Understanding and predicting ocean carbon uptake using coupled climate models: Recent achievements and open challenges

Guest Editors:
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The global ocean is a major sink of anthropogenic CO_2, significantly slowing the CO_2 increase in the atmosphere due to anthropogenic emissions. However, the absorption of excess greenhouse gases and the warming trend of our climate over the last few decades affect the ocean circulation, biogeochemistry and ecosystem structure. Those changes, in turn, may have positive feedbacks on atmospheric CO_2 concentrations through the slowdown of oceanic carbon uptake, further enhancing global warming. Therefore, feedbacks between the carbon cycle and climate represent a mechanism by which the overall climate sensitivity to radiative forcing may be amplified. The strength of these feedbacks depends on the complex interplay between physical and biogeochemical processes. These feedbacks remain a major uncertainty in climate simulations due to the number of processes and associated temporal and spatial scales involved and the difficulties of parameterizing them.

Ocean biogeochemistry in the fifth Coupled Model Intercomparison Project (CMIP5)

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Since model projections were used in the Intergovernmental Panel on Climate Change's (IPCC) First Assessment (Houghton et al. 1990), the trajectory of carbon dioxide (CO_2) has been a central player in climate projection and model intercomparison. Not until the Fifth Coupled Model Intercomparison Project (CMIP5; Taylor et al. 2012) in support of the IPCC 5th Assessment (IPCC 2013), however, were fully coupled climate-carbon cycle Earth system models mature and pervasive enough for explicit inclusion in the intercomparison. This article describes CMIP5 accomplishments and remaining challenges faced by the ocean biogeochemistry community for advancing coupled carbon-climate and marine ecosystem research.

Origins of CMIP5 ocean biogeochemistry

Since ocean biogeochemical general circulation models (OBGCMs; Sarmiento et al. 1993) began incorporating an explicit carbon cycle (Bacastow and Maier-Reimer 1990; Siegenthaler and Sarmiento 1993), global models of climate change response (Sarmiento and Le Quéré 1996; Bopp et al. 2001) and later, more 'intermediate' complexity models of coupled elemental cycles (Moore et al. 2004; Le Quéré et al. 2005) have been applied to the coupled carbon-climate problem. Typical OBGCM applications include tracking how much anthropogenic carbon uptake has occurred historically and its projection into the future, characterization of natural carbon cycle change, and description of ecosystem variability and change, all in the face of climate change.

IN THIS ISSUE

The Coupled Model Intercomparison Project phase 5 (CMIP5) provided coordinated sets of climate simulations with an interactive carbon cycling component that represented a unique and time-sensitive opportunity to assess the strength of the climate-carbon cycle feedbacks in a multi-model context. In 2012, a working group on "Oceanic carbon update in the CMIP5 models" - jointly sponsored by the US CLIVAR and OCB (Ocean Carbon Biogeochemistry) programs - set out to investigate differences among model predictions across multiple time scales and in different ocean basins and understand the representation of such feedbacks to possibly narrow uncertainties in the next generation of Earth system models. This effort culminated in a community workshop held in December 2014 in San Francisco, CA, entitled: "Ocean's Carbon and Heat Uptake: Uncertainties and Metrics."

This joint edition of the OCB and US CLIVAR newsletters is based on contributions spanning the range of topics covered at the workshop. It is representative of the challenges and advances across disciplines in modeling and understanding mechanisms, sensitivities, and feedbacks of ocean carbon uptake.

A variety of OBGCMs are now in use, and they exhibit fundamentally different representations of regional patterns in productivity and sensitivity to climate warming (Steinacher 2010). At their core are distinct ecological modeling strategies to distill the vast complexity of natural systems in the face of limited, imperfect information into a discrete set of mathematical representations of nutrient-phytoplankton-zooplankton interactions. The more sophisticated of these focus on phytoplankton functional groups (CO_2-fixers, N_2-fixers, silicifiers, calcifiers) to conduct multi-element biogeochemistry (Le Quéré et al. 2005), either through calibration to particularly well-known species from laboratory and field studies (e.g., Prochlorococcus, Trichodesmium, T. weissflogii, E. huxleya) or through stoichiometrically-constrained empirical functions (Dunne et al. 2007).

Assessment and attribution of biases in OBGCMs is challenged by a variety of factors. The biological and biogeochemical constraints and theories on which these models are based represent only a small amount of the overall ecosystem variability observed in nature. Thus modelers are afforded much latitude in prioritizing and parameterizing a given ecosystem's constraints based on process studies, field-, and satellite-based observations. Key uncertainties include mechanistic controls on euphotic zone nutrient consumption and degree of residual nutrient, particulate and dissolved organic matter passive and active transport, deviations in stoichiometry from Redfield (e.g., N_2 fixation), and remineralization scales through the twilight zone. Ecological uncertainties include the general controls, functional traits, adaptation limits on phytoplankton physiology, the predictability, phenology, and niche gaps in biodiversity, and the spectrum of trophic interactions. Beyond the biological factors themselves, many of the fundamental controls on ocean biogeochemistry are physical in origin, including atmospheric wind, freshwater and buoyancy forcing, and ocean physics and circulation. Similarly, external factors such as light, deposition, and river and sediment interactions may also be key to ecosystem function. Each model thus represents a consortium of expert decisions towards a highly idealized representation of the coupled physical, biogeochemical, and ecological system.

CMIP5 OBGCMs

With respect to marine ecology, the CMIP5 suite spans a range of phytoplankton species diversity and ecological interactions. The models consider a range of elemental cycles of carbon, nitrogen, oxygen, phosphorus, silicon, iron, alkalinity, and lithogenic material and different parameterizations of the processes that couple and decouple these elements, including gas exchange, primary production, ecosystem processing, particle sinking, dissolved organic matter cycling, atmospheric deposition, river input, scavenging, and sediment removal and supply. CMIP5 models include one to three of the following phytoplankton groups: Diatoms, picoplankton (Prochlorococcos), nanoplankton, flagellates, calcifiers, and diazotrophs (N_2-fixers). Most models distinguish between large diatoms and small phytoplankton and represents calcification implicitly as part of the small or large phytoplankton. Phytoplankton growth is limited by light, nutrients, and in most models by temperature (no temperature effect in HadGEM; The HadGEM2 development team 2011). Nitrogen is a limiting nutrient in all models; additionally many models consider limitation by iron, silicate, and sometimes ammonium and phosphate. Most models assume Redfield C:N:P Ratio but allow for varying Si, Fe and Chl:C ratios. A few models (e.g., GFDL-TOPAZ; Dunne et al. 2013 and PELAGOS; Vichi et al. 2007) enable deviations from the Redfield ratio. While the underlying equations of phytoplankton growth, temperature, and light limitation are similar among models, models follow either Michaelis-Menten or quota equations for nutrient limitation and exhibit vast differences in parameter values. CMIP5 models include one to three zooplankton types to representing different size classes - and use a variety of different grazing functional forms - resulting in food web dynamics

that differ greatly among models (Sailley et al. 2013). Particulate organic matter is produced during grazing and in some models by direct aggregation of phytoplankton. The equations describing particle sinking range from simplistic implementations of constant sinking speed and either a constant, temperature-, or depth-dependent remineralization (e.g., PELAGOS; Vichi et al. 2007) to more elaborate implementations, including particle aggregation, grazing of particles, mineral ballasting, or different particle sizes with different sinking speeds (e.g., IPSL-PISCES, Aumont et al. 2006; CESM-BEC, Moore et al. 2013; GFDL-TOPAZ Dunne et al., 2013).

Within several modeling centers, including GFDL (Dunne et al. 2012; 2013; Figure 1), GISS (Schmidt et al. 2014), IPSL (Dufresne et al. 2013), MPI (Ilyana et al. 2013), and others, alternative representations of the physical model underlying the biogeochemical algorithms were applied. These models were demonstrated to have vast differences in baseline simulation characteristics as a consequence of physics alone, as illustrated by particle export in GFDL's ESM2M (8 PgC/yr) being approximately 30% higher than that in the isopycnal coordinate ESM2G (5 PgC/yr), with similar overall fidelity and often opposing water column tracer biases (Dunne et al. 2013).

Comparison of CMIP5 OBGCM fidelity and sensitivity
Even in the face of such strong differences in baseline simulation (Figure 1; Dunne et al. 2013), anthropogenic carbon uptake across the CMIP5 suite of models (Figure 2; Frölicher et al. 2015) illustrates broad agreement at the 20% uncertainty level with relative dominance of the Southern Ocean in terms of uptake. Solubility and passive transport dominate CO_2 uptake along pathways of ocean gas exchange, surface ventilation, and interior propagation wherein this generation of model has demonstrated vast improvement over past generations of models in both carbon uptake (Doney et al. 2004; Matsumoto et al. 2004; Frölicher et al. 2015) and feedbacks (Arora et al. 2014; Friedlingstein et al. 2004). Key factors underlying this improved consensus across the CMIP5 suite likely include consistency in the implementation of aqueous geochemistry (algorithm of OCMIP2 (Orr et al. 2001) based on Millero et al. (1995)) and gas exchange (algorithm of Wanninkhof 1992), and in the representation of the large-scale ocean circulation.

Yet in the context of ocean acidification, surprises have arisen. Resplandy et al. (2013) demonstrated that the maximum acidification response to surface CO_2 forcing may somewhat quixotically be manifested in the subsurface as accumulation of the anthropogenic carbon signal in subtropical mode waters with naturally high levels of remineralized CO_2 combined with enhanced surface stratification and intensification and shoaling of the nutricline. Further work to identify key mechanisms in these models has demonstrated the importance of restratification and advection of interior temperature gradients that lead to strong divergence in the patterns of warming and anthropogenic CO_2 (Winton et al. 2013).
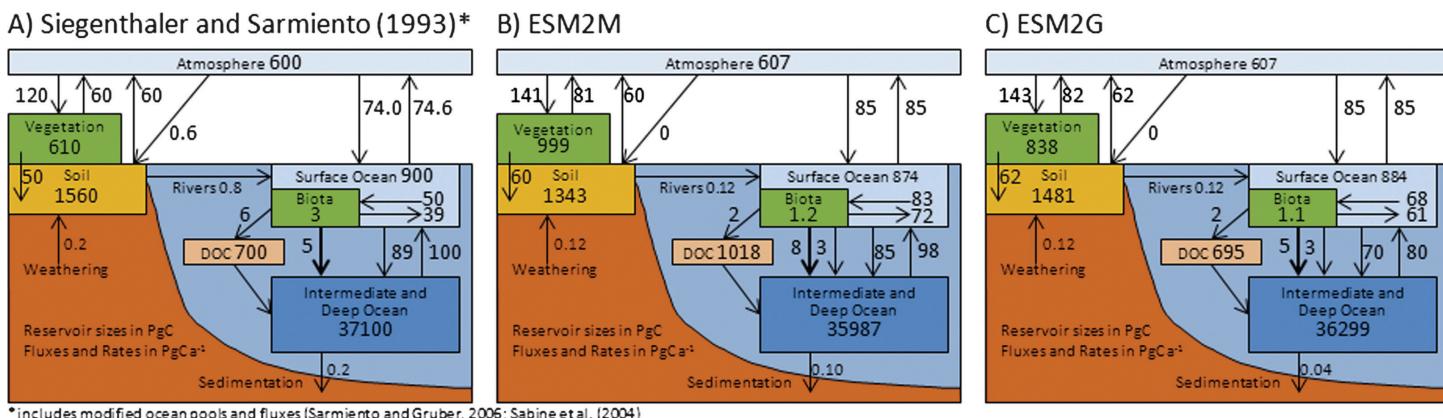

Figure 1: Global carbon cycle schematic comparison to estimate by Siegenthaler and Sarmiento (1993) with ocean additions from IPCC (Sabine et al. 2004). Reprinted from Figure 1 of Dunne et al. (2013).
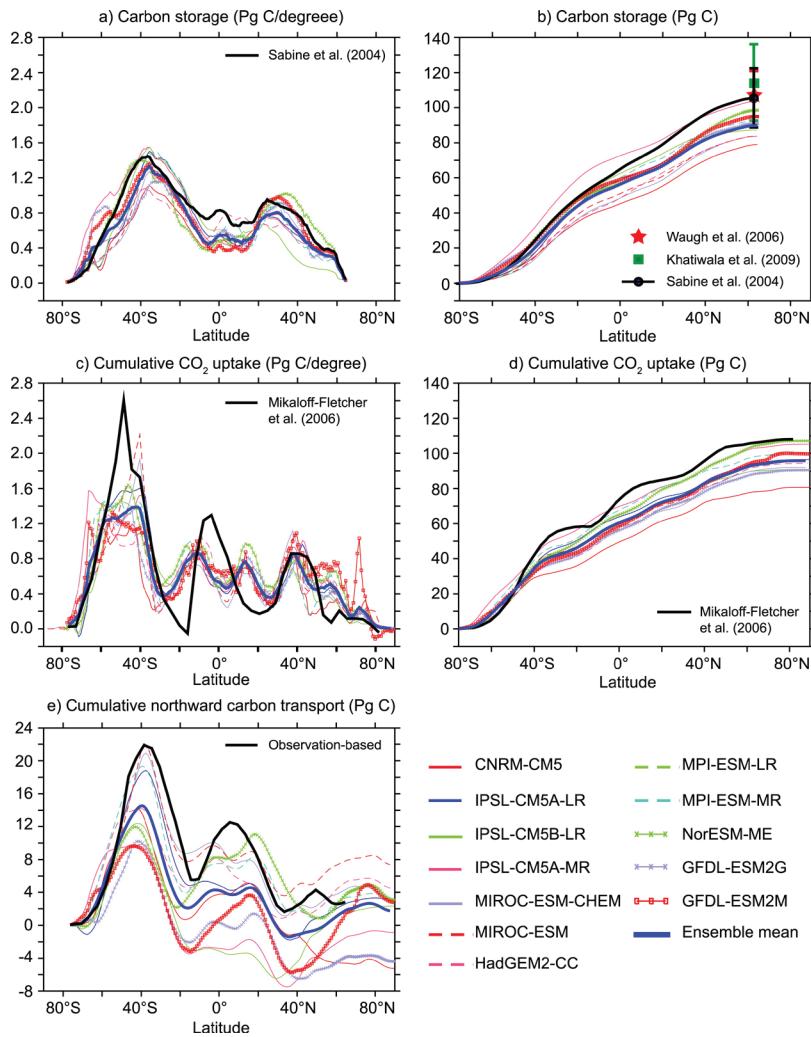
Figure 2: Changes in oceanic storage, uptake, and transport of anthropogenic carbon between 1870 (represented by mean of period 1861–80) and 1995 (represented by mean of period 1986–2005) simulated by 12 CMIP5 models. (a) Zonal integrated oceanic anthropogenic carbon storage, (b) zonal integrated oceanic anthropogenic carbon storage integrated from 90°S to 90°N such that the vertical scale goes from 0 at 90°S to the total storage at 90°N, (c) zonal integrated cumulative ocean anthropogenic CO_2 uptake, (d) zonal integrated cumulative ocean anthropogenic CO_2 uptake integrated from 90°S to 90°N such that the vertical scale goes from 0 at 90°S to the total uptake at 90°N, and (e) northward oceanic anthropogenic carbon transport. The transport of anthropogenic carbon is the divergence of the anthropogenic CO_2 uptake and the anthropogenic carbon storage. The observation-based estimate of oceanic anthropogenic carbon transport is the divergence of the anthropogenic carbon flux estimates of Mikaloff Fletcher et al. (2006) and the anthropogenic carbon storage estimates of Sabine et al. (2004). Anthropogenic carbon storage in (a) and (b) is given for the GLODAP dataset area only, which does not cover coastal regions and several marginal seas, most notably the Arctic, the Caribbean, and the Mediterranean Sea. Excluded regions from the GLODAP area account for 7% and 10% of the total anthropogenic carbon storage in the CMIP5 models and the observation-based estimates, respectively (Table 2 in Frölicher et al. 2015). Note that this has no impact when comparing results for the Southern Ocean (south of 30°S). Observation-based estimates are normalized to year 1994. Weighted mean estimates of inversion-based anthropogenic air–sea CO_2 fluxes are shown in (c) and (d). Reprinted from Frölicher et al. (2015).

In strong contrast to the apparent overall agreement among CMIP5 models in terms of anthropogenic carbon uptake (Figure 2), the ecological response in these models is highly uncertain. As demonstrated in Bopp et al. (2013), CMIP5 models do a far better job at representing regional patterns in sea surface temperature than biogeochemical parameters such as surface pH, subsurface ocean oxygen, and net primary production (NPP). Laufkötter et al. (2015) compared differences with representations of surface chlorophyll, the most directly measurable, biogeochemically relevant variable from satellites, which illustrated vastly different spatial patterns and inter-model variance. Also compared were the field-based climatologies of surface nutrients in the form of nitrate, for which models clustered well, and silicate, for which models diverged. Anderson et al. (2015) further demonstrated vastly different patterns in surface dissolved organic matter distributions. While Bopp et al. (2013) demonstrated that some potential ecosystem stressors such as sea surface temperature and pH undergo robust patterns of change under projections of future climate change, they further illustrated that similarly ecologically critical stressors, such as NPP and subsurface oxygen, undergo dramatic changes on the order of 50% in both the positive and negative directions, leading to vast uncertainty in the overall multi-stressor response.

At first order, warming increases stratification in CMIP5 models such that ventilation and nutrient supply to the euphotic zone decreases, NPP decreases, and phytoplankton composition shifts toward smaller size classes and the microbial loop (Cabre et al. 2014). Second, these models broadly experience a poleward expansion and slow-down of subtropical gyres, leading to a shoaling nutricline in the subtropical gyres and enhanced nutrients, hypoxia, and acidification in some areas (Bopp et al. 2013, Cabre et al. 2014). An intensified hydrological cycle and warming reduces North Atlantic overturning, leading to a shoaling northern subpolar Atlantic and deepening tropics (Winton et al. 2013). Projections for the bottom-up drivers of NPP changes (i.e., temperature, light and particularly

nutrient limitation) show a wide range of responses (Figure 3). As a result, between both models and regions, different mechanisms are responsible for the NPP changes. Uncertainties in sea ice projections and future NO_3 limitation lead to disagreement on Arctic NPP response (Vancoppenolle et al. 2013). In the subtropical gyres, few models show decreases in phytoplankton growth due to lower nutrient availability. In the majority of models, phytoplankton growth increases due to warming, despite lower nutrient concentrations. However, temperature-driven intensification of grazing pressure decreases biomass in most models, resulting in net decreases in NPP in almost all models (Laufkötter et al. 2015). Overall, a changing balance of processes creates intense regional structure in projected change that currently shows little consensus among CMIP5 models. Moving forward on ocean biogeochemistry from CMIP5 CMIP5 has made a massive amount of model output available to data

variables were also available on a uniform grid. One such candidate uniform grid is that used for the World Ocean Atlas (WOA13; e.g. Locarnini et al. 2013). Finally, the full model documentation and parameter values are often difficult if not impossible to obtain. This severely limits the ability to both analyze internal mechanisms and compare with previous analyses. While the data restriction is logistically hard to overcome and requires expert decisions on the list of variables requested by the MIP, model documentation could be significantly improved by requiring that it include an updated list of parameter values for participation in the MIP.

With respect to individual science research, moving forward from CMIP5 will involve a multi-pronged approach of application of existing models, exploration of process representation for baseline fidelity and sensitivity, refined development towards increased comprehensiveness, and increasing resolution. Near-term priorities for application of these models include: Multi-member ensembles for detection and attribution, centennial-millennial scales, idealized sensitivity, diverse impacts application and assessment of potential for predictability and integration with seasonal-decadal climate prediction efforts, and exploring opportunities for experimental biogeochemistry prediction. Sensitivity priorities include physiological responses to temperature, acidification, oxygen, macro- and micro-nutrient limitation, and combined multi-stressor responses. Comprehensiveness priorities

Figure 3: Zonal mean of the relative change in (a) temperature, (b) light, and (c) nutrient limitation factors for nine marine biogeochemistry models. Relative change is calculated as the 2081–2100 average divided by the 2012–2031 average. Based on Figure 6 of Laufkötter et al. (2015).

analysts through an easily accessible online data portal. Common data formats units, model grid descriptions, and variable names assist in the comparison of key variables. However, while descriptive comparison of key biogeochemical variables is straightforward and well supported in CMIP5, analysis of the underlying mechanisms is met with three main challenges. First, limitations in storage capacity severely curtail analysis necessary to understand the drivers of ecosystem changes, requiring either liberal use of correlation as indicator for causation (e.g., Cabre et al. 2014), or extensive re-calculation efforts, in which numerical inaccuracies are often unavoidable (Laufkötter et al. 2015). Second, the availability of output only on non-uniform grids forces analysis to be much more difficult than it would be if output for conservatively remapped

include going beyond closing the CO_2 cycle to fully comprehensive and internally consistent representation of aerosol, Fe, CH_4 and N cycles, and ecosystems. Finally, the ever-present challenge of resolution must be addressed to capture key mechanisms in regional atmosphere-land interactions, currents, and the mesoscale ocean for improved base state, change, and human and marine applications.

With respect to community engagement, moving forward from CMIP5 will involve a complementary multi-pronged approach. Past discontinuities in research support have been highly debilitating for long-term science investments such as carbon cycle science. In the face of the seeming agreement between models of ocean anthropogenic carbon uptake, the ocean carbon and acidification
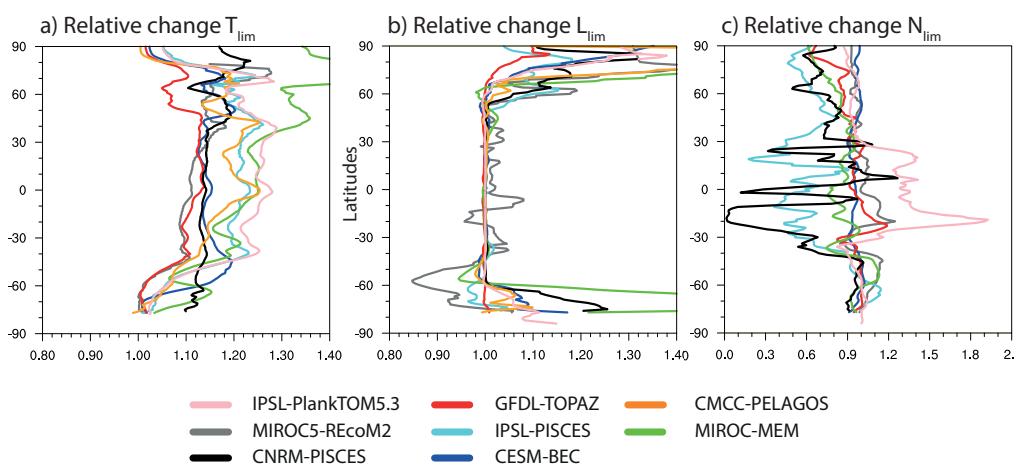
communities must better illustrate the remaining uncertainties requiring further research. For example, while CMIP5 models converged on ocean anthropogenic carbon uptake rates in general agreement with observational estimates, this consensus is limited to a set of very similar models with fixed biogeochemistry and parameterized mesoscale and submesoscale dynamics. There remain broad and critical gaps in this interdisciplinary and vigorous science field that will require long-term exploration of diverse modeling approaches for ocean circulation, physiological responsiveness, functional biodiversity, ecological thresholds, and other unresolved factors limiting robustness. The current generation of ocean carbon models must move past their currently simplistic, coarse, and similar parameterization to more adequately represent the spectrum of alternative approaches, future possible scenarios, and biodiversity in physiology and ecology. The relationship between ocean carbon uptake from a climate change feedback perspective and ocean acidification from an impacts perspective must remain vigilant and open across the broadening size and scope of the community. This will help to both maintain the high quality of carbon observations and analysis, while providing ever-more complex, species-specific, and site-specific information. Finally, as ocean carbon model analysis has grown into a broad analysis community, it must assure broad access and cooperation to maximize applicability of model intercomparison efforts to inform understanding of Earth system feedbacks and vulnerabilities.

References

Arora, V. K., G. J. Boer, P. Friedlingstein, M. Eby, C. D. Jones, J. R. Christian, G. Bonan, L. Bopp, V. Brovkin, P. Cadule, T. Hajima, T. Ilyina, K. Kindsay, J. F. Tjiputra, and T. Wu, 2013. Carbon–concentration and carbon–climate feedbacks in CMIP5 Earth system models. *J. Climate*, **26,** 5289-5314, doi: 10.1175/JCLI-D-12-00494.1.

Aumont, O. and L. Bopp, 2006. Globalizing results from ocean *in situ* iron fertilization studies. *Global Biogeochem. Cycles*, **20,** 1–15, doi:10.1029/2005GB002591.

Bacastow, R., and E. Maier-Reimer, 1990. Ocean-circulation model of the carbon cycle. *Climate Dyn.*, **4,** 95-125, doi: 10.1007/BF00208905.

Bopp, L., P. Monfray, O. Aumont, J.-L. Dufresne, H. Le Treut, G. Madec, L. Terray, and J. C. Orr, 2001. Potential impact of climate change on marine export production. *Global Biogeochem. Cycles*, **15,** 81–99, doi:10.1029/1999GB001256.

Bopp, L., L. Resplandy, J. C. Orr, S. C. Doney, J. P. Dunne, M. Gehlen, P. Halloran, C. Heinze, T. Ilyina, R. Séférian, J. Tjiputra, and M. Vichi, 2013. Multiple stressors of ocean ecosystems in the 21st century: projections with CMIP5 models. *Biogeosciences*, **10,** 6225-6245, doi:10.5194/bg-10-6225-2013.

Cabré, A., I. Marinov, and S. Leung, 2014. Consistent global responses of marine ecosystems to future climate change across the IPCC AR5 Earth System Models. *Climate Dyn.*, 1-28, doi: 10.1007/s00382-014-2374-3.

Doney, S.C., K. Lindsay, K. Caldeira, J.-M. Campin, H. Drange, J.-C. Dutay, M. Follows, Y. Gao, A. Gnanadesikan, N. Gruber, A. Ishida, F. Joos, G. Madec, E. Maier-Reimer, J.C. Marshall, R.J. Matear, P. Monfray, A. Mouchet, R. Najjar, J.C. Orr, G.-K. Plattner, J. Sarmiento, R. Schlitzer, R. Slater, I.J. Totterdell, M.-F. Weirig, Y. Yamanaka, and A. Yool, 2004. Evaluating global ocean carbon models: the importance of realistic physics. *Global Biogeochem. Cycles*, **18,** GB3017, doi:10.1029/2003GB002150.

Dufresne, J-L - Foujols, M. A. Denvil, , and 60 others, 2013. Climate change projections using the IPSL-CM5 Earth System Model: from CMIP3 to CMIP5. *Climate Dyn.* **40**: 2123-2165 doi:10.1007/s00382-012-1636-1.

Dunne, J. P., J. L Sarmiento, and A. Gnanadesikan, 2007. A synthesis of global particle export from the surface ocean and cycling through the ocean interior and on the seafloor. *Global Biogeochem. Cycles*, **21,** GB4006, doi:10.1029/2006GB002907.

Dunne, J. P., J. John, E. Shevliakova, R. J. Stouffer, J. P. Krasting, S. Malyshev, P. C. D. Milly, L. T. Sentman, A. Adcroft, W. F. Cooke, K. A. Dunne, S. M. Griffies, R. W. Hallberg, M. J. Harrison, H. Levy II, A. T. Wittenberg, P. Phillipps, and N. Zadeh, 2013. GFDL's ESM2 global coupled climate-carbon Earth System Models Part II: Carbon system formulation and baseline simulation characteristics. *J. Climate*, **26,** doi:10.1175/JCLI-D-12-00150.1.

Friedlingstein, P., P. Cox, R. Betts, L. Bopp, W. Von Bloh, V. Brovkin, P. Cadule, S. Doney, M. Eby, I. Fung, G. Bala, J. John, C. Jones, F. Joos, T. Kato, M. Kawamiya, W. Knorr, K. Lindsay, H. D. Matthews, T. Raddatz, P. Rayner, C. Reick, E. Roeckner, K.-G. Schnitzler, R. Schnur, K. Strassmann, A. J. Weaver, C. Yoshikawa, and N. Zeng, 2006. Climate-carbon cycle feedback analysis: Results from the C4MIP model intercomparison. J. *Climate,* 19, 3337-3353, doi:10.1175/JCLI3800.1.

Frölicher, T. L., J. L. Sarmiento, D. J. Paynter, J. P. Dunne, J. P. Krasting, and M. Winton, 2015. Dominance of the Southern Ocean in anthropogenic carbon and heat uptake in CMIP5 models. *J. Climate*, 28, 862–886, doi: 10.1175/JCLI-D-14-00117.1.

The HadGEM2 Development Team: Martin, G.M., Bellouin, N., Collins, W. J., Culverwell, I. D., Halloran, P. R., Hardiman, S. C., Hinton, T. J., Jones, C. D., McDonald, R. E., McLaren, A. J., O'Connor, F. M., Roberts, M. J., Rodriguez, J. M., Woodward, S., Best, M. J., Brooks, M. E., Brown, A. R., Butchart, N., Dearden, C., Derbyshire, S. H., Dharssi, I., Doutriaux-Boucher, M., Edwards, J. M., Falloon, P.

D., Gedney, N., Gray, L. J., Hewitt, H. T., Hobson, M., Huddleston, M. R., Hughes, J., Ineson, S., Ingram, W. J., James, P. M., Johns, T. C., Johnson, C. E., Jones, A., Jones, C. P., Joshi, M. M., Keen, A. B., Liddicoat, S., Lock, A. P., Maidens, A. V., Manners, J. C., Milton, S. F., Rae, J. G. L., Ridley, J. K., Sellar, A., Senior, C. A., Totterdell, I. J., Verhoef, A., Vidale, P. L. & Wiltshire, A., 2011. The HadGEM2 family of met office unified model climate configurations. *Geoscientific Model Development*, 4, 723-757, doi:10.5194/gmd-4-723-2011.

Houghton, J. T., G. J. Jenkins and J. J. Ephraums (Eds.), 1990. Climate Change: The IPCC Scientific Assessment. Cambridge University Press, Cambridge, Great Britain, New York, NY, USA and Melbourne, Australia 410 pp.

Ilyina T., K. D. Six, J. Segschneider, E. Maier-Reimer, H. Li, I. Núñez-Riboni, 2013. The global ocean biogeochemistry model HAMOCC: Model architecture and performance as component of the MPI-Earth System Model in different CMIP5 experimental realizations. *J. Adv. Model. Earth Sys.*, 5, 287-315 doi: 10.1002/jame.20017.

IPCC, 2013: Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change [Stocker, T.F., D. Qin, G.-K. Plattner, M. Tignor, S.K. Allen, J. Boschung, A. Nauels, Y. Xia, V. Bex and P.M. Midgley (Eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, 1535 pp, doi:10.1017/CBO9781107415324.

Laufkötter, C., M. Vogt, N. Gruber, M. Aita-Noguchi, O. Aumont, L. Bopp, E. Buitenhuis, S. C. Doney, J. Dunne, T. Hashioka, J. Hauck, T. Hirata, J. John, C. Le Quéré, I. D. Lima, H. Nakano, R. Seferian, I. Totterdell, M. Vichi, and C. Völker, 2015. Drivers and uncertainties of future global marine primary production in marine ecosystem model. *Biogeosciences Discuss.*, 12, 3731-3824, doi:10.5194/bgd-12-3731-2015.

Le Quéré, C., S.P. Harrison, I.C. Prentice, E.T. Buitenhuis, O. Aumont, L. Bopp, H. Claustre, L. Cortrim da Cunha, R. Geider, X. Giraud, C. Klaas, K.E. Kohfeld, L. Legendre, M. Manizza, T. Platt, R.B. Rivkin, S. Sathyendranath, J. Uitz, A.J. Watson and D. Wolf-Gladrow, 2005. Ecosystem dynamics based on plankton functional types for global ocean biogeochemistry models. *Global Change Biology*, 11, 2016-2040, doi:10.1111/j.1365-2486.2005.01004.x.

Locarnini, R. A., A. V. Mishonov, J. I. Antonov, T. P. Boyer, H. E. Garcia, O. K. Baranova, M. M. Zweng, C. R. Paver, J. R. Reagan, D. R. Johnson, M. Hamilton, and D. Seidov, 2013. World Ocean Atlas 2013, Volume 1: Temperature. S. Levitus, Ed., A. Mishonov Technical Ed.; NOAA Atlas NESDIS 73, 40 pp.

Matsumoto, K., J. L. Sarmiento, R. M. Key, O. Aumont, J. L. Bullister, K. Caldeira, J.-M. Campin, S. C. Doney, H. Drange, J.-C. Dutay, M. Follows, Y. Gao, A. Gnanadesikan, N. Gruber, A. Ishida, F. Joos, K. Lindsay, E. Maier-Reimer, J. C. Marshall, R. J. Matear, P. Monfray, A. Mouchet, R. Najjar, G.-K. Plattner, R. Schlitzer, R. Slater, P. S. Swathi, I. J. Totterdell, M.-F. Weirig, Y. Yamanaka, A. Yool, and J. C. Orr, 2004. Evaluation of ocean carbon cycle models with data-based metrics. *Geophys. Res. Lett.*, 31, L07303, doi:10.1029/2003GL018970.

Mikaloff Fletcher, S. E., N. Gruber, A. R. Jacobson, S. C. Doney, S. Dutkiewicz, M. Gerber, M. Follows, F. Joos, K. Kindsay, D. Menemenlis, A. Mouchet, S. a. Muller, J. L. Sarmiento, 2006: Inverse estimates of anthropogenic CO2 uptake, transport, and storage by the ocean. *Global Biogeochem. Cycles*, 20, GB2002, doi:10.1029/2005GB002530.

Millero., F. J., 1995: Thermodynamics of the carbon dioxide system in the oceans. *Geochimica et Cosmochimica Acta*, 59, 4, 661–677, doi:10.1016/0016-7037(94)00354-O.

Moore, J. K., K. Lindsay, S. C. Doney, M. C. Long, and K. Misumi, 2013. Marine ecosystem dynamics and biogeochemical cycling in the community Earth system model [CESM1(BGC)]: Comparison of the 1990s with the 2090s under the RCP4.5 and RCP8.5 scenarios. *J. Climate*, 26, 9291–9312, doi:10.1175/ JCLI-D-12-00566.1.

Orr, J. C., E. Maier-Reimer, U. Mikolajewicz, P. Monfray, J. L. Sarmiento, J. R. Toggweiler, N. K. Taylor, J. Palmer, N. Gruber, C. L. Sabine, C. Le Quéré, R. M. Key, and J. Boutin, 2001. Estimates of anthropogenic carbon uptake from four three-dimensional global ocean models. *Global Biogeochem. Cycles*, 15, 43-60, 2, doi: 10.1029/2000GB001273.

Resplandy, L., L. Bopp, J. C. Orr, and J. P. Dunne, 2013. Role of mode and intermediate waters in future ocean acidification: analysis of CMIP5 models. *Geophy. Res. Lett.*, 40, 3091-3095, doi:10.1002/grl.50414.

Sabine, C. L., R. A. Feely, N. Gruber, R. M. Key, K. Lee, J. L. Bullister,1 R. Wanninkhof, C. S. Wong, D. W. R. Wallace, B. Tilbrook, F. J. Millero, T.-H. Peng, A. Kozyr, T. Ono, and A. F. Rios, 2004. The oceanic sink for anthropogenic CO2. *Science*, 305, 367-371, doi: 10.1126/science.1097403.

Sailley, Severine, Vogt, M., S. Doney, M. Aita, L. Bopp, E. Buitenhuis, T. Hashioka, I. Lima, C. Le Quere, and Y. Yamanaka, 2013. Comparing food web structures and dynamics across a suite of global marine ecosystem models, *Ecologi. Model.*, 261-262, 43–57, doi:10.1016/j.ecolmodel.2013.04.006.

Sarmiento, J. L., R. D. Slater, M. J. R. Fasham, H. W. Ducklow, J. R. Toggweiler, and G. T. Evans, 1993. A seasonal three-dimensional ecosystem model of nitrogen cycling in the North Atlantic Euphotic Zone. *Global Biogeochem. Cycles*, 7, 417-450, doi: 10.1029/93GB00375.

Sarmiento, J. L., and C. Le Quéré, 1996. Oceanic carbon dioxide uptake in a model of century-scale global warming. *Science*, 274, 5291, 1346-1350, doi:10.1126/science.274.5291.1346

Schmidt, G. A., et al. 2014. Configuration and assessment of the GISS ModelE2 contributions to the CMIP5 archive. *J. Adv. Model. Earth Syst.*, 6, 141–184, doi:10.1002/2013MS000265.

Shuter B. J., 1978: Size dependence of phosphorus and nitrogen subsistence quotas in unicellular microorganisms. *Limnol. Oceanogr.* 23, 1248–55, doi: 10.4319/lo.1978.23.6.1248.

Siegenthaler, U., and J. L. Sarmiento, 1993. Atmospheric carbon dioxide and the ocean. *Nature*, 365, 119-125, doi: 10.1038/365119a0.

Taylor, K. E., R. J. Stouffer, and G. A. Meehl, 2012. An overview of CMIP5 and the experiment design. Bull.Amer. Meteor. Soc., 93, 485–498, doi: 10.1175/BAMS-D-11-00094.1.

Vichi, M., N. Pinardi, and S. Masina, 2007. A generalized model of pelagic biogeochemistry for the global ocean ecosystem. Part I: Theory. *J. Marine Syst.*, 64, 89–109, doi:10.1016/j.jmarsys.2006.03.006.

Wanninkhof, R., 1992. Relationship between wind speed and gas exchange over the ocean. *J. Geophy. Res.,* 97, 7373-7382, doi:10.1029/92JC00188.

Winton, M., S. M. Griffies, B. Samuels, J. L. Sarmiento, T. L. Frölicher, 2013. Connecting changing ocean circulation with changing climate. *J. Climate*, 26, 2268-2278, doi: 10.1175/JCLI-D-12-00296.1.

Ocean heat and carbon uptake in transient climate change: Identifying model uncertainty

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Global warming on decadal and centennial timescales is mediated and ameliorated by the ocean sequestering heat and carbon into its interior. Transient climate change is a function of the efficiency by which anthropogenic heat and carbon are transported away from the surface into the ocean interior (Hansen et al. 1985). Gregory and Mitchell (1997) and Raper et al. (2002) were the first to identify the importance of the 'ocean heat uptake efficiency' in transient climate change. Observational estimates (Schwartz 2012) and inferences from coupled atmosphere-ocean general circulation models (AOGCMs; Gregory and Forster 2008; Marotzke et al. 2015), suggest that ocean heat uptake efficiency on decadal timescales lies in the range 0.5-1.5 W m^{-2} K^{-1} and is thus comparable to the climate feedback parameter (Murphy et al. 2009). Moreover, the ocean not only plays a key role in setting the timing of warming but also its regional patterns (Marshall et al. 2014), which is crucial to our understanding of regional climate, carbon and heat uptake, and sea-level change.

This short communication is based on a presentation given by A. Romanou at a recent workshop, *Ocean's Carbon and Heat Uptake: Uncertainties and Metrics*, co-hosted by US CLIVAR and OCB. As briefly reviewed below, we have incomplete but growing knowledge of how ocean models used in climate change projections sequester heat and carbon into the interior. To understand and thence reduce errors and biases in the ocean component of coupled models, as well as elucidate the key mechanisms at work, in the final section we outline a proposed model intercomparison project named FAFMIP. In FAFMIP, coupled integrations would be carried out with prescribed "overrides" of wind stress and freshwater and heat fluxes acting at the sea surface.

Ocean's role in shaping the patterns and timing of temperature response in a warming world

Mechanisms of ocean heat uptake

What ocean processes control the efficiency of ocean heat uptake? Mixing (across and along isopycnal surfaces) was identified by Sokolov et al. (2003), who also found that this "effective diffusion" varies significantly with latitude, as being somewhat small in the tropics but fifty-fold larger at high latitudes. Huang et al. (2003) showed that heat penetration to the deep ocean could be mediated by changes in convection and eddy stirring. On the other hand, Knutti et al. (2008) did not detect notable sensitivity of ocean heat uptake to the rate of diffusive mixing in their model. In a study of many CMIP5 models, Kostov et al. (2014) showed that the modeled Atlantic meridional overturning circulation (AMOC) plays a large role in transient ocean heat uptake through its control of deep ocean ventilation. They found (see Figures 1a and b) that the AMOC depth sets the depth to which heat is sequestered, and hence the effective heat capacity of the ocean in transient climate change, and that the strength of the AMOC influences the sequestration rates. Therefore, the spread in heat uptake across the models could be largely explained by differences in their AMOC properties. The importance of the AMOC (Figure 1c) is perhaps to be expected, given that 50% of the net heat uptake in the global ocean occurs in the Atlantic north of 35°N. Distinguishing different oceanic processes, Exarchou et al. (2015) showed from global diagnostics of a suite of climate models that diapycnal diffusion (below the mixed layer) is the least important process in controlling heat uptake, as compared to mixed layer physics and convection and advection by mean circulation.
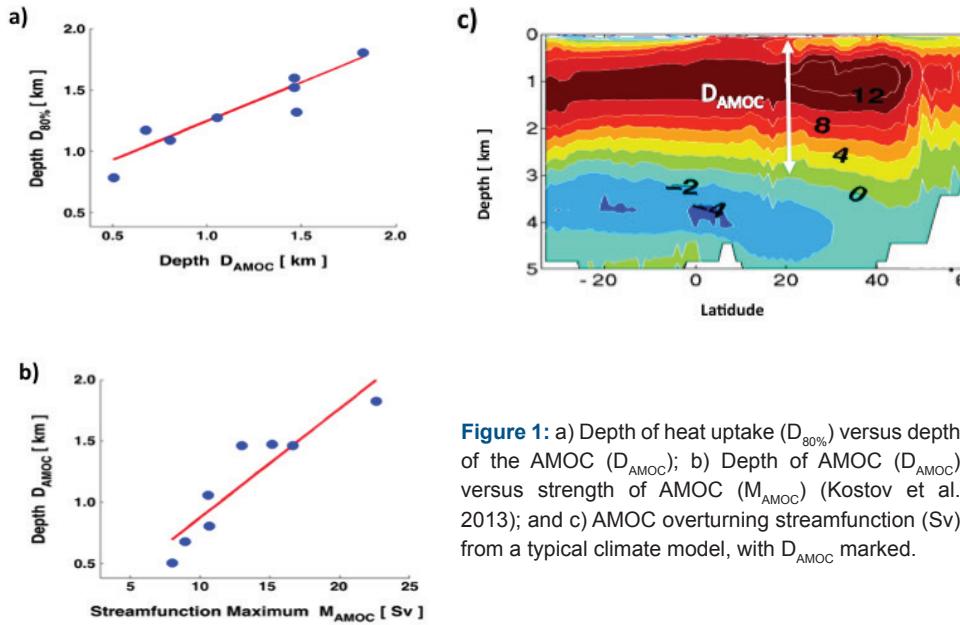
Figure 1: a) Depth of heat uptake ($D_{80\%}$) versus depth of the AMOC (D_{AMOC}); b) Depth of AMOC (D_{AMOC}) versus strength of AMOC (M_{AMOC}) (Kostov et al. 2013); and c) AMOC overturning streamfunction (Sv) from a typical climate model, with D_{AMOC} marked.

order, not due to regional feedbacks since they are kept constant and uniform in our override experiments. That said, Armour et al. (2013) and Rose et al. (2014) emphasize the importance of regional atmospheric climate feedbacks in setting the time-evolving pattern of surface warming and ocean heat uptake.

Transient CO_2 and tracer uptake

The ocean also plays an important role in CO_2 uptake, reducing the airborne greenhouse gas concentrations and thus the rate of atmospheric warming. It is not yet clear how the ocean sink of anthropogenic CO_2 will change in a warming world (Le Quéré et a.l 2009; Gloor et al. 2010). Observations indicate that the outgassing of natural CO_2 from the interior ocean has increased in the last few decades, particularly in the Southern Ocean, offsetting the anthropogenic sink. Some studies argue that this may be linked to an increase in the westerly winds blowing over the Southern Ocean, whereas other studies question whether increased outgassing is occurring. The net (natural +
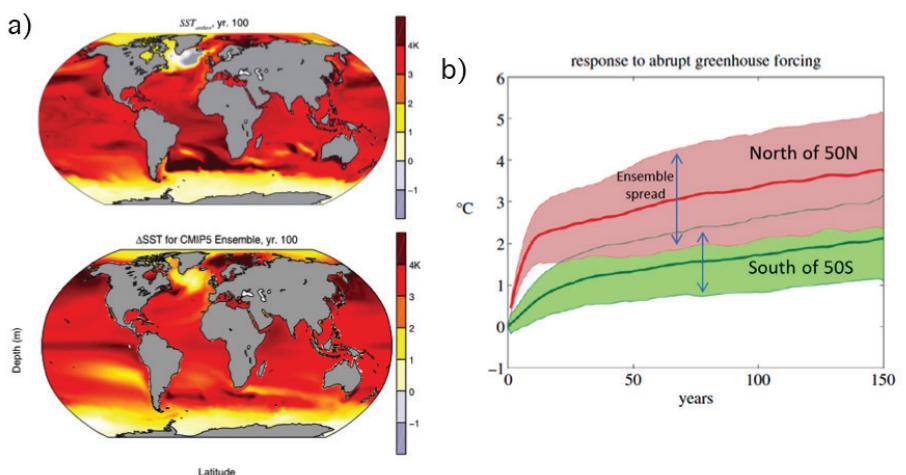
Spatial patterns and timing of SST anomalies

Marshall et al. (2014 a,b) employ a stand-alone ocean model run under Coordinated Ocean-ice Reference Experiment (CORE) forcing (Griffies et al. 2009) to study how ocean circulation shapes patterns of SST response in a warming world. They carry out "override" experiments, in which SST evolves in response to air-sea fluxes given by CORE, but augmented by a spatially uniform, constant-in-time downwelling radiative flux. Climate feedbacks are parameterized through an SST damping term at a rate that is constant in space and time. This setup, although highly idealized, is useful in investigating the role of the ocean in setting the patterns and timescales of the transient climate response. Despite the idealized model framework, both Arctic amplification and delayed warming signals in the North Atlantic and around Antarctica are captured, and in common with CMIP5 climate change experiments with complex coupled models (note the marked similarity between Figure 2a, from the override experiment, with Figure 2b from an ensemble of coupled CMIP5 models). We conclude that these patterns can largely be attributed to ocean rather than atmospheric processes. Similarly, the regional climate response is, to the first

Figure 2: a) (top) SST perturbation (SST_{anthro}) from a 100-year run of a stand-alone ocean with specified, spatially uniform downwelling radiation and a linear damping of SST at the sea surface (from Marshall et al. 2014a); (bottom) SST change after 100 years from CMIP5 model runs of $4\times CO_2$ forcing; b) SST conditional random fields for greenhouse gas emissions forcing computed from an ensemble of 15 CMIP5 models under quadrupling of CO_2. The Arctic is defined as north of 50° N (in red) and the Antarctic between 50° S and 70° S (in green). Thick lines denote the ensemble mean and the shaded area spans 1 s.d. (from Marshall et al. 2014b).

anthropogenic) CO_2 flux depends on the strength of the wind, upwelling, and the mixed-layer cycle of carbon and nutrients, and is thus directly related to ocean dynamics. Indeed, uptake of CO_2 in models varies substantially, mostly due to differences in physical parameterizations (structural uncertainty), increasing the uncertainty of future climate projections (Krasting et al. 2014). To address structural uncertainty, tracer uptake experiments, both realistic (CFC, SF_6, etc.) and idealized (ventilation-tracer, ocean age, and passive temperature-like tracers as in Marshall et al. 2014), can be used to highlight heat and carbon uptake processes. Figure 3, for example, shows a ventilation tracer set equal to one at the surface of the subpolar North Atlantic Ocean and subsequently integrated forward in time. The experiments only differ in the strength of the AMOC. We find that as the depth and strength of the AMOC grow, additional tracer is sequestered to greater depths (Romanou et al. *in prep*). Therefore, the AMOC controls not only the rate and depth of heat uptake, but also that of many tracers, including anthropogenic CO_2.

Proposed Flux-Anomaly-Forced Model Intercomparison Project (FAFMIP)

A coordinated model intercomparison project could provide very useful information about how the ocean component of coupled models contributes to uncertainty in climate change projections. A focus might be regional sea-level change, coupled with global and regional SST patterns, heat and carbon uptake, AMOC change, etc. Knowledge of which ocean processes and phenomena have a large model spread may help us evaluate and refine our models. Ideally, one might couple the same atmosphere to different ocean models, but this would be difficult to organize. Alternatively, one could parameterize atmospheric climate feedbacks with a simple parameter and run ocean-only models (as in Marshall et al. 2014), but this would fail to capture the richness and the regional detail of the feedbacks. A viable way forward, we think, is to use existing coupled control runs and add air-sea flux "overrides" - i.e., wind stress, evaporation-precipitation, heat fluxes – chosen to be representative of those induced by climate change.

Such experiments are proposed within the Flux-Anomaly-Forced Model Intercomparison Project (FAFMIP, http://www.met.reading. ac.uk/~jonathan/FAFMIP/). Each modeling group would adopt the same protocol and run experiments ascribing the same override fields, computed from ensembles of CMIP5 models perturbed by climate change. We would then attempt to assess the spread in the resulting AMOC, heat and carbon uptake, and patterns of sea-level change, both regionally and globally, and identify their causes. The community has some familiarity already with override experiments – e.g., freshwater forcing (Stouffer et al. 2006); wind forcing (Gent and Danabasoglu 2011); or both heat and freshwater forcing experiments (Zhang and Vallis 2013). Due to the dominance of heat flux-SST feedbacks, it is not yet clear how to carry out meaningful heat flux override experiments. This is currently under study (http://www.met.reading.ac.uk/~jonathan/FAFMIP/FAFMIP_method_heat.pdf).
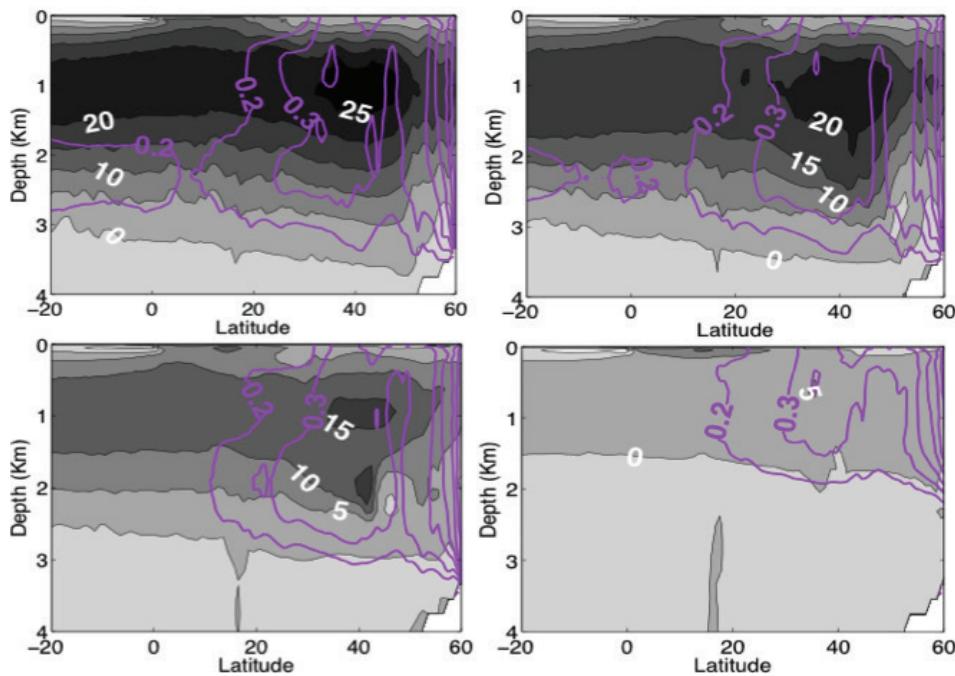

Figure 3: Zonally averaged section showing (purple contours) ventilation tracer concentration (from a stand-alone NASA GISS ocean run driven with CORE-1 forcing. The AMOC overturning stream-function (Sv) is also plotted in gray shading with white labels.

References

Armour, K. C., C. M. Bitz, and G. H. Roe, 2013. Time-varying climate sensitivity from regional feedbacks. *J. Climate*, **26,** 4518-4534, doi: 10.1175/JCLI-D-12-00544.1.

Exarchou, E., T. Kuhlbrodt, J. M. Gregory, and R. S. Smith, 2014. Ocean heat uptake processes: A model intercomparison. *J. Climate,* **28**, 887–908, doi: 10.1175/JCLI-D-14-00235.1.

Gent, P. R. and G. Danabasoglu, 2011. Response to increasing Southern Hemisphere winds in CCSM4. *J. Climate*, **24,** 4992-4998, doi: 10.1175/JCLI-D-10-05011.1.

Gloor M., J. L. Sarmiento, and N. Gruber, 2010. What can be learned about carbon cycle climate feedbacks from the CO_2 airborne fraction? *Atmos. Chem. Phys.*, **10,** 7739-7751, doi:10.5194/acp-10-7739-2010.

Gregory, J. M., and P. M. Forster, 2008. Transient climate response estimated from radiative forcing and observed temperature change. *J. Geophys. Res.*, **113**, D23105, doi:10.1029/2008JD010405.

Gregory J. M. and J. F. B. Mitchell, 1997. The climate response to CO_2 of the Hadley Centre coupled AOGCM with and without flux adjustment. *Geophys. Res. Lett.*, **24**, 1943-1946, doi: 10.1029/97GL01930.

Griffies, S. M., et al. 2009. Coordinated Ocean-ice Reference Experiments (COREs). *Ocean Model.*, **26**, 1–46, doi: 10.1016/j.ocemod.2008.08.007.

Hansen, J., G. Russell, A. Lacis, I. Fung, D. Rind, and P. Stone, 1985,. Climate response times: Dependence on climate sensitivity and ocean mixing. *Science*, **229**, 857–859, doi:10.1126/science.229.4716.857.

Huang, B., P. H. Stone, and C. Hill, 2003. Sensitivities of deep-ocean heat uptake and heat content to surface fluxes and subgrid-scale parameters in an ocean general circulation model with idealized geometry. *J. Geophys. Res.*, **108**, 1978-2012, doi:10.1029/2001JC001218.

Knutti, R. and Hegerl, G. C., 2008. The equilibrium sensitivity of the Earth's temperature to radiation changes. *Nature Geosci.*, **1**, 735–743, doi: 10.1038/ngeo337.

Kostov, Y., K. C. Armour, and J. Marshall, 2014. Impact of the Atlantic meridional overturning circulation on ocean heat storage and transient climate change. *Geophys. Res. Lett.,* **41**, 2108–2116, doi:10.1002/2013GL058998.

Krasting, J. P., J. P. Dunne, E. Shevliakova, and R. J. Stouffer, 2014. Trajectory sensitivity of the transient climate response to cumulative carbon emissions. *Geophys. Res. Lett.*, **41**, 2520–2527, doi:10.1002/2013GL059141.

Le Quéré, C, et al., 2009. Trends in the sources and sinks of carbon dioxide. *Nature Geosci.*, **2**, 831-836, doi:10.1038/ngeo689.

Marshall, J., J. R.Scott, K. C. Armour, J.-M. Campin, M. Kelley, A. Romanou, 2014. The ocean's role in the transient response of climate to abrupt greenhouse gas forcing. *Climate Dyn.*, **44,** 2287-2299, doi:10.1007/s00382-014-2308-0.

Marshall, J., K. C. Armour, J. R. Scott, Y. Kostov, U. Hausmann, D. Ferraiera, T. G. Shepherd, and C. M. Bitz, 2014. The ocean's role in polar climate change: Asymmetric Arctic and Antarctic responses to greenhouse gas and ozone forcing. *Phil. Trans. Royal Soc. 372*, doi: 10.1098/rsta.2013.0040.

Marotzke, J. and P. M. Forster, 2015. Forcing, feedback and internal variability in global temperature trends. *Nature*, **517**, 565–570.

Murphy, D. M, S. Solomon, R. W. Portmann, K. H. Rosenlof, P. M. Forster, and T. Wong, 2009. An observationally based energy balance for the Earth since 1950. *J. Geophys. Res.*, **114**, D17107, doi:10.1029/2009JD012105.

Raper, S. C. B., J. M. Gregory, and R. J. Stouffer, 2002. The role of climate sensitivity and ocean heat uptake on AOGCM transient temperature response. *J. Clim*ate, **15**, 124–130, doi: 10.1175/1520-0442(2002)015<0124:TROCSA>2.0.CO;2.

Rose, B. E. J., K. C. Armour, D. S. Battisti, N. Feldl, and D. D. B. Koll, 2014. The dependence of transient climate sensitivity and radiative feed- backs on the spatial pattern of ocean heat uptake. *Geophys. Res. Lett.,* **41**, doi:10.1002/2013GL058955.

Schwartz, S. E., 2012. Determination of Earth's transient and equilibrium climate sensitivities from observations over the twentieth century: Strong dependence on assumed forcing. *Surveys Geophys.*, **33**, 745–777, doi: 10.1007/s10712-012-9180-4.

Sokolov, A., C. Forest, and P. Stone, 2003. Comparing oceanic heat uptake in AOGCM transient climate change experiments. *J. Climate*, **16,** 1573-1582, doi: 10.1175/1520-0442-16.10.1573.

Stouffer, R. J. et al., 2006. Investigating the causes of the response of the thermohaline circulation to past and future climate changes. *J. Climate*, **19**, 1-23, doi: 10.1175/JCLI3689.1.

Zhang Y. and G. K. Vallis, 2013. Ocean heat uptake in eddying and non-eddying ccean circulation models in a warming climate. *J. Phys. Oceanogr.*, **43,** 2211-2229, doi: 10.1175/JPO-D-12-078.1

Are anthropogenic changes in the tropical ocean carbon cycle masked by Pacific Decadal Variability?

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Observed changes in the tropical Pacific carbon cycle

The tropical Pacific is the ocean's largest natural source of CO_2 to the atmosphere, thus playing a key role in the global carbon cycle (Takahashi et al. 2009; Gruber et al. 2009). Strong equatorial upwelling of carbon-rich thermocline waters causes partial pressure of CO2 (pCO_2) in the surface ocean to exceed that in the atmosphere. This $pCO2$ difference, surface ocean pCO_2 minus the atmospheric $pCO2$ (ΔpCO_2), drives outgassing of CO_2 into the atmosphere. Additional factors such as wind speed, and to a lesser degree salinity and temperature, modulate the CO_2 flux at the sea-air interface. Anthropogenic emissions continue to drive increasing atmospheric CO_2 concentrations. Understanding and predicting how sea-air CO_2 fluxes respond to this change is a major challenge in carbon cycle research. Observational evidence on the mechanisms driving changes in outgassing over the equatorial Pacific is inconclusive. Studies using the near-continuous observational record of ocean pCO_2 in the central equatorial Pacific show that since 1980, ocean pCO_2 has risen at about the same rate as atmospheric pCO_2 (Feely et al. 2006; Fay and McKinley 2013). This near-zero trend in ΔpCO_2 implies a near-zero trend in sea-air CO_2 flux. The sea-air CO_2 flux, however, has increased in this region, mainly driven by increases in wind speed (Feely et al. 2006).

Anthropogenic response

The Earth System Models (ESMs) participating in the 5th phase of the Coupled Model Intercomparison Project (CMIP5) and a 28-member ensemble of simulations conducted with the Community Earth System Model (CESM) show a robust decrease in ΔpCO_2 and sea-air CO_2 flux in the equatorial Pacific over the 50-year period of 2030 to 2079 as atmospheric CO_2 concentration rises (Figure 1). The following two mechanisms could explain this response: (1) Water in the equatorial thermocline is mostly isolated from the anthropogenic CO_2 perturbation in the atmosphere. When this water upwells to the surface, it is exposed to an atmosphere with ever-increasing CO_2 concentration, resulting in a negative trend in
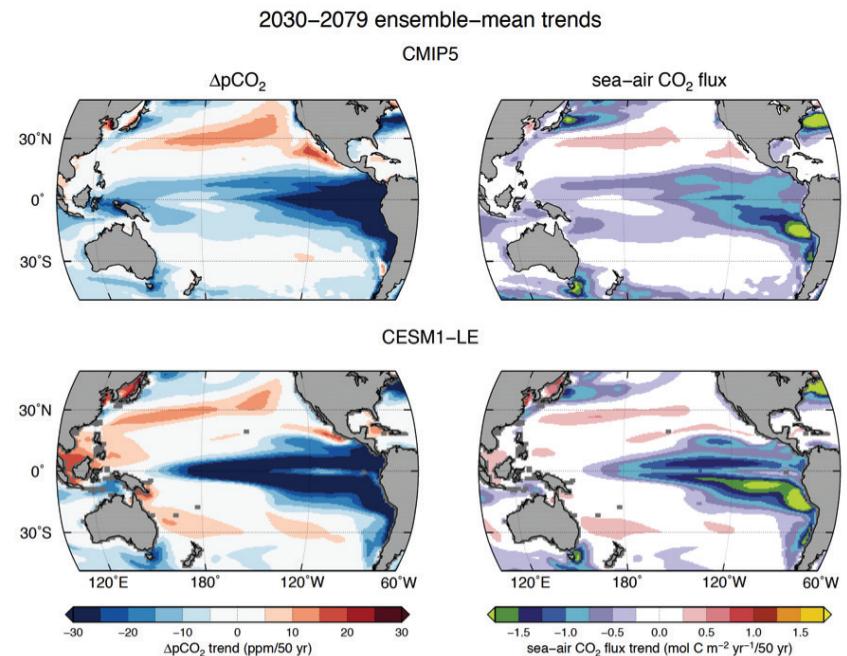

Figure 1: Ensemble-mean trends (2030-2079) in ΔpCO_2 (left; ppm/50 yr) and sea-air CO_2 flux (right; mol C m^{-2} yr^{-1}/50 yr) simulated by CMIP5 models (top) and the CESM1-LE (bottom). Positive sea-air CO_2 flux indicates increased outgassing.

ΔpCO$_2$ and sea-air CO$_2$ flux (Maier-Reimer and Hasselmann 1987). (2) Models also project reduced upwelling due to weaker equatorial easterly winds associated with a reduced Walker circulation in response to global warming (Vecchi and Soden 2007; DiNezio et al. 2009), which could drive decreases in ΔpCO$_2$ and sea-air CO$_2$ flux.

Impact of decadal climate variability

The fact that ΔpCO$_2$ has remained steady over the observation-rich historical period (1980-present) is inconsistent with the consensus among ESMs. Can these differences be reconciled? It is well known that climate variability associated with El Niño/Southern Oscillation (ENSO) can complicate the detection of anthropogenic changes (McKinley et al. 2004; Feely et al. 2006; Sutton et al. 2014). However, the effect of decadal variability has not been explored because the observational record is too short to span more than one realization of Pacific Decadal Variability (PDV) for a robust assessment.

The post-1980 period was characterized by a multi-decadal strengthening of the Pacific trade winds and an acceleration of the shallow overturning circulation and equatorial upwelling (McPhaden and Zhang 2004; Merrifield and Maltrud 2011). We hypothesize that during this period, stronger upwelling driven by strengthened trade winds led to increases in ΔpCO$_2$ and sea-air CO$_2$ flux that counteracted the decreases expected from the anthropogenic perturbation of atmospheric CO$_2$ concentration. Here, we test this hypothesis using an ensemble of simulations performed with CESM1, an ESM that simulates a realistic mean tropical carbon cycle as well as its seasonal and interannual variability (Long et al. 2013). The large number of realizations (28; hereafter referred to as the CESM1-LE; Kay et al. 2015) allows separation of internal decadal variability and externally forced changes.

For each realization of the CESM1-LE, we estimate the changes in both climate and biogeochemistry by computing linear trends over the period 1980-2014 when continuous observations of pCO$_2$ in the equatorial Pacific are available. We also focus on the central tropical Pacific defined by a modified

Niño-3.4m box (170°E-130°W, 5°S-5°N). The 28 simulations of the CESM1-LE show ΔpCO$_2$ changes ranging from -12.6 ppm to +5.6 ppm, suggesting that multi-decadal climate variability has a sizable impact during this 35-year period. The ensemble-mean (forced) change is -6.2 ppm, consistent in sign with the response of ΔpCO$_2$ to anthropogenic increases in atmospheric CO$_2$ discussed above. We extract two 9-member sub-ensembles, grouped according to the lower and upper terciles of the ΔpCO2 trends over the Nino-3.4m box. The lower-tercile sub-ensemble shows a pronounced decrease in ΔpCO$_2$ over the tropical Pacific and associated
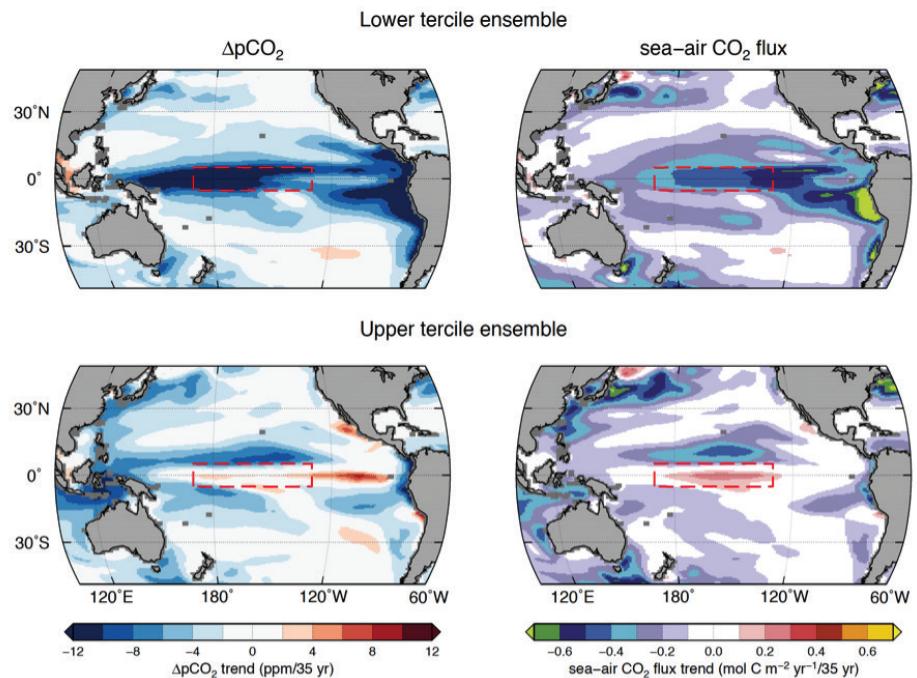

Figure 2: Trends (1980-2014) in ΔpCO$_2$ (left; ppm/35 yr) and sea-air CO$_2$ flux (right; mol C m^{-2} yr^{-1}/35 yr) simulated by the CESM1-LE, over grouped according to the lower and upper terciles of the pCO$_2$ trends over the Nino-3.4m region. The lower tercile ensemble (top) contains 9 simulations with the most negative Nino-3.4m ΔpCO$_2$ trends. The upper tercile ensemble (bottom) contains 9 simulations with negligible trends. The red box over the central equatorial Pacific indicates the Nino-3.4m region. Positive sea-air CO$_2$ flux indicates increased outgassing.

reduction in outgassing (Figure 2 top), while the upper-tercile sub-ensemble shows negligible changes in ΔpCO$_2$ over the equatorial Pacific, and a slight increase in CO$_2$ sea-air flux (Figure 2 bottom). Moreover, the former shows climate anomalies consistent with the positive phase of PDV (Figure 3 top), while the latter shows climate anomalies consistent with its negative phase (Figure 3 bottom). This suggests that wind-driven changes in equatorial upwelling
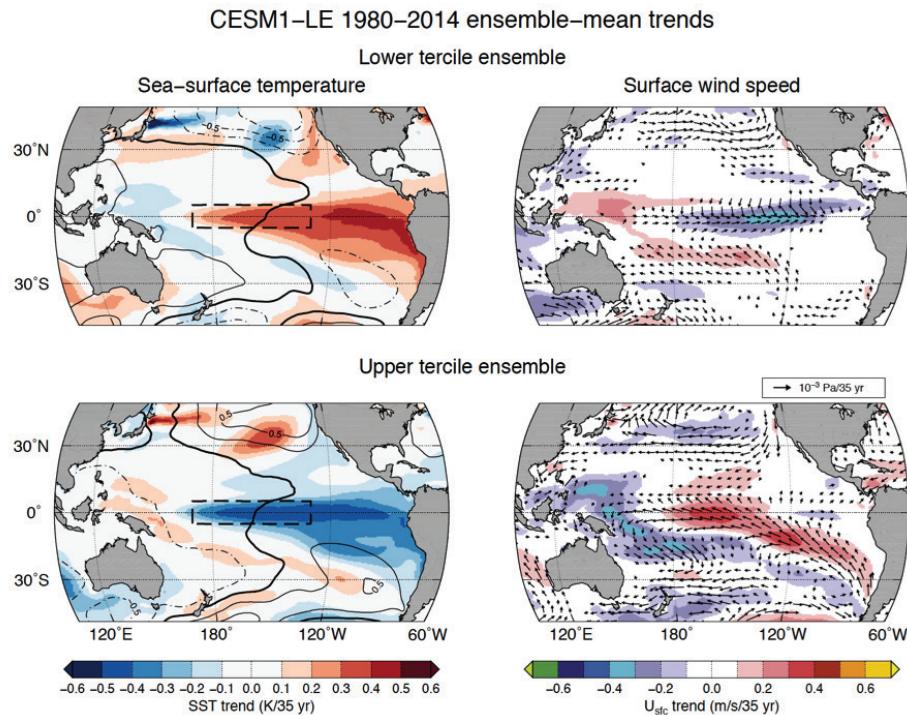
Figure 3: As in Figure 2, but for (left) sea-surface temperature (colors; K/35 yr) and sea-level pressure (contours) and (right) surface wind stress (vectors; 10-3 Pa/35 yr) and wind speed (colors; m s^{-1}/35 yr).

decadal changes in Pacific climate. Therefore, the steady $\Delta p CO_2$ trend seen in observations (Feely et al. 2006; Fay and McKinley 2013) could be indicative of an anthropogenic response; otherwise, $\Delta p CO_2$ should be increasing following the observed multi-decadal acceleration of the tropical circulation (McPhaden and Zhang 2004; Merrifield and Maltrud 2011). Furthermore, we cannot reject the model projections of decreasing tropical Pacific outgassing in response to increasing atmospheric CO_2. The anthropogenic response could be masked by decadal variability in Pacific climate.

We expect that these ideas will stimulate further efforts to reconcile observations and model projections. A next step is a full attribution of the effects of natural and anthropogenic influences on the tropical Pacific carbon cycle. How much of the observed $\Delta p CO_2$ change is anthropogenic, and how much is driven by the strengthening of the Pacific Ocean circulation? Could

associated with the positive and negative phases of PDV could have a considerable influence on trends in the tropical Pacific carbon cycle.

Towards detection of anthropogenic changes

The CESM1-LE shows a strong reduction in $\Delta p CO_2$ and outgassing when the PDV is trending positive (constructive effects of PDV and anthropogenic forcing). Conversely, it shows negligible changes in $\Delta p CO_2$ and a slight increase in outgassing when the PDV is trending negative (cancelling effects of PDV and anthropogenic forcing). The latter case is analogous to the changes observed during 1980-2014, when the Pacific Ocean has trended toward a negative PDV phase, characterized by stronger trade winds and stronger upwelling.

Within the context of the CESM1-LE, internally driven and forced trends can have similar magnitudes, suggesting that PDV can overwhelm the forced response in particular ensemble members. Translating this result to nature implies that equatorial outgassing could be already diminishing in response to increasing atmospheric CO_2. However, this signal has not emerged from the background of internal variability, particularly due to the ongoing multi-

observations be used to determine whether the carbon content of upwelled waters is increasing more slowly than atmospheric CO_2, as proposed by Maier-Reimer and Hasselmann (1987)? Will the reduction in outgassing vanish once the tropical thermocline fully equilibrates with the atmospheric CO_2? Answering these questions requires process-based understanding of the observed and simulated changes and would ultimately lead to reduced uncertainty in model projections (Friedlingstein et al. 2014).

Changes in the CO_2 sources and sinks are highly uncertain, and they could have a significant influence on future atmospheric CO_2 levels (Le Quéré et al. 2009). It is therefore crucial to reduce these uncertainties. For instance, a recent trend in the airborne fraction of the total emissions suggests that the growth in uptake rate of CO_2 sinks is not keeping up with the increase in CO_2 emissions (Canadell et al. 2007; Le Quéré et al. 2009). For how long will the ocean continue to increase its CO_2 uptake? A more complete understanding of the role played by the tropical Pacific in the global carbon cycle is critical to answering these important questions.

Methods

Earth System Models (ESMs) simulate coupled interactions among the atmosphere, ocean, land, as well as ocean ecosystems and chemistry, and the ocean and terrestrial carbon cycle. We use output from two different types of ESM ensembles, each of which addresses a key source of uncertainty. The first is a multi-model ensemble of simulations of 21st Century climate and biogeochemistry (BGC) change coordinated by CMIP5 and performed with 11 ESMs ran under the same external forcings defined by the RCP8.5 scenario. The models are: CESM1-BGC, MPI-ESM-LR, MPI-ESM-MR, HadGEM2-CC, HadGEM2-ES, IPSL-CM5A-MR, IPSL-CM5B-LR, IPSL-CM5A-LR, MIROC-ESM, GFDL-ES-M2G, GFDL-ESM2M. This ensemble was specifically designed to explore the effect of model (structural) uncertainty, although they also contain uncertainty due to internal variability. We use this ensemble to explore the robustness of the anthropogenic response. We focused on the period 2030-2079 because this is when the forced response of the global ocean carbon cycle is more pronounced.

The second ensemble consists of 28 simulations performed with one single model, in this case the Community Earth System Model Version 1 (CESM1). All the simulations in this large ensemble (CESM1-LE) were started at year 1920 and run under historical forcings until year 2005 and under RCP8.5 scenario from year 2006 to year 2100. A small random perturbation was applied to each simulation in the initial air temperature at year 1920, which causes them to simulate independent weather and internal climate variability. All 28 simulations, however, have the same anthropogenic response because of the common forcing. Thus this large initial-condition ensemble is ideally suited to study the interplay between anthropogenic changes and natural climate variability (Kay et al. 2015). The 28-member ensemble analyzed here is made of 24 simulations with BGC from the 30-member CESM1-LE presented in Kay et al (2015) plus 4 additional simulations following the same experimental protocol.

For each simulation of the CESM1-LE we estimate the changes in climate and BGC by computing linear trends over the period 1980-2014. We focus on this period because it corresponds to when there are continuous observations of pCO$_2$ over the equatorial Pacific. We average the trends over a modified Nino-3.4 region (hereafter "Nino-3.4m": 170°E-130°W 5°S-5°N) for two reasons: 1) this is where the observational network is densest and 2) this is where CESM1 exhibits the strongest forced ΔpCO$_2$ change (Figure 1 bottom). This box is zonally wider than the conventional definition in order to capture the full spatial pattern of the forced response. During 1980-2014 the magnitude of the simulated Nino-3.4m ΔpCO$_2$ changes range from -12.6 to 5.6 ppm, suggesting a large influence of natural variability. The ensemble-mean change is -6.2 ppm and the median change is -7.5 ppm consistent in sign with the anthropogenic reduction discussed for the 2030-2079 period.

References

Canadell, J. G. C. Le Quéré, M. R. Raupach, C. B. Field, E. T. Buitenhuis, P. Ciais, T. J. Conway, N. P. Gillett, R. A. Houghton, and G. Marland, 2007. Contributions to accelerating atmospheric CO$_2$ growth from economic activity, carbon intensity, and efficiency of natural sinks. *Proc. Natl. Acad. Sci.*, **104**, 18866–18870, doi:10.1073/pnas.0702737104.

DiNezio, P. N., A. C. Clement, G. A. Vecchi, B. J. Soden, B. J. Kirtman, and S.-K. Lee, 2009. Climate response of the equatorial Pacific to global warming. *J. Climate*, **22**, 4873–4892, doi: 10.1175/2009JCLI2982.1.

Fay, A. R., and G. A. McKinley, 2013. Global trends in surface ocean pCO$_2$ from in situ data. *Global Biogeochem. Cycles*, **27**, 541–557, doi:10.1002/gbc.20051.

Feely, R. A., T. Takahashi, R. Wanninkhof, M. J. McPhaden, C. E. Cosca, S. C. Sutherland, and M.-E. Carr, 2006. Decadal variability of the air-sea CO$_2$ fluxes in the equatorial Pacific Ocean. *J. Geophys. Res.*, **111**, doi:10.1029/2005JC003129.

Friedlingstein, P., M. Meinshausen, V. K. Arora, C. D. Jones, A. Anav, S. K. Liddicoat, and R. Knutti, 2014. Uncertainties in CMIP5 climate projections due to carbon cycle feedbacks. *J. Climate*, **27**, 511–526, doi: 10.1175/JCLI-D-12-00579.1.

Gruber, N., M. Gloor, S. E. Mikaloff Fletcher, S. C. Doney, S. Dutkiewicz, M. J. Follows, M. Gerer, A. R. Jacobson, F. Joos, K. Lindsay, D. Menemenlis, A. Mouchet, S. a. Muller, J. L. Sarmiento, and T. Takahashi, 2009. Oceanic sources, sinks, and transport of atmospheric CO$_2$. *Global Biogeochem. Cycles*, **23**, doi:10.1029/2008GB003349.

Kay, J. E., et al., 2015. The Community Earth System Model (CESM) Large Ensemble Project: A community resource for studying climate change in the presence of internal climate variability. *Bull. Amer. Meteor. Soc.*, doi:10.1175/BAMS-D-13-00255.1.

Le Quéré, C. et al., 2009. Trends in the sources and sinks of carbon dioxide. *Nature Geosci.*, doi: 10.1038/NGEO689.

Long, M. C., K. Lindsay, S. Peacock, J. K. Moore, and S. C. Doney, 2013. Twentieth-century oceanic carbon uptake and storage in CESM1(BGC). *J. Climate*, **26**, 6775–6800, doi:10.1175/JCLI-D-12-00184.1.

Maier-Reimer, E., and K. Hasselmann, 1987. Transport and storage of CO_2 in the ocean—An inorganic ocean-circulation carbon cycle model. *Climate. Dyn.*, **2**, 63–90, doi: 10.1007/BF01054491.

McKinley, G. A., M. J. Follows, and J. Marshall, 2004. Mechanisms of air-sea CO_2 flux variability in the equatorial Pacific and the North Atlantic. *Global Biogeochem. Cycles*, **18,** doi:10.1029/2003GB0002179.

McPhaden, M. J., and D. Zhang, 2004. Pacific Ocean circulation rebounds. *Geophys. Res. Lett.*, **31**, doi:10.1029/2004GL020727.

Merrifield, M. A., and M. E. Maltrud, 2011. Regional sea level trends due to a Pacific trade wind intensification. *Geophys. Res. Lett.*, **38**, doi:10.1029/2011GL049576.

Sutton, A. J., R. A. Feely, C. L. Sabine, M. J. McPhaden, T. Takahashi, F. P. Chavez, G. E. Friederich, and J. T. Mathis, 2014. Natural variability and anthropogenic change in equatorial Pacific surface ocean pCO_2 and pH. *Global Biogeochem. Cycles*, **28**, 131–145, doi:10.1002/2013GB004679.

Takahashi, T., et al., 2009: Climatological mean and decadal changes in surface ocean pCO_2, and net sea-air CO_2 flux over the global oceans. *Deep Sea Res.*, *Part II*, **56**, 554-577, doi:10.1016/j.dsr2.2008.12.009.

Vecchi, G. A., and B. J. Soden, 2007: Global warming and the weakening of the tropical circulation. *J. Climate*, **20**, 4316–4340, doi: 10.1175/JCLI4258.1.JCLI3631.1.

Present and projected climate variability at high latitudes and its impact on the ocean carbon cycle

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Given that the ocean carbon reservoir is about fifty times greater than that of the atmosphere, a small perturbation to the ocean could theoretically produce a spectacular change in atmospheric concentrations. So it might at first seem surprising that atmospheric carbon dioxide (CO_2) concentrations have been so stable over the last millennium. High-resolution ice cores suggest that multidecadal- to century-scale variability of atmospheric CO_2 was less than 10 ppm (~3.5% of background concentrations, Ciais et al. 2013), despite climate and ocean circulation variability. Although climate and ocean circulation variability yield regional fluctuations in the ocean carbon cycle that can confound the detection of trends, these ice cores suggest that the preindustrial (or "natural") ocean carbon cycle, when integrated globally, was largely in steady state. This might reflect compensations between underlying climate-driven changes in the solubility and biological components of air-sea carbon fluxes (Marinov et al. 2011). At the start of industrialization, anthropogenic emissions of CO_2 fundamentally altered this global steady state, as atmospheric concentrations began their rapid climb from about 270 ppm in the 18th century to their current concentration above 400 ppm. Throughout this time, the ocean has provided a major sink for anthropogenic CO_2, mitigating its radiative impact (Sabine et al. 2004). Yet the radiative impact of anthropogenic CO_2 remaining in the atmosphere has raised ocean temperatures, changed freshwater and alkalinity fluxes to the ocean, and altered large-scale ocean circulation patterns. Collectively, these changes are projected to influence both the natural carbon cycle and the uptake and storage of anthropogenic carbon as they continue into the future (Figure 1).
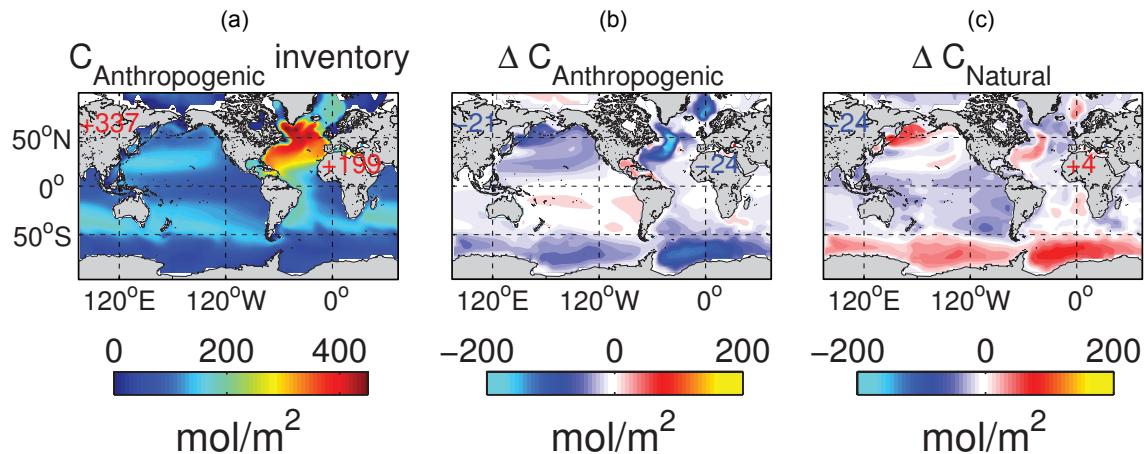
Figure 1: Three-member ensemble average of column-integrated DIC components from our coupled model, CM2Mc, averaged over the period 2081-2100. (a) Total anthropogenic carbon inventory in the ocean. Perturbation of carbon inventory due to climate change on (b) anthropogenic carbon and (c) natural carbon for the period 2081-2100 with respect to the preindustrial state. The climate change simulation was performed by prescribing historical+Representative Concentration Pathway 8.5 forcing (Meinshausen et al. 2011). Numbers over continents refer to total DIC gained or lost (PgC) for the Indo-Pacific (over Asia) and the Atlantic basins (over Africa) including the Southern Ocean. Climate change acts to reduce ocean storage of both anthropogenic and natural carbon.

Here we review recent work that exposes how climate variability and change at high latitudes influences the ocean storage and uptake of natural and anthropogenic carbon. Particularly, we focus on the Southern Ocean and North Atlantic, which provide the dominant ocean sinks for anthropogenic carbon and have very dynamic natural carbon cycles (Gerber and Joos 2010; Sabine et al. 2004; Gruber et al. 2009).

Southern Ocean

South of the Antarctic Polar Front (PF), westerly winds drive upwelling of old, relatively warm, nutrient- and carbon-rich Circumpolar Deep Water (CDW). In preindustrial times, the upwelled CDW released large quantities of natural carbon into the atmosphere. At present, anthropogenic emissions are rapidly decreasing the difference between the atmospheric and oceanic CO_2 partial pressures (pCO_2) with a resulting decrease in CO_2 degassing during CDW upwelling. Ekman transport of CDW north of the PF, together with air-sea interactions, result in the formation of intermediate and mode waters. At the surface, oceanic pCO_2, already reduced by degassing, is further lowered by phytoplankton uptake, which results in net CO_2 flux into the ocean north of the PF. Some CDW is also transported south towards the Antarctic continental shelf, providing source waters for Antarctic Bottom Water (AABW), which fills a large fraction of the global ocean volume.

Due to these ocean circulation patterns, the Southern Ocean south of 30oS is critically important for both setting the global strength of the natural ocean carbon pump and for determining atmospheric pCO_2 on long, equilibrium timescales (e.g., Marinov et al. 2006). It is responsible for about half of the annual global ocean uptake of anthropogenic carbon (Sabine et al. 2004; Gruber et al. 2009; Khatiwala et al. 2012), despite making up only a third of the ocean surface area. Here, we discuss two important modes of variability in the Southern Ocean and associated implications for the carbon cycle. The first is the Southern Annular Mode (SAM), the most important pattern of large-scale climate variability in the Southern Hemisphere middle and high latitudes which manifests as a variability in Southern Ocean westerlies. The second is the variability associated with deep Southern Ocean convection. Modeling studies show that variability in SAM and deep Southern Ocean convection results in strong variability in CO_2 fluxes in the subpolar (~40°S-55°S) and polar (~55°S-90°S) Southern Ocean regions, respectively (Resplandy et al. 2015).

SAM and carbon

There are strong links between subpolar variability associated with SAM and Southern Ocean CO_2 fluxes on interannual time scales (e.g., Lovenduski et al. 2007), which take on added importance as the combined effects of Antarctic ozone hole and greenhouse gas

warming have resulted in a more positive SAM, i.e., strengthened and poleward-shifted mid-latitude westerlies. This recent trend is expected to continue into the 21st century, though with uncertainty arising from inter-model variability across the current generation of Earth System Models (ESMs) that contributed to the Coupled Model Intercomparison Project Phase 5 (CMIP5; Swart and Fyfe 2012).

If the SAM mechanisms that influence the carbon cycle on interannual timescales also operate on longer timescales associated with climate change, the continuing trend toward positive SAM is expected to drive an increase in the upwelling of old, carbon-rich CDW south of the PF, and subsequent outgassing of the natural CO_2 flux to the atmosphere (Lovenduski et al. 2007), effectively reducing the global oceanic carbon sink. This mechanism has been proposed to explain the apparent saturation in the Southern Ocean sink for atmospheric CO_2 in recent decades (Le Quéré et al. 2007). This claim, based primarily on atmospheric inverse models and coarse resolution Global Circulation Models, has been heavily debated by the ocean carbon research community.

Trends in CO_2 uptake are hard to detect in the observations due to effects of autocorrelation and monthly variability. Majkut et al. (2014a) show that directly detecting changes such as the one associated with the recent saturation of the Southern Ocean CO_2 sink (-0.08 PgC yr^{-1} $decade^{-1}$) will require up to three decades of observations. Based on this assessment, most currently available data sets are not long enough to differentiate natural variability from the anthropogenically driven trends in CO_2 fluxes (Keller et al. 2012). In contrast to Le Quéré et al. (2007), Fay and McKinley (2013) argue that the influence of a positive trend in SAM has waned and the Southern Ocean carbon sink has regained strength since the early 2000s, following a 1990s slowdown. Majkut et al. (2014b) merge observations and model pCO_2 estimates to find increasing ocean carbon uptake south of 45°S for 1980–2009 and attribute this increase to surface ocean cooling, which offsets the expected response to increased winds.

CMIP5 models simulate a small negative effect of climate change (~5 PgC) on the ocean carbon uptake over the historical period (Fröelicher et al. 2015). The model-projected overall response of the carbon cycle to future climate change is uncertain. Climate-driven warming is acting against the intensified winds to stratify the Southern Ocean water column, reducing convective mixing and outgassing of deep ocean natural carbon (Sarmiento et al. 1998). Bernardello et al. (2014a) showed that over the 21st century, enhanced stratification and reduced deep-water mass formation in

both the North Atlantic and Southern Ocean promote increased storage of natural carbon in the ocean, particularly in high latitudes, and dominate over wind effects. However, thermal solubility effects decrease ocean carbon storage, particularly in low latitudes. The net effect of climate change in the Bernardello et al. (2014a) model analysis is an overall reduction in natural ocean carbon storage (-20 PgC) from 1860-2100 (Figure 1c). The climate-driven perturbation to the anthropogenic carbon (-45 PgC) is higher than the impact on natural carbon, and is due primarily to reductions in mid- to high-latitudes (Figure 1b).

Southern Ocean deep convection, AABW, and carbon

AABW formation sets the carbon, heat, and oxygen content of much of the deep ocean. Presently, AABW is formed at specific locations on the Antarctic continental shelf (Orsi et al. 1999). In the past, AABW was also known to form during open ocean deep convection events in the Weddell Sea (Gordon 1982; Killworth 1983; Carsey 1980), as observed for three consecutive winters in the 1970s. The current generation of climate models (CMIP5) forms AABW almost entirely through open ocean convection in the Weddell and Ross Seas, with little contribution from the continental shelf (Heuzé et al. 2013).

In 25 of the 33 CMIP5 ESMs, open ocean convection occurs as a natural oscillation in the preindustrial climate, with convective events occurring with different frequencies and durations (de Lavergne et al. 2014). This multi-decadal variability occurs in our control simulation of the GFDL CM2Mc model (Figure 2) with regular periodicity, and is similar to that observed on centennial timescales in the Kiel model (Martin et al. 2012; Latif et al. 2013; Martin et al. 2015). The system oscillates between convective periods (when heat and dissolved inorganic carbon (DIC) stored in the upper CDW (UCDW) are released to the atmosphere, melting the Antarctic sea ice) and non-convective intervals (when strong stratification isolates the surface from the warmer, DIC-rich UCDW below, decreasing atmospheric CO_2 and temperatures). Deep convective oscillations in the polar Southern Ocean promote large variability in CO_2 fluxes on multi-decadal timescales (Séférian et al. 2013; Bernardello et al. 2014b), contributing to the Southern Ocean dominance over the multidecadal global carbon flux variability in five CMIP5 models (Resplandy et al. 2015).

Strong increases in both surface heat and freshwater fluxes at Southern Ocean high latitudes are predicted under future climate forcing (Fyfe et al. 2012), with an expected increase in stratification. As a result, climate models show cessation of Southern Ocean open sea convection over the 21st century (de Lavergne et al.
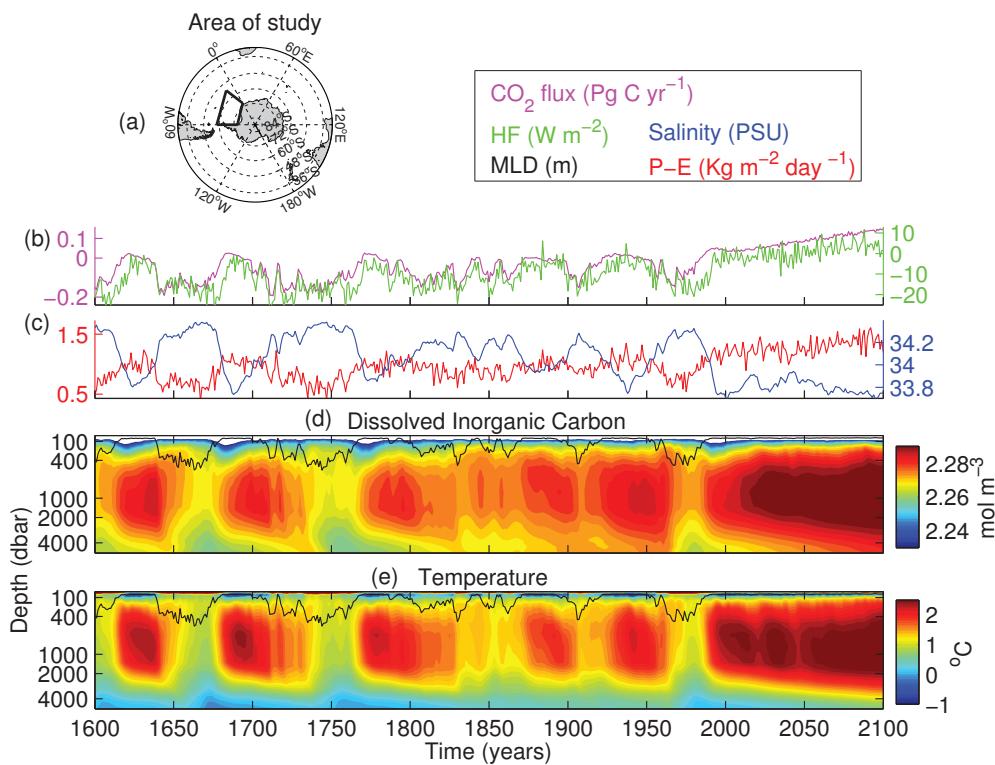
Figure 2: Regular Weddell Sea convection cycles in a 500-year segment composed of a preindustrial simulation (1600-1859) and a climate change simulation using historical + RCP8.5 forcing (1860-2100) in a coupled ocean-atmosphere model, CM2Mc. a) Area of study – dark black polygon; b) air-sea CO_2 and heat flux (HF) integrated over the study area; c) area-averaged precipitation minus evaporation (P-E) and salinity (0 to 50 m depth); d) area-averaged dissolved inorganic carbon (DIC); and e) area-averaged temperature and mixed layer depth (MLD). During convective years, the ocean loses heat and CO_2 to the atmosphere. Climate-induced freshening of the water column due to a trend in P-E after the 1970s stops convection and results in sub-surface storage of DIC and heat. Modified from Bernardello et al, (2014b).

of Southern Ocean deep convection could be an important source of inter-model spread for the projected future evolution of the carbon cycle.

North Atlantic

The North Atlantic is the next biggest ocean sink for anthropogenic carbon after the Southern Ocean (Sabine et al. 2004) and the most intense per unit area (Takahashi et al. 2009). Despite being one of our best-observed ocean basins, internal variability hinders the evaluation of climate change-driven trends. Here, the dominant mode of climate variability at the interannual time scale is the North Atlantic Oscillation (NAO), a fluctuation in the strength of the pressure gradient between the Icelandic low and Azores high (Hurrell 1995), which manifests as variability in the storm track and oceanic mixed layer depths (Dickson et al. 1996). The NAO has been linked to variability in North Atlantic carbon dynamics in observations (Gruber et al. 2002; Bates et al. 2002) and in modeling studies (Keller et al. 2012). Early speculation that a positive phase of the NAO could lead to a basin-wide increase in ocean carbon storage has been replaced with evidence for compensating responses between the subtropical gyre, where a positive phase of the NAO is linked to an enhanced carbon sink, and the subpolar gyre, where the opposite is true (Keller et al. 2012; Thomas et al. 2008). Thus, while NAO variability has confounded detection of trends in the oceanic uptake of anthropogenic CO_2 locally, it seems to have a small impact on uptake when averaged over the entire North Atlantic.

On longer time scales, the dominant mode of variability in the North Atlantic is expressed as swings in basin-average sea surface temperature (SST) of more than 0.4°K, with a period of 65-85 years, and is generally referred to as the Atlantic Multidecadal Oscillation (AMO) (Delworth and Mann 2000; Kushnir 1994). Though there

2014), with important implications for Southern Ocean carbon uptake and storage. As an example, Figure 2 shows enhanced storage of subsurface natural carbon and less natural carbon outgassing following the climate-driven shutdown of Southern Ocean convection in the GFDL CM2Mc model experiments. While convective shutdown increases Southern Ocean natural carbon storage, it decreases the Southern Ocean anthropogenic uptake. The cessation of open ocean convection in the Weddell Sea, which occurs in the model on average in year 1981 (Figure 2), is responsible for 22% of the Southern Ocean decrease in total (anthropogenic plus natural) ocean carbon uptake and 52% of the decrease in the anthropogenic component, despite the Weddell Sea representing only 4% of the area of the Southern Ocean (Bernardello et al. 2014b). Therefore, differences in representation

is ongoing controversy over the degree to which external forcing has played a role in the amplitude and timing of AMO variability (Booth et al. 2012; Zhang et al. 2013; Mann et al. 2014), general circulation models and paleoclimate proxy data collectively suggest that internal variability associated with the Atlantic Meridional Overturning Circulation (AMOC) is largely responsible for this low-frequency SST oscillation. The SST variability alone, regardless of its cause, creates fluctuations in solubility with consequences for anthropogenic CO_2 uptake (McKinley et al. 2011; Löptien and Eden 2010). However, because the AMO is driven largely by fluctuations in the large-scale circulation, there can be competing effects of circulation on DIC, such that the total trend in the rate of carbon uptake may be opposite the temperature-driven trend alone (Fay and McKinley 2013; McKinley et al. 2011).

North Atlantic internal variability complicates the detection of climate-driven trends. For instance, McKinley et al. (2011) show that purported trends in the North Atlantic anthropogenic carbon uptake, diagnosed using the difference between trends in atmospheric and ocean pCO_2, are sensitive to start and end year of the trend calculation (Figure 3). In their work, trends of oceanic pCO_2 match trends in atmospheric pCO_2 throughout the entire North Atlantic when the full length of the observational record is taken into account

(Figure 3a). However, when a shorter period is considered, the trend is regionally specific: the permanently stratified subtropical region sees increased uptake; the seasonally stratified subtropical region sees decreased uptake; and uptake in the subpolar region remains steady (Figure 3b). Thus, the observational record does not yet reveal any reduction in North Atlantic carbon uptake due to climate change, despite the contribution of rising SST to decreasing CO_2 solubility starting to emerge from background variability (Fay and McKinley 2013; Séférian et al. 2014, Majkut et al. 2014b).

In the coming decades, the AMOC is widely predicted to slow down (Stocker et al. 2013), with important implications for the storage of natural carbon and uptake of anthropogenic carbon. The AMOC slowdown is predicted to decrease the outgassing of natural carbon, as remineralized carbon accumulates in the subpolar North Atlantic and along the North Atlantic Deep Water pathway (Bernardello et al. 2014a; Sarmiento et al. 1998). This increase in natural carbon retained by the ocean, however, is more than offset by the reduction to the anthropogenic carbon uptake caused by the decreasing exposure of deep waters to the atmosphere reinforced by overall SST warming (Figure 1). Thus, in the coming decades, climate-driven changes in North Atlantic circulation and SST are likely to reduce the pace of oceanic uptake of anthropogenic carbon.
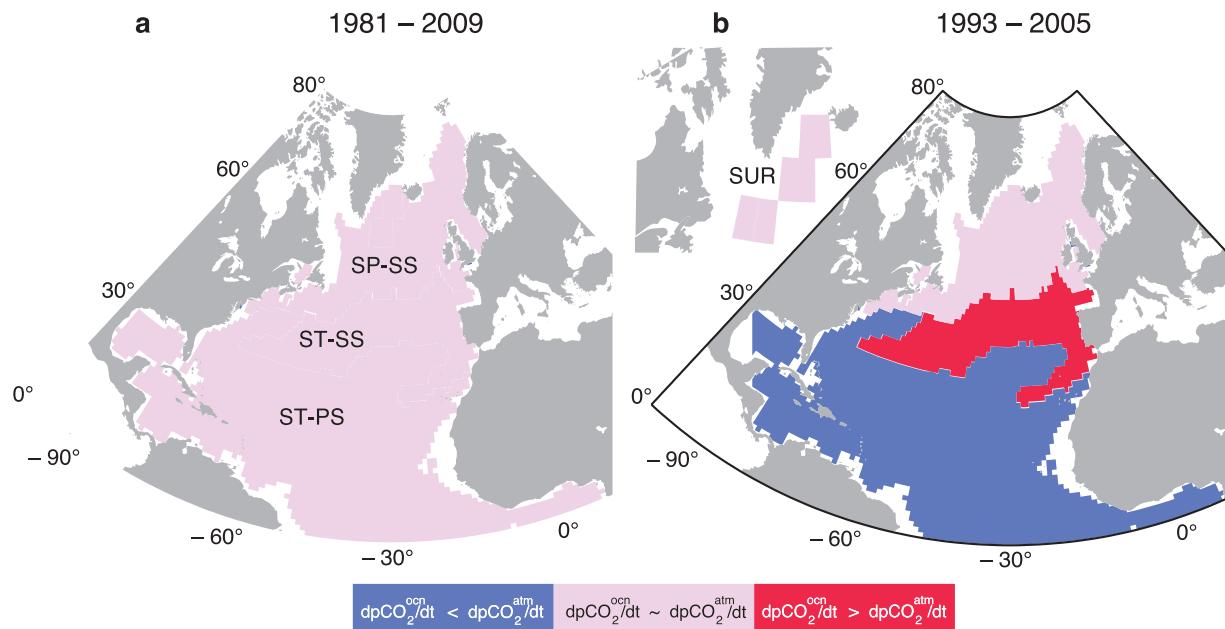

Figure 3: Trend in oceanic pCO_2 compared to atmospheric pCO_2 for two periods of different length: a) 1981-2009 and b) 1993-2005. Blue for oceanic pCO_2 trend less than atmospheric; pink for indistinguishable; and red for oceanic exceeding atmospheric. In this analysis, the North Atlantic is divided into three "biomes": subtropical - permanently stratified (ST-PS), subtropical - seasonally stratified (ST-SS), and subpolar - seasonally stratified (SP-SS). The inset shows a region of the SP-SS that had additional chemical data available. From McKinley et al. (2011).

Conclusions

Understanding the response of the Southern Ocean and North Atlantic carbon uptake and storage to changing climate is a prerequisite for predicting future atmospheric CO_2 concentrations. The lack of long-term observations has thus far hampered a complete understanding of the carbon cycle in these regions. At the same time, the current generation of climate models is still affected by critical issues like the incomplete representation of ice sheet and ice shelf dynamics. Coarse resolution can result in the net Southern Ocean meridional overturning and the natural Southern Ocean carbon storage being too sensitive to changes in wind-stress compared to eddy-permitting ocean models (e.g., Hallberg and Gnanadesikan 2006; Munday et al. 2014). Models must also overcome challenges in accurately representing the export of deep water from continental shelves and marginal seas. Model biases in Labrador Sea convection can result in unrealistic links between the NAO and AMOC, while biases in Weddell Sea convection might affect Southern Ocean decadal to centennial variability (Martin et al. 2015; Marinov et al., in prep.), with carbon cycle implications. Importantly, the inter-model spread for air-sea CO_2 fluxes and anthropogenic C inventories is largest in the Southern Ocean, where intense vertical exchange occurs (e.g., Orr et al. 2001; Matsumoto et al. 2004).

We argue that the marriage of modeling and observational approaches will continue yielding insight into variability and trends in the ocean carbon cycle. Already, we have learned a great deal about ocean physics through a hierarchy of modeling approaches, such as idealized models aimed at understanding mesoscale eddies (e.g., Morrison and McC. Hogg 2013), the development and testing of parameterizations to transport deep and bottom water from the shelves and marginal seas to the open ocean (e.g., Snow et al. 2015), and the comparison of climate responses across high- and low-resolution models (Bryan et al. 2014; Griffies et al. 2014; Winton et al. 2014). We expect the same gains by the continued deployment of such tools to carbon cycle questions. Likewise, long-term sustained observations in the high latitude oceans are critical to reliably document the global changes in carbon uptake, storage, and transport; separate natural variability from anthropogenic forcing; and evaluate the success of our models. The future deployment of ~200 Argo floats with biogeochemical capabilities in the Southern Ocean by the newly-established Southern Ocean Carbon and Climate Observations and Modeling (SOCCOM) project, and the upcoming monitoring of the full annual cycle of gas exchange in the Labrador Sea deep convective region through the Canadian program, Ventilations, Interactions and Transports Across the Labrador Sea (VITALS), are steps in the right direction. Adding biogeochemical sensors to existing ocean arrays that monitor the AMOC (e.g., the OSNAP and RAPID arrays) could likewise provide a rich data source to understand this critical component of our Earth system.

References

Bates, N. R., A. C. Pequignet, R. J. Johnson, and N. Gruber, 2002. A short-term sink for atmospheric CO2 in subtropical mode water of the North Atlantic Ocean. *Nature*, **420**, doi:10.1038/nature01253.

Bernardello, R., I. Marinov, J. B. Palter, J. L. Sarmiento, E. D. Galbraith, and R. D. Slater, 2014a. Response of the ocean natural carbon storage to projected twenty-first century climate change. *J. Climate*, **27**, 2033–2053, doi: 10.1175/JCLI-D-13-00343.1.

Bernardello, R., I. Marinov, J. B. Palter, E. D. Galbraith, and J. L. Sarmiento, 2014b. Impact of Weddell Sea deep convection on natural and anthropogenic carbon in a climate model. *Geophys. Res. Lett.*, **41**, 7262–7269, doi:10.1002/2014GL061313.

Booth, B. B. B., N. J. Dunstone, P. R. Halloran, T. Andrews, and N. Bellouin, 2012. Aerosols implicated as a prime driver of twentieth-century North Atlantic climate variability. *Nature*, **484**, 228–232, doi:10.1038/nature10946.

Bryan, F. O., P. R. Gent, and R. Tomas, 2014. Can Southern Ocean eddy effects be parameterized in climate models? *J. Climate*, **27**, 411–425, doi: 10.1175/JCLI-D-12-00759.1.

Carsey, F. D., 1980. Microwave observation of the Weddell Polynya. *Mon. Wea. Rev.*, **108**, 2032–2044, doi: 10.1175/1520-0493(1980)108<2032:MOOTWP>2.0.CO;2.

Ciais, P., C. Sabine, G. Bala, L. Bopp, V. Brovkin, J. Canadell, A. Chhabra, R. DeFries, J. Galloway, M. Heimann, C. Jones, C. LeQuere, R. B. Myneni, S. Piao, and P. Thorton, 2013. Carbon and other biogeochemical cycles. *Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change*. T.F. Stocker et al., Eds., Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA.

de Lavergne, C., J. B. Palter, E. D. Galbraith, R. Bernardello, and I. Marinov, 2014. Cessation of deep convection in the open Southern Ocean under anthropogenic climate change. *Nat. Climate Change*, **4**, 278–282, doi:10.1038/nclimate2132.

Delworth, T. L., and M. E. Mann, 2000. Observed and simulated multidecadal variability in the Northern Hemisphere. *Climate Dyn.*, **16**, 661–676, doi: 10.1007/s003820000075.

Dickson, R., J. Lazier, J. Meincke, P. Rhines, and J. Swift, 1996. Long-term coordinated changes in the convective activity of the North Atlantic. *Prog. Oceanogr.*, **38**, 241–295, doi:10.1016/S0079-6611(97)00002-5.

Fay, A. R., and G. A. McKinley, 2013. Global trends in surface ocean *p*CO2 from *in situ* data. *Global Biogeochem. Cycles*, **27**, 541–557, doi: 10.1002/gbc.20051.

Frölicher, T. L., J. L. Sarmiento, D. J. Paynter, J. P. Dunne, J. P. Krasting, and M. Winton, 2015. Dominance of the Southern Ocean in anthropogenic carbon and heat uptake in CMIP5 Models. *J. Climate*, **28**, 862–886, doi: 10.1175/JCLI-D-14-00117.1.

Fyfe, J. C., N. P. Gillett, and G. J. Marshall, 2012. Human influence on extratropical Southern Hemisphere summer precipitation. *Geophys. Res. Lett.*, **39**, doi: 10.1029/2012GL054199.

Gerber, M., and F. Joos, 2010. Carbon sources and sinks from an Ensemble Kalman Filter ocean data assimilation. *Global Biogeochem. Cycles*, **24**, doi: 10.1029/2009GB003531.

Gordon, A. L., 1982. Weddell deep water variability. *J. Mar. Res.*, **40**, 199–217.

Griffies, S. M., M. Winton, W. G. Anderson, R. Benson, T. L. Delworth, C. O Dufour, J. P. Dunne, P. Goddard, A. K. Morrison, A. Rosati, A. T. Wittenberg, J. Yin, and R. Zhang, 2014. Impacts on ocean heat from transient mesoscale eddies in a hierarchy of climate models. *J. Climate,* **28,** 952-977, doi:10.1175/JCLI-D-14-00353.1.

Gruber, N., C. D. Keeling, and N. R. Bates, 2002. Interannual variability in the North Atlantic ocean carbon Sink. *Science.*, **298**, 2374–2378, doi:10.1126/science.1077077.

Gruber, N., M. Gloor, S. E. Mikaloff Fletcher, S. C. Doney, S. Dutkiewicz, M. J. Follows, M. Gerber A. R. Jacobson, F. Joos, K. Lindsay, D. Menemenlis, A. Mourchet, S. A. Muller, J. L. Sarmiento, and T. Takahashi, 2009. Oceanic sources, sinks, and transport of atmospheric CO$_2$. *Global Biogeochem. Cycles*, **23**, doi: 10.1029/2008GB003349.

Hallberg, R. W., and A. Gnanadesikan, 2006. The role of eddies in determining the structure and response of the wind-driven Southern Hemisphere overturning: Results from the modeling eddies in the Southern Ocean (MESO) project. *J. Phys. Oceanogr.*, **36**, 2232–2252, doi: 10.1175/JPO2980.1.

Heuzé, C., K. J. Heywood, D. P. Stevens, and J. K. Ridley, 2013. Southern Ocean bottom water characteristics in CMIP5 models. *Geophys. Res. Lett.*, **40**, 1409–1414, doi:10.1002/grl.50287.

Hurrell, J. W., 1995. Decadal trends in North Atlantic oscillation: Regional temperatures and precipitation. *Science*, **269**, 676–679, doi: 10.1126/science.269.5224.676.

Keller, K. M., F. Joos, C. C. Raible, V. Cocco, T. L. Frolicher, J. P. Dunne, M. Gehlen, L. Bopp, J. C. Orr, J. Tjiputra, C. Keinze, J. Segschneider, T. Roy, and N. Metzl, 2012. Variability of the ocean carbon cycle in response to the North Atlantic Oscillation. *Tellus B*, **64**, doi:10.3402/tellusb.v64i0.18738.

Khatiwala, S., T. Tanhua, S. Mikaloff Fletcher, M. Gerber, S. C. Doney, H. D. graven, N. Gruber, G. A. McKinley, A. Murata, A. F. Rios, C. L. Sabine, and J. L. Sarmiento, 2012. Global ocean storage of anthropogenic carbon. *Biogeo. Discuss.*, **9**, 8931–8988, doi:10.5194/bgd-9-8931-2012.

Killworth, P. D., 1983. Deep convection in the World Ocean. *Rev. Geophys.*, **21**, 1, doi:10.1029/RG021i001p00001.

Kushnir, Y., 1994. Interdecadal variations in North Atlantic sea surface temperature and associated atmospheric conditions. *J. Climate*, **7**, 141–157, doi: 10.1175/1520-0442(1994)007<0141:IVINAS>2.0.CO;2.

Latif, M., T. Martin, and W. Park, 2013. Southern Ocean sector centennial climate variability and recent decadal trends. *J. Climate*, **26**, 7767–7782, doi: 10.1175/JCLI-D-12-00281.1.

Löptien, U., and C. Eden, 2010. Multidecadal CO2 uptake variability of the North Atlantic. *J. Geophys. Res.*, **115**, D12113, doi: 10.1029/2009JD012431.

Lovenduski, N. S., N. Gruber, S. C. Doney, and I. V Lima, 2007. Enhanced CO$_2$ outgassing in the Southern Ocean from a positive phase of the Southern Annular Mode. *Global Biogeochem. Cycles*, **21**, doi: 10.1029/2006GB002900.

Majkut J. D., B. R. Carter, T. L. Frölicher, C.O. Dufour, K. B. Rodgers, and J. L. Sarmiento, 2014a. An observing system simulation for Southern Ocean carbon dioxide uptake. *Phil. Trans. R. Soc. A,* **372**, doi: 10.1098/rsta.2013.0046.. Majkut, J. D., J. L. Sarmiento, and K. B. Rodgers, 2014b. A growing oceanic carbon uptake: Results from an inversion study of surface pCO$_2$ data. *Global Biogeochem. Cycles*, **28**, 335–351, doi: 10.1002/2013GB004585.

Mann, M. E., B. A. Steinman, and S. K. Miller, 2014. On forced temperature changes, internal variability, and the AMO. *Geophys. Res. Lett.*, **41**, doi: 10.1002/2014GL059233.

Marinov, I., A. Gnanadesikan, J. Toggweiler, and J. Sarmiento, 2006. The Southern Ocean biogeochemical divide, *Nature***, 441,** 964–967, doi:10.1038/nature04883.

Marinov, I., and A. Gnanadesikan, 2011. Changes in ocean circulation and carbon storage are decoupled from air-sea CO$_2$ fluxes. *Biogeosciences*, **8**, 505-513, doi:10.5194/bg-8-505-2011.

Martin, T., W. Park, and M. Latif, 2012. Multi-centennial variability controlled by Southern Ocean convection in the Kiel Climate Model. *Climate Dyn.*, **40**, 2005–2022, doi:10.1007/s00382-012-1586-7.

Martin, T., W. Park, and M. Latif, 2015: Southern Ocean forcing of the North Atlantic at multi-centennial time scales in the Kiel Climate Model. *Deep Sea Res. Part II Top. Stud. Oceanogr.*, **114**, 39–48, doi:10.1016/j.dsr2.2014.01.018.

Matsumoto, K., et al., 2004. Evaluation of ocean carbon cycle models with databased metrics. *Geophys. Res. Lett.,* **31**, doi:10.1029/2003GL018970.

McKinley, G. A., A. R. Fay, T. Takahashi, and N. Metzl, 2011. Convergence of atmospheric and North Atlantic carbon dioxide trends on multidecadal timescales. *Nat. Geosci.*, **4**, 606–610, doi:10.1038/ngeo1193.

Meinshausen, M., S. J. Smiht, K. Calvin, J. S. Daniel, M. L. T. Kainuma, J.-F. Lamarque, K. Matsumoto, S. A. Montzka, S. C. B. Raper, K. Riahi, A. Thomson, G. J. M. Velders, and D. P. P. van Vuuren, 2011. The RCP greenhouse gas concentrations and their extensions from 1765 to 2300. *Climate Change*, **109**, 213–241, doi:10.1007/s10584-011-0156-z.

Morrison, A. K., and A. McC. Hogg, 2013. On the relationship between Southern Ocean overturning and ACC transport. *J. Phys. Oceanogr.*, **43**, 140–148, doi:10.1175/JPO-D-12-057.1.

Munday, D. R., H. L. Johnson, and D. P. Marshall, 2014. Impacts and effects of mesoscale ocean eddies on ocean carbon storage and atmospheric pCO$_2$. *Global Biogeochem. Cycles*, **28**, 877–896, doi: 10.1002/2014GB004836.

Orr, J. C., E. Maier-Reimer, U. Mikolajewicz, P. Monfray, J. L. Sarmiento, J. R. Toggweiler, N. K. Taylor, J. Palmer, N. Gruber, C. L. Sabine, C. Le Quere, R. M. Key, and J. Boutin, 2001. Estimates of anthropogenic carbon uptake from four three-dimensional global ocean models, *Global Biogeochem. Cycles*, 15, 43– 60.

Orsi, A. H., G. C. Johnson, and J. L. Bullister, 1999. Circulation, mixing, and production of Antarctic Bottom Water. *Prog. Oceanogr.*, **43**, 55–109, doi:10.1016/S0079-6611(99)00004-X.

Le Quéré, C., C. Rodenbeck, E. T. Buitenhuis, T. J. Conway, R. Langenfelds, A. Gomez, C. Labuschagne, M. Ramonet, T. Nakazawa, N. Metzl, N. Gillett, and M. Heimann, 2007. Saturation of the southern ocean CO_2 sink due to recent climate change. *Science*, **316**, 1735–1738, doi: 10.1126/science.1136188.

Resplandy, L., R. Séférian, and L. Bopp, 2015. Natural variability of CO_2 and O_2 fluxes: What can we learn from centuries-long climate models simulations? *J. Geophys. Res. Ocean.*, **120**, 384–404, doi: 10.1002/2014JC010463.

Sabine, C. L., R. A. Feely, N. Cruber, R. M Key, K. Lee, J. L. Bullister, R. Wannikhof, C. S. Wong, D. W. R. Wallace, B. Tilbrook, F. J. Millero, T.-H. Peng, A. Kozyr, T. Ono, and A. F. Rios, 2004. The oceanic sink for anthropogenic CO_2. *Science.*, **305**, 367–371, doi:10.1126/science.1097403.

Sarmiento, J. L., T. M. C. Hughes, R. J. Stouffer, and S. Manabe, 1998. Simulated response of the ocean carbon cycle to anthropogenic climate warming. *Nature*, **393**, 245–249, doi:10.1038/30455.

Séférian, R., L. Bopp, D. Swingedouw, and J. Servonnat, 2013. Dynamical and biogeochemical control on the decadal variability of ocean carbon fluxes. *Earth Syst. Dyn.*, **4**, 109–127, doi:10.5194/esd-4-109-2013.

Séférian, R., A. Ribes, and L. Bopp, 2014. Detecting the anthropogenic influences on recent changes in ocean carbon uptake. *Geophys. Res. Lett.*, **41**, 5968–5977, doi: 10.1002/2014GL061223.

Snow, K., A. M. Hogg, S. M. Downes, B. M. Sloyan, M. L. Bates, and S. M. Griffies, 2015. Sensitivity of abyssal water masses to overflow parameterisations. *Ocean Model.*, **89**, 84–103, doi:10.1016/j.ocemod.2015.03.004.

Stocker, T. F.,et al., 2013. Technical Summary. *Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change*, T.F. Stocker et al., Eds., Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA.

Swart, N. C., and J. C. Fyfe, 2012. Observed and simulated changes in the Southern Hemisphere surface westerly wind-stress. *Geophys. Res. Lett.*, **39**, doi: 10.1029/2012GL052810.

Takahashi, T., et al., 2009. Climatological mean and decadal change in surface ocean pCO_2, and net sea-air CO_2 flux over the global oceans. *Deep. Res. Part II Top. Stud. Oceanogr.*, **56**, 554–577, doi:10.1016/j.dsr2.2008.12.009.

Thomas, H., A. E. Friederike Prowe, I. D. Lima, S. C. Doney, R. Wanninkhof, R. J. Greatbatch, U. Schuster, and A. Corbiére, 2008. Changes in the North Atlantic Oscillation influence CO_2 uptake in the North Atlantic over the past 2 decades. *Global Biogeochem. Cycles*, **22**, doi: 10.1029/2007GB003167.

Winton, M., W. G. Anderson, T. L. Delworth, S. M. Griffies, W. J. Hurlin, and A. Rosati, 2014. Has coarse ocean resolution biased simulations of transient climate sensitivity? *Geophys. Res. Lett.*, **41**, 8522–8529, doi: 10.1002/2014GL061523.

Zhang, R., T. L. Delworth, R. Sutton, D. L. R. Hodson, K. W. Dizon, I. M. Held, Y. Kushnir, J. Marhsall, Y. Ming, R. Msadek, J. Robson, A. J. Rosati, M. Ting, and G. A. Vecchi, 2013. Have aerosols caused the observed Atlantic Multidecadal Variability? *J. Atmos. Sci*, 70, 1135-1144, doi: 10.1175/JAS-D-12-0331.1.

The future of the Southern Ocean carbon storage in CMIP5 models

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The oceans represent the largest carbon reservoir relevant to climate on human timescales (Sabine et al. 2004). Within this reservoir, the Southern Ocean serves as the dominant player in ocean carbon uptake relative to other basins (Marshall and Speer 2012), owing to the strength of the vertical exchanges between surface and deep waters that characterize its circulation. It is therefore imperative to diagnose the future evolution of the Southern Ocean sink in order to predict the global ocean response to increasing atmospheric carbon dioxide (CO_2) levels.

The Southern Ocean and the carbon storage problem in CMIP5

As indicated in Dunne and Laufkötter (2015), CMIP5 has provided new insights on the evolution of the ocean carbon storage, but attribution and understanding of long-term behaviors are limited by the intrinsic difficulties in modeling the complexity of the ocean biogeochemistry and its multiple feedbacks.

Here, using a suite of Earth System Model (ESM) simulations participating in the Fifth Coupled Model Intercomparison Project (CMIP5; Taylor et al. 2012), we investigate the Southern Ocean past carbon inventory and future projections, and we discuss advantages and limitations of the stored model outputs (Ito et al. 2015). We concentrate on the twentieth and twenty-first century and consider the Representative Concentration Pathway 8.5 (RCP8.5) scenario (Meinshausen et al. 2011). The RCP8.5 projections are forced with emissions such that the radiative forcing induced by greenhouse gases reaches ~8.5 Wm^{-2} in 2100.

The subset of models under consideration include the:
- Community Earth System Model, or CESM (Long et al. 2013; Moore et al. 2013);
- Max Plank Institute model, or MPI-LR (Giorgetta et al., 2013);
- Two versions of the Geophysical Fluid Dynamics Laboratory (GFDL) Earth System Model, GFDL-ESM2G and GFDL-ESM2M (Dunne et al. 2013), which differ in their ocean module, specifically in the choice of vertical coordinate system for each component;
- Two versions of the Global Environment Model version 2, HadGEM2-ES and HadGEM2-CC (Collins et al. 2011), whereby HadGEM2-CC adopts a vertical extension of the atmospheric module from 38 to 60 layers but does not include the atmospheric chemistry scheme; and
- Three versions of the Institute Pierre Simon Laplace model, IPSL-A-LR, IPSL-A-MR, and IPSL-B-LR (Dufresne et al. 2013), whereby differences between IPSL-A-LR and -MR are limited to the resolution of the atmospheric component (1.875° × 3.75° in the LR (low resolution) and 1.25° × 2.5° in the MR (medium resolution)), and IPSL-B-LR implements a recently developed, physically based parameterization scheme for clouds and convection.

These models exhibit substantial spread in their equilibrium climate sensitivity, both in the magnitude of temperature increase in response to a doubling of CO_2 and in their representation of biogeochemical processes. Despite those differences, however, most models predict similar changes into the future. In all of the models, the Southern Ocean surface waters warm, sea ice decreases, surface and intermediate layers freshen, and the deep layer warms (Sallée et al. 2013; Meijers 2014). However, there are discrepancies among models in the magnitude of warming and freshening. For example, the sea surface temperature (SST) over the extratropical Southern Hemisphere (45°S-60°S) in the GFDL model increases by 1°C by

2100 when compared to the 1950-1960 average, independently of the version considered, while SST increases by more than 3°C in CESM (Figure 1). Each model family is characterized by its own temperature trend, with little variability across model versions, but not by a common SST mean state. For example, in IPSL, the mean SST from 1900-2100 over the extratropical Southern Hemisphere (45°S-60°S) is ~3.5°C in both IPSL-A realizations and close to 7°C in the IPSL-B-LR (with the new convective scheme) realization.
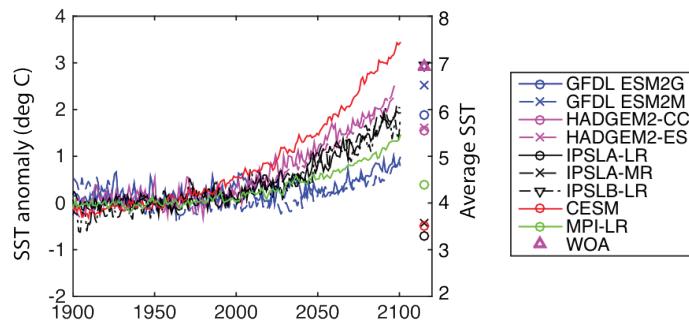
Figure 1: Area-weighted annual mean sea surface temperature (°C) over the extratropical Southern Hemisphere oceans (45°S-60°S). Average (1950-1960), and anomalies (1900-2100) calculated relative to the 1950-1960 mean sea surface temperature in all CMIP5 models considered and in the World Ocean Atlas.

The observed warming in the models is linked to a slowdown in the formation of Antarctic bottom water (AABW; de Lavergne et al. 2014), which limits ocean uptake of atmospheric CO_2 (Sarmiento et al. 1998). On the other hand, upper ocean circulation associated with the formation of key mode and intermediate water masses is predicted to intensify (Waugh et al. 2013) due to stronger near-surface winds (Thompson et al. 2011). This is shown in Figure 2 with CESM and IPSL displaying the smallest and largest changes, respectively. Stronger winds have been shown to have contrasting effects on ocean carbon uptake. By increasing vertical mixing, they can increase the subduction of carbon into the thermocline and its transport equatorward (Ito et al. 2010); however, stronger winds can also increase outgassing of carbon-rich deep waters to the atmosphere (Lovenduski et al. 2013).

According to coupled global climate model (CGCM) projections, competing physical changes in the buoyancy and momentum forcing will therefore

affect the carbon uptake in the Southern Ocean in the future. These physical changes must be considered together with changes in the ocean biological response. It is worth noting that the response of the ocean to increased greenhouse gases portrayed by current state-of-the-art CGCMs neglects changes in eddy activity due to resolution constrains. It has been hypothesized that on decadal time scales, the effect of eddies may rival that of wind variability in the Southern Ocean (Boning et al. 2008; Meredith et al. 2012) and is likely to represent the largest source of uncertainty in current model projections.

In the analysis of the evolution of ocean carbon uptake, two major contributions to the carbon inventory of an ocean basin must be considered: the "preformed" carbon and the "regenerated" carbon. The former is sequestered via physical processes and is transported from surface to depth in the form of dissolved inorganic carbon (DIC), including the anthropogenic DIC. Regenerated carbon, on the other hand, results from biological processes (i.e., photosynthesis and the subsequent formation of organic material). Organic material sinks and is remineralized back into inorganic carbon at depth, representing storage of CO_2 via the biological pump. Regenerated carbon is not a quantity commonly stored in models, but can be derived from the oxygen deficit relative to the atmospheric saturation (whenever this variable is available) under the assumption of a constant elemental stoichiometric ratio. A smaller, but still significant amount of regenerated carbon is sequestered through the formation of calcium carbonate deep-
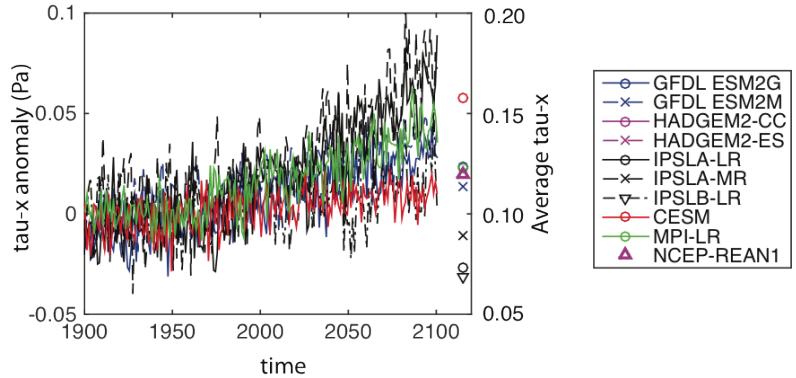

Figure 2: Area-weighted annual mean of zonal wind stress (Pa) over the extratropical Southern Hemisphere oceans (45°S-60°S). Average zonal wind stress (1950-1960) and zonal wind stress anomalies (1900-2100) calculated with respect to the 1950-1960 zonal wind stress mean in all CMIP5 models considered and in the NCEP reanalysis.
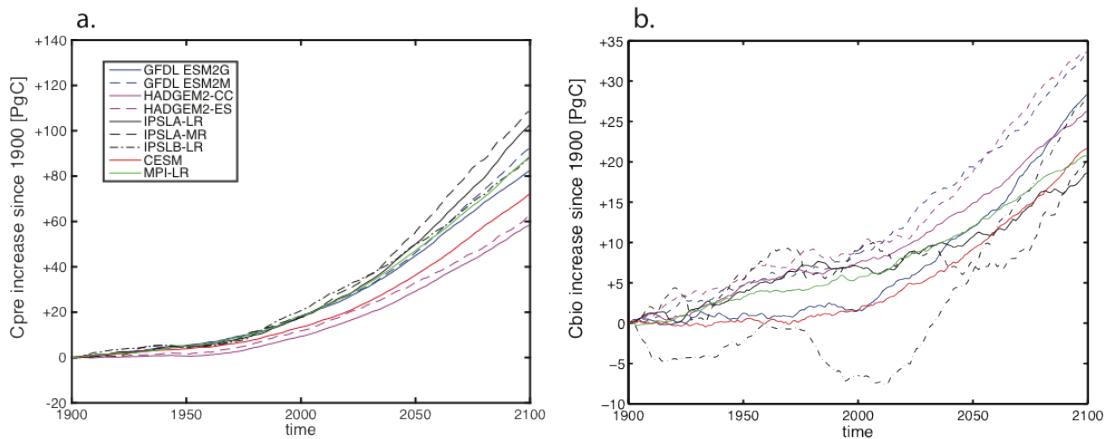
Figure 3. Southern Ocean carbon inventory change (in PgC) since 1900 in the subset of the CMIP5 archive analyzed; a) preformed carbon; b) regenerated carbon.

sea carbonates, predominantly by calcitic coccoliths and planktonic foraminifera (Milliman 1993). This contribution can be evaluated via a calculation of excess alkalinity in the subsurface waters (Brewer, 1978).

CMIP5 projections indicate that both preformed and regenerated carbon inventory in the Southern Ocean will increase in the future, at least up to 2100 (Figure 3; Ito et al. 2015). The preformed carbon inventory increases from 60 (HadGEM2) to 110 (IPSL-A-MR) PgC between 1900 and 2100. Such increase takes place primarily in the upper thermocline and reflects the surge in atmospheric CO_2. Each model's regenerated carbon inventory depends on its representation of ocean biological processes, so it is not surprising to see large model-model differences, given the distinct ecological modeling strategies applied by each model family (Dunne and Laufkötter 2015). Nonetheless, all models predict increased biological activity towards the end of the 21st century compared to present and past conditions, with highest accumulations in the Southern Ocean (Figure 4). The multi-model median inventory increase is 26 PgC, with the minimum of 18 PgC found in IPSL-A-LR and the maximum of 33 PgC in HadGEM2-ES and GFDL-ESM2M.

Moving forward

The analysis of a sample of CMIP5 models has revealed that the ability of the Southern Ocean to store CO_2 will continue to increase during this century, in agreement with recent investigations by Bernardello et al. (2013), Meijers
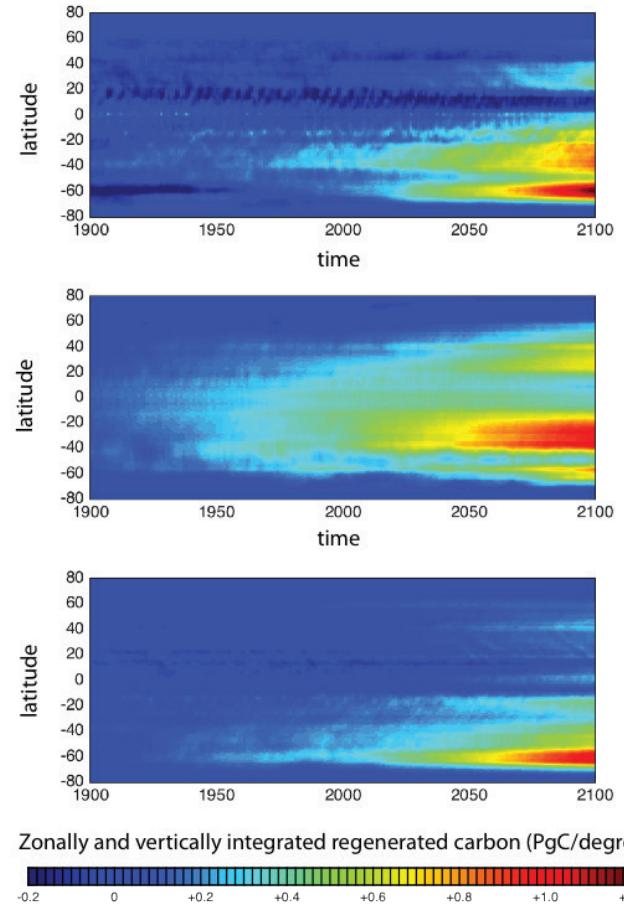

Figure 4: Zonally and vertically integrated annual mean regenerated carbon anomaly (PgC/degree) from 1900-2100 relative to 1860 in three of the models analyzed. Top: GFDL-ESM2M; Middle: IPSL-A-LR; Bottom: MPI-LR.

(2014), and de Lavergne et al. (2014). However, the same models predict the opposite trend for the global uptake, i.e., a slowdown of atmospheric CO_2 uptake by the ocean (Doney et al. 2014).

Using only CMIP5 integrations, it is not possible to quantify the relative contributions of physical and biological processes, or to estimate the degree of nonlinearity of the interactions between processes. Experiments that include passive tracers such as CFCs and SF_6 could help constrain the roles of physical advection and mixing, and should be included in the next model intercomparison effort. Additionally, sensitivity experiments in which perturbations to physical or biological states are introduced in a controlled manner (e.g., Ito et al. 2015) represent an essential tool to improve mechanistic understanding. Such exercises could help identify strengths and weaknesses of CGCM projections and should be prioritized in the CMIP framework and timeline.

As a final note, the figures presented here suggest that the time-varying rate of change of various quantities may be significantly different across models, more so than the global time integral, for which a better convergence is achieved. Furthermore, the uncertainty associated with the inter-model spread in Southern Ocean carbon storage is greater than the uncertainty in the global average of ocean carbon uptake (Arora et al. 2013). This points to the existence of different compensating effects between basins in the various models and to the need to investigate the regional expression of carbon uptake at regional scales. In the few available integrations continuing out to 2300, those regional divergences amplify, severely eroding the global inter-model agreement. An unresolved question that must be answered before any consensus can be achieved pertains to the origin of those differences. Do they result from differences in the representation of mean state patterns, ventilation, and uptake, or are they linked to the spatial and temporal characteristics of the decadal variability modes?

References

Arora, V. et al. 2013. Carbon–concentration and carbon–climate feedbacks in CMIP5 Earth System Models. *J. Climate*, **26**, 5289-5314, doi: 10.1175/JCLI-D-12-00494.1.

Bernardello, R., I. Marinov, J. B. Palter, J. L. Sarmiento, E. D. Galbraith, and R. D. Slater, 2013. Response of the ocean natural carbon storage to projected twenty-first-century climate change. *J. Climate*, **27**, 2033-2053, doi:10.1175/JCLI-D-13-00343.1.

Boning, C. W., A. Dispert, M. Visbeck, S. R. Rintoul, and F. U. Schwarzkopf, 2008. The response of the Antarctic Circumpolar Current to recent climate change. *Nat. Geosci*, **1**, 864-869, doi:10.1038/ngeo362.

Brewer, P. G., 1978. Direct observation of the oceanic CO_2 increase. *Geophys. Res. Lett.,* **5**, 997-1000, doi: 10.1029/GL005i012p00997.

Collins, W. J., et al. 2011. Development and evaluation of an Earth-System model – HadGEM2, *Geosci. Model Dev.*, **4**, 1051-1075, doi:10.5194/gmd-4-1051-2011.

de Lavergne, C., J. B. Palter, E. D. Galbraith, R. Bernardello, and I. Marinov, 2014. Cessation of deep convection in the open Southern Ocean under anthropogenic climate change. *Nat. Climate Change*, **4**, 278-282, doi:10.1038/nclimate2132.

Doney, S.C., L. Bopp, and M.C. Long, 2014. Historical and future trends in ocean climate andbiogeochemistry. *Oceanography,* **27,** 108–119, doi:10.5670/oceanog.2014.14.

Dufresne, J. L., et al. 2013. Climate change projections using the IPSL-CM5 Earth System Model: from CMIP3 to CMIP5, *Climate Dyn.*, **40,** 2123-2165, doi:10.1007/s00382-012-1636-1.

Dunne, J. P., et al. 2013. GFDL's ESM2 Global Coupled Climate–Carbon Earth System Models. Part II: Carbon system formulation and baseline simulation characteristics*. *J. Climate*, **26**, 2247-2267, doi:10.1175/JCLI-D-12-00150.1.

Dunne, J. P., and C. Laufkötter, 2015. Ocean biogeochemistry in the Fifth Coupled Model Inter-comparison Project (CMIP5), *US CLIVAR Variations*, this edition.

Giorgetta, M. A., et al. 2013. Climate and carbon cycle changes from 1850 to 2100 in MPI-ESM simulations for the Coupled Model Intercomparison Project phase 5. *J. Adv. Mod. Earth Sys.*, **5,** 572-597, doi:10.1002/jame.20038.

Ito, T., M. Woloszyn, and M. Mazloff, 2010. Anthropogenic carbon dioxide transport in the Southern Ocean driven by Ekman flow. *Nature*, **463**, 80-85, doi:10.1038/Nature08687.

Ito, T., A. Bracco, C. Deutsch, H. Frenzel, M. Long and Y. Takano, 2015. Sustained growth of the Southern Ocean carbon storage in a warming climate. *Geophys. Res. Lett.,* doi: 10.1002/2015GL064320.

Long, M. C., K. Lindsay, S. Peacock, J. K. Moore, and S. C. Doney, 2013. Twentieth-century ocean carbon uptake and storage in CESM1(BGC). *J. Climate,* **26**, 6775-6800, doi:10.1175/JCLI-D-12-00184.1.

Lovenduski, N. S., M. C. Long, P. R. Gent, and K. Lindsay, 2013. Multi-decadal trends in the advection and mixing of natural carbon in the Southern Ocean. *Geophys. Res. Lett.*, **40**, 139-142, doi:10.1029/2012gl054483.

Marshall, J., and K. Speer, 2012. Closure of the meridional overturning circulation through Southern Ocean upwelling. *Nat. Geosci.*, **5,** 171-180, doi:10.1038/ngeo1391.

Meijers, A. J. S., 2014. The Southern Ocean in the Coupled Model Intercomparison Project phase 5. *Phil. Trans. Roy. Soc. A*, **372,** doi:10.1098/rsta.2013.0296.

Meinshausen M., S. J. Smith, K. Calvin, J. S. Daniel, M.L.T. Kainuma, J. F. Lamarque, K. Matsumoto, S. A. Montzka, S. C. B. Raper, K. Riahi, A. Thomson, G. J. M. Velders, and D. P.P. van Vuuren, 2011. The RCP greenhouse gas concentrations and their extensions from 1765 to 2300. *Climte Change,* **109**, 213–241, doi: 10.1007/s10584-011-0156-z.

Meredith, M. P., A. C. N. Garabato, A. M. Hogg, and R. Farneti, 2012. Sensitivity of the Overturning Circulation in the Southern Ocean to Decadal Changes in Wind Forcing. *J. Climate*, **25**, 99-110, doi:10.1175/2011jcli4204.1.

Milliman, J. D., 1993. Production and accumulation of calcium carbonate in the ocean: Budget of a nonsteady state. *Global Biogeochem. Cycles,* **7**, 927-957, doi: 10.1029/93GB02524.

Moore, J. K., K. Lindsay, S. C. Doney, M. C. Long, and K. Misumi, 2013. Marine ecosystem dynamics and biogeochemical cycling in the Community Earth System Model CESM1(BGC). *J. Climate*, **26,** 9291-9321, doi: 10.1175/JCLI-D-12-00566.1.

Sabine, C. L., R. A. Feely, N. Gruber, R. M Key, K. Lee, J. L. Bullister, R. Wanninkhof, C. S. Wong, D. W. Wallace, B. Tilbrook, F. J. Millero, T.-H. Peng, A. Kozyr, T. Ono, and A. F. Rios, 2004. The oceanic sink for anthropogenic CO_2. *Science*, **305**, 367-371, doi:10.1126/science.1097403.

Sallée, J. B., E. Shuckburgh, N. Bruneau, A. J. S. Meijers, T. J. Bracegirdle, Z. Wang, and T. Roy, 2013. Assessment of Southern Ocean water mass circulation and characteristics in CMIP5 models: Historical bias and forcing response. *J. Geophys. Res. - Oceans*, **118**, 1830-1844, doi: 10.1002/jgrc.20135.

Sarmiento, J. L., T. M. C. Hughes, R. J. Stouffer, and S. Manabe, 1998. Simulated response of the ocean carbon cycle to anthropogenic climate warming. *Nature*, **393**, 245-249, doi:10.1038/30455.

Taylor, K.E., R, J., Stouffer, and G. A. Meehl, 2012. An overview of CMIP5 and the experiment design. *Bull. Amer. Meteor. Soc.,* **93**, 485–498, doi: 10.1175/BAMS-D-11-00094.1.

Thompson, D. W. J., S. Solomon, P. J. Kushner, M. H. England, K. M. Grise, and D. J. Karoly, 2011. Signatures of the Antarctic ozone hole in Southern Hemisphere surface climate change, *Nat. Geosci.*, **4**, 741-749, doi:10.1038/ngeo1296.

Waugh, D. W., F. Primeau, T. DeVries, and M. Holzer, 2013. Recent changes in the ventilation of the Southern Ocean. *Science*, **339,** 568-570, doi:10.1126/science.1225411.

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ANNOUNCEMENTS

New US CLIVAR Working Group: Arctic Change and Possible Influence on Mid-latitude Climate and Weather

The dramatic retreat of perennial Arctic sea ice has been a wake-up call to the climate community that climate change may not necessarily be slow and steady or its impacts only of consequence in the far off future. The newly revealed open waters of the Arctic Ocean and the collapse of warm-season snow cover are known to have profound impacts on the energy balance of the Arctic. And just as heating anomalies in the tropics can influence weather around the globe, large heating anomalies in the Arctic basin may have ripple effects at lower latitudes, especially across the industrialized countries and population centers of the Northern Hemisphere. This Working Group will focus on better understanding the coupling between Arctic variability and mid-latitude weather.

The main objectives of the working group are:
1. Assess and synthesize existing knowledge on the links between Arctic climate change and mid-latitude weather variability including weather extremes;
2. Identify key questions and knowledge gaps, with a particular an attention on physical processes and scale interactions considering the relatively short time period and multiple components included in the hypothesized linkages;
3. Propose or recommend targeted measurements that will allow better understanding of Arctic climate variability and surface-atmosphere coupling;
4. Provide a preliminary assessment of the ability of current models to reproduce the correct relationship between Arctic and mid-latitude weather and climate variability. Small sample size in the observations remain a challenge, therefore modeling studies are needed to test for significance;
5. Coordinate our efforts with those of other national and international programs, such as SEARCH (Study of Environmental Arctic Change), CliC (Climate and Cryosphere), and IASC (International Arctic Science Committee), by including their members among our WG, in teleconferences, and possibly joint meetings; and
6. Inform funding agencies through US CLIVAR Interagency Group and the IARPC (Interagency Arctic Research Policy Committee) of opportunities for advancing scientific understanding of Arctic influences on mid-latitude climate.

Working Group Members

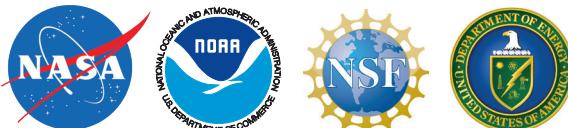|			
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