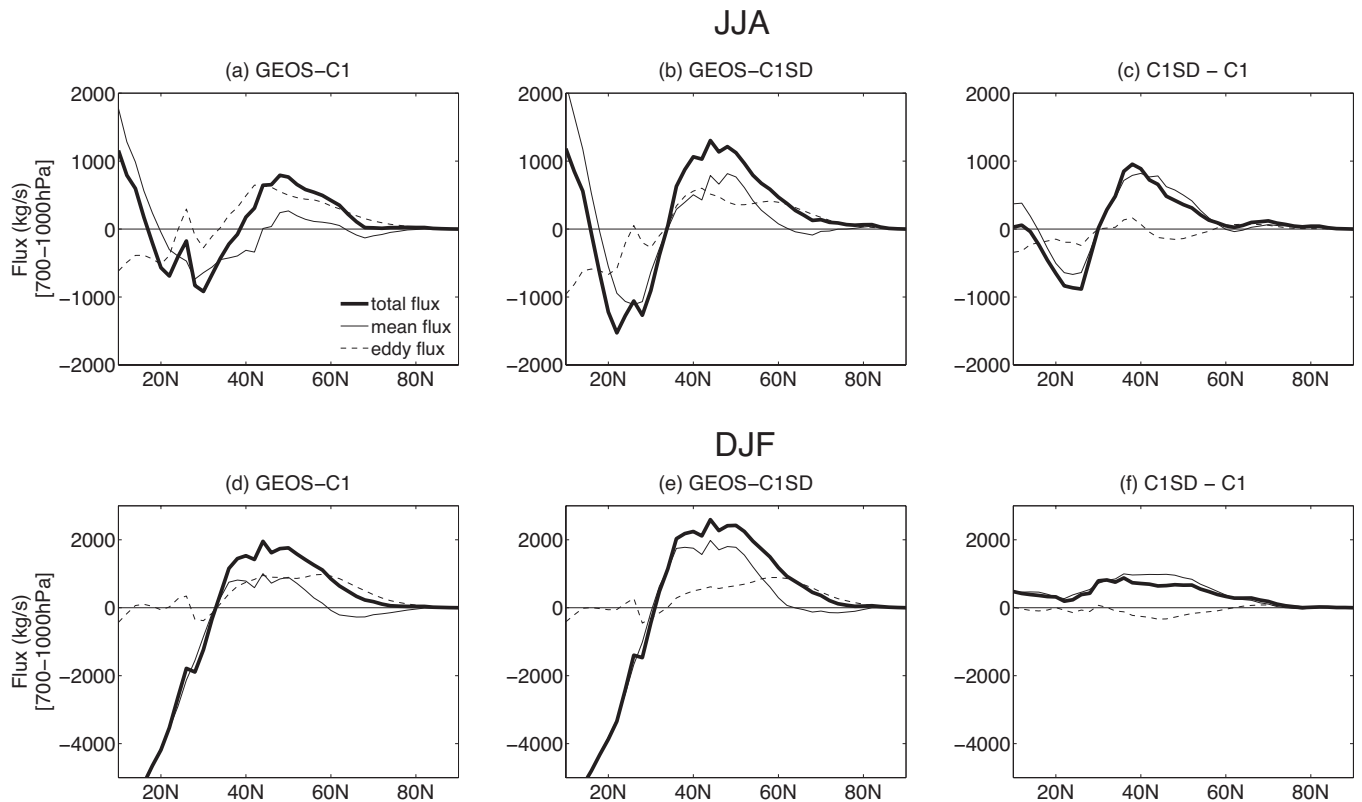


Figure 8. Tracer flux diagnostics showing the total flux $\langle \overline{v \chi_{\text{CO50}}} \rangle$ (heavy), zonal mean flux $\langle \overline{v \overline{\chi_{\text{CO50}}}} \rangle$ (light, **solid**) and eddy flux $\langle \overline{v' \chi'_{\text{CO50}}} \rangle$ (light, **dashed**) of CO50 (vertically integrated from 1000 hPa to 200 hPa; positive means northward; units: kg/s) in (a) GEOS-C1, (b) GEOS-C1SD, and (c) the difference GEOS-C1SD – GEOS-C1 during summer. (d,e,f) are similar to (a,b,c) but during winter. Results are based on the daily GEOS model output.

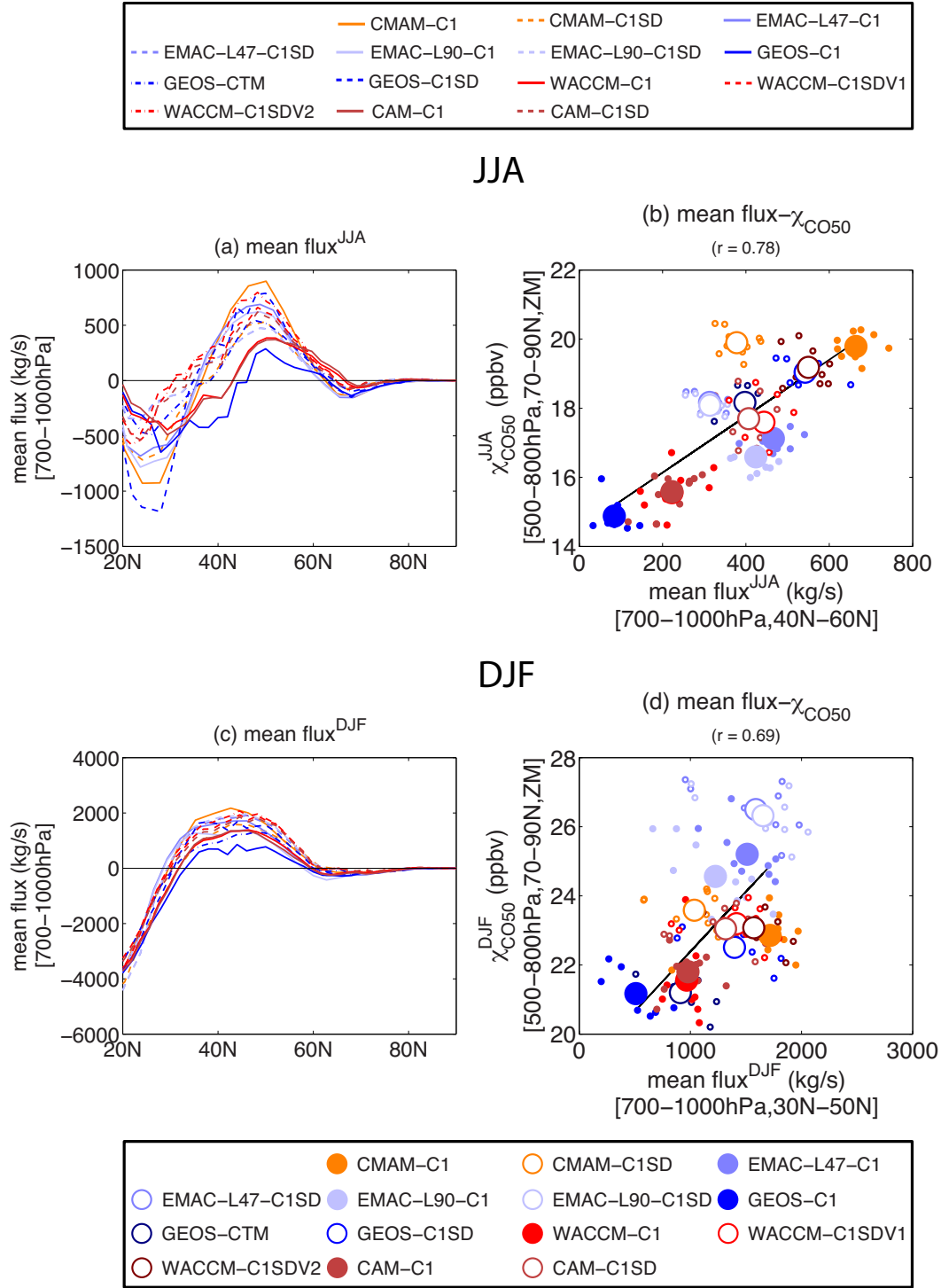


Figure 9. Approximated zonal-mean flux using monthly output: (a,c) latitudinal profile of $\langle \bar{v} \overline{\chi_{CO50}} \rangle$ (units: kg/s); (b,d) correlation between the zonal-mean flux (latitudinal average of 40°N - 60°N **during JJA** and 30°N - 50°N **during DJF**) and the Arctic CO50 concentration (vertical average of 500 hPa - 800 hPa, latitudinal average of 70°N - 90°N, and zonal mean (ZM), units: ppbv), during JJA in (a,b) and during DJF in (c,d).

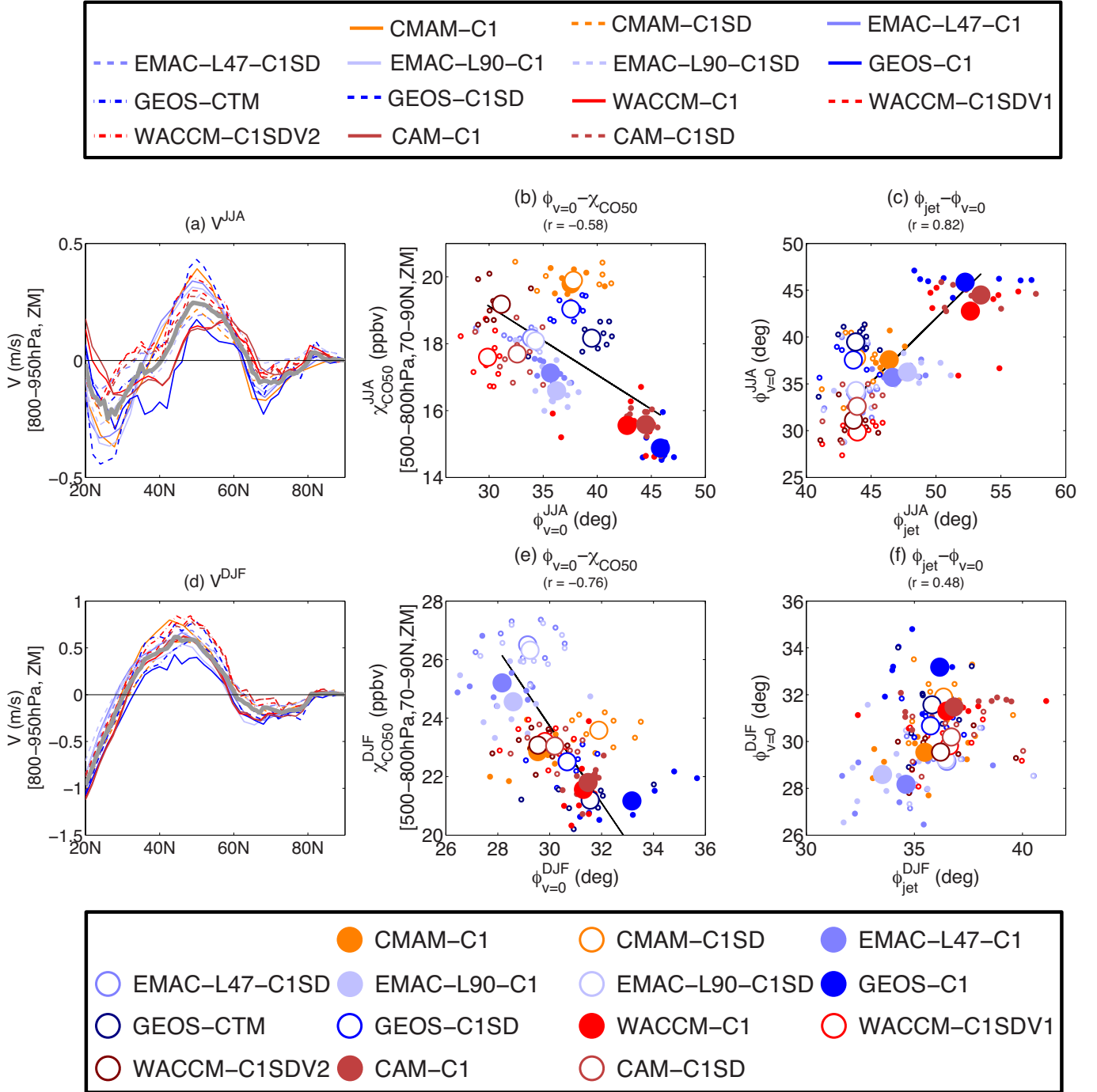


Figure 10. Similar to Fig. 9 in (a,b,d,e), but for low-level (800 hPa - 950 hPa) zonal-mean meridional wind v and $\phi_{v=0}$ marking the latitude for low-level v switching from southward flow at south to northward flow at north. The correlations between $\phi_{v=0}$ and jet location ϕ_{jet} are further shown in (c,f).

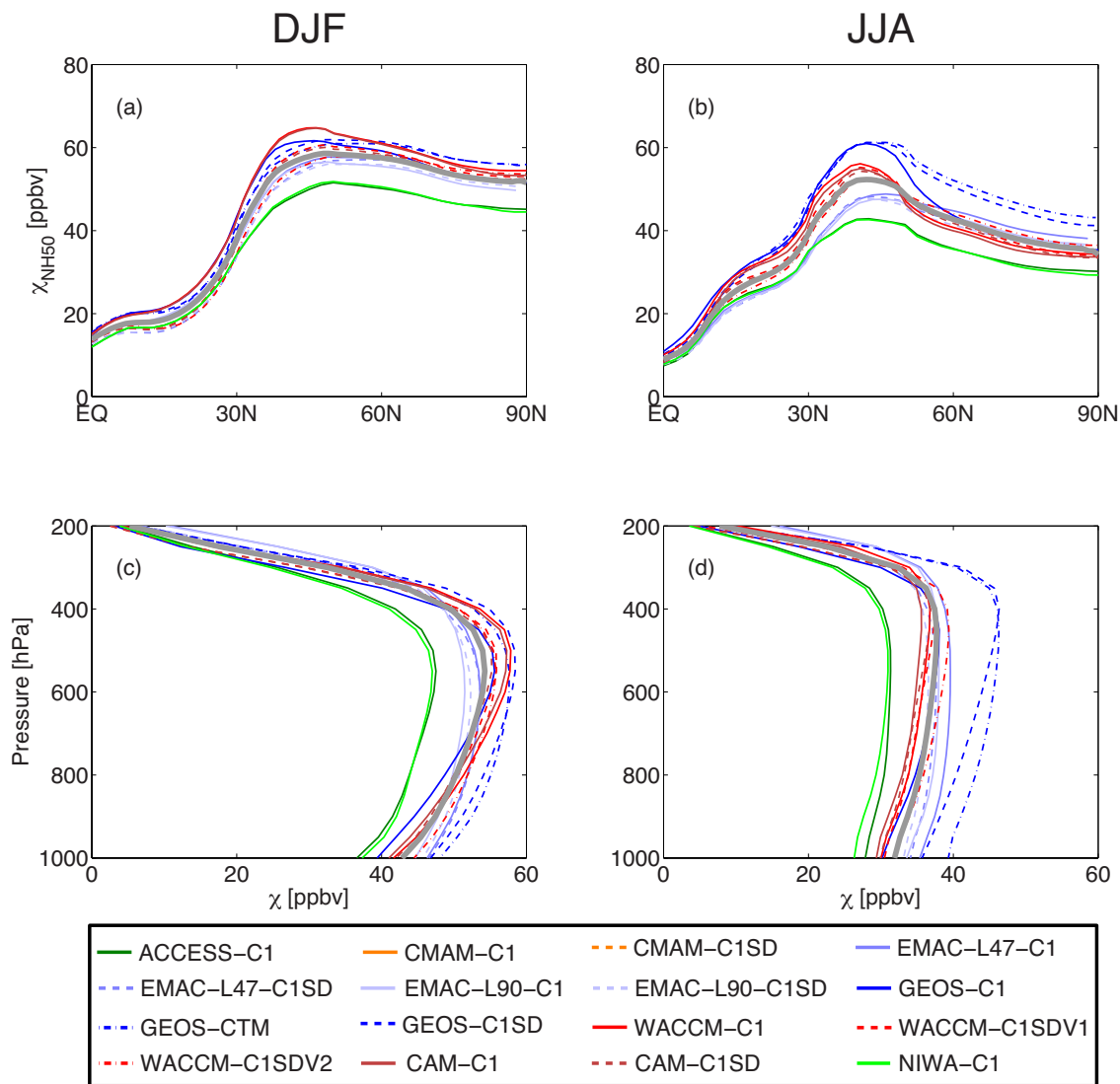


Figure 11. As in Fig. 2 except for NH₅₀.

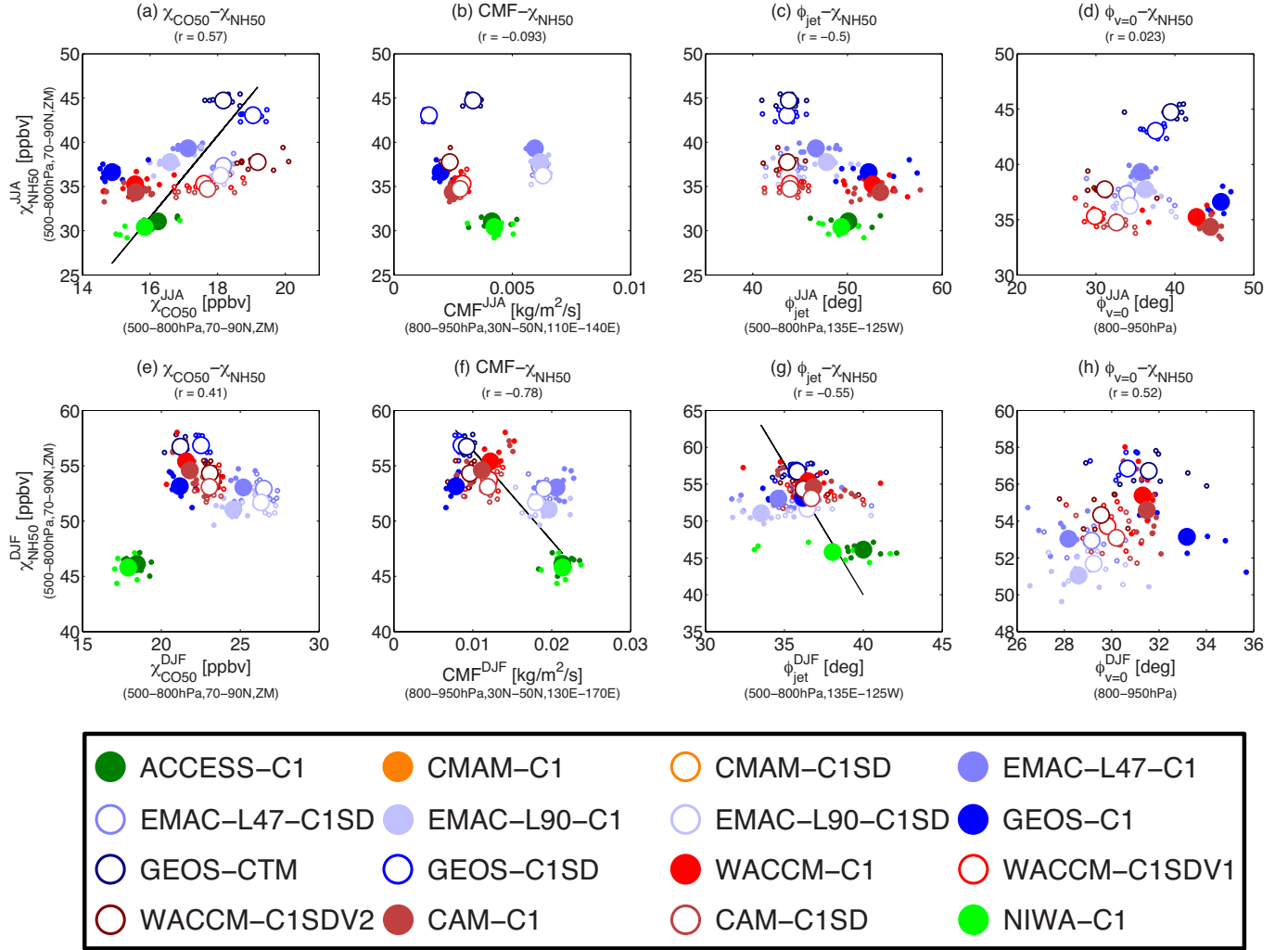


Figure 12. (a,e) Tracer-tracer correlation between NH50 and CO50 over the Arctic (500 hPa - 800 hPa, 70°N-90°N, ZM), (b,f) as in Fig. 3, (c,g) as in Fig. 6, and (d,h) as in Fig. 10(b,e), except y-axis is replaced as the Arctic NH50 concentration. Results for JJA are shown in (a,b,c,d), while those for DJF are shown in (e,f,g,h)

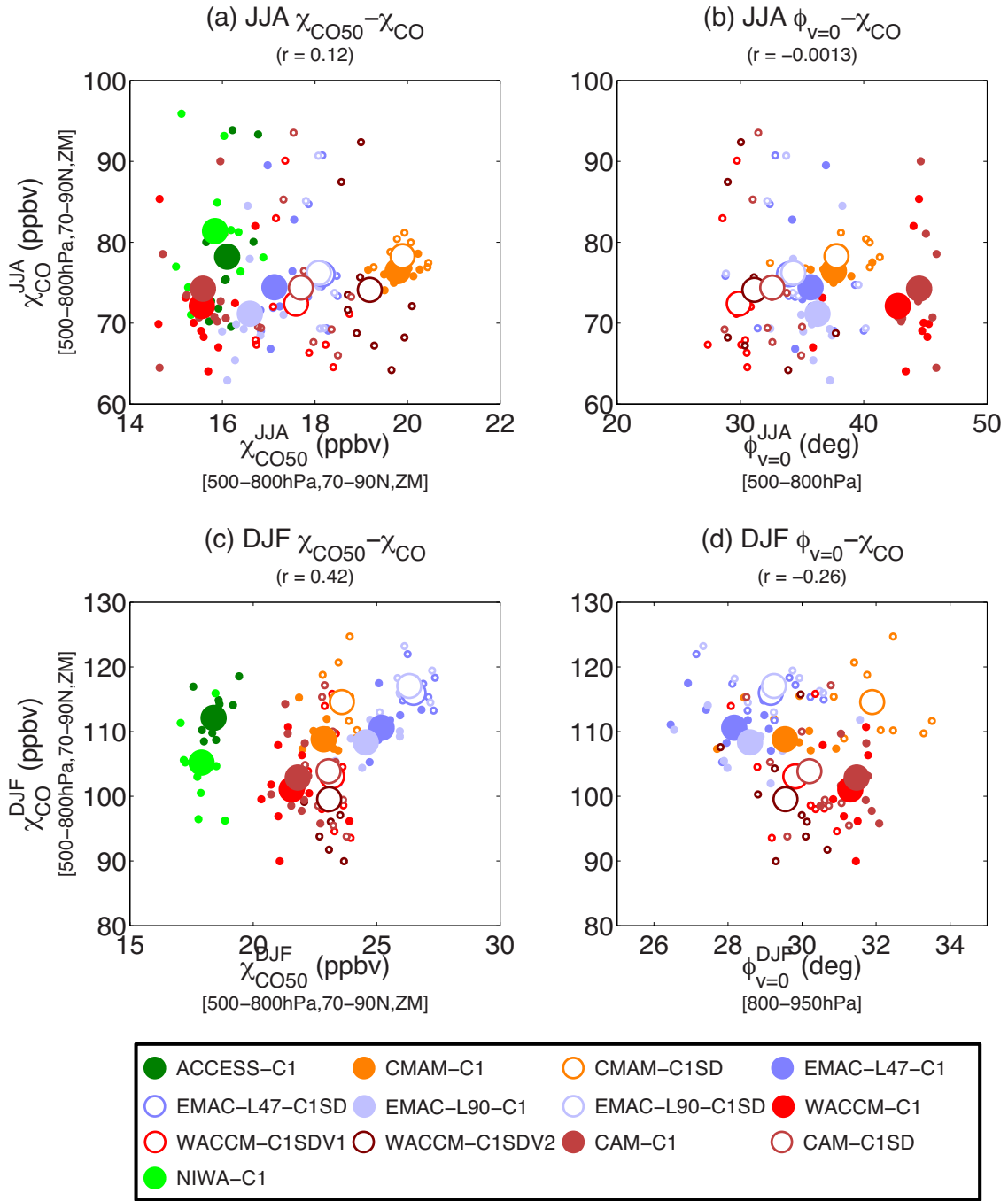


Figure 13. Similar to Fig. 12(a,c,e,g), but for correlations: (a,c) between χ_{CO} and χ_{CO50} , and (b,d) between $\phi_{v=0}$ and χ_{CO} ; during DJF in (a,b) and JJA in (c,d).