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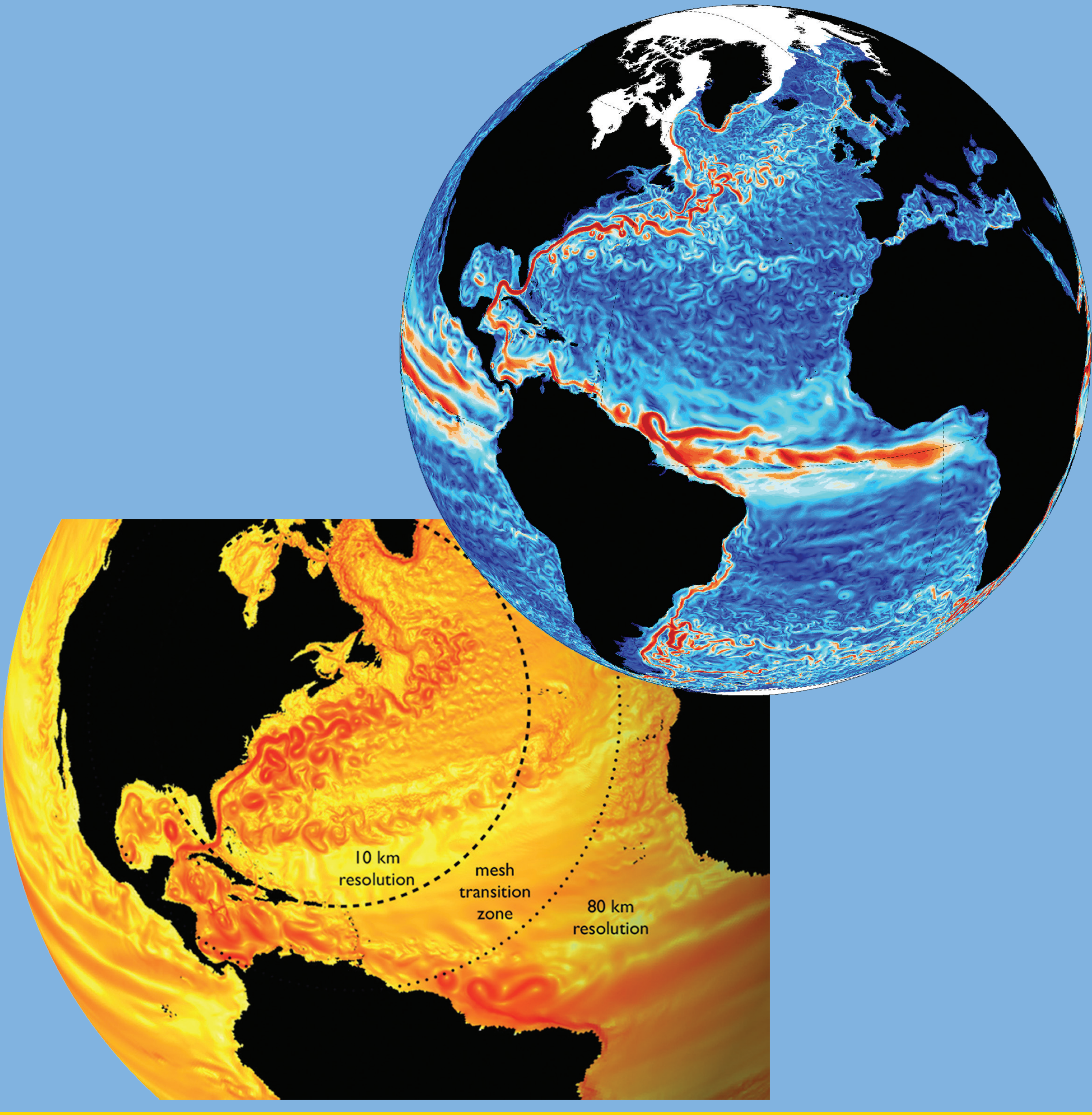

CLIVAR Ocean & Climate: Variability, Predictability and Change is the
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Editorial

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It is with great pleasure, as one of my first acts as incoming Executive Director of the ICPO, that I introduce this excellent Special Issue of Exchanges, based on the outcomes of the OMDP Workshop on High Resolution Ocean Climate Modelling that was held on April 7-9 2014 in Kiel, Germany. It also provides the opportunity to introduce to you some recent developments in the evolution of the CLIVAR project.

The CLIVAR Scientific Steering Group (SSG) at its 20th Session in May 2013, reviewed the overall direction of CLIVAR in the context of recent developments in the WCRP and decided to rename the project: CLIVAR – Climate and Ocean: Variability, Predictability and Change. This went hand in hand with a renewed emphasis on the role of the oceans in the coupled climate system and also with a new organizational scheme for CLIVAR. In addition to the traditional Panels and Working Groups, the SSG saw a need for a mechanism to rapidly organize research around new topics that would benefit from international coordination and cooperation and that could be expected to yield results on a relatively short time-horizon. These topics are referred to as CLIVAR "Research Foci" (short descriptions of current topics can be found at http://www.clivar.org) and teams of scientists are developing science and implementation plans. These will be further refined at the Pan-CLIVAR meeting in The Hague, Netherlands, 14-18 July 2014 and presented to the SSG at its meeting in November of this year. The SSG will consider proposals for new Research Foci as topics arise. Several of the CLIVAR Research Foci are also directly related to the WCRP Grand Science Challenges (http://www.wcrp-climate.org/index.php/grand-challenges).

The CLIVAR Ocean Model Development Panel (previously WGOMD) provides leadership in the wider WCRP context on issues related to modelling the ocean as a component of the climate system, addressing ocean modelling issues arising from CLIVAR and WCRP panels and working groups. WGOMD activities have included (1) the Coordinated Ocean-ice Reference Experiments (CORE), (2) contributions towards ocean analysis of CMIP5 simulations, (3) the development of the CLIVAR Repository for Evaluating Ocean Simulations (REOS), and (4) the organization of international science workshops. As a "panel" rather than a "working group", the OMDP will take on some additional responsibilities and revised Terms of References will be presented at the November 2014 Session of the CLIVAR SSG.

High-resolution ocean modelling is required for many scientific climate studies and practical applications, including regional climate information on sea level and extremes. Mesoscale eddies are certainly the most spectacular and most energetic dynamics in the ocean, but the workshop discussions also included discussion on all climate-relevant processes that are deeply impacted by ocean model resolution. These include the representation of western boundary currents and major fronts, emerging modes of air-sea interactions, including air-sea coupling in upwelling areas, exchanges with marginal seas, and processes on continental shelves.

The workshop outcomes, many of which are summarized in this Special Issue make important contributions to the goals of the all the CLIVAR Research Foci. High resolution ocean climate modelling will contribute to the improved representation of some key physical processes, for example, air-sea interaction and feedbacks, oceanic mesoscale and sub-mesoscale variability, upwelling, and boundary currents that are at the heart of these Research Foci topics. The topic is also highly relevant for the WCRP Grand Challenges, particularly those seeking regional information on small spatial scales, thus requiring improved understanding and simulation of physical processes on these scales.

Special thanks are in order to the guest editors of this issue, Drs Anne Marie Treguier and Gokhan Danabasoglu and to the many contributing authors, and to Dr Anna Pirani, newly appointed Editor of CLIVAR Exchanges.

Participants of the WGOMD Workshop on High Resolution Ocean Climate Modelling held on April 7-9 in Kiel Germany. See Treguier et al (2014, this issue) for an overview and the articles in this Special Issue contributed by workshop invited speakers. The workshop organisers with to thank the generous support provided by the cluster 'The Future Ocean' and the GEOMAR Helmholtz Centre for Ocean Research Kiel, WCRP and CLIVAR, ONR, NASA, NSF, NOAA, and DOE, U.S. CLIVAR.

Figures on cover page. Top: Snapshot (1992/12/01) of surface current speeds simulated by the NEMO ocean/sea-ice model in the global ORCA12 configuration. The spatial resolution of ORCA12 varies from 9.2 km at the equator to 2.5 km at high latitudes. This 50-year simulation was performed in the framework of the DRAKKAR project at CINES, the GENCI computing center located in Montpellier (grant X201301727). Figure courtesy of J.-M. Molines, CNRS-LGGE, Grenoble, France. Bottom: Figure 2 in article by S. Danilov & T. Ringler, on page 46: A snapshot of fluid kinetic energy at a depth of 100 m from a MPAS-Ocean, global, multi-resolution simulation. The mesh resolution is 10 km with the region of the North Atlantic containing the Gulf Stream, North Atlantic Current and Northwest Corner. Away from these currents of interest, the mesh transitions smoothly to 80 km.

CLIVAR WGOMD Workshop on high resolution ocean climate modelling: outcomes and recommendations

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The CLIVAR Working Group on Ocean Model Development (WGOMD) organised the "High Resolution Ocean Climate Modelling Workshop" in Kiel, Germany, during April 7-9, 2014. The workshop gathered 60 international scientists, with attendance intentionally limited to enable lively interactions and discussions among the participants. Thanks to the help of the GEOMAR team, talks and discussions were broadcast online and followed each day by over 30 additional scientists around the world.

This special issue of CLIVAR Exchanges is devoted the workshop topics, with a collection of short articles prepared by many of the invited speakers. Further information about the workshop, including the full presentations and abstracts, is available at http://www.clivar.org/wgomd/highres.

1. Workshop objectives

Ocean observations and numerical models reveal a turbulent ocean, full of eddies and thin energetic currents at the so-called "mesoscale" (spatial scales of the order of 10-200 km). These eddies and swift boundary currents are responsible for much of the global heat and salt transports. They control the mechanisms of spreading of deep waters, preconditioning of deep convection, subduction and CO_2 sequestration, and the stratification of the surface layers, thus also impacting the global biogeochemical cycles. They modulate air-sea interactions through their effects on air-sea heat fluxes and wind stress. The complexity of mesoscale dynamics is a challenge for parameterisations: those presently used in low resolution, i.e., non-eddy-resolving, climate models are relatively crude and neglect non-local effects. All these considerations motivate the push toward fully "eddy resolving" ocean components for coupled earth system models.

As a preliminary note, let us define a few words, which have a clear meaning for the workshop participants but may seem otherwise a bit vague for the scientific community in general:

- "coarse resolution" or "low resolution" designates the ocean component of climate models with the typical resolution used in CMIP5 (1° grid). These models do not allow the development of mesoscale eddies because they do not resolve the relevant dynamical scale (Rossby radius wavelength, of order 200-300 km between 20-40 degrees of latitude). It is assumed that one needs at least 12 grid points to represent a wavelength, equivalent to two grid points per Rossby radius (see Hallberg, 2013, figure 1).
- "eddy permitting" models are those with typical grid scales of 1/4°, where baroclinic eddies are allowed to grow but are marginally resolved;
- in "eddy resolving" models the full dynamics and life cycle of baroclinic eddies should be represented realistically. Presently models in the 1/10° resolution range claim to be "eddy resolving" in the subtropical and tropical latitudes, but this is still a matter of opinion. The comparison of these higher resolution models is underway and the terminology may evolve in the future.

A few climate modelling centres have begun to run ocean components at "eddy resolving" resolutions (1/10°); many centres will use "eddy permitting" ocean models in the 1/4° resolution range for the next CMIP exercise. Meanwhile, ocean modellers are also exploring the "submesoscale" (kilometre scale) range, simulating the processes by which eddies are dissipated through filamentation or shear instabilities, as well as other dynamical phenomena such as internal waves. These recent evolutions in ocean modelling set the stage for this workshop. The overarching goal was to bring together modelling groups conducting high-resolution ocean – sea-ice and / or fully coupled earth system simulations as well as groups working in process orientated studies / theory, in order to review the state-of-science, including new sensitivities and processes emerging in these high resolution simulations.

We note that the workshop is strongly aligned with the science needs of several WCRP Grand Challenges and CLIVAR Research Foci, e.g., dynamics of regional sea level variability; dynamics of upwelling; decadal variability and prediction, towards the provision of climate information at regional scales.

2. Overview of the presentations and discussions

The workshop was organised in three parts. It started with an introductory session with review talks on the state of current understanding, some of which are summarised in this issue. Griffies presented the advances that have taken place since the 2009 WGOMD Workshop on Ocean Mesoscale Eddies in Exeter, U.K. The problem of large scale biases in temperature and salinity arises when asking the question "are coupled climate models with eddies more scientifically useful than low resolution models?". Both Griffies and Hasumi conclude that despite remarkable improvements in the circulation, high resolution is not an automatic remedy that eliminates large-scale model biases (Griffies, 2014 and Hasumi, 2014, this issue).

The impacts of ocean eddies and of boundary currents on the atmospheric circulation are a very active field of research. Dewar and Small reviewed recent progress; Justin Small summarised the outcomes of the CLIVAR Frontal Scale Air-Sea Interaction Workshop held in Boulder 2013, http://www.cgd.ucar.edu/events/fsasi-workshop/. It is now widely accepted that the strong gradients in western boundary currents have an impact on the variability of the troposphere (Kelly et al. 2010, Kwon et al, 2010), although the vertical penetration and strength of the air-sea coupling mechanisms at different space and time scales (e.g., Sheldon and Czaja 2013) and how far such local response extends into basin-scale atmospheric circulation response are still being investigated. The interpretation of the atmospheric momentum response is especially complex, with different processes at play (pressure gradients, vertical mixing, etc.) and with a dependency of sea surface temperature (SST) – wind coupling on numerical parameters and time and space scales. The intimate coupling between the atmosphere and ocean raises questions about our current approaches to force ocean models (see Deremble and Dewar, 2014, this issue).

In the second part of the workshop, about 20 groups carrying out high resolution ocean simulations relevant for climate presented the current status of their simulations, the scientific questions they are applying high-resolution simulations to, and the main challenges they see to progress in high-resolution modelling. Many groups are already running coupled ocean – sea-ice – atmosphere simulations at resolutions of order 10 km or are planning to do so in the near future: e.g., GDFL, NCAR and DOE (USA), U.K. MetOffice, MPI Germany, JMA/MRI and MIROC (Japan), ARC (Australia), EC-Earth European consortium, CMCC (Italy). Some groups, e.g., JAMSTEC, HYCOM, Drakkar, and MIT, are venturing into the submesoscale range (1/20° to 1/48°), either globally or by using nesting techniques. As all the presentations are available on the workshop web site, here we provide only a brief summary of the challenges that were identified during these presentations:

- Model spin up and initialization: with costly high resolution ocean models, this becomes a burning issue. Which techniques can make initialization shorter, less computationally expensive? How to initialise coupled climate forecasts with eddy permitting or resolving ocean components?
- Parameterisations: which subgrid-scale parameterisations are adequate? How should they be modified depending on ocean model resolution? What is the appropriate choice of parameterisations and numerical schemes?
- Forcing the ocean/coupling to the atmosphere: what are the appropriate methods, e.g., for calculations of wind stress and freshwater fluxes? What are the best datasets?
- Technical issues: how much data from the high resolution simulations needs to be stored for model analysis? Is regridding or coarsening stored data a way forward? Where do we stand in terms of code scalability on massively parallel machines? How do we develop more efficient analysis tools?
- Strategies regarding resolution and regionalisation: how to choose the resolution in the ocean and in the atmosphere for coupled simulations? Do we need to increase resolution globally or rather use nesting techniques? How to ensure a better representation of local processes (e.g., marginal seas, shelf seas)?
- Model validation and testing: how to tune parameters and test parameterisations when ocean models become computationally very expensive? Do we have adequate observational datasets to validate high resolution ocean models?

The third part of the workshop was devoted to four topical sessions in which recent advances were presented and the above challenges were discussed in more detail.

Ocean physical processes, air-sea interactions, and their parameterisations

Speakers in this session presented processes that are now being investigated in submesoscale-resolving models at the kilometre scale: processes related to tides, such as the generation of internal tides and their role in vertical mixing (see Arbic et al. 2014, this issue); wave drag arising from flow-topography interactions (Nikurashin); vertical velocities driven by submesoscale dynamics and their impacts on biogeochemistry (Hogg). These processes are part of the ocean energy cycle. As pointed out by Nikurashin, kinetic energy is dominated by surface-intensified mesoscale eddies while the dissipation is concentrated at the bottom. Eden presented a new framework for energetically consistent ocean models (Eden and Olbers 2014), separating the kinetic energy of transient motions into eddies, internal waves, and turbulence. Despite such advances in theoretical frameworks, the precise quantification of energy pathways in the ocean is still in its infancy.

Allowing eddies in the ocean components of climate models reveals new air-sea coupling mechanisms, such as the influence of the Kuroshio Extension variability on the Pacific storm tracks (Taguchi). Latif and Zhou presented a new estimate of the tropospheric response to mid-latitude SST anomalies in the Pacific Ocean. These preliminary results, regarding the influence of daily SST variability, are puzzling, and confirm that more investigation with high resolution ocean-atmosphere models are needed in order to build a synthetic, consensus view.

In ocean modelling, “high resolution” is a moving target. The need to simplify and unify numerical tools to deal with large scale to submesoscale ocean motions motivates the concept of “scale-aware physical parameterisations” as discussed by Fox-Kemper (2014, this issue). The Leith (1996) closure, based on the concept of enstrophy cascade, gives more promising results than Smagorinsky or hyperviscosity schemes, as shown in a recent intercomparison of two-dimensional turbulence by Pietarila-Graham and Ringler (2013). Further work on the extension of this scheme to anisotropic and quasi-geostrophic regimes is underway, with the goal of defining parameterisations that would be valid over a large range of spatial scales without the need for retuning of parameters (Fox Kemper, 2014, this issue).

A large portion of the distinct water masses of the global ocean is formed in marginal seas and enter the deep ocean through narrow passages by so-called “overflow processes”. A review of the state of the art is given by Chassignet (2014, this issue): many challenges remain even in high resolution models, especially in the most widely used level-coordinate models. Last but not least in the zoo of oceanic processes, instabilities associated with ocean – sea-ice coupling emerging at high resolution have been presented and discussed (Hallberg, 2014, this issue).

3. New technical and methodological challenges for ocean simulations

High resolution models bring new questions regarding vertical and horizontal grids. Based on idealised test cases using different models (MIT-GCM, ROMS, MOM, GOLD), with different numerical schemes and vertical coordinate systems, Ilicak et al (2012) found a strong link between spurious diapycnal mixing and the lateral grid Reynolds number. Their conclusion was that lateral viscosity, an often overlooked parameter, had to be high enough in order to keep the grid Reynolds number low. Adcroft presented different types of Arbitrary Lagrangian Eulerian (ALE) coordinates that seem to be the way to go to provide flexibility for the vertical grid of ocean models. Regarding horizontal (e.g., longitude-latitude) grids, new developments are all aimed at flexible grid refinement allowing an increase of spatial resolution in the regions of interest. One example is the MPAS (Model for Prediction Across Scales) dynamical core developed jointly by the National Center for Atmospheric Research (NCAR) and the U.S. Department of Energy (DoE) Los Alamos National Laboratory (LANL) (Ringler et al., 2013). This model has an unstructured grid based on Voronoi tesselations and has been tested in global configurations, although it is not quite ready to be run in the forced ocean – sea-ice configuration, following the CORE protocol (Maltrud). Other global models based on unstructured meshes are developed at AWI; a recent review is provided by Danilov (2013); see also Danilov et al., 2014, this issue.

Eddying ocean simulations pose a new challenge because of the emergence of intrinsic variability caused by oceanic instabilities (Penduff et al., 2011; also Penduff et al, 2014, this issue). Together with the increased synoptic variability in high resolution atmospheric models, these extra modes of variability in the ocean make ensemble simulations necessary to study cause and effect relationships in the Earth's climate. Both high resolution and ensemble simulations will require new strategies for data storage and analysis in the near future.

The conclusions of the session on "Downscaling" are presented in the papers by Biastoch et al., 2014, Tatebe et al. 2014 (this issue).

4. Workshop conclusions and recommendations

All participants noted the challenge and need for making progress with costly high resolution models in a coordinated way. One thought-provoking idea was put forward: instead of having multiple, never satisfactory "eddy permitting" solutions, why not team together and produce a single "truly eddy resolving" global solution that would serve as the baseline to tune parameterisations in coarse climate models? The conclusion was that such an effort is not yet feasible, because any single model solution is bound to be flawed in some respects (e.g., due to insufficient knowledge of the atmospheric forcing) and will require parameterisations (we are still far from direct numerical simulations in the ocean). Given the need for multiple simulations, coordination between groups at the international level is the best way to ensure that resources are used in the most efficient way. The CORE-I protocol for climatological simulations (Griffies et al. 2009) and the CORE-II protocol for interannual simulations (Danabasoglu et al. 2014) have demonstrated usefulness of coordinated efforts and groundwork laid out by them should be applicable to high resolution ocean models in the near future, although improvements will be required in the atmospheric datasets (e.g., higher resolution) and in the forcing protocols (e.g., simulating more coupled processes in the atmospheric boundary layer while keeping the atmospheric forcing close to an observed state).

The development of shared modules for parameterisations appears a sound way both to pool resources for development and to allow rigorous testing of subgridscale physics in more than one model framework. An interesting example is the CVMix (Community ocean Vertical Mixing) project aiming to provide dynamical-core independent modules for vertical mixing parameterisations used in ocean modelling community. CVMix is a collaborative effort between NCAR, GFDL, and LANL. It will facilitate intercomparison of model solutions at the process level.

The need for resolution to build climate scenarios at the regional scale is thus unavoidable, and there is at present no single strategy to tackle this problem: unstructured meshes, nesting and high resolution uniform grids all need to be developed and tested further. Although "scale aware parameterisations" may help modellers best exploit their simulations in different resolution regimes - from coarse to "eddy permitting" to "eddy resolving" to "submesoscale resolving", it was pointed out that the main added value of resolution in ocean models is not related to eddy dynamics. Rather, solutions become more realistic as the bathymetry is better resolved, as key passages, marginal seas, overflows and river runoffs are better represented.

High resolution coupled climate models offer new opportunities and may help us tap into an unexpected potential for predictability in the climate system. This is exemplified by the recent study of Scaife et al. (2014). These results are puzzling and a theoretical explanation is still lacking. Clearly, more research is needed with different high resolution models to understand the mechanisms that give rise to climate predictability, at seasonal to decadal time scales.

Finally, the workshop participants expressed their concern regarding the difficulty to keep ambitious model developments (> 10 years) going despite the short funding cycles (3-4 years). Another concern is the difficulty to offer adequate training and attractive careers to young ocean modellers, given the incompatibility between model developments that may require many years to reach fruition and high rates of publication now becoming the norm for advancement of early career scientists. We encourage international bodies such as CLIVAR and WCRP Modelling Advisory Council (WMAC) to address these issues.

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A review on ocean resolution dependence of climate biases in AOGCMs

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1. Introduction

Resolution of climate models has steadily increased. Of the 23 coupled atmosphere-ocean general circulation models (AOGCMs) listed in the fourth assessment report of the Intergovernmental Panel on Climate Change (IPCC; IPCC, 2007), the horizontal grid size for the ocean components is 1.5° or coarser in about half the number. In the recently published fifth assessment report (AR5, IPCC, 2013), on the other hand, more than half of the 50 listed AOGCMs adopt ocean horizontal grid size of 1° or finer. We can access products of these AOGCMs via the Coupled Model Intercomparison Project (CMIP) database to assess the dependence of modelled climate on ocean resolution. However, most AOGCMs in IPCC AR5 remain in the regime of eddy-less ocean, i.e., oceanic mesoscale eddies are not represented therein. It is a well established understanding that ocean models exhibit qualitatively different behaviour depending on whether or not their resolution is sufficiently fine to represent mesoscale eddies (Hecht and Hasumi, 2008). In this regard, we are yet unable to investigate the most interesting part of the ocean resolution dependence based on the IPCC/CMIP experiments.

Ocean models come into the eddy-rich regime when the horizontal grid size becomes finer than 0.5°, at which resolution mesoscale eddies are marginally represented (or "permitted") at low and mid latitudes. Mesoscale eddies are "resolved" in most oceanic regions when the horizontal grids are refined to ~0.1°. These two categories, eddy-permitting and eddy-resolving, are collectively referred to as "eddying" hereafter. There are two AOGCMs listed in IPCC AR5 whose ocean component is in the (marginally) eddying regime. Apart from the IPCC/CMIP frame work, some more AOGCMs are being used in the eddying ocean regime. We can now find several among such AOGCMs whose control climate is comprehensively described in published documents and compared with that of lower ocean resolution versions of the same model suites. Based on these published documents, this article briefly reviews the dependence of climate biases of AOGCMs on ocean resolution.

2. Models

Five AOGCMs with eddying ocean components are picked up here, whose oceanic and atmospheric resolution is listed in Table 1. In the papers cited in Table 1, climate biases of the eddying models are compared with those of the non-eddying version models (whose resolution is also listed in Table 1). It should be noted that the configuration of the atmospheric component (including resolution) is the same between the eddying and non-eddying versions in some model suites (CCSM and MPI-ESM) while it is not the case in the others (GFDL CM, HiGEM, and MIROC). It should also be noted that a number of aspects, other than resolution, are different among these five models, such as the method and length of spin-up integration, the configuration (the year at which climate forcing is fixed) and duration of control integration, and eddy parameterisation used in the ocean component.

3. Resolution dependence of climate biases

Biases in sea surface temperature (SST) are the most important and indicative measure for the performance of AOGCMs. Their global magnitude and large scale patterns are controlled mostly by surface and cloud albedo. In this regard, sea ice distribution could significantly influence the large scale SST biases. Eddying resolution significantly improves representation of local oceanic features, especially in the regions of strong influence of western boundary currents and coastal upwelling and in the equatorial Pacific. In such regions, high ocean resolution helps considerably to reduce SST biases. Surface current representation also affects sea ice distribution and thereby SST. The sea ice-covered high latitudes are also the region of deep water formation, which is the start point of the global-scale deep overturning circulation and play a role in the global-scale heat transport. Each of these aspects is presented below. The features described below are all taken from the papers listed in Table 1 although they are not cited every time.

3.1 Regions of strong influence of western boundary currents

Eddying resolution leads to a drastic improvement in reproducibility of oceanic western boundary currents. For example, the Kuroshio runs along the south of Japan with a width of ~100 km, separates from the eastern coast of Japan at ~35°N, and continues to flow eastward as the Kuroshio Extension. Such a feature is generally well captured by eddying ocean models, while the Kuroshio in non-eddying ocean models is unrealistically broader and overshoots to the north, flowing along the east coast of Japan to ~40°N, and the Kuroshio Extension becomes a weaker current and flows far to the north of the actual position (Hasumi et al., 2010). Since the Kuroshio and the Kuroshio Extension transports a large amount of heat from lower latitudes, this overshooting causes a conspicuous high SST bias in AOGCMs extending eastward from the Japan's coast along ~40°N. A high SST bias at non-eddying resolution and its correction at eddying resolution are a common feature among these selected models.

Another conspicuous source of SST bias associated with western boundary currents is the behaviour of the Agulhas Current. In reality, its retroflection at the southern tip of Africa prevents direct intrusion of the Agulhas Current water into the South Atlantic. Non-eddying ocean models often fail in reproducing the retroflection and simulate a large subtropical gyre connecting the Indian and South Atlantic Oceans. This misrepresentation also deteriorates the Antarctic Circumpolar Current (ACC) and associated ocean fronts in AOGCMs. The SST biases induced by these failures of non-eddying models are commonly improved in the eddying models. The path of the ACC is also heavily dependent on resolution. One of the typical failures of non-eddying models, as with the ACC, is the Malvinas Current flowing northward along the Argentine coast, resulting in a cold SST bias near the southern end of Argentina and a warm SST bias to the north. These biases are also commonly reduced in the eddying models.

On the other hand, the models commonly exhibit a dipole of SST biases at mid latitudes of the North Atlantic (SST is too high along the east coast of America and too low offshore), both at non-eddying and eddying resolution. This is due to the ill-representation of the path of the Gulf Stream. It is pointed out that the separation of the Gulf Stream is affected by a number of factors, such as representation of the North Atlantic deep western boundary current, and its realistic representation is not automatically guaranteed even at ~0.1° resolution (Bryan et al, 2007). The highest resolution model, CCSM, does a better job in this regard, but the separation point of the Gulf Stream is still a bit too far to the north of the actual latitude.

3.2. Regions of coastal upwelling

Non-eddying models commonly fail in reproducing coastal upwelling systems and induce warm SST biases in those regions. Eddying resolution helps reduce these biases in some cases, but other factors are also important.

For example, warm SST biases over the coastal upwelling zones west of America (both North and South) are corrected in GFDL CM but not in MPI-ESM. The fundamental cause of these coastal upwelling systems are the persistent along-coast winds, and these winds are strongly controlled by the orographic effect of high mountains. The surface topography is often heavily smoothed in low resolution atmospheric models for the sake of numerical stability, so orographically-controlled winds are underrepresented in such models. In this regard, MPI-ESM employs the same, relatively low resolution atmospheric model both in non-eddying and eddying AOGCMs, while the atmospheric resolution of the eddying version GFDL CM is significantly higher than its non-eddying version (and MPI-ESM). Therefore, resolution of atmospheric models seems to be of more importance in this case. However, the orographic control of winds is not the sole source of problem, as understood by the next example.

All the five eddying AOGCMs suffer from too high SST over the Benguela upwelling system west of Africa, unimproved compared with non-eddying versions. It has been pointed out by various studies that stratocumulus clouds over the upwelling regions are also very important for SST via their influence on radiation. Furthermore, there is a feedback loop among winds, clouds, radiation, and SST over the upwelling regions, and an error in any of these factors ends up in enhancing the SST bias (e.g., Grodsky et al., 2012). Current AOGCMs, whether or not their ocean component is eddying, generally suffer from this vicious cycle, especially in the case of the upwelling region west of Africa. The SST bias in the Benguela upwelling system is transported and deteriorates the SST reproducibility in the equatorial Atlantic region, too.

Table 1. List of AOGCMs and their horizontal resolution

Model	High res. grid size		Low res. grid size		Reference
	Ocean	Atmosphere	Ocean	Atmosphere	
CCSM	0.1°	0.625°×0.5°	~1°×0.5°	0.625°×0.5°	Kirtman et al. (2012)
GFDL CM	1/4°	~50 km	~1°	~200 km	Delworth et al. (2012)
HiGEM	1/3°	1.25°×0.83°	~1.5°	1.875°×1.25°	Shaffrey et al. (2009)
MIROC	1/4°×1/6°	0.5625°	~1.5°	2.8125°	Jungclaus et al. (2013)
MPI-ESM	0.4°	1.875°	~1.5°	1.875°	Sakamoto et al. (2012)

3.3. Equatorial Pacific

In contrast to the case of the equatorial Atlantic region, high ocean resolution is known to greatly help improve reproducibility of the equatorial Pacific region (e.g., Roberts et al., 2009). Non-eddying AOGCMs commonly simulate too low SSTs in the Pacific equatorial upwelling region, and this bias is commonly corrected in the eddying AOGCMs. The improvement is primarily due to the success in reproducing tropical instability waves (TIWs) that transport heat meridionally towards the equator. Non-eddying models also tend to simulate a too fast South Equatorial Current, which deteriorates the cold bias. TIWs also transport momentum and make the South Equatorial Current slower, leading to a further reduction of the cold bias in the eddying AOGCMs.

Variability of the Pacific equatorial SST is also significantly improved by eddying resolution. The region of ENSO-related high SST variability, which lies from the Peruvian coast to around the International Date Line, tends to separate from the Peruvian coast and extend too much westward in non-eddying AOGCMs. This bias is commonly corrected among the eddying AOGCMs. Some eddying AOGCMs exhibit improvement of the ENSO skewness (asymmetry of the magnitude of SST anomalies between warm and cold events) because TIWs are more actively generated for cold events and reduce negative SST anomalies, but this is not the case in other eddying AOGCMs.

3.4. Sea ice and deep water formation

Some models exhibit drastic changes of sea ice cover between the non-eddying and eddying resolution versions, but such a difference is not necessarily due to ocean resolution. Model parameters, especially the albedo of ice/snow surface in this case, are often differently tuned between non-eddying and eddying resolution versions. But there is a common improvement caused by ocean resolution among the eddying AOGCMs in the northern hemisphere: a retreat of sea ice edge on the Atlantic side of the Arctic Ocean (or in the Barents Sea). It is due to the intrusion of the Norwegian Atlantic Current, a continuation of the Gulf Stream. This significant change of sea ice extent could further affect the climate over a wider region. Indeed, in the case of CCSM, which uses the identical atmospheric model setup between the non-eddying and eddying versions, SST is significantly higher in the eddying resolution version over a wide area of the northern hemisphere.

In the southern hemisphere, some models exhibit persistent large open ocean polynyas off the Weddell and Ross Seas, which is not the case in reality, irrespective of ocean resolution. Such open ocean polynyas indicate that deep convection takes place and deep waters are formed there, which is not the case in reality, either. Other models do not show such a feature, but it does not mean that the sites and processes of deep water formation are correct therein. It seems that no fundamental improvement is achieved as to the deep water formation around Antarctica in the eddying AOGCMs relative to the non-eddying versions.

On the other hand, the deep water formation sites in the northern North Atlantic are significantly improved in the eddying AOGCMs. The deep convection occurring in the Greenland and Labrador Seas is preconditioned by the doming up of deep isopycnals, which is associated with the wind-driven cyclonic subpolar gyre. The deep convection is also heavily affected by the boundary currents flowing around Greenland (the East and West Greenland Currents which transport fresh Arctic water, and the Irminger Current, which transports warm Atlantic subtropical water). Representation of these gyre and boundary currents is considerably improved by high ocean resolution. However, intensity of the deep water formation, or the depth to which waters of higher density sink, is strongly controlled by factors other than oceanic processes (namely, cold air break, freshwater input at high latitudes, etc.), so it is not necessarily improved in the eddying AOGCMs even if the deep water formation sites become better represented.

3.5. Deep oceanic structures

Resolution dependence of deep oceanic structures in AOGCMs is not discussed in detail in the referenced documents. These documents compare the global-scale deep meridional overturning circulation (MOC) between the non-eddying and eddying versions, and all seem to show a reduction of the Atlantic MOC intensity by employing eddying resolution. However, the reason is unclear and may be different among the models. Eddying models still have deficiencies in properly representing deep water formation, as noted above. We need to extend our understanding on the various processes involved in deep water formation, such as down-sloping motion of dense water and entrainment, to make a good assessment on the resolution dependence of MOC representation.

Although it is not discussed in the referenced documents (except for only brief notes), the representation of mode and intermediate waters may be significantly improved in the eddying resolution models. Formation of these waters heavily depends on a number of oceanic features that are significantly improved by eddying resolution, such as the position of oceanic fronts and subduction induced by mesoscale eddies. Mode and intermediate waters are considered to play important roles in climate variability of interannual to multi-decadal time scales. It is important to investigate this further.

4. Summary

Eddying ocean resolution leads to a significant reduction of climate biases across different models in some aspects, such as those around western boundary currents and in the equatorial Pacific region. But there still remain biases uncorrected by eddying ocean resolution, such as those in the coastal upwelling regions and at deep water formation sites. For these yet-uncorrected biases, different models sometimes look to show different behaviours in terms of dependence on ocean resolution. It is difficult to identify the causes of such biases from these existing model experiments because they are differently designed and analysed. Well-coordinated model experiments and metrics for analysis would help improve our understanding on the dependence of climate biases on ocean model resolution.

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Climate modelling with an energetic ocean mesoscale

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1. A new era with mesoscale eddying climate models

The climate science community is entering an exciting and challenging new era in which climate models are starting to admit a portion of the ocean mesoscale spectrum (e.g., Yokohata et al. (2007), Shaffrey et al. (2009), Roberts et al. (2009), Bryan et al. (2010), McClean et al. (2011), Farneti et al. (2010), Delworth et al. (2012), Kirtman et al. (2012), Bryan et al. (2014), Winton et al. (2014), Griffies et al. (2014)). Such climate models have evolved from compelling efforts in regional and quasi-global ocean forced simulations, such as those from Semtner and Chervin (1992), Smith et al. (2000), Maltrud and McClean (2005), and Hallberg and Gnanadesikan (2006) (see Hecht and Hasumi (2008) for a review). Quite simply, mesoscale eddy-rich ocean simulations look more like the ocean, thus enabling a new breed of meaningful comparisons between simulations and observations (e.g., McClean et al. (2006), Penduff et al. (2010), Petersen et al. (2012)). Consequently, such models entrain passionate interest from observational oceanographers, thus facilitating valuable interactions between sub-disciplines of ocean and climate science. Additionally, mesoscale eddying models hold the potential to remove much of the uncertainty related to mesoscale eddy parameterisations required at the coarse resolution.

In this note, we introduce certain problems and prospects related to mesoscale eddying ocean models, with a focus on their role in global climate simulations. Among the problems with ocean mesoscale eddying climate models, it is important to acknowledge their extreme cost. Given the long time scales of ocean circulation, it remains rare to see all but the most idealised eddying ocean simulations run to thermodynamic equilibrium. Furthermore, such models are hugely taxing on data archives, analysis tools, and the patience of analysts who struggle to manipulate the huge datasets. Difficulties developing and analysing such models have kept them largely in the hands of a few of large modelling centres, with little opportunity for extensive model tuning available with coarser models. Even so, scientific progress is happening, and model development experiences are spreading.

2. Ocean model development with an active ocean mesoscale

We aim to present a suite of issues related to the development and analysis of climate models possessing an active ocean mesoscale. Space limitations warrant omissions of important topics, though many are addressed in other articles within this issue of CLIVAR Exchanges. If this document provokes discussion, indeed disagreement, then it has served its purpose.

2.1. Concerning the needs of grid resolution

There are many flavours to the question about grid resolution, and we introduce some of the issues here.

2.1.1. Rossby radius and mesoscale eddies

In a milestone paper, Smith et al. (2000) presented a simulation of the Atlantic at 0.1° resolution. They made a compelling case that eddy energetics approached that of the satellite measures, motivating them to conclude that the mesoscale field was resolved, at least in the middle latitudes. They based their resolution requirement partly on noting what grid spacing is needed to represent the first baroclinic Rossby radius, with this scale related to the eddy scale. Note that the Rossby radius, R, is related to the baroclinic Rossby wavelength, λ, via $R = \lambda / 2\pi$. If the grid spacing Δ satisfies $\Delta < R$, then $2\pi\Delta < \lambda$, which is the traditional criteria for resolving a wave on a discrete grid. Hallberg (2013) chose the more conservative criteria in which $2\Delta < R$. The associated grid spacing from Hallberg's analysis extends from roughly 100 km in the deep tropics to finer than 2 km in the high latitude shelf regions (see Figure 1 in Hallberg, 2013)). The 0.25° Mercator spacing used in such models as ORCA025 (Barnier et al., 2006) or GFDL-CM2.5 (Delworth et al., 2012) resolves the Rossby radius in most deep ocean regions equatorward of 30°. This is the resolution targeted by certain groups for use in climate simulations during the next 5-10 years. In contrast, the 0.1° Mercator spacing used by Smith et al. (2000) and GFDL-CM.6 (Delworth et al., 2012), which resolves the Rossby radius for most regions equatorward of 50°, remains too expensive on today's computers for use in routine centennial-scale climate simulations.

Although there is a connection between the Rossby radius and mesoscale eddy scales, it is unclear what resolution is required for ocean simulations to reach numerical convergence, whereby key features of the simulations, such as mesoscale eddy contributions to tracer budgets, remain unchanged upon refining the grid. Additionally, in many regions, second or higher baroclinic modes are important, with these modes having finer horizontal and vertical scales. This issue raises the further question about the vertical, with refined vertical resolution required to accurately capture the enhanced vertical motions and structure within eddy features. However, there are few studies that consider the needs of vertical resolution in concert with the refinements of horizontal spacing. Given such uncertainty, we recommend eschewing the term "eddy resolving". Instead, we prefer "mesoscale eddy-active" or mesoscale eddy-rich", or the more generic "mesoscale eddying". Such terms offer no pretence that the simulation "resolves" mesoscale features, while they acknowledge the presence of transient fluctuations associated with a geostrophically turbulent flow.

2.1.2. Spurious dianeutral mixing

Related to the question of grid resolution is the issue of spurious dianeutral mixing associated with discrete advection errors. As discussed by Griffies et al. (2000), the fundamental issue is that in regions where geostrophic eddies are most active, tracer variance cascades preferentially to the grid scale. Dissipation methods are required to keep the simulation from becoming a sea of grid noise. Tracer dissipation is introduced either via advection schemes (e.g., upwind biasing) or closure operators (e.g., biharmonic diffusion).

In general, dissipation in the tracer equation introduces dianeutral mixing. The numerical aim is to keep this mixing far lower than physical ocean mixing, which can be quite small in the pycnocline (Ledwell et al., 1993; Ledwell and Watson, 1998; Ledwell et al., 2011; MacKinnon et al., 2013). One approach is to employ an isopycnal vertical coordinate, where levels of spurious diapycnal mixing are quite small (Ilicak et al., 2012). For non-isopycnal models, spurious mixing is generally reduced with the use of high order advection methods (e.g., Hill et al., 2012). Nonetheless, poorly resolved features will lead to increased spurious dianeutral mixing and/or the introduction of unphysical extrema. More precisely, Ilicak et al. (2012) emphasised the role of the grid Reynolds number and the representation of features near the grid scale in determining the scale of spurious mixing.

2.1.3. Air-sea interactions

There is a growing number of studies that suggest air-sea interactions at the frontal scale are quite active throughout the World Ocean (e.g., Chelton et al. (2004),Minobe et al. (2008), Frenger et al. (2013)). According to the analysis of Bryan et al. (2010), capturing such interactions in simulations requires matching the fine resolution in the ocean with a similarly fine atmospheric resolution. This conclusion adds further cost to the requirements of accurately representing the impact of ocean mesocales within climate simulations.

2.2. Interactions with the land-sea boundaries

One of the most striking enhancements associated with grid refinement is associated with clear improvements to the land-sea boundaries, fine scale channels and throughflows. Many of these regions are of leading importance for their impacts on transport of mass and tracers between ocean basins and marginal seas (see Sprintall et al., 2013, for a review). Despite attempts (e.g., Section 3.5 of Griffies et al. (2005)), it is very difficult to accurately parameterise transport through unresolved channels. Refining the grid spacing is a straightforward way to allow for fluid to move through, say, the islands of the Canadian and Indonesian Archipelogos; the Strait of Gibraltar, etc. So long as the large-scale density gradients are sensible (not always the case!), experience indicates that the net transport through such regions is not far from observed. However, getting the net mass transport correct is not sufficient to ensure the transport occurs within the proper density classes.

The land-sea geometry is fractal (Mandelbrot, 1967), with more geographical scales appearing at each scale of grid refinement. Relatedly, grid refinement does not ameliorate fundamental numerical questions about the representation of topography and the adjacent flow. For example, details for how topography is discretised can have sizable impacts on the large-scale circulation (e.g., Roberts and Wood (1997), Adcroft et al. (1997), Adcroft and Marshall (1998), and Pacanowski and Gnanadesikan (1998)). One promising approach to removing, or at least reducing, sensitivities comes from Adcroft (2013), who proposes an objective means to generate model topography so to capture in a statistical sense important subgrid scale topographic features. Furthermore, Barnier et al. (2006) and Penduff et al. (2007) identified the importance of how momentum transport is discretised next to topography, particularly in so far as boundary current separation is concerned (see also Chassignet and Marshall (2008), Deremble et al. (2012), and Quartly et al. (2013)).

Along with lateral flow next to boundaries, there are important regions of the ocean where strong downslope flow occurs, with such regions important for the production of abyssal watermasses (Legg et al., 2009). Many ocean models fail to accurately capture these overflow regions (e.g., Roberts and Wood (1997),Winton et al. (1998), Tang and Roberts (2005), Legg et al. (2009), Danabasoglu et al. (2010), Zhang et al. (2011), Bates et al. (2012), and Danabasoglu et al. (2014)). The grid constraints from Winton et al. (1998) suggest that we need both very fine horizontal and vertical resolution. Horizontal resolutions meeting these constraints (i.e., 2 km in subpolar gyre) are only just now being realised with realistic regional models (Behrens, 2013), with the resulting dense overflow signal better maintained. As a result, there is a more realistic Deep Western Boundary Current core and a dense Denmark Straight Overflow Water in the Labrador Sea, as well as a thicker North Atlantic Deep Water overturning cell. As an alternative to enhancing such resolution in level coordinate models, there is good evidence that isopycnal models perform better for overflow processes than other vertical coordinates (Legg et al., 2006). Such models naturally enhance vertical degrees of freedom in regions of strong baroclinicity, such as overflow regions, and they minimise spurious mixing. Chassignet further discusses this topic in this edition of CLIVAR Exchanges.

2.3. Representing and parameterising the ocean mesoscale

The time mean and transient mesoscale fields are of leading importance for the lateral and vertical transport of tracers and momentum. In particular, the importance of lateral, more specifically poleward, heat transport has long been highlighted in ocean climate studies Bryan, 1996; Jayne and Marotzke, 2002). Upwelling zones near land-sea boundaries are also more adequately represented with resolution sufficient to represent mesoscale boundary currents, thus admitting important mechanisms for ecosystem dynamics. Additionally, there is a growing appreciation for the role that both transient eddies and time mean currents play on the vertical transport of ocean heat (Gregory (2000), Gnanadesikan et al. (2005), Wolfe et al. (2008), Gregory and Tailleux (2011), Delworth et al. (2012), Morrison et al. (2013), Hieronymus and Nycander (2013), Palter et al. (2014), Zika et al. (2014), Griffies et al. (2014)). Both lateral and vertical ocean transport play a huge role in determining how the ocean participates in the climate system, affecting such emergent processes as transient climate sensitivity and steric sea level rise (e.g., Gregory (2000), Kuhlbrodt and Gregory (2012), Church et al. (2013)).

Ocean models with an active transient mesoscale have routinely been used to benchmark eddy parameterisations. However, such studies are largely restricted to idealised configurations. Realistic global configurations with an active eddy field enable a more complete scientific assessment of eddy parameterisations for the ocean climate system, with efforts along these lines provided by Farneti et al. (2010), Gent and Danabasoglu (2011), Bryan et al. (2014), and Griffies et al. (2014).

Even so, as emphasised by Hallberg (2013), we are a long way from routinely running global models with the 2~km resolution required to resolve the Rossby radius in the high latitudes. Parameterisations will thus remain essential in regions where the grid does not resolve the Rossby radius. The studies from Delworth et al. (2012) and Griffies et al. (2014) illustrate this point by considering the GFDL-CM2.5 climate model, whereby the 0.25° ocean component has no mesoscale eddy parameterisation, other than to employ a frictional dissipation scheme to absorb the downscale enstrophy cascade. This model suffers from large biases in the high latitudes (e.g., very deep mixed layers associated in part with weak eddy restratification). Additionally, the relatively weak transient eddy fluctuations bias the vertical transport of heat in the middle latitude gyres relative to the finer 0.1° ocean in the GFDL-CM2.6 climate model. In brief, the vertically upward eddy heat transport in GFDL-CM2.5 is not sufficient to compensate for the downward heat transport by the mean flow. The result is a relatively cold upper 200 m and warm intermediate depth (200 m-1500 m). Simulations with a 0.1° ocean in GFDL-CM2.6 significantly reduce these biases by providing a more powerful mesoscale field rendering an increase in vertically upward eddy heat transport. Eddies therefore act as a regulator for the extent that heat (and other tracers) can enter the ocean interior. Accurately representing or parameterising this vertical transport are important for addressing questions such as ocean heat uptake and sea level rise.

Nearly all coarse resolution ocean climate models base their eddy parameterisations on the paradigm of neutral diffusion proposed by Solomon (1971) and Redi (1982), and the eddy-induced advection of Gent and McWilliams (1990) and Gent et al. (1995). Recent developments on eddy closure suggest the importance of depth dependence to the eddy diffusivity (see Abernathey et al. (2013) for a review), and emphasise the distinction between a diffusivity used for tracers versus buoyancy (though see Fox-Kemper et al. (2013) for a different perspective). Furthermore, there are suggestions that eddy diffusivities should be based on a mesoscale eddy kinetic energy budget, such as that proposed by Eden and Greatbatch (2008) and Marshall and Adcroft (2010). Following the suggestion of Hallberg (2013), the best of these schemes should be used in those regions that do not resolve the Rossby radius, whereas the parameterised eddy fluxes should be set to zero in regions resolving the Rossby radius with two or more grid points.

In addition to the neutral diffusion and eddy-advection approaches commonly used in coarse models, recent work on ``stochastic backscatter'' emphasises the importance of returning kinetic energy dissipated near the grid scale (i.e., through biharmonic friction) to the larger scales, thus respecting the inverse energy cascade of quasi-geostrophic turbulence (Kitsios et al., 2013; Mana and Zanna, 2014; Jansen and Held, 2014. These ideas, as well as the resolution function approach from Hallberg (2013), hold some promise for parameterising unresolved eddies within a model that admits some eddies. Further ideas for such Mesoscale Ocean Large Eddy Simulations (MOLES) are discussed by Fox-Kemper and Menemenlis (2008).

No global climate model employs a parameterisation for the time mean mesoscale currents associated with strong boundary currents, such as in the subpolar North Atlantic. Instead, parameterisations focus on impacts from transient eddies. However, boundary currents are of leading order importance for shaping the sea ice edge in the high latitudes, which in turn impacts high latitude climate through air-sea fluxes. One idea for enhancing such currents, promoted by Holloway (1992) (see Maltrud and Holloway (2008) for an application in an eddying model), has been to parameterise the interaction between eddies and topography using a modification to the standard friction operator. However, this scheme has not been widely used in the global climate modelling community, with Appendix E of Danabasoglu et al. (2014) summarizing problems with its use in the subpolar North Atlantic.

3. Climate model development with an active ocean mesoscale

There are many stages encountered when building a global mesoscale eddying ocean model. One of the initial stages involves the stimulating "wish list" discussions in which one dreams about the new science possibilities available with an eddying ocean model. Upon working through various compromises and completing a massive amount of configuration development, a prototype model is finally built and short (few years) simulations are performed. This stage typically engenders elation. Compelling animations are often produced to help garner further excitement as well as resources necessary to continue the research. Thereafter, the tough questions arise as drift and associated biases, sometimes huge, are uncovered. Because of the expense of rerunning the model with altered configuration details, there are relatively few opportunities to uncover mechanisms for the drift and biases. In short, ocean eddying simulations are tough to tune, and their "slippery" fluid is notoriously difficult to tame.

Nonetheless, there are some successes. We highlight here efforts with the climate models GFDL-CM2.5 and GFDL-CM2.6 as documented by Delworth et al. (2012) and Griffies et al. (2014), where relatively stable simulations reveal novel and compellingly realistic representations of the ocean in a coupled climate model run for multiple centuries. Notably, the 0.1° ocean component in GFDL-CM2.6 was developed only after a tremendous amount of experience was garnered with GFDL-CM2.5 using a less expensive 0.25° ocean component. The development of GFDL-CM2.5 provided important lessons and insights, many engineering in nature, into climate modelling with an eddying ocean. Attempts to go straight from a one-degree climate model to the 0.1° would have side-stepped these lessons and perhaps failed to produce a respectable GFDL-CM2.6. Additionally, there are many applications where the 0.25° is of scientific utility, particularly as a means to understanding an ocean with a nontrivial mesoscale field (Penduff et al., 2010). Advances in eddy parameterisations in this eddy-active model also lend promise to the utility of the 0.25° ocean configurations.

The stepwise hierarchical approach followed at GFDL is not universally practiced. For example, NCAR/DOE and MPI development efforts jumped from their one-degree ocean climate models straight to 0.1°. Arguments for this approach are largely based on considering the 0.25° ocean to present a poor rendition of the mesoscale, whereas the 0.1° is far more accurate. However, this argument can certainly continue to even finer resolutions. Scientific applications are many even with an imperfect model. Furthermore, from an engineering perspective, there is little evidence that jumping straight to the 0.1° ocean has resulted in a successful climate model sans rather large climate drifts and associated watermass biases (e.g., McClean et al. (2011), Kirtman et al. (2012), Bryan et al. (2014)). We conjecture that a stepwise approach, involving a relatively inexpensive eddying ocean component (e.g., 0.25°), is the more prudent strategy for developing an eddy-rich (e.g., 0.1°) climate model.

4. Closing comments

Assuming continued growth in computational power and software able to exploit the ever changing hardware, we can expect that global climate models with an active, and perhaps even a very rich, ocean mesoscale eddy field will become commonplace over the next decade. Even with the many difficulties still facing climate model development with an eddying ocean, community knowledge and experience are expanding. One useful means for such knowledge growth is through workshops such as that held in Kiel, Germany in April 2014, sponsored by the WCRP/CLIVAR Working Group on Ocean Model Development (WGOMD). As the community continues to expand this base of knowledge and experience, we can expect climate models with an energetic mesoscale field to be providing increasingly realistic and trustworthy simulations out to the multiple-century time scale. They will also provide a far more valuable means to downscale towards the even finer resolutions required for addressing questions about climate change at the regional and coastal scales. It is thus a very exciting time, where the needs for robust climate simulations are merging with the interests of oceanographers who are stimulated by simulations containing features observed with modern observational platforms.

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An overview of CheapAml: An atmospheric boundary layer for use in ocean only modelling

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1. Introduction

As more attention is drawn toward oceanic low-frequency variability, there is a need to run longer ocean simulations at high resolution. An important associated consideration is the method by which the ocean is forced. Reanalysis atmospheric data is often limited to ~60 years and comes at low resolution (generally ~1 degree). On one hand, imposing an atmospheric state without any knowledge of the ocean state is appealing because it maintains the ocean model close to a realistic state. On the other hand, this strategy might damp intrinsic modes of variability and air-sea interaction at small scales is completely lost.

A way to tackle this problem is to use a full high resolution coupled model, but this comes at a price in terms of computing resources. Deremble et al. (2013) proposed a third option, i.e. forcing the ocean with a thermodynamically active atmospheric boundary layer model that responds to the model Sea Surface Temperature (SST). This model, namely CheapAML, was inspired by the earlier work of Seager et al. (1995) and uses a prescribed wind field to advect the atmospheric temperature and humidity. The latter are also modified by air-sea fluxes. In this short note, we first review the main equations of the model and illustrate its utility via a simple example.

2. Main equations

The basic assumptions of CheapAML are that atmospheric reanalysis variables like humidity and temperature are accurate on large scales and of these the least sensitive to ocean surface structure is (nominally ten meter) velocity (u). We thus accept atmospheric velocity as a known and develop equations governing the atmospheric tracer fields of temperature and water. This shortcut avoids the complexities of atmospheric dynamics and instead concentrates on thermodynamics. There are two fundamental equations solved by CheapAML. The first is the equation for atmospheric boundary layer potential temperature T

$$\frac{\partial T}{\partial t} = -u.\nabla T + \nabla\cdot(K_T \nabla T) + \frac{SH + F_{ol}^{\uparrow} - F_l^{\downarrow} - F_l^{\uparrow}}{\rho_a c_p H}, \qquad (1)$$

with

$$SH = C_{dh}|u|(SST - T) \qquad (2a)$$
$$F_{ol}^{\uparrow} = \varepsilon\sigma SST^4 \qquad (2b)$$
$$F_l^{\downarrow} = \frac{1}{2}\varepsilon\sigma T^4 \qquad (2c)$$
$$F_l^{\uparrow} = \frac{1}{2}\varepsilon\sigma[T(z_l)]^4, \qquad (2d)$$

with SH the sensible heat flux computed with a bulk formula, $F_{ol}^{\uparrow}$ the upward longwave emitted by the ocean, $F_l^{\downarrow}$ the atmospheric downward longwave, and $F_l^{\uparrow}$ the atmospheric upward longwave. In principle, to get the right radiative fluxes, one needs a full atmospheric model (multiple layers). The parameterisation of the optical depth is done by adjusting the temperature at which the long wave is emitted. This parameter is the height z_l in Eq.(2d). See Deremble et al. (2013) for a full description of this equation.

The second equation governs atmospheric water content

$$\frac{\partial q}{\partial t} = -u.\nabla q + \nabla\cdot(K_q \nabla q) + \frac{E - F_q^{\uparrow} - \lambda P}{\rho_a H}, \qquad (3)$$

where q is atmospheric specific humidity and

$$E = C_{de}|u|(q_s^{SST} - q) \qquad (4a)$$
$$F_q^{\uparrow} = \mu C_{de}|u|q \qquad (4b)$$
$$P = \frac{\rho_a H^T}{\tau}(q - 0.7q_s)\frac{w}{w_0}, \qquad (4c)$$

with E the evaporation computed with a bulk formula, $F_q^{\uparrow}$ the entrainment at the top of the boundary layer and P the precipitation. Equation (4c) is a proxy for the precipitation occurring in the entire atmosphere. Only a fraction λ is occurring at the top of the boundary layer since we are considering a sub-cloud layer (see Deremble et al., 2013, for details).

3. Example

Let us illustrate the utility of this model with a simple example. Our purpose is to show that we obtain better values for atmospheric forcing when we use CheapAML, rather than imposing an atmospheric state decorrelated with the ocean state. To demonstrate this, we select two random years 2000 and 2005. All data in this example come from ERA-Interim (Dee et al., 2011). The reanalysis SST pattern in each of these years is different and is representative of a certain state of the ocean. Moreover the reanalysis atmospheric state in 2000 is 'consistent' with the SST in 2000. In a forced ocean model, given the chaotic nature of the ocean, the SST in model year 2000 will differ from that of the real ocean, particularly at small scales, and will no longer 'agree' with the imposed atmospheric structure.

To illustrate the error realised in that case, we plot in Fig.1 the mean surface air temperature in 2005 minus the mean surface air temperature in 2000 from the reanalysis data. This map exhibits warmer and cooler regions distributed all over the globe. In fact, these regions are well correlated with the SST difference between 2005 and 2000 (not shown). The mean value of this map (excluding land and parts of the ocean covered by ice) is 0.3 K and the standard deviation 0.6 K. Similar conclusions are obtained with humidity but are not presented here for brevity. We compare the error obtained in that case with the error that we would get when

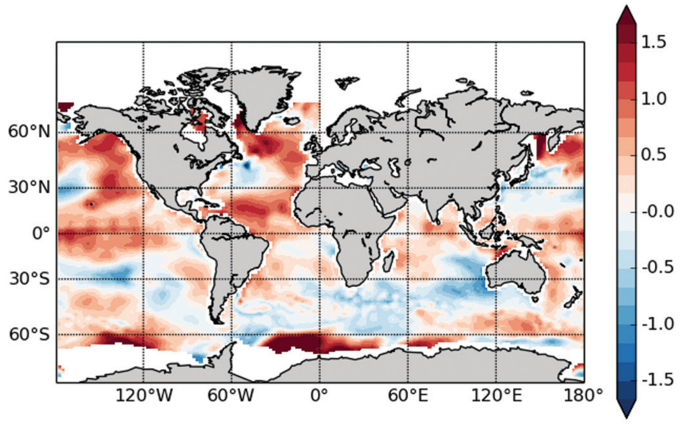

Figure 1: T(2005) - T(2000). Units: ˚K. The colour bar is limited to a maximum anomaly of 1.5 K for visibility, but the magnitude of the anomaly is more than 1.5 in certain places. The regions where sea-ice is present is masked.

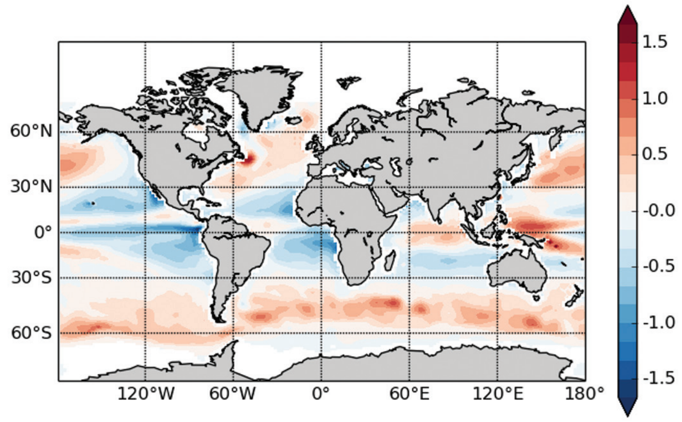

Figure 2: T(AML) - T(2000). CheapAML experiment with the wind of 2000 and land temperature of 2000

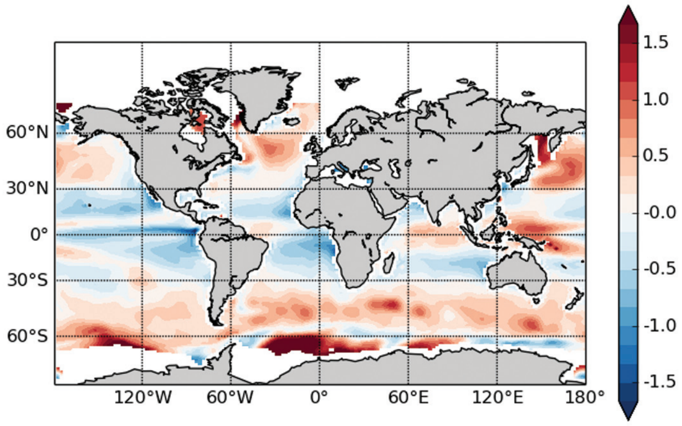

Figure 3: T(AML) - T(2000). CheapAML experiment with the wind of 2005 and land temperature of 2005

using CheapAML. We run CheapAML with a prescribed SST (2000, monthly mean). We use the wind of 2000 and over the continents the atmospheric variables are restored to the values of year 2000. The atmospheric variables are updated every 6 hours. We use a constant atmospheric boundary layer height of 1000 m and an 'optical depth' of z_i = 100 m. Figure 2 is the difference between the mean temperature reconstructed by CheapAML and T(2000). The visual comparison with Fig. 1 illustrates that the magnitude of the anomaly is lower for the latter case. The mean anomaly is in fact 5x10-3 K and the standard deviation is 0.4 K. The pattern of this anomaly resembles the climatological cloud fraction (not shown). An advanced radiation scheme with a cloud parameterisation would certainly decrease this bias. Nevertheless the comparison argues in favour of CheapAML which is able to reconstruct an atmospheric temperature that matches the SST. In this example however, we used the atmospheric wind and land boundary condition of 2000, that is in agreement with the underlying SST.

To unambiguously assess the validity of CheapAML, we run the model with the SST of 2000, the wind of 2005 and the atmospheric land boundary conditions of 2005. The map of the mean temperature obtained by CheapAML minus the temperature in 2000 is shown in Fig.3. This map is reminiscent of the map in Fig. 2: we recognise the same patterns with the same amplitude that we attribute again to the missing physics in our model (mostly clouds). We attribute the several warm biases near Antarctica to a different position of the ice cover in 2005 which affects substantially the surface atmospheric temperature. The mean value of the anomaly is 0.1 K and the standard deviation is 0.5 K.

This simple example emphasises the advantages of moving from the traditional practice of applying a prescribed atmospheric field to an ocean model to applying an atmospheric boundary layer model. CheapAML permits the atmospheric field in adapt in a realistic way to underlying SST. The icing on the cake is the minimal cost associated with this model in terms of implementation and computing resources.

Amongst the most significant concerns we have heard voiced about using CheapAML comes when considering ensemble experiments. In this case, due to intrinsic ocean variability, the literal ocean-atmosphere exchange varies between ensemble members and this raises issues about whether individual realizations can be properly thought of as a controlled ensemble. We would suggest in response the way to consider ensemble generation is to think of the exchange variability as a natural consequence of the evolution. The classification of a set of experiments as an 'ensemble' then becomes an issue of using identical winds, land temperatures and solar radiation, i.e. the elements to force CheapAML to allow it to force the ocean.

4. Conclusions

The value of this model is to capture part of the non-local feedback of the ocean surface on air-sea exchanges, while stopping well short of computing a full coupled ocean-atmosphere model. We believe that for an oceanic model, it is preferable to use CheapAML than to prescribe the temperature and humidity (or fluxes) from a reanalysis data set: as soon as the oceanic state deviates from the observed state, the reanalysis temperature and humidity fields and the oceanic state are not related anymore. The computational cost of using CheapAML is minimal, and does not materially increase the execution time of the model run. Furthermore, CheapAML captures the 'weather' impacts of the atmosphere on air-sea exchange with improved fidelity relative to its predecessor, Seager et al. (1995).

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DRAKKAR: developing high resolution ocean components for European Earth system models

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1. Introduction

DRAKKAR is a consortium of European ocean modelling teams. It was "created to take up the challenges of developing realistic global eddy-resolving/ permitting ocean/sea-ice models, and of building an ensemble of high resolution model hindcasts representing the ocean circulation from the 1960s to present" (quoting the DRAKKAR Group, 2007, in a CLIVAR Exchanges paper where the DRAKKAR strategy was presented for the first time). Now in the second decade of its existence, the DRAKKAR Group is active and thriving, and it is now timely to present recent developments and future plans in this special issue of CLIVAR Exchanges.

DRAKKAR was initiated when a group of leading ocean modelling teams in Europe decided to use common global ocean-ice model configurations based on the NEMO platform in order to explore the ocean variability forced by the atmosphere, at time scales from seasonal to multi-decadal, with the highest possible spatial resolution. High resolution is required to resolve narrow boundary currents and energetic mesoscale eddies, which are ubiquitous in the world ocean. High resolution global simulations have considerable added value compared to regional models, because eddies and unstable jets can transfer eddy energy across ocean basins and from one basin to the next. Such features are difficult to reproduce in limited-area models - unless high resolution (both temporal and spatial) boundary conditions are used at the open boundaries, which again requires high resolution global simulations. The DRAKKAR simulations are well-suited both for the analysis of the global ocean variability forced by the atmosphere, and as boundary conditions for regional models. Furthermore, the multi-decadal forced simulations carried out by DRAKKAR are necessary steps to assess ocean model configurations as trustworthy components of earth system models, in preparation for - or as companion to - the CORE multi-century forced ocean-ice simulations (Coordinated Ocean Reference Experiments, e.g., Danabasoglu et al., 2014).

2. Global ocean simulations: from 1/4° to 1/12°

The three key ingredients of DRAKKAR global configurations are
- the NEMO modelling platform (Madec, 2008), based on the OPA ocean code and the sea-ice model LIM;
- the ORCA tripolar global grid, which is almost isotropic (it is refined as the cosine of the latitude in the Southern hemisphere and also refined poleward in the Northern hemisphere). The ORCA grid allows a good representation of the Arctic Ocean and its exchanges with the Pacific and Atlantic oceans (e.g., Lique et al., 2009)
- the DRAKKAR forcing sets (DFS, Brodeau et al., 2010).

These atmospheric forcings follow the methodology of Large and Yeager (2010) but use ECMWF atmospheric variables instead of NCEP. Within DRAKKAR, simulations with DFS are compared with simulations using the Large and Yeager forcing, following the CORE protocol (Griffies et al., 2012).

The first DRAKKAR simulations were carried out in 2006 used ORCA025, a global configuration with resolution 1/4° at the equator (DRAKKAR Group, 2007; see also Barnier et al., 2006, and Penduff et al., 2007). The model configuration was shared with MERCATOR-Ocean, the French operational oceanography agency. MERCATOR-Ocean tested ORCA025 in operational mode; the DRAKKAR teams developed it further, regarding numerical schemes, parameterisations and bathymetry, which benefitted not only research projects but also the operational forecasts. Recognizing the need for increased vertical resolution in the surface layers (of order 1m to represent the diurnal cycle), the UK Met Office and MERCATOR-Ocean operational centres and the DRAKKAR Group jointly agreed on a new vertical grid with 75 levels. ORCA025 simulations carried out by DRAKKAR partners have made possible a large number of scientific studies about, for instance, the global and regional variability of the ocean circulation, heat and salt content; the oceanic response to atmospheric modes of variability; eddy dynamics and their impacts on biogeochemical cycles and biology. About 50 peer-reviewed articles per year based on DRAKKAR ORCA025 simulations or DFS forcings have been published between 2011 and 2013.

The ORCA025 global configuration is now the ocean component of high resolution coupled climate models (Scaife et a.l, 2014), and it is anticipated that a large number of European climate centres will use it for CMIP6, at least for short term scenarios (time scales of a few decades).

Although the ORCA025 grid allows for the development of baroclinic instability and large eddies, such as Agulhas rings, it has long been known that even higher resolution is required to represent faithfully the mid-latitude western boundary currents systems such as the Gulf Stream and Kuroshio and their recirculation gyres (Smith et al, 2000). This has been recognised by the operational centres, and in 2009 MERCATOR-Ocean brought the 1/12° ORCA12 configuration into pre-operational mode (Hulbert et al., 2009). The use of such a costly ocean model for multi-decadal simulations was a challenge that the DRAKKAR group decided to take up in a coordinated fashion. Long experiments (10 to 30 years) were carried out by three DRAKKAR teams in 2012, followed in 2013 and 2014 by a climatological simulation (84 years) and a long interannual simulation (from 1958 to 2012).

3. Recent results

The DRAKKAR ORCA12 simulations are a unique opportunity to investigate the variability of the Atlantic Meridional Overturning Circulation (AMOC), the eddy or small scale mechanisms that contribute to its variability, and its stability. Indeed, there was concern that coarse resolution climate models may be at odds with observations for the representation of a key property of the AMOC: its stability with respect to an increased freshwater input in the northern high latitudes (Hawkins et al., 2011). Following the introduction of the AMOC bistability concept by Ramstorf (1996), modellers have focussed on a specific indicator: FTov, the salinity anomaly transport (or conversely, freshwater anomaly) by the AMOC at the entrance of the South Atlantic along 30°S. When the AMOC imports fresh waters into the Atlantic, an increase in AMOC leads to a reduction in Atlantic Ocean salinity (negative feedback on the AMOC) while in the opposite case, a stronger AMOC imports more salt into the northern latitudes, inducing more dense water formation (a positive feedback and possible instability of the AMOC). Using four ORCA12 simulations and two ORCA025 simulations, Deshayes et al. (2013) showed that the ORCA12 simulations are in better agreement with observations than ORCA025 simulations or coarse resolution climate models in terms of the FTov indicator. This is because the AMOC maximum is located at a more realistic depth in relation to the water masses at 30°S. These results confirm that the present -day AMOC could be in the bi-stable regime. These results hint that future coupled scenarios with high resolution ocean components could predict more dramatic future changes of the AMOC, compared to the present coarse resolution scenarios.

Just like this stability issue, the AMOC variability requires more than a single simulation in order to understand which events are robust, and what is their significance. Blaker et al. (2014) use a total of seven simulations; six ORCA025 simulations with different forcings and one ORCA12 simulation, in order to investigate the dramatic minimum in AMOC observed during the winter 2009-2010 by the RAPID array (McCarthy et al., 2012), which was followed by a similar event the next winter. The two simulations covering the observational period (up to 2011) reproduce the weakening events, demonstrating that these events are directly forced by the atmosphere, as shown in Figure 1. The long time series provided by the DRAKKAR simulations allow Blaker et al. (2014) to look for historical analogues of these weakening events. At least two pairs of events have occurred in the past, and these events show substantial meridional coherence.

ORCA12 simulations give us a unique opportunity to investigate in detail the mechanisms of AMOC variability. Analysis of the RAPID array have shown that wind forcing is a driver at the seasonal scales, and the dynamics near the eastern boundary have a strong impact on the AMOC at 26°N. ORCA12 provides a four-dimensional context to these observations, and shows how the relationship between the wind stress curl and the density structure is influenced by small scale features such as the Canary Islands (Duchez et al., 2014). Other studies are targeted at the North Atlantic subpolar gyre, considering the circulation in the boundary currents (Marzocchi et al., 2014) or wind-forced variability in relation to atmospheric weather regimes (Barrier et al., 2014). Hughes et al. (2014) have recently investigated the "Southern mode" of variability of the Antarctic circumpolar transport in both ORCA025 and ORCA12. The availability of many simulations and also the contrast between the interannual simulations and those forced by a repeated

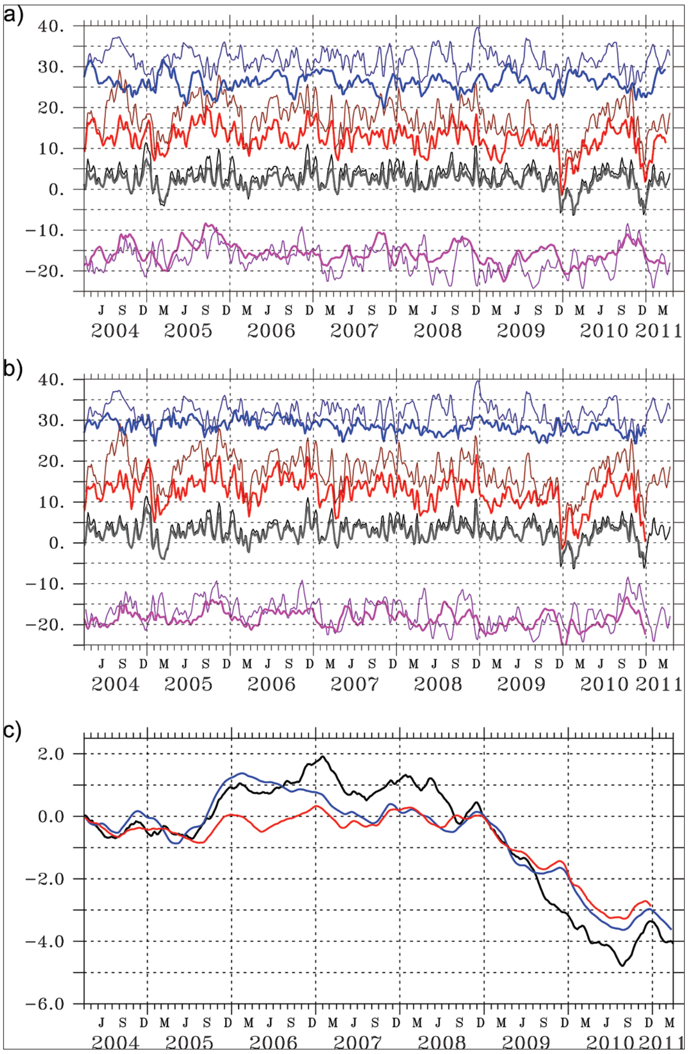

Figure 1: a, b) Comparison of RAPID observations of the AMOC (thin lines McCarthy et al., 2012) with an ORCA025 (a) and ORCA12 (b) simulation (Blaker et al., 2014). The time series of the total AMOC (red line) is the sum of three components: Gulf Stream transport (blue), Ekman transport (black) and the upper mid-ocean transport (pink). All time series are in units of Sv (Sverdrup). Note that the two events of weak AMOC at the end of years 2009 and 2010 are well simulated in ORCA12. c) Cumulative upper mid-ocean transport anomalies (following Bryden et al., 2014) for the RAPID observations (black), ORCA025 (blue) and ORCA12 (red). Units are Sv/year. The good agreement between ORCA025/ORCA12 and RAPID shows that the similar AMOC evolutions seen in model and observations in a) and b) are not just due to Ekman transports but also to the model's ability to capture a large fraction of the long-term evolution of the upper mid-ocean transport.

seasonal cycle (climatological runs) is the key to a quantification of the "intrinsic" ocean variability (Penduff et al., 2011, Hirschi et al., 2013). These studies initiated with ORCA025 are currently being pursued using ORCA12 and will also benefit from ensemble strategies at lower resolution (see Penduff et al., 2014, this issue).

Long ORCA12 simulations (especially the unique, 84-years long climatological simulation) are well suited for an evaluation of the impact of transient ("eddy") correlations on the global meridional transports of heat and salt. Regarding the global salt balance, until last year the only estimate of the eddy contribution dated from the nineties and relied on a 0.5° numerical model. The new calculation carried out with ORCA12 by Treguier et al. (2014) demonstrates that eddy velocity-salinity correlations are responsible, on average, for about half the flux of salt out of the subtropical gyres that is required to balance the excess evaporation in these regions, as shown in Figure 2.

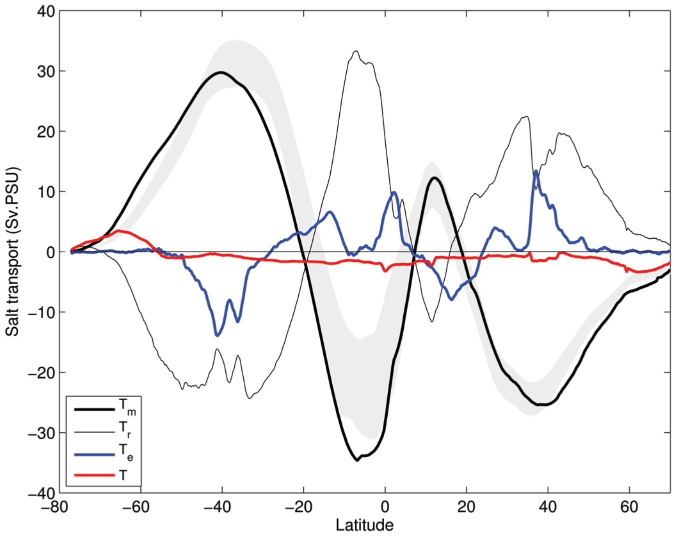

Figure 2: Decomposition of the meridional transport of salt in the climatological ORCA12 simulation, averaged zonally over the global ocean (Treguier et al., 2014). The total transport (T, red line) is close to zero, reflecting the equilibrium of the salinity distribution in this long experiment. The non-zero transport north and south of 60° is due to ice-ocean fluxes. This total transport results from the near-cancellation of three contributions: 1) the salt transport by the net ocean mass flux (Tm, thick black line) which is directly forced by the evaporation/precipitation balance and which compares well with the observations (Large and Yeager, grey shading), 2) the compensating transport by time-mean recirculations, overturning and gyres (Tr, thin black line), and 3) the salt transport by transient "eddy" velocity-salinity correlations (Te, blue line), which is as important as the mean in the mid latitudes (40°N and 40°S).

4. Conclusions and future plans

While global simulations at 1/10° to 1/12° resolution have been around for more than a decade, the possibility to run multiple multi-decadal experiments with such models is recent. In the past four years the DRAKKAR Group has run such simulations with the global 1/12° ORCA12 model based on the NEMO European modelling platform. The dramatic improvement of the western boundary current systems will certainly lead to great advances when these models are coupled to the atmosphere and to biogeochemical cycles. However, we have also found new biases and drifts that arise at high resolution. For example, although the Gulf Stream separation is generally improved, the path of the North Atlantic current is not yet robust between the different ORCA12 simulations and in some cases appears less realistic than in the lower resolution configurations. Extensive tests carried out by the DRAKKAR Group suggest that despite its fine grid, ORCA12 is still very sensitive to the sub-grid scale parameterisations and numerics. Further work is needed to improve the representation of bathymetry, and to choose the optimal vertical resolution for ORCA12. The improvement of the global ORCA12 model will benefit from the exploration of even higher resolution regimes in key regions, using the AGRIF refinement package (e.g., Talandier et al., 2014; Biastoch et al., 2014, this issue).

It is the genuine collaboration across all the DRAKKAR modelling groups, and the planning and execution of coordinated sets of model integrations, that greatly enhances our ability to achieve rapid scientific advances with such high-resolution models. ORCA12 simulations are documented on the web site www.drakkar-ocean.eu and the results are available upon request.

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HYCOM high-resolution eddying simulations

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1. Introduction

Over the past 15 years, a broad partnership of institutions has collaborated on developing and demonstrating the performance and application of eddy-resolving, real-time global and basin-scale ocean prediction systems using the HYbrid Coordinate Ocean Model (HYCOM). This paper is intended to briefly summarise the current status of global high resolution eddy-rich or "eddying" HYCOM simulations and their relevance to high resolution coupled ocean-atmosphere-ice climate simulations. In this context, eddy-rich or "eddying" means that the numerical simulations are eddy-resolving over most of the domain and include an energetic mesoscale eddy field.

None of three main vertical coordinates currently in use (level, isopycnal, or terrain-following) provides universal utility (Griffies et al., 2000; Chassignet et al., 2006), and hybrid approaches have been developed in an attempt to combine the advantages of different types of vertical coordinates in optimally simulating the ocean. The term "hybrid vertical coordinates" can mean different things to different people: it can be a linear combination of two or more conventional coordinates or it can be truly generalised, i.e., aiming to mimic different types of coordinates in different regions of a model domain (Bleck, 2002; Adcroft and Hallberg, 2006). The hybrid or generalised coordinate ocean models that have much in common with isopycnal models are POSEIDON (Schopf and Loughe, 1995) and HYCOM (Bleck, 2002; Chassignet et al., 2003; Halliwell, 2004). Other generalised vertical coordinate models currently under development are the Model for Prediction Across Scales (MPAS; Ringler et al., 2013) and MOM6 (http://www.gfdl.noaa.gov/ocean-model). The default configuration of HYCOM is isopycnic in the open stratified ocean, but makes a dynamically smooth transition to terrain-following coordinates in shallow coastal regions and to fixed pressure-level coordinates in the surface mixed layer and/or unstratified seas. In doing so, the model takes advantage of the different coordinate types in optimally simulating coastal and open-ocean circulation features.

2. Global HYCOM high-resolution eddying simulations

Much of the impetus for integrating high-resolution eddying global numerical simulations comes from the US Navy's interest in advanced global ocean nowcasting/forecasting systems (Metzger et al., 2014a). Within the framework of the multinational Global Ocean Data Assimilation Experiment (GODAE) and under the sponsorship of the National Ocean Partnership Program (NOPP), a broad-based partnership of institutions participated in the development of high resolution data assimilation systems (Chassignet et al., 2009) that were eventually transitioned for operational use by the U.S. Navy at the Naval Oceanographic Office (NAVOCEANO), Stennis Space Center, MS, and by the National Oceanic and Atmospheric Administration (NOAA) at the National Centers for Environmental Prediction (NCEP), Washington, D.C. The current operational global ocean forecast system at NAVOCEANO is referred to as the Global Ocean Forecast System 3.0 (GOFS 3.0). It will be upgraded to GOFS 3.1 in the summer of 2014 to include 3D Variational (3D-VAR) data assimilation methodology, increased vertical resolution in the upper ocean, and two-way coupling to the Los Alamos sea ice model (CICE - Hunke and Lipscomb, 2008). The current HYCOM GOFS configuration has an equatorial resolution of 0.08° (1/12.5°or ~ 9 km near the equator, ~7 km at mid-latitudes, and ~ 3.5 km near the North Pole). The horizontal resolution will be increased in 2017 to 0.04° (~3.5 km at mid-latitudes) with the addition of tidal forcing (see paper by Arbic et al. 2014, this Issue).

The impact of going to 1/25° horizontal resolution was assessed by Thoppil et al. (2011) who compared the modelled eddy kinetic energy (EKE) with long-term observations from surface drifters, geostrophic currents from satellite altimetry, subsurface floats and deep current meter moorings. Adequately representing mesoscale eddies is key to simulating the mean circulation since the surface and abyssal ocean circulation are strongly coupled through the energy cascades that vertically redistribute the energy and vorticity throughout the entire water column. Although the present generation of eddy-resolving global OGCMs at 1/10° resolve the dominant eddy scale, at this resolution the models significantly underestimate the EKE in the abyssal ocean (i.e., depths greater than 3000 m) (Scott et al., 2010). The 1/12.5° HYCOM is deficient in EKE in both the upper and abyssal ocean by ~21% and ~24% respectively compared to surface drifting buoys and deep current meters. Increasing the model resolution to 1/25° significantly increases the surface and the abyssal EKE to levels consistent with the observations, and clearly demonstrates the need for better representation of upper ocean EKE as a prerequisite for strong eddy-driven abyssal circulation.

Because of the large heat and freshwater transports and interaction with the atmosphere, the Atlantic Meridional Overturning Circulation (AMOC) plays a fundamental role in establishing the mean state and variability of the Earth's climate. Xu et al. (2014) analysed an interannual 1/12.5° global HYCOM simulation forced with the three-hourly, 0.5° Navy Operational Global Atmospheric Prediction System (NOGAPS, Rosmond et al., 2002) to investigate the driving mechanisms behind the variability of the AMOC transport during 2004-2012. The model results are in very good agreement with the RAPID observations at 26.5ºN (see Smeed et al. (2014) for the latest results). This is true not only for the total AMOC transports, but also for its components (the Florida Current, the mid-ocean, and Ekman transports). The model simulates well the observed AMOC variability at 26.5o N on intraseasonal, seasonal, and interannual time scales, as well as the observed long-term decrease in the AMOC and the Florida Current transports (3 Sv over 2004-2012). At 41ºN, however, the agreement between the model results and the Argo-based observations is mostly due to the Ekman transport and the geostrophic transport is approximately six months out of phase. Mielke et al. (2013) also found a similar phase shift in the seasonal variability of the geostrophic transport between the observations at 41ºN and the global simulation of von Storch et al. (2012). This lack of agreement between models and observations suggest that either the models do not adequately represent the ocean dynamics at 41ºN and/or the Argo floats are not able to sample adequately the AMOC variability at that latitude. Xu et al. (2014) show that both observations and model results exhibit higher AMOC variability on seasonal and shorter time scales than on interannual and longer time scales. On intraseasonal and interannual time scales, the AMOC variability is often coherent over a wide latitudinal range, but no overall consistent coherent pattern between the Equator and 70ºN can be identified on any of these time scales (a,c). On seasonal time scales (Figure 1b), the AMOC variability exhibits two distinct coherent regimes north and south of 20ºN, the boundary between the North Atlantic subtropical and tropical gyres, due to different wind stress patterns and variability in the tropics and subtropics. These results highlight the importance of the surface wind in driving the AMOC variability.

3. Reanalysis

The US Navy operational global ocean forecast GOFS model (Metzger et al., 2014) is driven by atmospheric fields from the Navy operational numerical weather prediction model, Navy Global Environmental Model (NAVGEM), and prior to 2013, NOGAPS. As with other operational models, the forecast systems are continually modified and improved. Thus, it is difficult to obtain a consistent evaluation of the model performance over a long period of time. To address this issue, the Naval Research Laboratory has performed a reanalysis of the 1/12.5º global HYCOM forced by the NCEP Climate Forecast System Reanalysis (CFSR) fluxes (Saha et al., 2010, 2014) using the 3D-VAR data assimilation scheme of the Navy Coupled Ocean Data Assimilation (NCODA, Cummings, 2005; Cummings and Smedstad, 2013) for the period 1993 to 2012. The chosen time period spans the modern satellite altimeter era. The altimeter sea surface height (SSH) data provide the largest and most consistent set of observations to constrain the mesoscale eddy circulation in the model. The CFS Reanalysis ends in 2009, but the same model, CFSv2, was used operationally to extend the forcing data set to 2012. In addition to the HYCOM reanalysis, a twin simulation without data assimilation was performed. Over the reanalysis period, the observing system changed substantially. The number of satellite altimeters varied from two to four over the 20-year period and, after 2000, Argo began to provide an increasing number of vertical profiles of temperature and salinity in place of XBT temperature profiles.

The reanalysis was completed in February 2014. As noted by Thoppil et al. (2011), the 20-year CFSR non data assimilative simulation underestimates the EKE at all depths in the ocean compared to the historical surface drifters, geostrophic altimetric EKE, subsurface floats and deep current meters. Preliminary analyses show that data assimilation increases the EKE in both the surface and deep ocean by 10%. However, the reanalysis EKE is still weaker than the observed surface drifter EKE and deep current meter EKE by ~10%. Both the simulation and reanalysis reproduce the 2004-2012 observed AMOC variability. The observed AMOC from the RAPID array has a -0.40 Sv/year trend over the eight years, while the reanalysis and simulation have slightly weaker trends of -0.32 Sv/year and -0.26 Sv/year, respectively. However, none of these interannual trends are significant at the 95% confidence level. Over the 20-year (1993-2012) period, the trends in the AMOC are much weaker at -0.03 Sv/year for the reanalysis and -.14 Sv/year for the simulation. For the 20-year period, the reanalysis mean AMOC is slightly weaker, 19.4 Sv, than the non data assimilative simulation, 19.8 Sv. However, the variability of the AMOC in the reanalysis is greater than the simulation. For the RAPID array period, the reanalysis and simulation AMOC are larger than observed by 2.2 Sv for the reanalysis and 0.9 Sv for the non data assimilative simulation.

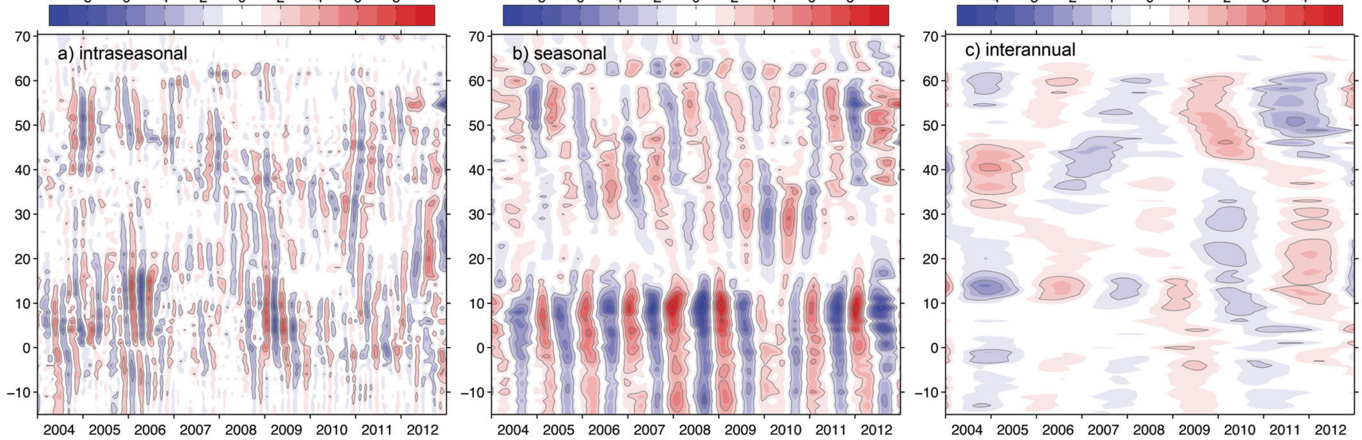

Figure 1. AMOC variability in the 1/12.5º global HYCOM as a function of time and latitude on a) intraseasonal, b) seasonal, and c) interannual time scales. (adapted from Xu et al., 2014).

4. Earth System Prediction Capability (ESPC) program

The Earth System Prediction Capability (ESPC) inter-agency program (Eleuterio and Sandgathe, 2012) was established in 2010 as a coordinated effort to improve collaboration across the sponsored environmental research and operational prediction communities in the US for the development and implementation of improved physical earth system prediction. The ultimate goal of ESPC is to create a high-resolution extended range, coupled atmosphere, ocean, wave, land, and ice to provide more accurate and longer range predictions at the weather-climate interface. The initial Operational Capability is targeted for 2018 (Metzger et al., 2014b) with the following individual components: HYCOM (ocean), WW3 (waves), NAVGEM (atmosphere), NAAPS (aerosol), NAVGEM-LSM (land), and CICE (ice). At that time, daily 10-day forecasts will be performed with a 41 layer 1/25o HYCOM, 70 level T639 (20 km) NAVGEM and 1/8° WW3. Weekly 30-day forecasts using a reduced resolution ocean and wave model and an ensemble of 90-day forecasts with reduced resolution atmosphere, ocean and wave models will also be performed. The component models will be coupled through a mediator layer using the Earth System Modeling Framework (ESMF)/National Unified Operational Prediction Capability (NUOPC) protocols.

For the air-sea momentum exchanges, the momentum flux will include the ocean surface velocity in the shear across the surface, which was surmised by McClean et al. (2011) to improve the performance of the ocean model. Including the ocean shear in the momentum flux improves the penetration of the western boundary currents into the ocean basins, the distribution of EKE at the surface, and the size of the eddy driven recirculation gyres as shown in Figure 2. When using the traditional wind stress estimates from the numerical weather prediction models, the Gulf Stream and Kuroshio do not penetrate as far into the ocean as observed by surface drifters and the recirculation gyre does not extend far enough to the east. Including the ocean surface currents in the wind stress estimation increases the eastward penetration and size of the recirculation gyre. For the Agulhas, as it the case for many ocean models in that resolution range (McClean et al., 2011), too many Agulhas rings are generated that follow a northward pathway into the South Atlantic and the Agulhas Return Current eddies are too weak. Including the ocean shear in the wind stress reduces the number of Agulhas rings and widens their pathways into the South Atlantic and increases the strength and number of Agulhas Return Current eddies.

Preliminary tests of the fully coupled T359 NAVGEM, 1/12.5° HYCOM, and CICE have been performed for the DYNAMO (international experiment to study the initiation of the Madden-Julian Oscillation (MJO)) intensive observation period of November 2011 and for the summer Arctic melt and Antarctic freeze period of July 2011. The 30-day forecasts of the stand-alone atmospheric model were unable to

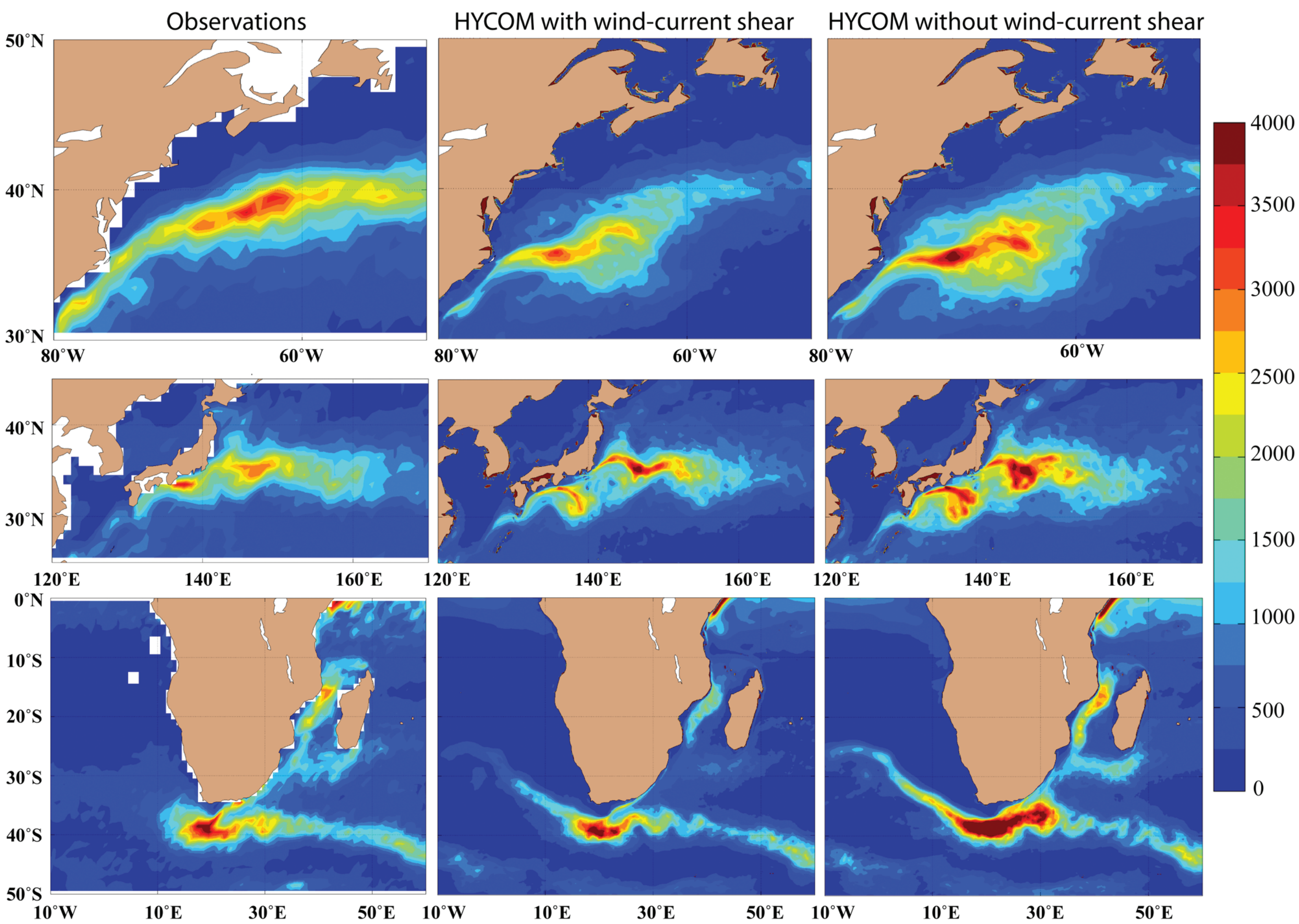

Figure 2. 5-year average surface eddy kinetic energy in cm2s-2 from observations (left panels), 1/12.5° HYCOM simulation that includes the wind-current shear (centre panels), and 1/12.5° HYCOM simulation that does not include the wind-current shear (right panels). The panels from top to bottom show the western boundary current systems of the Gulf Stream, Kuroshio, and Agulhas Current, respectively.

reproduce the onset of the late November MJO, the eastward propagation of the MJO, and the observed tropical rainfall patterns. The ESPC coupled model, on the other hand, developed an MJO in late November, but the eastward propagation was too weak and significant differences in the rainfall patterns remained. For the ocean, the global sea surface temperature rms error between the coupled model and the standalone analysis remained below 1 ºC for the 30-day period. Further testing with new convection schemes in the atmosphere are underway. Climate simulations at that resolution will not be feasible for the foreseeable future, but a great deal can be learned about the coupled model behaviour through these short-term experiments.

Acknowledgements

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Ensembles of eddying ocean simulations for climate

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1. Context

The recent IPCC's Fifth Assessment Report (AR5) provides an extensive description of the past evolution, present state, and projected future of the climate system. The modelling contribution to the assessment is based in particular on a large multimodel ensemble of historical and scenario simulations performed within the Coupled Model Intercomparison Project version 5 (CMIP5). The fully-coupled numerical models used for these climate simulations were implemented with a typical resolution of 1-2° in both the atmosphere and ocean components. At these spatial resolutions, atmospheric models spontaneously generate synoptic weather chaotic variability and associated turbulent fluxes that strongly influence larger time and space scales. Simulated ocean dynamics, however, remain laminar in most of these climate simulations: the oceanic "weather" associated with mesoscale eddies is unresolved, and only few of its impacts on larger scales are parameterised, in a rather crude way. Eddying ocean/sea-ice models, implemented at ¼° (and eventually finer) resolutions, will be routinely used in future generations of coupled climate and Earth system models. Indeed, the explicit representation of oceanic eddies improves the consistency of simulated dynamics and yields more realistic results in simulations driven by realistic atmospheric forcing functions (e.g. Penduff et al., 2010). This paper focuses on the nature of interannual-to-decadal variability in eddying ocean/sea-ice simulations with prescribed atmospheric forcing, which greatly differs from that in the laminar regime.

2. Intrinsic low-frequency variability in the eddying ocean

Since the pioneering simulations by Holland and Haidvogel (1981) and Cox (1987), idealised studies have shown that the ocean circulation driven by constant atmospheric forcings can spontaneously generate low-frequency (1-10 year) intrinsic variability when resolution (i.e. mesoscale eddy activity) increases. This generic behaviour was proven robust to the addition of seasonal forcing (Sushama et al., 2007; Shimokawa et al., 2010). Two main nonlinear paradigms have been proposed to explain this phenomenon. Spall (1996), Dewar (2003), Berloff et al. (2007), and Arbic et al. (2012) have shown that intrinsic low-frequency variability may be directly driven by mesoscale turbulence, i.e. through nonlinear interactions between the ambient potential vorticity (PV) field, eddies and the low-frequency variability, or through the temporal inverse cascade of kinetic energy driven by eddy-eddy interactions. Other authors, such as those reviewed in Dijkstra and Ghil (2005), used Dynamical System Theory approaches to show that mesoscale turbulence may "just" favour transitions of the oceanic circulation between multiple large-scale equilibria set by the mean PV field. It is very likely that these two paradigms are mutually compatible and provide complementary views on this phenomenon.

The contribution of low-frequency intrinsic variability in the real ocean cannot be estimated from observations alone since it is fully entangled with atmospherically-forced as well as coupled variability. This contribution is generally estimated from pairs of multi-decadal Ocean General Circulation Model (OGCM) simulations: one "fully-forced" hindcast, driven by the full range of atmospheric timescales over several decades, designed to mimic oceanic observations; and one climatological simulation, driven by a perpetual annual atmospheric cycle, designed to isolate the intrinsic low-frequency variability. Following this approach and Taguchi et al (2007)'s regional study, Penduff et al. (2011) showed at global scale that when mesoscale eddies are represented (at ¼° resolution), a substantial fraction of the interannual variance of Sea-Level Anomaly (SLA) found in a fully-forced hindcast remains present in the climatological simulation (without interannual forcing). Consistently with idealised studies, most of this intrinsic interannual SLA variance was found in mesoscale-active mid-latitude regions, i.e. in western boundary currents systems, and in the Antarctic Circumpolar Current (ACC).

In the framework of the CHAOCEAN project[1], these analyses of intrinsic low-frequency SLA variability are being extended to higher resolution simulations (1/12°) and to other climate-relevant variables (sea-surface temperature SST, Atlantic Meridional Overturning Circulation AMOC), with a focus on large spatial scales. Figure 1a shows that at scales larger than about 1000km, a substantial fraction of the SST interannual variance is spontaneously generated by the ocean in the very same regions where air-sea heat fluxes are the largest on Earth. This fraction is about 25-40% in the Gulf Stream and Kuroshio, and reaches 65-80% in the Agulhas Retroflection region. South of Africa, the intrinsic, large-scale interannual SST standard deviation reaches about 0.5°C. Figure 1b presents the same fraction for interannual AMOC variance: in the South Atlantic up to 40-60% of the fully-forced AMOC variability is actually intrinsic in our eddying simulations; interestingly, the 1/4° and 1/12° results do not differ much, suggesting that so-called "eddy-permitting" resolution captures the essential dynamics at work, at least in mid-latitudes. OGCM simulations are thus consistent with idealised studies: the nature of interannual variability in the eddying regime (1/4° or 1/12°) is different from that in laminar ocean models used in CMIP exercises, where the impact of intrinsic processes on SLA, SST or AMOC variances are barely noticeable (e.g. red line in Figure 1b).

1. Four-year research project funded by the Ocean Surface Topography Science Team: http://alturl.com/x8qs8

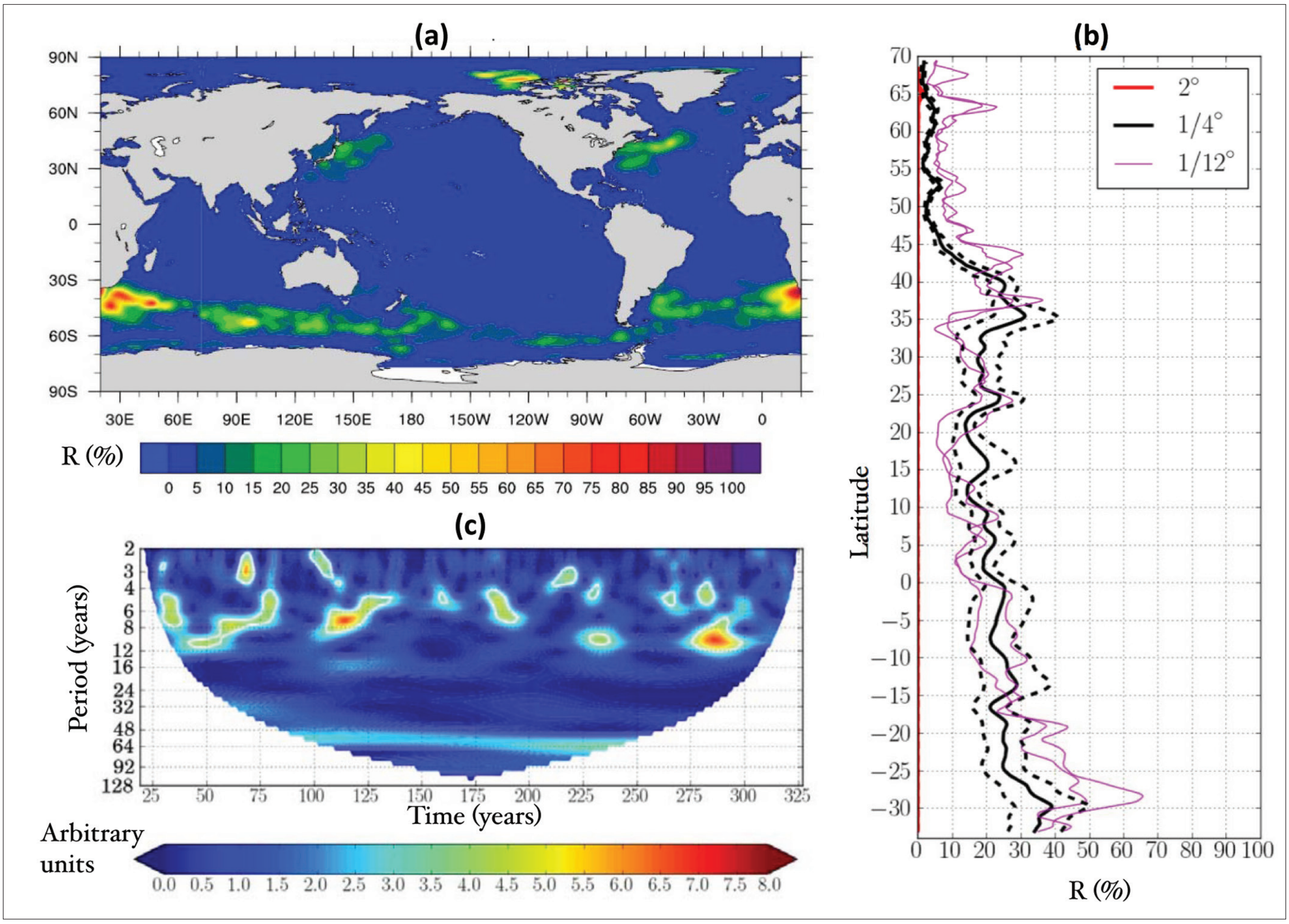

Figure 1: Panels (a) and (b) show ratios (R in %) between intrinsic and total interannual variances, deduced from seasonally- and fully-forced global ocean/sea-ice Drakkar simulations, respectively. (a) Ratio R is shown for large-scale sea-surface temperature (SST) from global 1/12° simulations: spatial scales smaller than 12° have been filtered out before computing variances. (b) Ratio R for Atlantic Meridional Overturning Circulation (AMOC), shown as an ensemble mean and spread for ¼° simulations (black lines), and as two independent estimates for 1/12° simulations (magenta lines); 2° simulations (red) are almost devoid of intrinsic variability. (c) Normalised wavelet power spectrum of the AMOC intrinsic variability timeseries at 26°N throughout the 327-year seasonally-forced, global ¼° Drakkar OGCM simulation. To appear in Sérazin et al. and Grégorio et al.

3.Stochastic character of the low-frequency variability

The idealised studies mentioned above also highlight the chaotic character of the low-frequency, pure intrinsic variability (i.e. simulated with seasonal or constant forcing). In other words, sensitivity to initial conditions and tendency for intermittent, random behaviour in nonlinear regimes are not restricted to mesoscale motions, but also concern low-frequency modulations of the circulation. Figure 1c shows that this property is also found in a realistic context: this wavelet power spectrum, computed from a 327-year 1/4° OGCM seasonally-forced simulation, reveals the intermittent and broadband character of the interannual AMOC pure intrinsic variability at 26°N.

An important question then comes to mind: does the pure (seasonally-forced) intrinsic low-frequency variability remain chaotic when a low-frequency atmospheric forcing is prescribed at the ocean surface? Recent investigations provide preliminary, although contrasting, answers. On the one hand, Pierini (2014) compared idealised shallow-water simulations of the North Pacific driven by constant and interannually-varying winds, and concluded that the low-frequency variability of the Kuroshio is chaotic in the former case but may become deterministic in the latter. On the other hand, the structure and magnitude of the interannual AMOC intrinsic variance generated by our global 1/4° OGCM under seasonal forcing does not differ much (in structure and intensity) from its counterpart estimated by Hirschi et al. (2013) under interannual forcing with the same model. In other words, interannual forcing might excite pure intrinsic variability modes (as also shown by O'Kane et al., 2013), but the interannually-forced low-frequency oceanic variability is very likely to remain stochastic in the eddying regime.

There are thus various indications that in forced eddying OGCM simulations, the large-scale, low-frequency, mid-latitude ocean variability is less deterministic (i.e. more sensitive to initial conditions) than in the laminar regime, and that eddy-related non-linearities give a stochastic flavour to low-frequency variability. This concerns in particular the large-scale, low-frequency SST and AMOC variability that can have direct and indirect impacts on the atmosphere and climate. This stochastic character advocates for a probabilistic study of the low-frequency oceanic variability in the eddying regime.

4. A global ensemble of multi-decadal oceanic hindcasts

The main goal of the OCCIPUT project[2] is to provide the first probabilistic description of the global oceanic evolution over the last decades, i.e. based on probability density functions (PDFs). Our plan is to force 50 eddying (1/4°) global ocean/sea-ice simulations with realistic atmospheric forcings over the period 1958-2013, given perturbed initial conditions. Our integration strategy, including the initialization, spin-up and ensemble generation, is being currently designed using regional reduced-sized ensembles, in close collaboration with the DRAKKAR Group[3]. During the beginning of the integration, we plan to activate within each member a stochastic parameterisation (of e.g. subgrid-scale density gradients, Brankart, 2013) to generate an ensemble spread. Ongoing tests confirm a priori expectations from instability theories: the ensemble spread, i.e. intrinsic variability, emerges at the scale of mesoscale eddies and grows fastest in frontal regions (Figures 2a, 2b, 2c). As expected from inverse cascade studies (Arbic et al., 2014), the space and time scales of intrinsic variability progressively increase: Figure 2d exhibits an inter-member difference that has reached Rossby wave scales (i.e. 1000km and 1-year) after 3 years away from the main fronts, i.e. at 26°N. These tests also show that switching off stochastic perturbations after a few months does not hamper the subsequent growth of the ensemble spread: mesoscale nonlinearities ultimately control the spin-up of intrinsic variability, as well as (presumably) its cascade toward larger scales and its magnitude. In other words, we plan to stochastically perturb the OCCIPUT ensemble members for 1 year or so, then let the spread grow, reach interannual and regional scales and its non-linearly saturated amplitude. This strategy should provide us with a data set of a few decades over which we plan to describe and study the spatial structure and variability of climate-relevant ensemble statistics in key regions: e.g. ensemble PDF of oceanic heat content, ensemble average (deterministic part) of Kuroshio transport, ensemble spread (stochastic part) of global SST, ensemble covariances (turbulent fluxes), etc.

5. Implementation, model outputs, expected outcomes

The OCCIPUT members are currently adapting the ORCA025 global 1/4° ocean/sea-ice model configuration (Barnier et al., 2006) for ensemble integrations and implementing the system on the CURIE supercomputer[4] . The 50 members will be integrated simultaneously as one executable over a few thousands of processors, so that specific ensemble (inter-member) statistics can be computed online, possibly at every timestep. Such statistics may include e.g. timestep estimates of (exact) turbulent heat fluxes across key sections, deciles of global SST distribution at hourly resolution, etc. The full 3-dimensional 50-member state

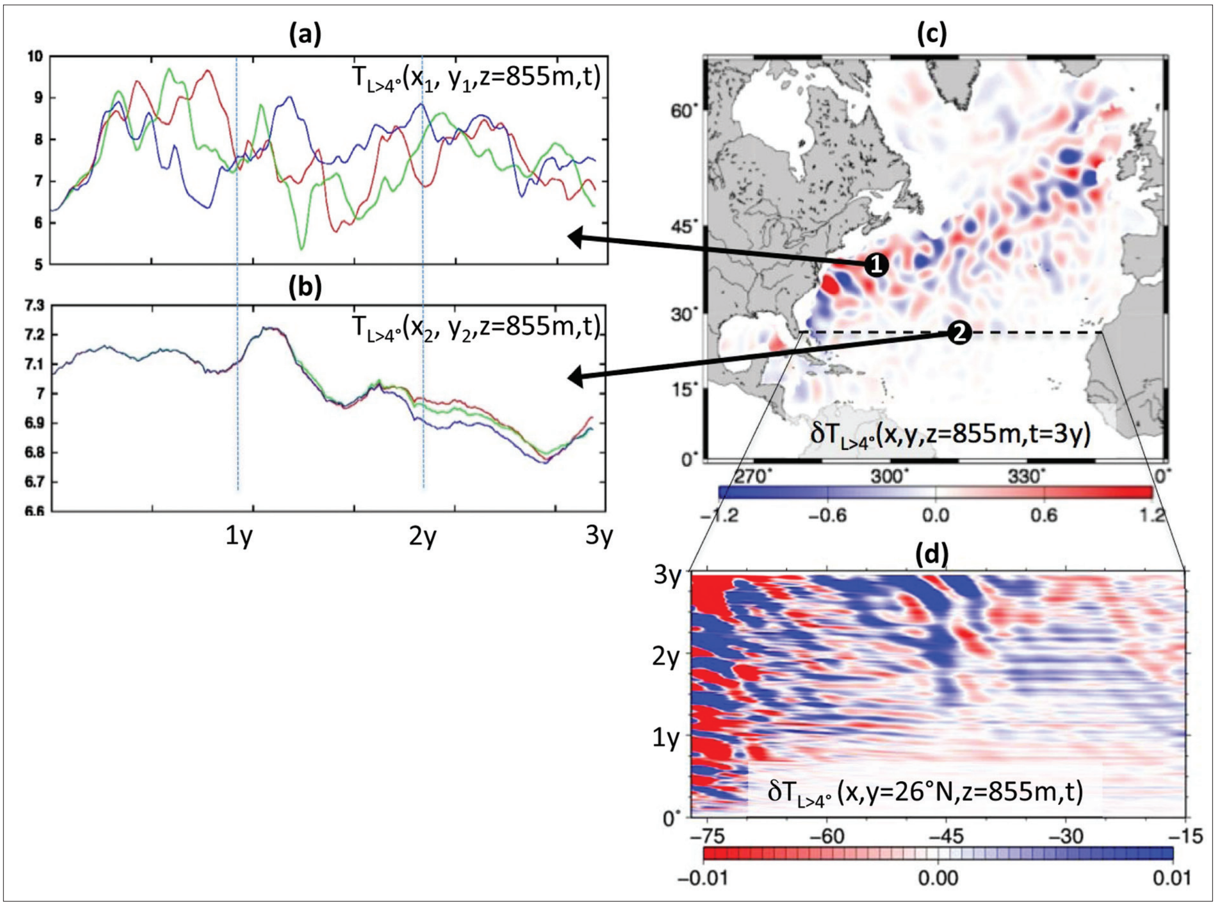

Figure 2: Subsurface temperature (T at 855m depth, °C) in a 3-member, fully-forced North Atlantic NEMO ¼° ensemble simulation with stochastic parameterisation (Brankart et al 2013); only scales larger than 4° are considered. Left panels show 3-year temperature timeseries from the 3 members at locations 1 (a) and 2 (b) indicated on the right. Right panels show the distribution of the inter-member T difference after 3 years as a function of longitude and latitude (c), and as a function of longitude and time along 26°N (d).

2. Four-year research project funded by ANR: http://alturl.com/ivfkg
3. who are developing the ORCA025, global ¼° ocean/sea-ice configuration: http://www.drakkar-ocean.eu/

4. http://www-hpc.cea.fr/en/complexe/tgcc-curie.htm. An estimated amount of 15 million core hours shall be requested on this machine through a PRACE proposal in fall 2014; we hope to produce the ensemble integration in 2015-2016.

vector will be archived as successive monthly averages, and selected 2-dimensional fields every 5 days; these data will allow offline computations of additional ensemble statistics (PDFs, variances, covariances) at 5-daily, monthly, or interannual timescales of e.g. sea-level anomalies, bottom pressure, components of the AMOC at 26°N or other observed latitudes, etc.

The system will also generate ensemble synthetic observations during the integration, i.e. 50 simultaneous collocated counterparts of the ENACT-ENSEMBLE temperature/salinity profile database (1958-present) or of along-track altimeter timeseries (e.g. Topex/POSEIDON, Jason, etc). Such data will allow a thorough, objective observation-based assessment of the probabilistic simulation using metrics defined in the context of weather ensemble prediction (e.g. rank histograms, Brier scores; see Toth et al., 2003). Such synthetic data may also provide relevant information about the representativeness of actual observations.

6. Conclusion

Academic models have illustrated the chaotic behaviour of the ocean circulation at high Reynolds number, not only in terms of mesoscale turbulence but also in larger (double-gyre or ACC-like) current systems, up to decadal timescales. Unlike laminar ocean models used in most current climate projections, eddying OGCMs also spontaneously generate a substantial interannual-to-decadal intrinsic variability under repeated seasonal forcing, with a stochastic character, a marked signature on AMOC, and on SST in regions where air-sea fluxes have the largest amplitude. Whether and how this ocean-driven low-frequency intrinsic variability may ultimately impact the climate predictability at various time and spatial scales is an important but unsettled question.

Before answering this question, it is useful to better assess the stochastic character of the low-frequency ocean variability, with a focus on climate-relevant indexes. The OCCIPUT project aims at performing in 2015 a 50-member ensemble of 1/4° global ocean/sea-ice hindcasts driven by the same 1958-present atmospheric forcing. The initial ensemble spread is expected to grow, cascade toward long space and time scales, and saturate in amplitude. We expect this eddying ensemble to provide a probabilistic description of the ocean state and evolution over the last decades, and a measure of the actual constraint exerted by the atmosphere on the past interannual-to-decadal ocean variability. New results are also expected from the computation of ensemble instead of temporal statistics (variances, covariances) for the estimation of time-dependant turbulent fluxes, and from the use of ensemble synthetic observations to probabilistically assess ocean simulations or the representativeness of observational datasets.

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Inserting tides and topographic wave drag into high-resolution eddying simulations

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1. Introduction

We report here on two of our recent development efforts for global high-resolution ocean models, undertaken in the HYbrid Coordinate Ocean Model (HYCOM; Chassignet et al., 2009). The simulations discussed here are "eddying" (Hecht and Hasumi, 2008), meaning that they include an energetic mesoscale eddy field. In one of our efforts (Arbic et al., 2010, 2012; Timko et al. 2012, 2013; Shriver et al. 2012, 2014; Richman et al., 2012), we have inserted tides into global eddying simulations. Tidal flows in a high-resolution global model with realistic rough topography generate internal tides-internal waves of tidal frequency. Similarly, geostrophic flows over rough topography generate internal lee waves (Bell, 1975). It is not yet possible to resolve internal lee waves in global models, but the impact of internal lee waves can be parameterised. In the other effort reported here, we have inserted parameterised topographic lee wave drag into global eddying simulations that do not include tides, with a preliminary focus on the resulting model energy budget (Trossman et al., 2013).

Global simulations of the tides and the atmospherically forced eddying general circulation have long been done separately. Global general circulation models began to resolve mesoscale eddies in the 1990's (e.g., McClean et al., 1997). Since then eddying models have become increasingly realistic (e.g., Hecht and Hasumi, 2008). Global modelling of internal waves is a newer endeavour. In the first global simulations of internal tides (Arbic et al., 2004; Simmons et al., 2004), only tidal forcing was employed, and the stratification was taken to be horizontally uniform. The insertion of tides into atmospherically forced general circulation models allows for internal tide propagation in a realistic horizontally varying stratification, for interactions between internal tides and eddies, and for the co-existence of tides and near-inertial waves, another important class of internal waves (e.g., Simmons and Alford, 2012). In global high-resolution models with simultaneous atmospheric and tidal forcing, the internal wave spectrum is beginning to be resolved, just as mesoscale eddies began to be resolved in global models two decades ago. Our simulations with simultaneous tidal and atmospheric forcing are being used to address a variety of scientific and operational questions.

The global energy budget has been a topic of great recent interest, largely because the mixing occurring when energy is dissipated may exert a strong control on the oceanic meridional overturning circulation (Munk and Wunsch, 1998). The mechanisms underlying the dissipation of the eddying oceanic general circulation are still under investigation. Candidate dissipation mechanisms include the transfer of energy from geostrophic flows to submesoscale motions in the upper ocean (e.g., Capet et al., 2008), bottom boundary layer drag (Sen et al., 2008; Arbic et al., 2009), and breaking of internal lee waves generated by geostrophic flows over rough topography (e.g., Naveira-Garabato et al., 2004). Nikurashin and Ferrari (2011) and Scott et al. (2011) estimated the globally integrated energy flux of geostrophic flows into internal lee waves over rough topography, and found it to be a substantial fraction of the ~1 TW of energy put by the wind into geostrophic flows (e.g., Wunsch, 1998, and others). Both the Nikurashin and Ferrari (2011) and Scott et al. (2011) estimates were "offline"; they utilised bottom flows from eddying simulations that did not employ topographic wave drag as they ran. In Trossman et al. (2013) we inserted inline topographic lee wave drag into eddying HYCOM simulations, thus ensuring feedbacks between the model bottom stratification and flow fields and the topographic internal lee wave drag.

The HYCOM simulations shown here are forced by output from the Navy Operational Global Atmospheric Prediction System (NOGAPS); going forward the simulations will be forced by the Navy Global Environmental Model (NAVGEM). The simulations shown here are run on a combination Mercator and tripolar grid with an equatorial grid resolution of 0.08° (8.9 km) and are coupled to a sea ice model at high latitudes. Global HYCOM simulations at 0.04° (4.5 km) have also been performed, but are not discussed here (see article by Chassignet et al. 2014, this issue). The model reproduces the general circulation of the global ocean, the strength and variability of the western boundary currents, such as the Gulf Stream and Kuroshio, and the mesoscale eddies generated by instabilities of the major currents (Thoppil et al., 2011).

2. Insertion of tides into eddying simulations

In our HYCOM simulations forced by both atmospheric fields and tides (Arbic et al., 2010, 2012), we include the four largest semidiurnal tidal constituents (M2, S2, N2, and K2) and the four largest diurnal tidal constituents (K1, O1, P1, and Q1). We use the parameterised topographic internal wave drag scheme of Garner (2005), modified to limit the maximum decay rates to (9 hours)$^{-1}$. The wave drag parameterisation represents the drag resulting from the generation and subsequent breaking of unresolved small-vertical scale internal waves by tidal flow over rough topography. Because the wave drags acting on tidal versus non-tidal motions have different strengths (Bell, 1975), tidal and non-tidal bottom flows are separated using a 25-hour running boxcar filter. Presently, the scalar approximation (Ray, 1998) is used for the self-attraction and loading term (Hendershott, 1972).

Barotropic tidal sea surface elevations in HYCOM have been compared to measurements made from tide gauges (Shum et al., 1997) and state-of-the-art data-assimilative barotropic tide models (Egbert et al., 1994). In Shriver et al. (2012) we also compared the modelled internal tide perturbation to sea surface height (SSH) - computed from a spatial high-pass of the total amplitude of tidal SSH - to internal tide perturbations computed from along-track satellite altimeter data (Ray and Byrne, 2010). Figure 1 shows global maps of the M2 internal tide amplitude in HYCOM versus along-track altimeter data. The hotspots around locations such as Hawai'i, the Aleutian Islands, the Tuamotu Archipelago in the tropical central South Pacific, and Madagascar match up reasonably well in the two maps. Furthermore, the root-mean-square (rms) perturbation magnitudes over the five hotspot regions delineated by boxes in the bottom panel of Figure 1 agree with the altimeter rms values to within about 20%. Comparisons of tidal currents, and their vertical structure, in HYCOM versus historical moored observations were made in Timko et al. (2012; 2013).

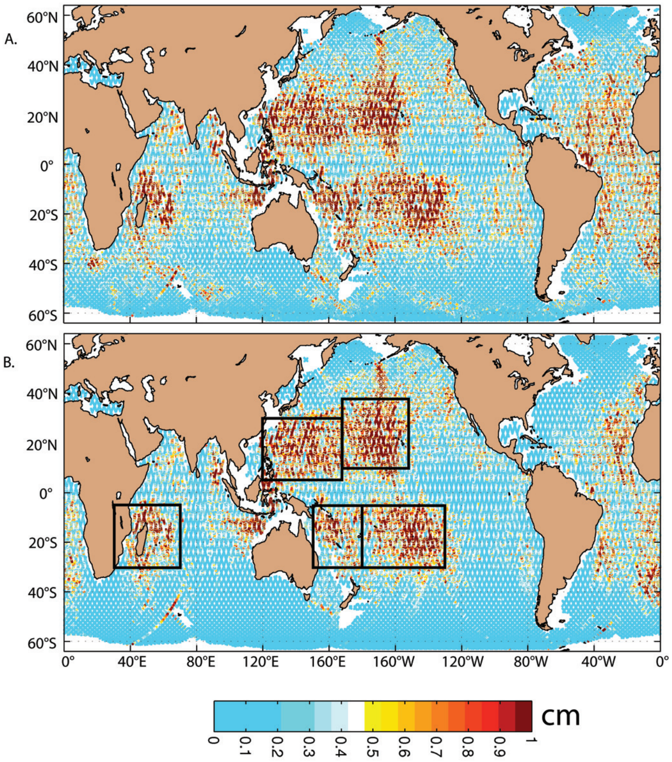

Figure 1. The internal tide amplitude of the principal lunar semidiurnal constituent M2 from the (a) altimetric-based and (b) HYCOM tidal analyses. The five subregions denoted by black boxes in (b) are used to compute area-averaged amplitudes in Shriver et al. (2012). From Shriver et al. (2012).

We have used the HYCOM tidal simulations to distinguish between tidal and non-tidal contributions to the wavenumber spectrum of SSH (Richman et al., 2012). The slope of the wavenumber spectrum is of great theoretical interest, as it contains clues about the dominant dynamics of low-frequency flows (e.g., LeTraon et al., 2008; Xu and Fu, 2011). Because of the 10-day repeat track time, current state-of-the-art TOPEX/JASON satellite altimeters alias tides into longer periods (Parke et al., 1987). In contrast, because our model output is written at hourly intervals, we can easily separate low- from high-frequency signals in HYCOM. In regions such as the Kuroshio, where low-frequency motions are more energetic than tidal motions, the wavenumber spectrum of SSH is dominated by low-frequency motions. However, in regions where internal tides are energetic, for instance near Hawai'i, high-frequency motions dominate the high-wavenumber end of the wavenumber spectrum. The results of Richman et al. (2012) imply that internal tides will have to be accurately removed from data taken by the planned high-resolution wide-swath satellite altimeter (Fu et al., 2012) before low-frequency oceanic motions can be investigated. The accuracy with which internal tides can be removed from altimeter signals depends on the degree of non-stationarity of the internal tides. Internal tide non-stationarity in HYCOM is investigated in Shriver et al. (2014).

3. Insertion of topographic wave drag into eddying simulations

In Trossman et al. (2013) we again use the topographic wave drag parameterisation of Garner (2005), this time applied to low-frequency motions. The Trossman et al. (2013) simulations did not include tides. For simplicity, we refer to the Garner (2005) scheme here as a "wave drag" scheme even though it includes effects of topographic blocking, which yields low-level turbulence, as well as lee wave production and breaking. Maps of energy dissipation by quadratic bottom drag and by the parameterised topographic lee wave drag, both averaged over one year, are shown in Figure 2. As in the "offline" estimates of Nikurashin and Ferrari (2011) and Scott et al. (2011), as well as the offline estimates done in Trossman et al. (2013), the inline estimates shown in the bottom panel of Figure 2 indicate substantial lee wave drag dissipation in the Southern Ocean. The globally integrated internal lee wave drag dissipation in the inline Trossman et al. (2013) estimates is 0.4 TW, comparable to that seen in the offline estimates of Nikurashin and Ferrari (2011), Scott et al. (2011), and Trossman et al. (2013). However, as anticipated, there is indeed a strong feedback between the internal lee wave drag and the model flows. Adding wave drag to HYCOM substantially changes the modelled near-bottom eddy kinetic energy and stratification fields (Trossman et al., 2013).

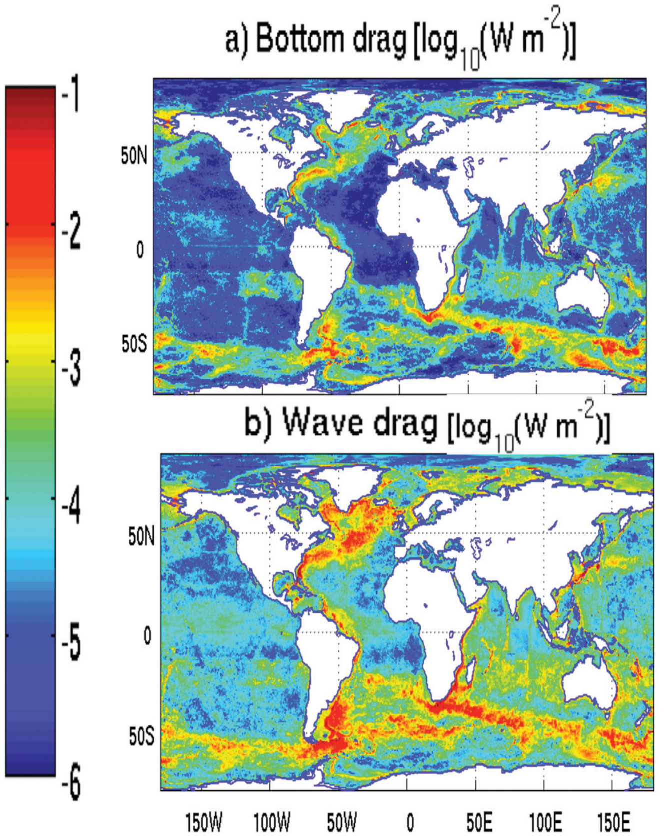

Figure 2. Log10 of the bottom and wave drag terms in the energy budget of 1/12 eddying HYCOM simulation with , where each drag term has units [W m-2]: (a) quadratic bottom boundary layer drag and (c) parameterised internal lee wave drag. Adapted from Trossman et al. (2013).

4. Ongoing and planned work

In ongoing and planned work with the HYCOM tide simulations, we are comparing the SSH frequency spectra in HYCOM versus tide gauges, as a measure of the accuracy of the partition of modelled low- versus high-frequency energy (Savage et al., paper in revision). We are estimating tidal aliasing errors in altimetrically-derived low-frequency wavenumber spectra and spectral fluxes (Savage et al., paper in preparation). We are also comparing the kinetic energy frequency spectra in HYCOM versus moored current meter records (Müller et al., paper in preparation). We are comparing both low- and high-frequency temperature variances (which are related to available potential energies) to temperature variances in moored historical records (Luecke et al. and Bassette et al., respective papers in preparation). To further elucidate the energetics of internal tides, we are preparing global maps of the baroclinic tidal energy fluxes and barotropic-to-baroclinic tidal energy conversions (Ansong et al. and Buijsman et al., papers in preparation). The latter work is part of the National Science Foundation-funded Climate Process Team project "Collaborative Research: Representing internal-wave driven mixing in global ocean models", which focuses on improving estimates of internal-wave mediated mixing in the ocean, and is led by Professor Jennifer MacKinnon of the Scripps Institution of Oceanography. Finally, our long-term goals for the HYCOM tides work include improvement of tidal accuracy through the inclusion of data assimilation and better estimates of the self-attraction and loading term, and usage of the global HYCOM tidal solution to force regional and coastal models at their open-water boundaries.

In ongoing and planned work on parameterised topographic lee wave drag, we will investigate whether the inclusion of wave drag into eddying models improves the comparison of the models with observations. We will also compare the dissipation predictions of the Garner (2005) and Bell (1975) schemes with dissipation inferred from microstructure observations. Finally, we will investigate the impact of a more complete parameterisation of the vertical deposition of internal lee wave drag on the energetics and abyssal circulation in eddying models. These projects are all led by co-author Trossman (papers in preparation).

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Overflow parameterisations in climate models

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1. Introduction

Flows across shallow sills and through straits control the distribution of the water masses in the deep ocean. Yet, the credibility of present climate models is limited by their ability to represent processes that occur on scales smaller than the model grid scale (currently typically 100 km) such as overflows (Legg et al., 2009). A proper representation of overflows in numerical models must adhere to two principles: 1) minimization (or understanding) of numerically induced dianeutral mixing (also called "spurious" mixing), and 2) a decent parameterisation of unresolved processes. The practical implementation of parameterisations is directly linked to the vertical coordinate choice (Griffies et al., 2000a). Numerically induced dianeutral mixing arises in fixed-coordinate (level or terrain-following) models because of advective truncation errors and horizontal/isosigma diffusion tensors. Isopycnal models by definition have no spurious diapyncal mixing and small dianeutral mixing as long as isopycnals do not deviate far from neutral surfaces. The challenge for fixed coordinate models is to reduce the numerically-induced mixing to levels that are below observations (Griffies et al., 2000b; Marchesiello et al., 2009).

In the case of overflows, Ilicak et al. (2012) state that in order to reduce spurious dianeutral mixing, the grid Reynolds number should be small enough to suppress the grid noise and resolve the flow. This is consistent with Legg et al. (2008), who argued that the bottom frictional boundary layers should be resolved to avoid spurious entrainment. However, the small scale nature of overflow processes with horizontal and vertical length scales (order 1 km and 10 m, respectively) requires finer horizontal and vertical resolutions than what is commonly used in climate models (100 km and 50-200 m, respectively). In level coordinate models with such a coarse resolution, the flows over staircase topography tend to have excessive convective entrainment, resulting in deep waters that are too light and that remain too shallow (Roberts et al., 1996; Winton et al., 1998). Several approaches have been proposed that attempt to reduce this bias with mixed results (Treguier et al., 2012): artificial modification of the model's bottom topography, bottom boundary layer parameterisations, and streamtube models. Implementation of these approaches turned out to be quite sensitive to small changes in the topography and models grids. The most successful implementation to date in a climate model is the parameterisation used by the NSF-DOE Community Earth System Model (CESM), which is based on the Marginal Sea Boundary Condition (MSBC) of Price and Yang (1998). When Nordic Sea overflows are parameterised, Danabasoglu et al. (2010) report a significant improvement in the penetration depth of the North Atlantic Deep Water (NADW), reducing the chronic, shallow penetration depth bias in level coordinate models. They also show bias reductions in the deep temperature and salinity distributions in the North Atlantic with parameterised overflows.

Even at high horizontal and vertical resolutions, reduction of the spurious dilution of dense waters in level coordinate models is only achieved when high viscosity is used to render the simulated flow more laminar (Ilicak et al., 2012). Terrain-following coordinate models at fine resolution do not exhibit as much spurious mixing as the level coordinate models (Legg et al., 2006; Ilicak et al., 2012), but at the present the use of this vertical coordinate for climate modelling purposes is very limited (see Griffies et al. (2000a) and Marchesiello et al. (2009) for a discussion). It is, however, ideal for the representation of bottom boundary layers (important for the large scale ocean circulation primarily as sinks of momentum and for mixing) since it can enforce very high resolution near the bottom and incorporate high-order turbulence closure schemes (Hallberg, 2000; Legg et al., 2006). The fact that, in contrast to level (and to some extent terrain-following) coordinate ocean models, an isopycnal coordinate model inherently has too little mixing, meaning that there is a need for parameterisations of entrainment due to shear-driven mixing in this class of models. To date, most entrainment

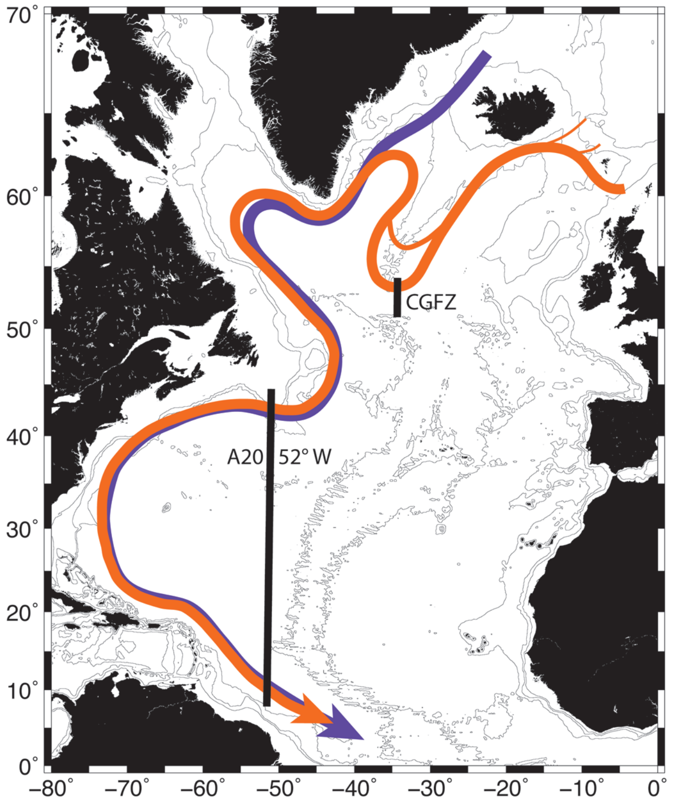

Figure 1: A schematic flow pattern of the Denmark Strait overflow water (purple) and the Iceland-Scotland overflow water (orange) in the deep western boundary current of the North Atlantic Ocean. The black vertical lines denote two meridional sections along which observed and modelled salinity/tracer distribution are shown (Figure 2).

parameterisations are based on either a critical Froude number as in the work of Price and Baringer (1994) (e.g., Hallberg, 2000; Xu et al., 2006; Jackson et al., 2008), or a subcritical Froude number (e.g., Cenedese and Adduce, 2010), or second-order turbulent closures (e.g., Ilicak et al., 2009).

2. Modelling overflows in the North Atlantic

Dense water formed in the Nordic Seas and the Arctic Ocean enters the deep North Atlantic via two overflow systems over the Greenland-Scotland Ridge (Figure 1). To the west, the Denmark Strait overflow water (Jochumsen et al., 2012) flows through the Denmark Strait and continues down the continental slope of the western subpolar North Atlantic (e.g., Schott et al., 2004). To the east, the Iceland-Scotland overflow water takes a more complex pathway, involving flows over the Iceland-Faroe Ridge and out through the Faroe Bank Channel. The Iceland-Scotland overflow water then continues to flow southwestward along the northwestern slope of the Iceland Basin (Saunders, 1996) and westward through the Charlie-Gibbs fracture zone (CGFZ) (Saunders, 1994). The Iceland-Scotland overflow water through the CGFZ subsequently turns northward into the Irminger Sea. Some Iceland-Scotland overflow water also enters the Irminger Sea by flowing westward across the Reykjanes Ridge north of CGFZ (Xu et al., 2010). In the Irminger Sea, the Iceland-Scotland overflow water joins and overrides the Denmark Strait overflow water on its way south toward the equator (Saunders, 2001). The Denmark Strait overflow water and Iceland-Scotland overflow water are two important components of the NADW, which also includes Labrador Sea

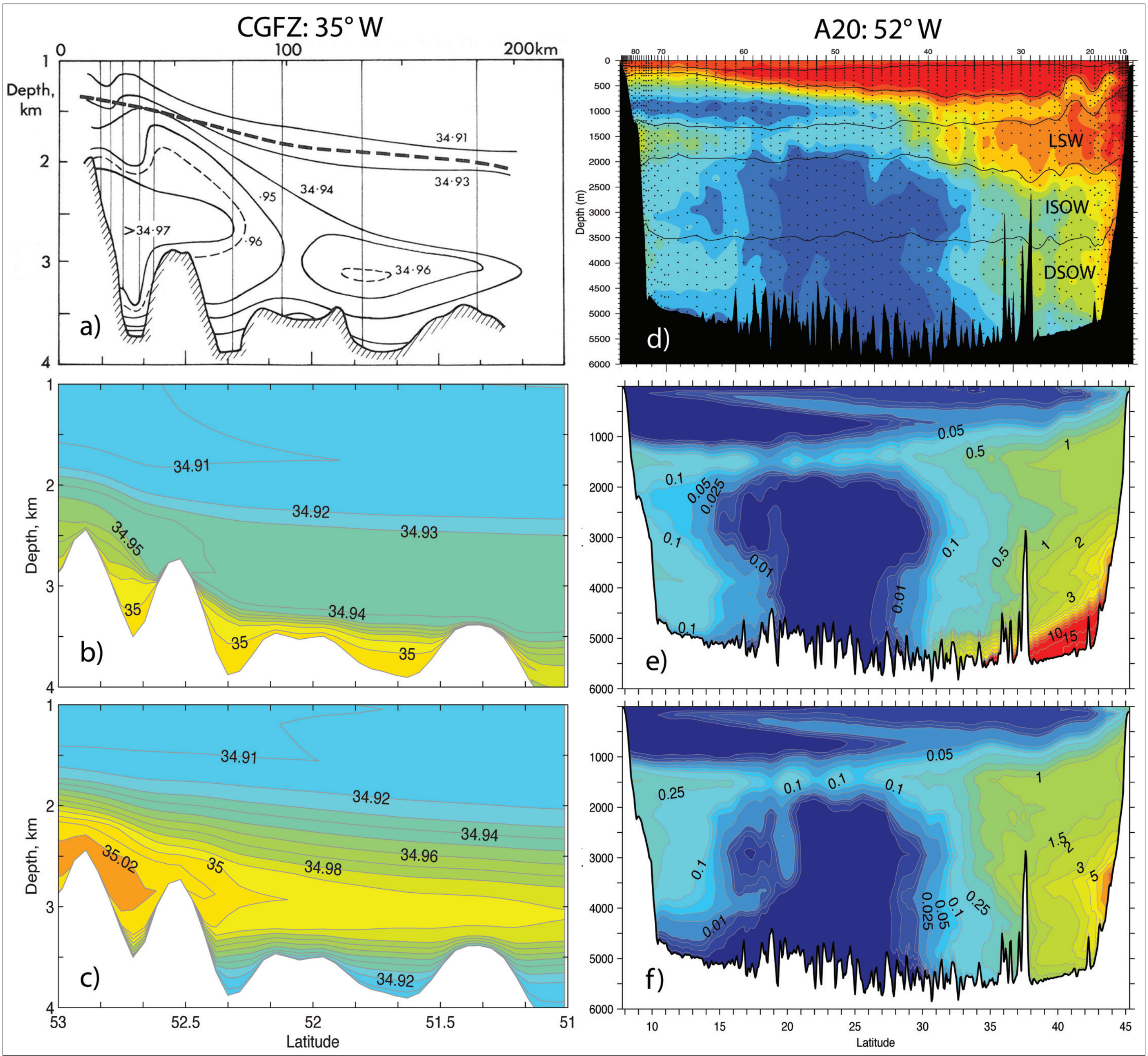

Figure 2: (a-c) Salinity distribution along a meridional section across the Charlie-Gibbs Fracture Zone at 35°W, and (d-f) tracer distribution along the WOCE section at 52°W in the western North Atlantic. (a) and (d) are observed salinity from Saunders (1994) and CFC concentration from Hall et al. (2004), respectively. (b) and (e) are modelled salinity and passive tracer concentration (x100) in the HYCOM simulations using KPP; (c) and (f) are the same as (b) and (e), but with the Xu et al. (2006) entrainment mixing parameterisation.

Water (LSW) and modified Antarctic Bottom Water (AABW). Knowledge of the detailed circulation pathways and volume transports of these overflow water masses is therefore fundamental for a general description of the Atlantic meridional overturning circulation. Moreover, the overflow waters can influence decadal (and longer time scale) climate variability, primarily through their impacts on the Labrador Sea stratification (Yeager and Danabasoglu, 2012).

Here we discuss an example of how one can represent these overflows in eddy-rich or "eddying" numerical simulations, meaning that they are eddy-resolving over most of the domain and include an energetic mesoscale eddy field. The grid spacing in these simulations also allows for a reasonable representation of the topography of the overflows. Using the HYbrid Coordinate Ocean Model (HYCOM, Bleck, 2002; Chassignet et al., 2003) configured for the North Atlantic with 1/12° resolution, Xu et al. (2010) showed that while the model overflow water from the Greenland-Scotland Ridge to Irminger Sea was in reasonable agreement with the observed volume transports as well as temperature and salinity characteristics, the modelled Iceland-Scotland overflow water through the CGFZ exhibited a salinity maximum near the bottom, whereas in observations the salinity maximum is located above the bottom at 2500-3000 m (Saunders, 1994; see Figure 2a). The default vertical coordinate configuration of HYCOM is isopycnic in the open stratified ocean, but it makes a dynamically and geometrically smooth transition to terrain-following coordinates in shallow coastal regions and to fixed pressure-level (mass conserving) coordinates in the surface mixed layer and/or unstratified open seas. In doing so, the model takes advantage of the different coordinate types in optimally simulating coastal and open-ocean circulation features (Chassignet et al., 2006). In HYCOM, the default shear-driven mixing parameterisation is that of the K-profile parameterisation (KPP, Large et al., 1994), which results in insufficient diapycnal mixing for overflow entrainment process (Xu et al., 2006) (Figure 2b,e). Since this parameterisation was developed to represent upper-ocean physics primarily, it is not surprising that it does not work in a different regime where the length scales and velocity shears driving the turbulence are different (Jackson et al., 2008).

The impacts of shear-driven mixing parameterisations on the Iceland-Scotland overflow water and Denmark Strait overflow water in the North Atlantic are evaluated in twin experiments, one with the default KPP and the other with the entrainment parameterisation of Xu et al. (2006). Both simulations have a horizontal resolution of 1/12° and 64 layers in the vertical. They are forced with climatological atmospheric forcing from the European Center for Medium range Weather Forecasting (ECMWF) reanalysis ERA40 for 20 years. Figure 2a,b,c compare the observed and modelled salinity distributions along a meridional section near 35°W across the CGFZ (Figure 1). The observations (Saunders, 1994) are based on a conductivity-temperature-depth (CTD) survey in 1988 while the model results are time average over the last 5 years. The observed high-salinity Iceland-Scotland overflow water, defined as salinity greater than 34.94, occupies a depth range from 1500–3500m (Figure 2a) which is well represented in the simulation with entrainment mixing parameterisation (Figure 2c), but not in the simulation with KPP (Figure 2b) where the highest salinity is found at bottom as in Xu et al. (2010). The Iceland-Scotland overflow water has a high salinity signature due to entrainment of the warm and salty upper North Atlantic Water as it flows into the Iceland Basin. It can be easily distinguished from the relatively low salinity of the Denmark Strait overflow water within the Irminger and Labrador Seas, but further south in the subtropical western North Atlantic the Denmark Strait overflow water overrides the modified AABW, which also has low salinity. In order to identify the model Denmark Strait overflow water without ambiguity, a passive tracer is injected in the numerical models north of the Denmark Strait sill. In Figure 2d,e,f, the distribution of the passive tracer is compared to observed chlorofluorocarbons (CFCs) (e.g., Hall et al., 2004) along a meridional section in the western North Atlantic near 52°W (Figure 1). This section (WOCE A20) has been surveyed several times and the basic structure has not changed. The high concentration of CFC near 3500-4000 m is the Denmark Strait overflow water core. As for the salinity fields, the HYCOM simulation with entrainment mixing parameterisation exhibits a tracer distribution that agrees with the observed CFCs, whereas the simulation with KPP show highest tracer concentration near the bottom. Note the discrepancy between models and observations in the upper layers is due to the fact that the model tracer was directly injected into the Denmark Strait overflow water below 400 m at the Denmark Strait sill whereas the CFCs enters the ocean through winter time convection. Figure 2 highlights the importance of the shear-driven mixing parameterisation in representing the structure of water mass distributions in the deep ocean. It is important to note that the density difference between the Denmark Strait overflow water and AABW is subtle (less than 0.02 kg m^{-3} in σ_2 in the subtropical North Atlantic), making it challenging to accurately model.

3. Discussion

The above results demonstrate that it is possible to parameterise the water mass transformations associated with overflows. The shear-driven mixing parameterisation of Xu et al. (2006) worked well for this specific configuration and class of model, but it is unclear how it would hold for coarser or finer horizontal resolutions – any shear mixing parameterisation needs to take into account changes in the magnitude of the vertical shear induced by any increase/decrease in the horizontal grid spacing (most of the time also associated with a change in numerical viscosity). However, shear-driven mixing is not the only physical process that controls the transport of dense water through overflows and its mixing with the ambient water. Hydraulic control, interactions with narrow canyons, mesoscale eddies, tides, and bottom friction also play a role (Legg et al., 2008). Bottom boundary layers are not resolved in current climate models and are seldom parameterised (Killworth, 2003). Legg et al. (2006) showed how important it is to properly capture the homogenization of tracers induced by the mixing driven by frictionally driven shear.

The ability of isopycnic coordinate models to have a decent representation of overflows is one of the reasons behind the development of a new generation of generalised vertical coordinate ocean models: Model for Prediction Across Scales (MPAS; Ringler et al., 2013) and MOM6 (http://www.gfdl.noaa.gov/ocean-model). Furthermore, the horizontal and vertical grid resolutions in the next generation of high-resolution climate simulations will still be coarser than what is needed to resolve the overflow physics. One approach that could be used to circumvent many of the issues raised in this article would be to apply very high-resolution nested grids in overflow regions (Fox and Maskell, 1996). Despite its potential, this approach has not yet been implemented in climate models, primarily because of technical challenges associated with three-dimensional two-way grid interactions and computing cost. Recent developments behind two-way grid interaction software (AGRIF and OASIS) and new computer architectures (GPUs, MICs) may provide an impetus for the inclusion of high resolution nested grids in the near future. MPAS with its multi-resolution approach is also well suited to test this approach.

Acknowledgements

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Numerical instabilities of the ice/ocean coupled system

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Abstract

The dynamics of floating ice and the ocean are tightly coupled. Both participate in the same external gravity waves, and the wind-driven Ekman transport spans both ice and water. However, for historical reasons, many coupled sea-ice/ocean models have traditionally introduced approximations that inhibit aspects of this dynamical coupling. These approximations also allowed sea-ice and the oceans to be treated as completely separate components that can be stepped forward in time with limited interactions. However, as these approximations are being relaxed to allow coupled ice-ocean models to realistically depict a more complete range of physical phenomena, groups around the world are experiencing the emergence of numerical ice/ocean coupling instabilities, especially at high horizontal resolution. This note describes several distinct dynamic ice-ocean coupling instabilities, how they can be diagnosed, and discusses options for eliminating these instabilities.

1. An historical perspective on the numerical evolution of coupled ice-ocean models

External gravity waves in the ocean are roughly 100 times faster than the flow speeds or internal gravity waves. Using the short time step required to explicitly resolve these fast external waves for the full 3-dimensional ocean currents and structure leads to a very slow model, so many early ocean-climate models made a rigid-lid approximation (Bryan, 1969). With a rigid lid, the ocean's depth-integrated flow is forced to be non-divergent, so precipitation and evaporation are treated as virtual salt fluxes ($F_{Salt} = - S_0 F_{Water}$) that approximately capture their impact on salinity instead of being treated as fluxes of fresh water. Similarly, sea-ice was treated with virtual salt fluxes for consistency with the ocean's rigid lid approximation; in this approximation, sea-ice has a mass for evaluating its accelerations and thermal properties, but it does not exert any pressure loading on the ocean.

If sea-ice does not exert any pressure on the ocean, the dynamic connections between the sea-ice and ocean occur via stresses and heat or fresh water fluxes, and the coupled interactions occur on timescales of hours to days, rather than the timescales of minutes to hours associated with external gravity waves. Moreover, without the tight dynamic response, sea-ice can be treated as a completely separate component from the ocean, and as a result sea ice models (e.g., CICE, LIM, or SIS) could be developed independently of the ocean models to which they are coupled.

Ocean climate models have long-since abandoned the rigid lid approximation in favour of a split-explicit (e.g., in GOLD, HyCOM, MOM, NEMO and ROMS), split-implicit (e.g., in POP), or 3-dimensionally implicit (e.g., in MITgcm) time stepping schemes to handle the external gravity waves. Many ocean climate models have also migrated from fixed-size Z-coordinates to mildly stretched Z*, pressure, isopycnal or hybrid coordinates that allow for modest or substantial movement of the ocean's surface. These changes in the ocean models' algorithms permit the use of "natural" fresh water fluxes between the ocean and ice (instead of virtual salt fluxes) and allow for the sea ice to exert a pressure that displaces the ocean's surface.

Traditionally, coupled models sequentially update the ice, atmosphere, and ocean with the latest properties of each component used in the fluxes applied to the next. On massively parallel computers, however, such "sequential coupling" yields a wall-clock throughput that is roughly half as fast a model that uses "concurrent coupling", in which the atmosphere and ocean are simultaneously updated on a separate set of processors, each using fluxes based on the properties of the other component at the start of a timestep. Sea ice has strong thermodynamic interactions with the atmosphere, while the fast gravity wave dynamic coupling with the ocean has traditionally been disabled in coupled models, so it is not uncommon for the sea-ice to be updated along with the atmosphere. However, as shown here, concurrent (i.e., lagged) coupling between the ocean and sea-ice exacerbates coupling instabilities.

The instabilities described below can be aggravated as vertical and horizontal ocean model resolutions become finer, due to the physics of sea-ice. An ice pack acts to rapidly redistribute stresses across its extent and can act to damp the dynamical instabilities. As a model's resolution increases, though, it is increasingly able to represent thick floes that become unlocked from a melting ice-pack.

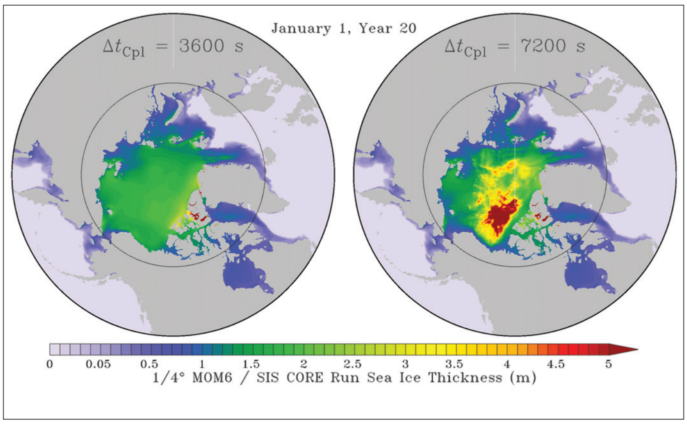

Figure 1: Sea-ice thickness at the start of the 20th year of two "Common Ocean Reference Experiment" (Griffies et al., 2009) forced ¼° resolution MOM6 / SIS global sea-ice ocean models, which differ only in their coupling timesteps. The run on the left uses a 1 hour coupling timestep, while that on the right uses a 2 hour step. The ocean and ice models use their own, shorter timesteps for their dynamics and thermodynamics, and these are identical between the two runs. The much thicker ice and irregular patterns in the central Arctic are a direct consequence of the coupled instabilities described here.

One example of a coupled ice-ocean instability is illustrated in Fig. 1, where the sea ice thickness is shown after a 20-year spinup of two variants of a ¼° resolution ice-ocean model. The two models differ only in their coupling timesteps. With a 1 hour timestep, the currents are relatively smooth and steady, and the sea ice has a sensible thickness. With a 2 hour coupling timestep, there are strong small-scale convergent and divergent flows that vary with high frequencies, which cause the ice to aggregate and thicken in some grid cells while leaving open water in adjacent cells where new ice can freeze. In the central Arctic, the model with the 2-hour timestep is also mechanically mixing from the surface down into the Atlantic water, significantly degrading the Arctic watermass structure. The coupled model does not become so unstable that it "crashes" in this case, but neither does it provide a physically plausible sea-ice field. We have encountered other examples where an ice-ocean instability is catastrophic, sometimes emerging within a very few time steps after decades or centuries of stable simulation. A sea ice thickness field that is very similar to the left panel can also be obtained by artificially reducing the viscous drag coefficient between the ice and ocean, or by "levitating" the sea ice and not applying the pressure from the weight of the sea ice to the ocean. These results provide strong circumstantial evidence for a mixed stress / gravity wave coupled time-stepping instability that severely degrades the physical credibility of the Arctic in this coupled model. To illustrate these instabilities, idealised models that exhibit them separately will be presented in the following sections.

2. Idealised illustrations of several ice-ocean coupling instabilities

2.a. Non-rotating stress instabilities

If the stress between the ocean and ice is a function of the speed of the two media, a numerical instability can arise if a semi explicit calculation of the stress is used. This can be illustrated with a simplified two-equation toy model:

$$\frac{\partial u_i}{\partial t} = -\frac{\lambda}{H_i}(u_i - u_o) \qquad \frac{\partial u_o}{\partial t} = \frac{\lambda}{H_o}(u_i - u_o) \tag{1}$$

Here u_i and u_o are the velocities of the ice and ocean, and H_i and H_o are the masses per unit area of the ice and ocean. The stresses in (1) are approximated with a simple linear drag law with coefficient λ, which has units of a velocity times a density; this could be a linearised proxy for a quadratic drag law or it could be a vertical integral of a laminar viscous drag. If λ is based on a linearization of a quadratic drag law, for a typical ocean flow speed of order 3 cm/s and a nondimensional drag coefficient of 3x10-3, λ would be of order 10-2 kg m^{-2} s^{-1}. In turn, $H_{i,o} = \rho_{i,o} h_{i,o}$, where h_i and h_o are the thicknesses, and ρ_i and ρ_o are the densities of the ice and ocean, respectively. Ice is rigid, so hi can be interpreted literally as the thickness of the ice, whereas ho is best interpreted as the thickness of an actively turbulent surface boundary layer or a laminar viscous surface layer over which the stress is distributed in the ocean.

The numerical analysis of the stability of this system of equations is particularly simple, since the two physical modes are a steady uniform flow of the ice and ocean, and a damped sheared flow between the ocean and ice. A standard Von Neumann stability analysis (e.g., Durran, 1999) leads to a quadratic equation for the numerical amplification factors of the two modes. The analysis of other dynamic instabilities can be obtained from similar analysis, but their derivation is more mathematically involved, resulting in quartic equations, so the key conditions for their growth will simply be quoted here without detailed derivation.

If the two velocities are stepped forward explicitly with a timestep, Δt, the evolution of the velocities from time n to time n+1 is given by:

$$\frac{u_i^{n+1} - u_i^n}{\Delta t} = -\frac{\lambda}{H_i}\left(u_i^n - u_o^n\right) \qquad \frac{u_o^{n+1} - u_o^n}{\Delta t} = \frac{\lambda}{H_o}\left(u_i^n - u_o^n\right) \tag{2}$$

Solving (2) for $u_{i,o}^{n+1}$ shows that with each timestep, the velocity difference between the ice and ocean are reduced by a factor of $A = \left(1 - \lambda\Delta t/H_o - \lambda\Delta t/H_i\right)$. For a short enough timestep and sufficiently thick ice and ocean, this is the correct result. However, if the sea ice or ocean are thin enough that A < -1, the velocities will oscillate and grow with each timestep and diverge. The ocean boundary layer is typically thick enough that this stress instability does not occur. By contrast, sea ice can be very thin when it starts to grow, making a fully explicit coupling unstable for any timestep. For example, with a 1 hour coupling timestep, the explicit coupling described by (2) would be unstable whenever the ice is thinner than about 2 cm.

Because thin ice invariably occurs somewhere, the ocean / sea-ice stress is invariably calculated implicitly with the sea-ice velocity. The toy coupling equation then becomes

$$\frac{u_i^{n+1} - u_i^n}{\Delta t} = -\frac{\lambda}{H_i}\left(u_i^{n+1} - u_o^n\right) \qquad \frac{u_o^{n+1} - u_o^n}{\Delta t} = \frac{\lambda}{H_o}\left(u_i^{n+1} - u_o^n\right) \tag{3}$$

With (3), the velocity differences between the ocean and ice are reduced by a factor of $A_{semi-implicit} = \left(\frac{1 - \lambda\Delta t/H_o}{1 + \lambda\Delta t/H_i}\right)$ with each timestep, which can be stable (damped) for any thickness of the sea ice, provided that the ocean boundary layer is thick enough that $\lambda\Delta t < 2H_o$. If both the ice and ocean velocities are treated implicitly in the stress calculation, the time stepping scheme is unconditionally stable, with an amplification factor of $A_{implicit} = \left(\frac{1}{1 + \lambda\Delta t/H_i + \lambda\Delta t/H_o}\right)$. Unfortunately, calculating the stress implicitly in both the ice and ocean is tough to implement if the ice and ocean are treated as separate components.

Finally, when the ice and ocean are treated as separate components that are stepped forward simultaneously on their own sets of processors (the "concurrent coupling" discussed earlier), the stresses between the ocean and ice must be lagged relative to each other, as in

$$\frac{u_i^{n+1}-u_i^n}{\Delta t}=-\frac{\lambda}{H_i}\left(u_i^{n+1}-u_o^n\right)$$
$$\frac{u_o^{n+1}-u_o^n}{\Delta t}=\frac{\lambda}{H_o}\left(u_i^n-u_o^{n-1}\right) \quad (4)$$

The same stress is applied to the ice and ocean here for conservation, and all of the stresses are calculated implicitly with the ice velocity. The lagging in (4) gives an even more restrictive limit on the stability of the timestep and ocean thickness than (3) by a factor of 2.

2.b Inertial stress instabilities

In a rotating system, there is an additional mode of linear instability when the stress depends on an explicit or lagged ocean velocity. This instability can be illustrated in a single-layer ocean driven by an atmospheric wind stress based on the difference between the atmospheric and ocean speeds. In this case the velocity equation is

$$\frac{\partial u}{\partial t}+ifu=\frac{\lambda}{H}\left(u_{Atm}-u^n\right) \quad (5)$$

where the real part of the velocity, u, is eastward, and the imaginary part is northward. There is a steady Ekman solution to (5); the velocity anomalies, u', from this steady solution evolve as

$$\frac{\partial u'}{\partial t}+ifu'=-\frac{\lambda}{H}u'^n \quad (6)$$

The time integration of the inertial oscillations is internal to the ocean model, but if it is assumed to be exact, (6) can be integrated forward in time analytically to give

$$u'\left(t^{n+1}\right)=\left[e^{-if\Delta t}+i\frac{\lambda}{Hf}\left(1-e^{-if\Delta t}\right)\right]u'\left(t^n\right)=Au'\left(t^n\right) \quad (7)$$

The stability of the explicit timestepping in (7) is given by the magnitude of A:

$$\|A\|^2=1-2\frac{\lambda}{Hf}\sin(f\Delta t)+2\left(\frac{\lambda}{Hf}\right)^2\left(1-\cos(f\Delta t)\right) \quad (8)$$

This scheme is unstable (i.e., $\|A\|^2>1$) when

$$\sin\left(|f|\Delta t\right)<\frac{\lambda}{H|f|}\left[1-\cos(f\Delta t)\right]$$

or

$$\Delta t>\sim\tfrac{1}{2}\pi/|f|$$

in the limit of a thick ocean. In a global ocean model with explicit stresses and sequential coupling, the largest stable coupled time step is slightly less than 3 hours, although the exact limit also depends on how inertial oscillations are stepped within the ocean model. If the ice and ocean (or atmosphere and ocean) are stepped concurrently on different sets of processors, the stresses are lagged by an additional timestep, so that

$$\frac{\partial u'}{\partial t}+ifu'=-\frac{\lambda}{H}u'^{n-1} \quad (9)$$

The stability in this case involves two roots and a quadratic equation, but it is still analytically tractable in the limit of weak stresses (i.e. $\lambda/H|f|<<1$), for which case the physical root gives

$$\|A_{Conc.}\|^2\approx 1-2\frac{\lambda}{Hf}\left[2\cos(f\Delta t)-1\right]\sin(f\Delta t)+O\left(\left(\lambda/_{Hf}\right)^2\right) \quad (10)$$

In the limit of weak stress, this first goes unstable when $\cos(f\Delta t)\approx\frac{1}{2}$, or $\Delta t\approx\pi/3|f|$, or a little less than 2 hours for a global model. Physically what is happening with this instability is that by the time the stress is applied, the velocity has rotated enough that, averaged through the time interval, it has reversed direction and the stress acts to amplify the velocity anomalies instead of damping them. The instability takes the form of spontaneously growing inertial oscillations.

Technically, this instability can apply whenever the stress at the ocean surface depends on a lagged ocean velocity. However, the atmospheric contributions to the stresses vary sufficiently quickly that they often do not reinforce the same phase of the inertial oscillations. By contrast, sea-ice packs do exhibit inertial oscillations, and this instability is usually manifest as growing inertial oscillations in ice-covered regions in both sea-ice and ocean. This instability can be suppressed if the ocean model uses a semi-implicit or predictor-corrector time-stepping scheme that damps inertial oscillations sufficiently strongly.

2.c Coupled gravity wave instabilities and "surfing icebergs"

In a mixture of ice and water, both ice and water participate in external gravity waves. With physically realistic sea-ice models that exchange water with the ocean, ice-ocean models should represent this role of ice in the gravity wave dynamics. However, the traditional practice of treating sea ice and the ocean as dynamically distinct components has often led to the neglect of the role of ice in gravity waves, leading to a coupled instability for sufficiently thick ice that is not locked up in a pack.

In cases where the ice is in free-drift, but exerts hydrostatic pressure on the ocean, the coupled gravity waves in the ice-ocean system can be represented using a two-layered shallow water system. This is exactly the same idealised model as is commonly used to study the stability of various approaches to the split time-stepping of the external and internal gravity waves in the ocean (e.g., Higdon and Bennett, 1995; Hallberg, 1997; Shchepetkin and McWilliams, 2003; Hallberg and Adcroft, 2009). The stability of this system can be ascertained examining the magnitude of the roots of a quartic equation, and the procedure follows the previous studies closely. Sequentially coupled gravity waves are marginally stable (depending on how the gravity wave terms are stepped in the sea-ice and ocean models). In contrast, without the damping effects of an ice-pack, concurrent coupling leads to growing instabilities with a wavelength (L) dependent growth over each timestep by a factor of

$$\|A_{Conc.}\| \approx 1+\frac{\rho_{Ice}}{\rho_{Ocn}}\left(1-\cos\left(\frac{2\pi}{L}\sqrt{gh_{Ice}}\Delta T\right)\right)\left(1-\cos\left(\frac{2\pi}{L}\sqrt{gh_{Ocn}}\Delta T\right)\right)$$
$$\approx 1+\frac{2\pi^2}{L^2}\frac{\rho_{Ice}}{\rho_{Ocn}}gh_{Ice}\Delta T^2 \text{ for } \frac{2\pi}{L}\sqrt{gh_{Ice}}\Delta T << 1 \text{ and } \cos\left(\frac{2\pi}{L}\sqrt{gh_{Ocn}}\Delta T\right)=-1 \quad (11)$$

where ΔT is the coupling timestep, g is the gravitational acceleration, and ρ_{Ice}, ρ_{Ocn}, h_{Ice} and h_{Ocn} are the densities and thicknesses of the ice and ocean, respectively. These instabilities grow fastest at the shortest wavelengths and for thicker ice. Because of the strong wavelength dependence, these instabilities are more readily manifest in finer spatial resolution ice-ocean models. Applying time filtering of the gravity wave coupling terms can reduce the growth rate of this instability, but does not eliminate it altogether.

The horizontal stresses in a solid sea-ice pack are very effective at damping convergent motions, and can overwhelm the growth of this coupled gravity wave instability. However, these stresses drop off rapidly when there are gaps in the sea-ice. In models with multiple sea-ice thickness categories, it is not uncommon in the springtime or summer for the thinner categories to melt away while the thicker categories persist. In our experience at GFDL, the most prominent coupled gravity wave instabilities occur in just this situation of partial coverage by thick ice that has become unlocked from the pack, especially in channels that inhibit the radiation away of the gravity wave energy.

Icebergs provide an especially vivid illustration of this coupled gravity instability. Martin and Adcroft (2010) added point-icebergs to GFDL's coupled climate models. When their weight was felt in the ocean's pressure field and sequential coupled timesteping was used, the icebergs would experience a pressure-driven acceleration based on their location from the previous timestep. Occasionally, when the conditions were just right, the icebergs would spontaneously "surf" across the ocean at a gravity wave speed, accelerated by the pressure "echo" from their previous location. This behaviour can be avoided by not applying the hydrostatic pressure of the icebergs to the ocean, and instead "levitating" the icebergs just above the ocean's surface. However, the physically correct solution to this instability would be to embed the sea-ice and icebergs in the ocean model, so that they can participate in the gravity wave dynamics in a coherent and consistent manner.

When virtual salt fluxes are used, the ice has no mass for the purpose of calculating hydrostatic pressure and the gravity waves occur in the ocean alone, and the coupled gravity wave instability cannot occur. Similarly, if ice does exchange water with the ocean, but the pressure from the weight of the ice is not felt by the ocean, the coupled gravity wave instability is precluded. However, both approaches have significant drawbacks. Virtual salt fluxes either lead to incorrect density changes due to brine rejection or they lead to nonconservation of salt when freezing and melting occur at different ocean salinities. Omitting the pressure of ice on the ocean leads to spurious vortex stretching (and the resultant circulations) where melting or freezing occurs. Both preclude ice grounding and can require artificial limits to inhibit the excessive growth of sea-ice in isolated embayments (eventually to thousands of meters thickness in some cases in GFDL models).

3. Discussion and summary

Sea ice and the ocean are tightly dynamically coupled. After all, ice is just frozen water, and this tight coupling can easily be illustrated by gently sloshing a glass of ice-water back and forth. Modelling the sea-ice and ocean as though they are dynamically separate can lead to several types of numerical instabilities when the two components are coupled. This note has briefly described two such instabilities and their symptoms, although in an actual coupled climate model, the numerical coupled instabilities that arise are likely to be a combination of the idealised archetypes described here.

The coupled inertial-stress instability is reasonably well known, as it can occur even when the ice-ocean fluxes are treated with virtual salt fluxes instead of as water fluxes, and the sea-ice exerts no pressure on the ocean. To avoid the inertial-stress instability with dynamically separate ice and ocean models, the coupling must occur with a timestep that is shorter than a fraction of the inertial period, or the inertial oscillations in the ice and ocean must be artificially damped. Alternatively, a longer timestep can be used for coupling with the atmosphere, provided that the stresses between the sea-ice and ocean and their inertial oscillations are integrated jointly. Because the fundamental timescale for this instability is the inertial period, it is not likely to become more problematic as the horizontal resolution of the ocean increases, although it can be accentuated by refining the vertical resolution of the ocean model.

The coupled gravity-wave instability can arise when sea-ice is treated as having mass and exerts pressure on the ocean, but the gravity wave terms in the ice and ocean models are time-stepped separately. It is strongly dependent on the model's horizontal resolution, in the form of the period of a grid-scale gravity wave based on the ice thickness compared with the coupled time step, and is therefore most likely to arise in high-resolution coupled models. It is most likely to arise in situations where there are patches of thick ice (or icebergs) that are not locked into a solid pack. This instability can be avoided by "levitating" sea-ice above the ocean, but doing so introduces artificial circulations and limits the scientific questions the model can address.

There are additional thermodynamic instabilities that can arise from treating the ice and ocean as separate, lagged components. The heat equation can exhibit instabilities that are similar to the non-rotating momentum equation. There may be appropriate mathematical tricks that can control these instabilities. Ultimately, though, the most reliable solution for avoiding the whole range of coupling instabilities is probably to design coupled ice / ocean models that more closely emulate the true dynamics of the physical system, and use numerical approaches that tightly integrate the high-frequency dynamics of the ocean and sea-ice. This more comprehensive and robust approach will take significant efforts, but several of the world's leading high-resolution sea-ice/ocean model development groups who were represented at the 2014 WGOMD-sponsored meeting in Kiel are moving in this direction.

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Principles and advances in subgrid modelling for eddy-rich simulations

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1. Introduction

One of the challenges of global and large-scale ocean modelling is the comparatively small scale of the turbulent features in comparison to the dimension of the ocean basins. The largest of these features, mesoscale eddies, usually form from hydrodynamic instabilities near the first baroclinic Rossby deformation scale. Recent evidence from coupled models and satellites shows many significant effects, such as eddy effects on air-sea coupling (Bryan et al., 2010; Frenger et al., 2013), near-inertial coupling (e.g., Jochum et al., 2013), and low-frequency eddy-driven decadal variability (e.g., Berloff et al., 2007). Parameterisations of the mechanisms behind these processes are likely to be complex or are presently unknown, and it is this sort of dynamics that drive interest in high-resolution climate modelling (e.g., McClean et al., 2011; Delworth et al., 2012). These coupled models come at a much greater cost, and there is more difficultly in diagnosing causal linkages between model physics and model biases, than even high-resolution ocean-only models. Thus, it is paramount to reduce the amount of tuning and number of unjustified parameters in these eddying ocean models.

Figure 1 shows the progression in resolution of the ocean model component of coupled Earth System Models reported in the IPCC versus the range of Rossby scales across the globe (Chelton et al., 1998). Computational capability increases exponentially: Moore's (1965) scaling predicts a resolution doubling every 6 years. However, the highest-resolution coupled Atmosphere-Ocean Models (AOMs) and Earth System Models (ESMs) refine slower due to increasing model complexity and numerical challenges (6.9 and 10.2 years to double resolution, respectively). Much higher resolution models exist for limited-duration basin or coastal applications, but they generally do not include the coupled dynamics mentioned and cannot be grouped with these other model types. The extrapolation in Figure 1 predicts decades to a century before climate models routinely fully-resolve mesoscale and submesoscale turbulence, with the majority of ESMs at only eddy-permitting resolution at most latitudes.

When models are eddy-permitting or even eddy-rich, it is important to tailor optimal subgridscale (SGS) closures to the physical and numerical setting. The tradition in coarse-resolution models is to fully parameterise all eddy effects, and in fine resolution models to turn off all physical parameterisations of eddy effects and minimise numerical closures (e.g., Delworth et al., 2012). Hallberg (2013) suggests that the scale at which this transition occurs can be selected dynamically and automatically during the course of a simulation, so that changes in stratification and latitude are handled smoothly in a physically-meaningful manner. Similarly, Fox-Kemper et al. (2011) feature a gridscale-dependent amplification factor for a submesoscale physics parameterisation, which extinguishes the parameterisation if mixed layer depth and stratification change so as to make the parameterised features resolved.

The key distinction between coarse- and fine-resolution in terms of subgridscale closure is whether there is a scale separation between the gridscale and the largest eddy

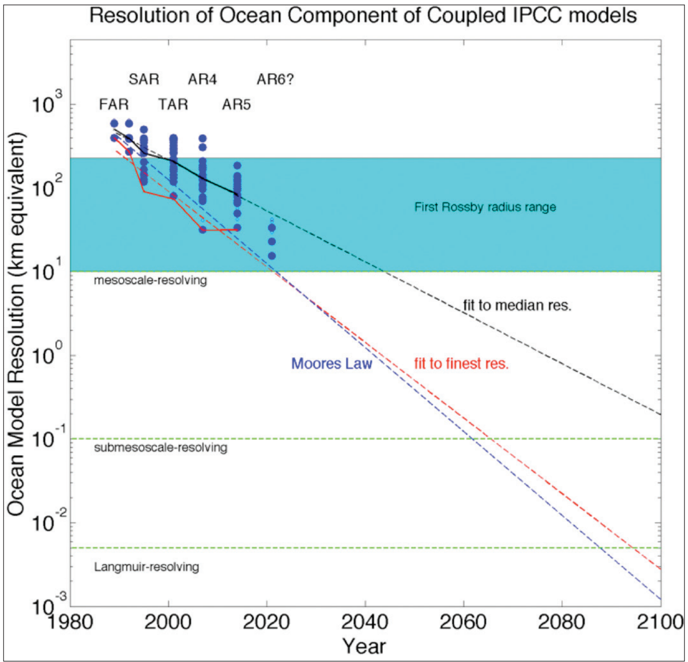

Figure 1: Estimate of the effective nominal horizontal resolution of ocean model components for primary baseline and climate change scenarios as reported in the IPCC reports by year of publication. Exponential fits to the median, finest-resolution, and a Moore's (1965) law estimate are shown; the doubling of resolution occurs every 10.2, 6.9, and 6 years, respectively. Standards for "resolving" turbulence types and the first baroclinic deformation radius range (Chelton et al. 1998) are also indicated. Sixth assessment report (AR6) high-resolution estimates based on present prototypes are indicated, but not fitted.

scale. If there is a separation, then closures depend only parametrically on flow variables and are independent of fine adjustments to the gridscale, and thus the closures are called parameterisations. If there is no scale separation between the resolved flow and the largest eddy scale, then the model falls into the category of Large Eddy Simulations (LES). Ocean models where the gridscale lies between the largest mesoscale eddy scale and the smallest can be called Mesoscale Ocean Large Eddy Simulations (MOLES). For example, complete resolution of baroclinic instability on all vertical modes (mesoscale and submesoscale) remains distant in global models (based on Figure 1), and so part of the eddy effects should still be included in a MOLES closure. In MOLES, the closures, or subgridscale (SGS) models, should depend on fine adjustments of the gridscale versus physically-important scales (be "scale-aware") and may take advantage of sampling the statistics of the resolved eddies to inform the SGS model (be "flow-aware"). The combination of the two guiding principles of SGS models avoids "double-counting" the largest eddies in both the resolved flow and the parameterisation. When a scale separation exists in some regions and not in others, a hybrid of the parameterisation extinguishing approach exemplified by Hallberg (2013) can be used to transition to MOLES SGS models.

Scale-aware and flow-aware SGS modelling began with Smagorinsky's (1963) viscosity scaling for three-dimensional (3d) turbulence in an inertial range. The upper panel of Figure 2 schematises his approach. Energy injection is expected to occur on large scales and then cascade through an inertial range to smaller scales. In Kolmogorov's (1941) idealization, energy flux (ε) through the scales of the inertial range (k_I) is constant and independent of the viscosity, which becomes important only at a smaller viscous scale (k_D). If grid resolution is insufficient to resolve the small scale dissipation processes, a larger value of viscosity is selected that depends upon the gridscale and the resolved flow of energy toward small scales. While there are sound objections to Kolmogorov's idealization, it is a useful framework for estimating the scaling laws needed for SGS models. By ensuring consistency, Smagorinsky devised a robust, scale-aware formulation of viscosity that has been used extensively. Early viscosity scalings for eddy-rich modelling followed Smagorinsky in selecting flow dependence on the deformation rate (Griffies and Hallberg, 2000; Willebrand et al., 2001), and depend on a power of the gridscale and a tunable coefficient to provide harmonic or biharmonic viscosities at all latitudes.

However, the large-scale ocean is too shallow and stratified to have 3d turbulence. Two-dimensional (2d) and quasigeostrophic (QG) turbulence, more applicable to the ocean dynamics on scales near the deformation radius, feature two conserved quantities - energy and enstrophy in 2d, or energy and potential enstrophy in QG - which lead to two distinct inertial ranges (Kraichnan, 1967; Charney, 1971; Figure 2 lower panel). The flow of energy to the largest scales depends on an energy flux, while the flow toward the smallest scales depends instead on a (potential) enstrophy flux. Energy and enstrophy injection into this cascade is considered to take place at intermediate scales where hydrodynamic instabilities are most active - i.e., near the appropriate deformation radius in mesoscale and submesoscale eddy-rich models. Leith (1996) and Fox-Kemper and Menemenlis (2008) suggest and implement, respectively, a SGS closure for eddy-rich ocean models based on the enstrophy cascade in 2d turbulence. It is important to note differences from the Smagorinsky-school closures: the flow-awareness differs and the scale-awareness differs. The Leith viscosity is proportional to the vorticity gradient instead of the deformation rate, so it is one differential order higher. The Leith harmonic viscosity depends on the third power of gridscale instead of the first (Willebrand et al., 2001) or second (Smagorinsky, 1963); and similarly is one to two powers of gridscale larger for biharmonic viscosity (Fox-Kemper and Menemenlis, 2008).

The differences in flow-awareness and scale-awareness make the Leith scaling more scale selective than the

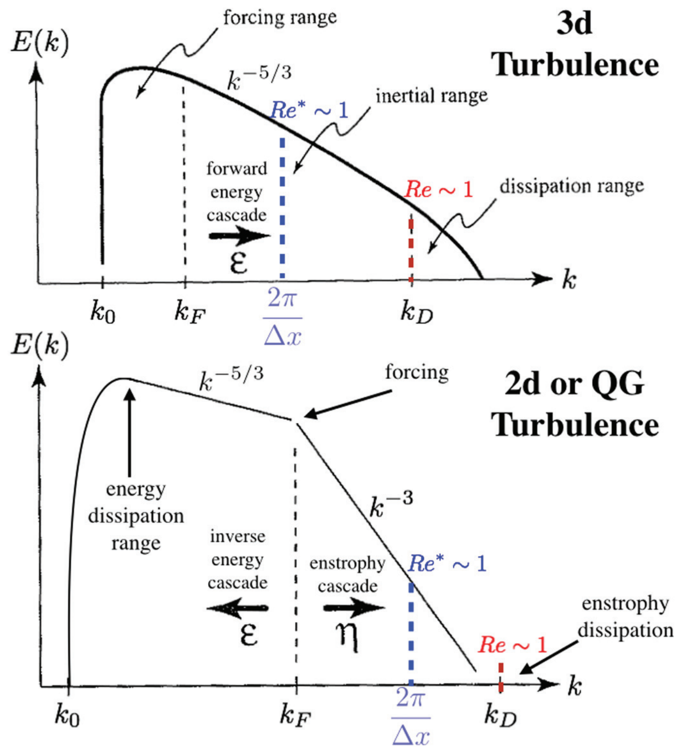

Figure 2: Comparison of the inertial range theories of Kolmogorov, Kraichnan, and Charney for inertial ranges in 3d (upper), 2d, and QG (lower) turbulence.

Smagorinsky scaling. Additional scale-selectivity was a goal of both Smagorinsky-school studies, which provoked the choice of biharmonic operators. Testing at variable resolutions in 2d turbulence simulations (Pietarila Graham and Ringler, 2013) and 3d Boussinesq models in the QG regime (Bachman and Fox-Kemper, in prep) show that the degree of scale selectivity in the Leith scheme (harmonic and biharmonic) is accurate across a broad range of MOLES resolutions. For example, recent global simulations using the MITgcm at 1/54th of a degree (Menemenlis, pers. comm.) used the same biharmonic Leith scheme described for a 1/8th degree run (Fox-Kemper and Menemenlis, 2008) without retuning of any viscosity or diffusivity parameters.

It is difficult to observe ocean eddy statistics, especially covariances such as eddy fluxes, directly (e.g., Flierl and McWilliams, 1977, find that long timeseries are needed). Thus, while it would be ideal to base closures directly on observations, it is unlikely to be possible. Indirect effects of eddies and other results of eddy fluxes (boundary current separation location, sea surface height variance, meridional heat transport, mooring velocity statistics, etc.) are presently the best way to check high-resolution models for consistency.

2. Choosing a subgridscale (SGS) Model for Mesoscale Ocean Large Eddy Simulations (MOLES)

In building or selecting SGS closures for high-resolution models, there are a number of considerations. Many difficulties can be avoided, and some choices bring many benefits.

A brief list of avoidable issues follows. Many SGS closures seek to avoid competition for energy sources and double-counting of eddy effects between the eddies that are resolved and handled by the SGS closure. Any parameters that arise in the SGS scheme should be dimensionless, so that as gridscale or physical setting changes the models can be applied without retuning. The dimensionless parameters, and the theoretical and physical motivation for the closure, should not be extended for use when the gridscale dynamics do not resemble those of the motivating principles. Nor should the parameters be tweaked to remove discrepancies that stem from new physical mechanisms. In oceanography, where the very weak abyssal turbulence preserves the distinctive mix of tracers in each for centuries, spurious mixing by SGS closures and numerical errors is a worry (Veronis, 1975; Griffies et al., 2000; Ilicak et al. 2012).

Careful selection of SGS closures heeds the following list of good practices. Clear theoretical and physical motivation for the closure should exist, and a clear statement of these principles should be available. With such a basis, ready improvement and evaluation of the model is straightforward. Where known asymptotic limits exist, a connection should be made, with the approach to this limit following observed scaling relationships. The SGS scheme should have robust numerical performance, so that linear and nonlinear instabilities are sufficiently damped (preferably in approximation to how they are damped in the real world). Closures can draw from past experience, but continual evaluation and comparisons versus contrasting or new ideas in controlled tests (e.g., idealised experiments where a high-resolution "truth" run is possible) reveals many deficiencies that are easily addressed. These tests are only meaningful if it is possible for a closure to fail (Popper, 1998). Convergence of key metrics with increasing resolution is ideal as a meaningful test. Finally, as already emphasised, in MOLES SGS models, scale- and flow-awareness are advantageous in the ocean, where heterogeneity of flow and gridscale are common.

MOLES differ substantially from traditional LES in the dynamical regime present at the gridscale. LES methods usually assume isotropic, homogeneous turbulence in models with unity aspect ratio. MOLES feature 2d or quasi-2d turbulence in a strongly anisotropic gridscale, which may or may not match the anisotropy of the flow features. For example, does the Burger number of the grid respect the f/N value everywhere in the flow? Furthermore, in stratified, rotating flow viscosity is not enough: eddy effects have been shown to resemble viscosity (Smagorinsky, 1963), diffusivity (Redi, 1982), advection (Gent & McWilliams, 1990), and dispersion (Nadiga & Bouchet, 2011), among other possibilities.

3. Recent Progress on SGS Models

Bachman and Fox-Kemper (in prep) improve on the ideas in Fox-Kemper and Menemenlis (2008) in a practical SGS model combining the best aspects of the Leith 2d scheme and the Gent-McWilliams and Redi parameterisations. This adiabatic model very naturally converts between 2d and QG scalings as the flow evolves, and specifies matched eddy-induced viscosity, diffusivity, and advection. This scale- and flow-awareness based on gridscale Rossby, Richardson, and Burger numbers tends toward a 2d enstrophy cascade or a QG potential enstrophy cascade as the regime transitions from 2d to QG physics.

Recent work on parameterisations (Smith and Marshall, 2009; Ferrari and Nikurashin, 2010; Abernathey et al., 2010, 2013; Eden, 2011; Fox-Kemper et al., 2013; Reckinger et al., in prep) emphasises enhancement or suppression of eddy mixing by flow and stratification effects, such as shear dispersion and critical layers. An exciting aspect of MOLES is that such effects may not need to be explicitly addressed - they will naturally arise from the resolved eddy interactions and therefore be carried into the SGS model. This conjecture is testable in future diagnoses of MOLES. Of course, the cost of global MOLES means that new parameterisations of these processes for coarse-resolution models will be used in most models for decades at least.

Proposing stochastic closures has also been active recently (Grooms and Majda, 2013; Porta Mana and Zanna, 2014; Jansen and Held, 2014). These models are naturally combined with MOLES closures, as has long been the case in LES stochastic backscatter schemes (e.g., Leith, 1990).

Other approaches to MOLES SGS modelling show promise in 2d and QG models. Chen et al. (2011) update the Sadourny (1975) anticipated potential vorticity method to scale-awareness. Nadiga and Bouchet (2011) show equivalence between LES methods in the 2d turbulence or MOLES regime. San and Staples (2011, 2013) adapt image-processing techniques as SGS models. Pietarila Graham and Ringler (2013) perform an excellent comparison of many of these models in an idealised 2d turbulence setting. Bachman and Fox-Kemper (in prep.) compare many of the common MOLES closures in 3d Boussinesq models in the QG regime.

4. Conclusions

Computational and theoretical advances have allowed the recent wave of realistic, high-resolution ocean modelling. Operational oceanographic models and high-resolution climate models are increasingly potentially important applications for society, but they are only as reliable and robust as the closures and numerics that they rely upon.

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Large-scale ocean modelling on unstructured meshes

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1. Introduction

Ocean circulation modelling helps to gain understanding of the ocean's role in the changing climate. Most modelling studies are performed with traditional ocean circulation models formulated on structured meshes, which warrants numerical efficiency. The complexity of basin geometry and the need to incorporate eddy motions on various scales, or the need to include physical processes that involve small scales (such as overflows or coastal upwelling zones) serve as motivation to studies at increasingly refined meshes. Such high-resolution simulations require immense computational resources and storage. Rather commonly, the focus of simulations is on a particular area, and in such a case resolution in structured mesh models can be refined locally through nesting in order to spare resources (see, e. g. Debreu and Blayo, 2008). Nesting is available, for example, with NEMO and ROMS, and many models use it routinely in a build-in form.

A novel approach is offered by models formulated on unstructured meshes. While such models are common in coastal studies where geometrical complexity of coastlines and the need to resolve estuaries leave hardly any other choice, these models are only starting to be applied to simulate large-scale circulation. Unstructured meshes provide multi-resolution functionality and can accommodate multiple areas of arbitrary form with refined resolution, as dictated by practical tasks. Additionally, unstructured meshes can be aligned with coastlines or the continental break. This approach is offered by Finite-Element Sea-ice Ocean circulation Model (FESOM) (Wang et al., 2008, 2014) and MPAS-Ocean (Ringler 2013), and other developments, such ICON (see ICON website) or the new core at AWI (Danilov 2012).

Compared to traditional nesting, the main advantage of using unstructured meshes is their unlimited refinement factor, lack of spurious reflections because of smooth transitions and consistent solution, and the ease of using: refinement is only the matter of mesh design. Their main drawback is their larger computational load per degree of freedom, and the fact that their time step is defined by the smallest element. Although unstructured-mesh models remain slower than their structured-mesh counterparts, at least one finite-volume implementation lags only by a factor about 3 (see Ringler et al., 2013), which is fully acceptable if one accounts for the relative immaturity of these models and the possibility to invest their degrees of freedom where needed. Time step limitation is not an issue in applications where the refined region contains the majority of the mesh nodes (see, e.g., Ringler et al., 2013).

It is therefore believed that models capable of working on unstructured meshes may be convenient and more optimal for certain tasks of large-scale ocean modelling. There is hope that they will contribute, for example, to reaching more realism with respect to simulation of overflows or dense water production in setups intended for climate studies. While unstructured grid models may not fully replace models formulated on structured meshes, the ability to invest the resources where necessary warrants their place in ocean modelling. It is also believed that with advancement of computer technology and further optimization the difference in computational efficiency between structured- and unstructured-mesh models will further decrease.

Finite-Element Sea-ice Ocean circulation Model (FESOM) is the first model designed to work on unstructured meshes on large-scales. MPAS-Ocean is more similar to traditional global ocean models with its finite-volume discretisation and accompanying computational efficiency. The rest of this note deals with examples illustrating benefits of multi-resolution, preceded by a brief resort to mathematics.

2. Mathematical challenges

The development of models capable of working on unstructured meshes faces challenges on both numerical and computer science sides, which are subjects of ongoing research. While the latter are largely related to indirect memory addressing, the former have a geometrical origin and are explained below.

Unstructured meshes may be composed of various polygons. Most popular are triangular meshes due to their flexibility in varying resolution or fitting the mesh to the details of domain geometry (FESOM). By connecting circumcentres or centroids of triangles one can obtain dual quasi-hexagonal meshes (a special variant is used by MPAS-Ocean). Simplest co-located discretisations like that used in FESOM need to be stabilised against pressure modes. Staggered discretisations on triangular or dual meshes usually support families of spurious numerical modes. These modes arise because of disparity between the numbers of degrees of freedom used to represent velocity and pressure. They have a geometrical origin: on triangular meshes the ratio of vertices to cells to edges is approximately 1:2:3. So if pressure is at vertices and velocity at cells, there are twice more velocities than needed (FV cell vertex discretisation in Ringler and Randall (2002) and Danilov 2012). If pressure is at cells (triangles) and normal velocities at edges, there are too many pressure degrees of freedom (ICON). On dual quasi-hexagonal meshes (MPAS-Ocean) with pressure at centres and normal velocities at edges, there are 1.5 times more velocities than needed. For mathematical details see Danilov (2013) and references therein. It is important to note that the various spurious modes resulting from the mismatch in degrees of freedom are not equally problematic; some are challenging (e.g. ICON) while others are benign (e.g. MPAS-Ocean). This is the reason why the utility of a particular discretisation is determined not solely by the ability to accurately resolve physical perturbations (waves) but also severity of the spurious modes. Eliminating them can be more difficult than on regular quadrilateral meshes, and sometimes requires special algorithmic solutions.

Ongoing research seeks new discretisations with less numerical artefacts.

3. Practical examples

Multi-resolution (unstructured-mesh) models are gradually becoming a reality. FESOM participates in the CORE-II intercomparison project, demonstrating that on meshes typical for current climate models it manages to maintain the large-scale ocean circulation with a degree of realism typical for structured-mesh models despite its very different numerical core (see Ocean Modelling, virtual special issue on CORE-II). MPAS explores the impact of refinement on eddy statistics, and demonstrates the feasibility of global eddy-resolving simulations on unstructured meshes (Ringler et al., 2013). Particular examples below illustrate the potential of multi-resolution models.

3.1 Freshwater transport through the Canadian Arctic Archipelago

Figure 1 shows the mesh configured to study the Arctic Ocean freshwater circulation (Wekerle et al., 2013) with FESOM. Arctic Ocean presents a large freshwater reservoir owing to high net precipitation and runoff of numerous rivers. The freshwater exchange of the Arctic Ocean with the North Atlantic happens partly through the Fram Strait and partly through the straits of the Canadian Arctic Archipelago (CAA). The excessive freshwater storage in the Arctic is due to precipitation and runoff of numerous rivers, and the outflow is largely driven by the pressure difference between the Arctic and the North Atlantic. Since the path of freshwater lies in the vicinity of main convection sites, the redistribution of freshwater between the Fram Strait and the CAA has global implications, in particular on the strength of the meridional overturning circulation (MOC).

The transport of freshwater through the CAA is mostly associated with volume transport. Most of the current climate models simulate it by artificially increasing the width of major channels (Parry Channel and Nares Strait). While this may be sufficient to simulate the mean transport, it is not necessarily so for the transport variability. The study by Wekerle et al., (2013) explores these issues with FESOM by comparing the performance of two global model versions, the control configuration with 24 km resolution in the CAA, which is common for climate models, and the other one with 5 km resolution in the CAA. Both resolve the Arctic Ocean with 24 km, getting coarser in the rest of the ocean except for the vicinity of coastlines. The fine resolution of the CAA (see Figure 1) adds about 50% nodes to the surface mesh and limits the time step to be about 10 min, yet the model is still fast enough to allow for multidecadal simulations.

The comparison between the simulation results and available observational data by Wekerle et al., (2013) demonstrates that the fine resolution run is indeed able to simulate the transport variability in closer agreement with the observational data. Further analysis shows that the freshwater transported through the CAA stays confined to the coast in the Labrador Current, while the increased salinity in the Eastern and Western Greenland Currents in the fine-resolution run leads to an increased mixed layer depth in the Labrador Sea, which, in turn, increases the strength of the MOC. The analysis also shows that the variability in freshwater transport is driven by the large-scale atmospheric pressure system. The global impact of the regional improvement indicates the potential of unstructured meshes in climate simulations.

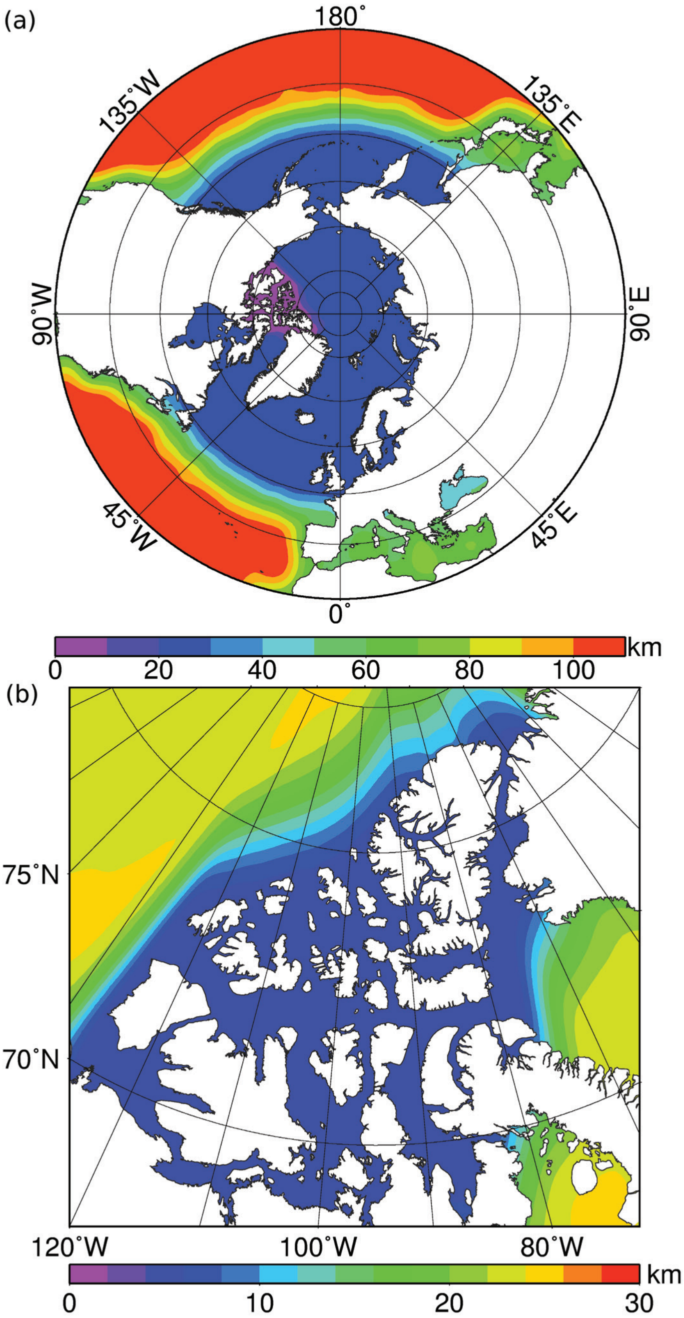

Figure 1: Horizontal resolution of the mesh with the CAA region refined used by Wekerle et al (2013): (a) view in the stereographic projection and (b) zoomed into the CAA region.

3.2. MPAS-Ocean

While mesoscale eddies play an important role in setting the climate of the ocean, globally resolving these eddies with climate system models is still a computationally demanding endeavour. Multi-resolution ocean models allow for the opportunity to resolve mesoscale eddies in regions of interest while parameterising mesoscale eddies elsewhere. Figure 2 (front page, bottom left) shows fluid kinetic energy at a depth of 100 m from a global MPAS-O simulation. The grid resolution is 10 km within a portion of the North Atlantic and transitions to 80 km elsewhere. Mesoscale eddies are well-represented within the 10 km region.

Ringler et al. (2013) compares and contrasts global, quasi-uniform simulations to global, multi-resolution simulations with enhanced resolution in the North Atlantic. The simulations are evaluated based the mean and variance of sea-surface height as compared to observations. The primary finding is that mesoscale eddy dynamics are simulated as well in the North Atlantic with the multi-resolution mesh as with the globally uniform mesh. Furthermore, the computational burden of the multi-resolution simulation was only 15% that of the globally uniform simulation. This finding has been confirmed in more idealised simulations of mesoscale eddy dynamics. Looking forward, opportunities exist to further enhance resolution within the specific regions of North Atlantic to study the dynamics of Gulf Stream separation, controls on the Northwest Corner and, more broadly, the importance of sub-mesoscale dynamics.

4. Conclusions

Multi-resolution models formulated on unstructured meshes have matured over recent years. They promise a convenient and economical approach to research inquiring into regional dynamics in the context of large-scale global ocean circulation. We see their potential in facilitating downscaling or being used to learn about functioning of certain aspects of local dynamics. Ongoing research seeks the ways of further improving their computational efficiency which is, however, already sufficient for many practical applications. The accumulating practical experience makes data processing on unstructured meshes or setting up of the models proper easier, making these models even more appealing.

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MIROC5 with a nested ocean component focused on the western North Pacific

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1. Introduction

In addition to the externally-forced projections, previously assessed by the Intergovernmental Panel on Climate Change (IPCC-AR4), predictions of internal climate variability on decadal timescales were one of the new challenges addressed by the climate modelling community for assessment by the IPCC-AR5. For this purpose, we initialised the ocean state of our climatic model, MIROC, by assimilating observed ocean hydrographic data. We paid a special attention to ocean initialization because the subsurface ocean, which varies slowly and is isolated from the atmosphere, could function as a memory for decadal climate predictions. Retrospective ensemble predictions using MIROC reveal multi-year predictability for some specific aspects of climate variability, such as the Pacific Decadal Oscillations (PDO) and the stepwise climate shift that occurred in the late 1990s over the Pacific (Mochizuki et al., 2010; Chikamoto et al., 2012).

For the decadal climate predictions, we used three versions of MIROC, lower and higher-resolution versions with the same physical package and a lower resolution version but with updated physics. The above-mentioned prediction skill is not assured among all of the versions. In particular, prediction skill of basin-wide climate variability is worse in the higher-resolution version than in lower resolution versions. We can raise some possible reasons for the deteriorated skill, such as, (i) insufficient model parameter tuning to improve model climate and variability, (ii) limited number of ensemble members, (iii) rapid error growth due to stronger nonlinearity. The first two are constrained mainly by the larger computational resources required for the higher-resolution model.

High-resolution modelling does not always guarantee advantages in ensemble climate predictions, but its importance has been increasingly recognised, especially, in terms of active air-sea interactions in the western boundary current (WBC) regions. It is reported that the sea surface temperature (SST) fronts in the WBC regions modulates the activity and tracks of wintertime storms and strength of precipitation. Furthermore, wintertime climatology of large-scale atmospheric circulations can be much influenced by the SST fronts (e.g., Nakamura et al., 2004; Minobe et al., 2008). Oceanic mesoscale features also play important roles in water mass exchanges across the wind-driven gyre boundaries and show significant influences on water-mass formation in the WBC regions (e.g., Ishikawa and Ishizaki 2009).

As described here, high-resolution modelling is required in order to represent realistic physical processes in an eddying ocean regime that affect gyre-scale water-mass formation and atmospheric circulations. However, high-resolution models cannot be easily applied to global-scale climate predictions because of their computational costs. We have developed a two-way nested ocean model, where a fine-resolution regional model is connected to a global ocean general circulation model (OGCM), for better representation of model climate and variability and for better climate predictions by resolving the above-mentioned physical processes in the WBC regions with less computational costs than global high-resolution models. In this short article, we show the ocean current structure, water-mass distribution, and resultant atmospheric responses represented in a global OGCM/climate model interactively connected with a regional ocean model.

2. Model description

A regional ocean model for the western North Pacific (15°N–55°N, 115°E–180°) is embedded in a global OGCM with two-way interaction. The regional model has the horizontal resolution of $0.1° \times 0.1°\cos\theta$, where θ is the latitude, and is formulated on the spherical coordinate system. The global model is formulated on the tripolar coordinate system and has the horizontal resolution of $0.5° \times 0.5°\cos\theta$ in the spherical coordinate portion (south of 63°N). There are 50 vertical levels in common. The COCO ocean model (Hasumi, 2006) is used for both the regional and global model. Details of the two-way nested model are described in Kurogi et al. (2013). This two-way nested ocean model is also incorporated into MIROC5 (Watanabe et al., 2010). Hereafter, the global COCO/MIROC5 with the regional ocean model is called as COCOn/MIROC5n.

In the present study, COCOn and COCO are driven by the CORE normal-year forcing which consists of a repeat annual cycle of atmospheric boundary condition data (Large and Yeager 2009). MIROC5 is integrated over a thousand years under the pre-industrial forcing and the 100-yr-long data are analysed here. MIROC5n is also driven under the pre-industrial forcing. The last 30-yr-long data of the total 70-yr-long integration of MIROC5n are analysed.

3. Reproducibility of North Pacific Intermediate Water

North Pacific Intermediate Water (NPIW), characterised by a salinity minimum centred at the $26.8\sigma\theta$ isopycnal surface, is known to spread widely in the subtropical North Pacific (e.g., Yasuda 1997). Cold and fresh subarctic water is transported from the subarctic to the subtropical region by an eddy-induced lateral transport and the Oyashio intrusion near the east coast of Japan. Then, the water of subarctic origin is mixed with warm and saline subtropical water through isopycnal mixing and is modified to form NPIW. This formation process of NPIW means that subarctic water rich in nutrients and CO2 is transported and isolated into the subsurface and intermediate layers of the subtropical gyre (e.g., Tsunogai et al., 1993).

Figure 1 shows instantaneous sea surface height (SSH) and salinity on the 26.8σθ isopycnal surface at the 15th model year. In COCOn, the Kuroshio south of Japan separates from the Japan coast and continues to the eastward Kuroshio Extension (KE). The KE meanders and accompanies mesoscale eddies spawned from the unstable current, and it bifurcates into the northern and southern branches around 35°N, 155°E. The Oyashio along the east coast of Japan flows southward beyond 40°N and then turns to the northeast. These features of the currents to the east of Japan are consistent with observations. On the other hand, in COCO, the Kuroshio south of Japan overshoots beyond 40°N, meets the Oyashio, and flows eastward with a much weaker front than in COCOn. In general, the current structure is diffuse in COCO.

The southward intrusion of the Oyashio and active eddies in COCOn can supply subarctic water to the east of Japan and feed NPIW. The successful separation of the Kuroshio south of Japan inhibits an excess transport of saline subtropical water to the north of the separation latitude. Consequently, in the subtropical region to the south of 40°N, the salinity minimum of NPIW is realistically maintained in COCOn.
It is notable that the salinity outside the nested domain is influenced by the features inside the nested domain described above, owing to the horizontal advection by the KE beyond the dateline. It is a clear advantage of the two-way interaction.

4. Feedback process among the western boundary currents, wintertime storms and the westerly jet

The better representation of the western boundary currents (WBCs) also brings SST differences between COCOn and COCO and also between MIROC5n and MIROC. A significant

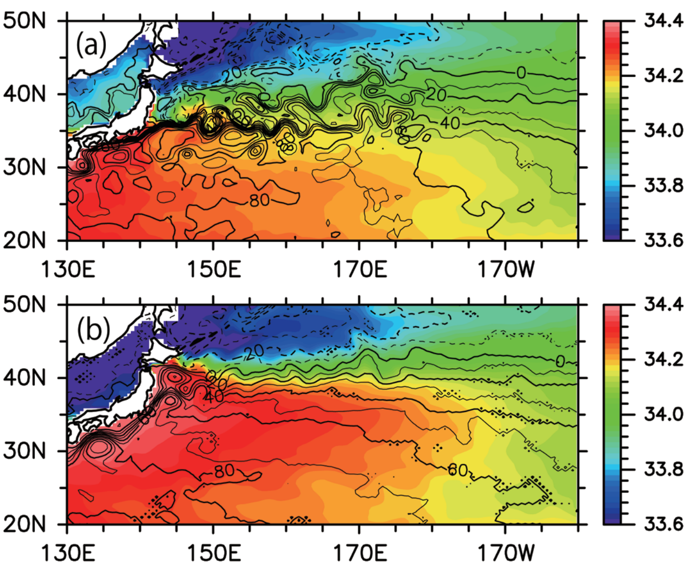

Figure 1. (a) Salinity on the 26.8σθ isopycnal surface (shades; unit is psu) and SSH (contours; cm) in COCOn. (b) Same as (a), but for COCO. These are the snapshots at January 1st of the 15th model year.

SST difference is commonly found in our OGCM and the coupled model experiments to the east of Japan.

The climatological mean wintertime SST difference between MIROC5n and MIROC5 is shown in Figure 2a. The SST difference to the east of Japan is positive, meaning that the SST there is warmer in MIROC5n than in MIROC. This positive anomaly can be explained by the larger northward and eastward heat transports by the Kuroshio south of Japan and the KE in MIROC5n than in MIROC5. This is also the case in the OGCM experiments. The subarctic front around 42°N in MIROC5n, which retreats northward compared to that in MIROC5, is also responsible for the positive anomaly. The meridional heat transport above the 1500 m depth integrated from 140°E and 180° is larger in MIROC5n than in MIROC5 around 40°N by 0.2 PW (Figure 2b), consistent with the positive SST anomaly. On the other hand, the SST adjacent to the Japan coast north of 35°N is lowered. This negative anomaly is due to the southward intrusion of the coastal Oyashio, as discussed in the previous section.

The larger amount of heat transported from the south to mid-latitudes in MIROC5n is released from the ocean to the atmosphere through surface turbulent heat fluxes. Sensible and latent heat fluxes are larger to the east of Japan in MIROC5n than in MIROC5 by over 50 Wm^{-2} and 100 Wm^{-2}, respectively (not shown). The larger heat release in MIROC5n enhances the atmospheric surface baroclinicity and leads to the larger vertical shear of zonal wind through the thermal-wind relation. As a result, the wintertime storm track activity (STA), which is remarkable along the meridional SST gradient associated with the KE, becomes larger in MIROC5n than in MIROC5, and the maximum of STA is shifted northward (Figures 2a and 2c). Note that STA is defined as the meridional eddy heat flux at the 850hPa level and that an 8-day high-pass filter is used to extract transient eddies.

On the other hand, there is a negative SST anomaly along the west coast of North America (Figure 2a). Although the figure is not shown, such a negative anomaly is not found in the OGCM experiments. Associated with the enhanced STA in MIROC5n, the upward eddy heat transport and poleward eddy momentum transport becomes larger in MIROC5n than in MIROC5 in the troposphere over the subarctic North Pacific (not shown). Through the eddy-mean flow interactions, the core speed of the westerly jet around 35°N becomes smaller while the wind speed of the westerly jet around 45°N becomes larger in MIROC5n than in MIROC5. Accordingly, the westerly jet is shifted northward accompanying a positive pressure height anomaly over the subarctic North Pacific (Figure 2d).

These changes of the wind field also manifest as changes of the surface Ekman transport. The northward (southward) surface Ekman transport on the eastern (western) side of the positive height anomaly becomes smaller in MIROC5n than in MIROC5. The negative SST anomaly along the west coast of North America can be explained by the weaker northward Ekman transport of relatively warm subtropical water.

On the other hand, to the west of the height anomaly, the weaker southward Ekman transport of relatively cold subarctic water raises the SST to the east of Japan. That is, the pre-existing positive SST anomaly to the east of Japan, which is originally induced by the larger heat transport in the WBC region, is amplified, suggesting that the surface baroclinity over the subarctic North Pacific could be enhanced again. The above-mentioned processes imply a possible feedback among the WBCs, STA, and the westerly jet.

Influence of the SST fronts in the WBC regions or the Southern Ocean on the wintertime STA has been examined extensively, using atmospheric general circulation models (AGCM). The surface heat flux difference across the fronts works to maintain the surface baroclinicity, and thus the STA is strengthened around the fronts (e.g., Nakamura et al. 2008; Taguchi et al. 2009). Regarding eddy-mean flow interactions between the wintertime storms and the westerly jet, the enhanced wintertime STA is suggested to lead to a poleward shift of the jet core. These are consistent with the present study. While previous AGCM experiments imposed prescribed SST profiles with sharp or smoothed cross-frontal gradients to extract roles of the oceanic fronts, the present climate model experiments are free from the SST constraint, and roles of the fronts have been examined directly. Further, since the modification of the westerly jet can alter the SST distribution, there would be a possible feedback among the WBCs, the STA, and the westerly jet, as presented above.

Latif and Barnett (1994) referred to similar processes within the context of the phase transition of the decadal climate variability over the North Pacific. They pointed out that the weakened westerly jet, which follows positive SST anomalies in the North Pacific mid-latitudes and the resultant enhanced STA, excites cold oceanic Rossby waves. The Rossby waves act to weaken the subtropical gyre as the waves reach the western boundary of the North Pacific, and thus the positive SST anomalies gradually disappear. By this means, the decadal variability is self-sustained. In the present climate model experiments, the northward shift of the westerly jet leads to a weakening of negative wind-stress curl impinging on the subtropical gyre and results in spinning-down of the gyre. Then, the northward heat transport by the Kuroshio and its extension could be reduced for reaching an equilibrium climate state. Detailed processes are being examined.

5. Summary and Discussion

We have developed a global climate model interactively connected with a regional ocean model. Embedding a fine resolution model of a targeted region in a global climate model has an advantage of representing realistic oceanic current systems, mesoscale eddies, and air-sea interactions with less computational costs than global high-resolution models.

As shown in the present study, a climate model is a powerful tool for science-oriented studies. Although results are not shown here, the nested grid OGCM used in this study reproduces the bimodal variation of the Kuroshio path south of Japan with realistic duration and frequency (Kurogi et al., 2013). The reproducibility of the bimodal path may bring predictability of the regional climate around Japan. For example, Nakamura et al., (2012) demonstrated that the Kuroshio path influences the cyclone tracks along Japan and snow fall in Tokyo area.

There are remaining issues to be addressed for long-term integration of climate models with regional nesting. At the lateral boundaries where the regional and global models contact each other, heat and water across the boundaries are not conserved strictly. Also, the area-integrated surface heat and freshwater fluxes passed from the atmosphere over the nested domain are not consistent between the regional and global models. For seasonal-to-interannual climate predictions by means of initialization techniques, which do not always guarantee conservation of heat and water, the above-mentioned inconsistent fluxes may not be of great concern. However, the inconsistency can induce non-negligible model climate drift for centennial projections under the increase of anthropogenic global warming gases. We are trying to solve these problems in order to construct our official model whose results will be contributed for assessment by the next IPCC report.

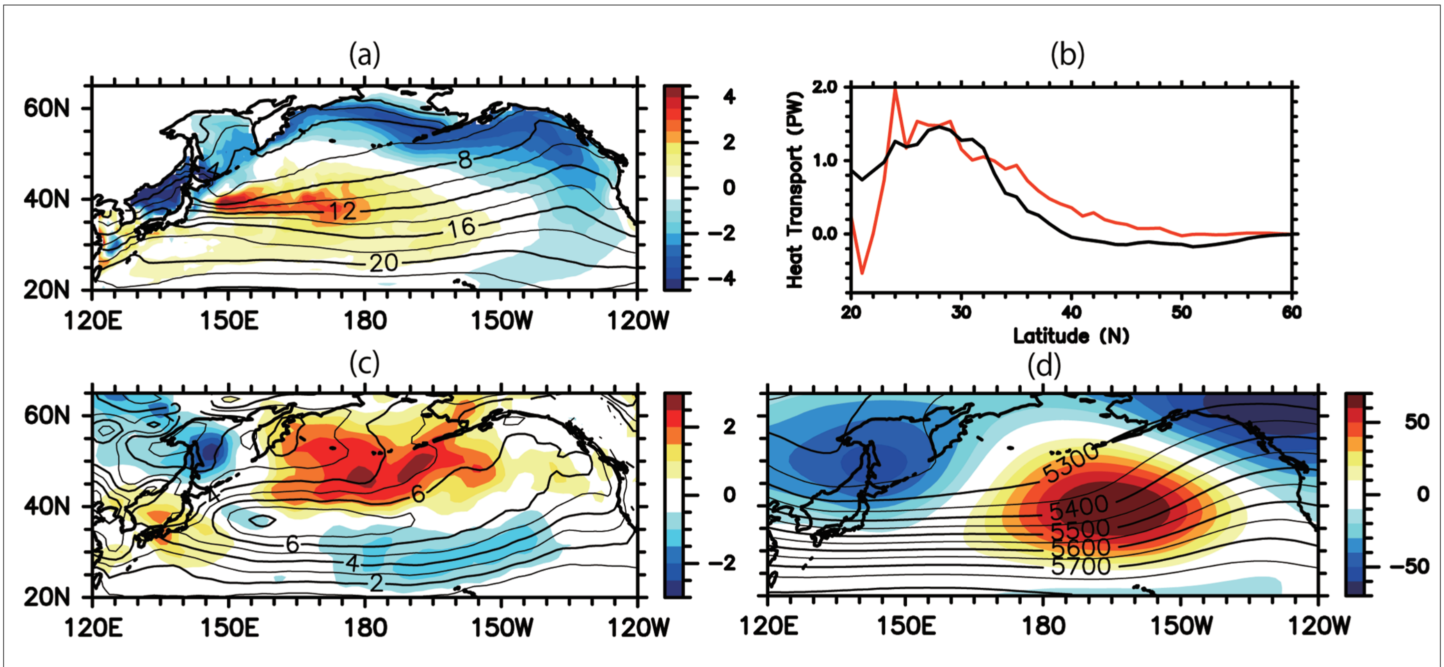

Figure 2. (a) Difference of the wintertime (DJF) SST between MIROC5n and MIROC (shades; ˚C) and the SST in MIROC5 (contours; ˚C). (b) Meridional heat transport integrated from 140°E to 180° for MIROC5n (red line) and MIROC5 (black line). (c) Same as (a), but for the wintertime STA (˚Cms-1). (d) Same as (a), but for the wintertime 500 hPa height (m).

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Nested ocean modelling

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The choice between resolution and coverage is one of the oldest challenges in ocean modelling. Early grand-challenge applications (Semtner and Chervin 1988) addressed it with massive computing power. However, more than 25 years hence, the challenge remains for a large number of applications including both regional and global models. Global extent is necessary for budget analyses of large-scale flows and the simulation of the variety of drivers of the thermohaline circulation. However, for global models, resolving mesoscale eddy variability remains elusive in the high-latitudes due to the reduction of the baroclinic Rossby Radius (Hallberg 2013). Similarly, unresolved coastal processes lead to significant errors in the representation of both eastern and western boundary currents in global ocean models. In lieu of unaffordable global high-resolution, two alternative approaches exist: unstructured meshes (see Danilov et al., 2014, this issue) and nesting. Here, we describe how modern nesting capabilities offer possibilities for targeted local resolution refinement. However, as outlined below, grid refinement is more than a computational compromise.

Early implementations of nested ocean models worked only for a limited time range; typically, within a few simulated weeks numerical noise from the boundaries contaminated the interior solution (Fox and Maskell 1996). The introduction of accurate interpolation/averaging techniques helped overcome some of these issues. A prime example is the Adaptive Grid Refinement in FORTRAN (AGRIF; Debreu et al. 2008), which has been implemented in a variety of models (NEMO, ROMS) and was introduced for a configuration of the Labrador Sea (Chanut et al., 2008). AGRIF works on constant refinement factors of the parent grid and couples it with the child grid at every timestep of the parent grid. Due to numerical stability considerations leading to a reduced timestep (often similar to the grid refinement factor) in the high-resolution child grid, the overhead of running the model on the parent grid is typically less than 10. This is aided by efficient and conservative interpolation techniques (Debreu and Blayo 2008). Another attractive feature of the AGRIF framework is that it permits two-way coupling between the parent and child grids. Supported by an effective on-the-fly interpolation of atmospheric forcing fields, setting up an AGRIF configuration is relatively straightforward. It basically requires the generation of topography and initial conditions and the adaptation of the resolution-dependent parameters and parameterisations. Figure 1 shows an example series of six actual configurations that are currently used at GEOMAR, some as part of PhD projects. Within the DRAKKAR community (DRAKKAR Group, 2014, this issue), a similar NEMO-based series exists (e.g., Jouanno et al. 2008; Talandier et al., 2014).

In the AGRIF framework, the nested configuration is a regional model that explicitly simulates its sidewall boundary conditions. These boundary conditions are performed at all timescales resolved by the timestep (typically minutes to hours) of the parent model. Thus, it is not required to prescribe the outside ocean at any particular temporal resolution (Herzfeld et al., 2011). In the inner nest, Rossby and Kelvin Waves with outside origin are implicitly simulated by propagation from the parent grid. The nested configuration allows for the interior solution to evolve more independently than in a prescribed open boundary condition situation. On the other hand, it also removes some of the control on the configuration that is available through specific changes of the open boundary condition (e.g., Döscher et al., 1994; Gerdes et al., 2001).

The AGRIF configuration allows for a flexible testing of nested configurations. For example, sensitivity as well as an assessment of regional resolution effects can easily be tested by reducing or expanding the area of the nested domain. Biastoch et al. (2008b) compared two nested configurations in order to understand the effect of upstream variability on the exchange between the Indian and Atlantic oceans, the Agulhas leakage. One configuration extended to the Mozambique Channel at mesoscale eddy resolving resolution (thus simulating Mozambique eddies), and the other omitted the Channel and was designed at a coarser base grid resolution. This setup permitted studying the significant differences due to the mesoscale flow patterns south of Africa. Despite the documented influence on the shape and timing of Agulhas rings, it is also worth noting that the difference in the simulated Mozambique Channel flow left the amount of Agulhas leakage unaffected.

A comparison of the two-way nested configuration with the coarser standalone parent model permits an evaluation of the far-field effect of the regional mesoscale dynamics on the large-scale circulation. Biastoch et al. (2008a) isolated the imprint of mesoscale Agulhas dynamics on the Atlantic Meridional Overturning Circulation (AMOC) and the northward signal propagation by coastal shelf waves. This up-scaling approach offers a potential to systematically isolate key regions of mesoscale activity and their downstream influence. As another example, Figure 2 shows the western boundary current regime at 26°N, simulated at 1/4° resolution. Compared to the standalone version (ORCA025), the one with a high-resolution nest north of 30°N (VIKING20, Behrens (2013), see Figure 1) features an improved North Atlantic Deep Water (NADW) component. In particular the southward transport of lower NADW (below 3000 m) has increased (16.1 Sv, compared to 6.4 Sv in ORCA025) and is more comparable to observations (12.6 Sv, Johns et al., 2008). Main reason for this increase is the improved simulation of the Denmark Strait Overflow (Fischer et al., 2014), leading to a downward expansion of the NADW-overturning cell throughout the Atlantic Ocean.

A more flexible framework that does not require the parent and child grids to be collocated, nor the parent and child models to be the same is described in Curchitser et al. (2011) and Curchitser et al. (2014). Such a new framework offers the advantage of optimizing grid and model design in specific regions. The framework has been implemented in the NCAR-CESM using the POP and ROMS global and regional models, respectively. In ocean-only hindcast mode, ROMS, which

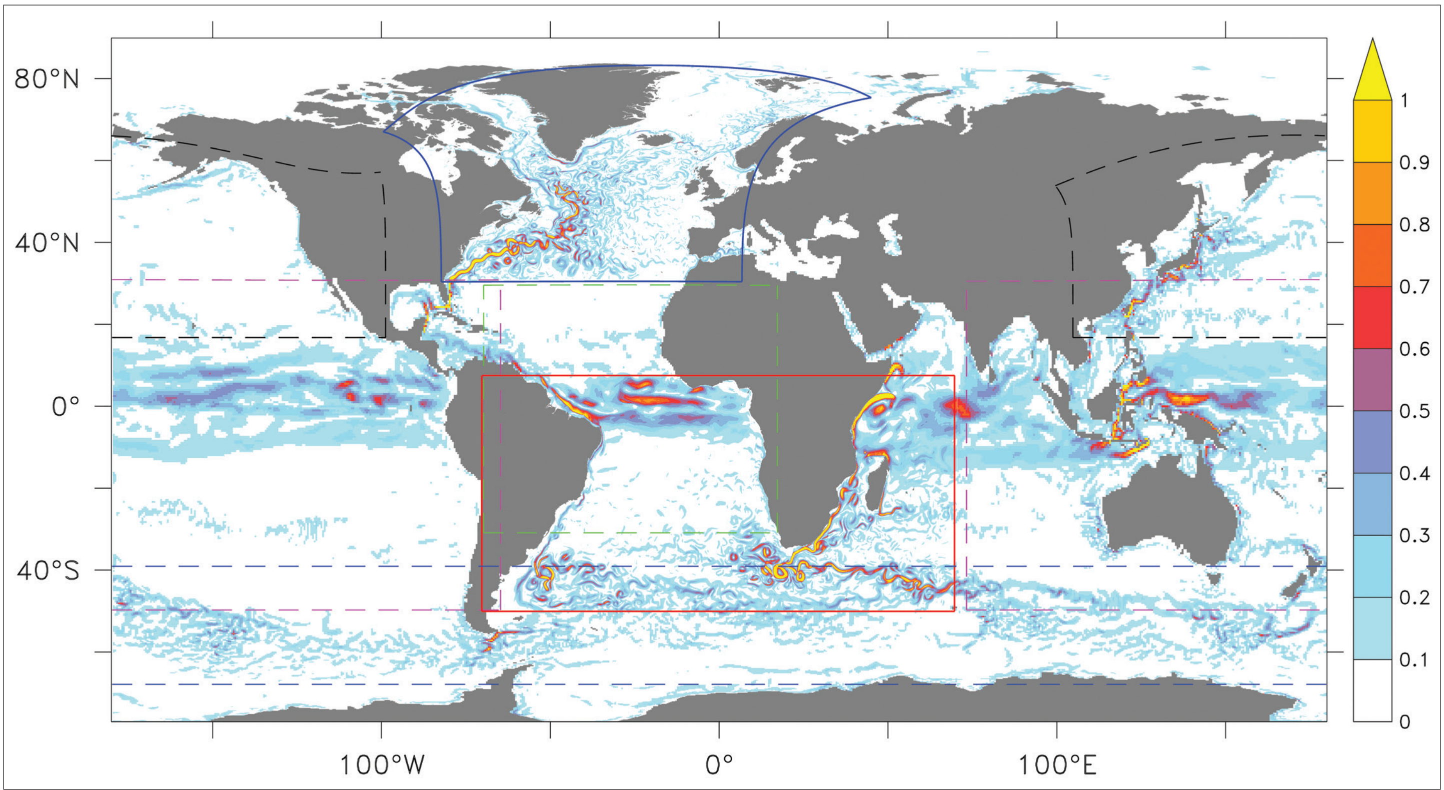

Figure 1: AGRIF configurations at GEOMAR. Shown is near-surface speed (in m/s) in a global ocean/sea-ice configuration at 1/2° resolution (ORCA05) and two high-resolution nests covering the northern North Atlantic (VIKING20 at 1/20°, blue (Behrens 2013)) and the Agulhas / South Atlantic (INALT01 at 1/10°, red (Durgadoo et al., 2013)). Other nested configurations cover the tropical Atlantic (TRATL01, 1/10°, green dashed (Duteil et al. 2014)) and Pacific (TROPAC01, 1/10°, purple dashed (van Sebille et al., 2014)), the North Pacific (NPAC01, 1/10°, black dashed (Behrens et al., 2012)) and the Southern Ocean (ORION12, 1/12°, blue dashed). Due to the tripolar ORCA base grid, the curved northern boundaries of VIKING20 and NPAC01 follow constant grid lines.

was designed as a coastal model has shown significant skill in modelling boundary currents (e.g., Kang and Curchitser 2013). The embedding such a model in a global configuration allows for isolation of specific processes affecting both the local climate representation and potential up-scaling effects. The additional flexibility comes at the expense of boundary condition simplicity. Full radiation boundary conditions are needed to pass information from the parent to the child grid.

Figure 3 shows an example of an embedded regional ocean model in the northwest Atlantic using the CESM multi-scale ocean configuration. The global ocean is POP at 1o, the regional model is ROMS at 7km resolution. The ocean models are two-way coupled. A merged ocean SST is passed to the global atmosphere at each coupling timestep, typically at a daily frequency (R. Dussin, pers. comm.). Though conspicuous features such as the Gulf Stream separation are not as skilful as in the pure ROMS hindcast of Kang and Curchitser (2013), there is an improvement over the coarse resolution model. This configuration is now being used to explore the effects of the Gulf Stream position and correction of some of the global model biases on downstream climate. Though significant effort is being devoted to the development of high-resolution global models, they remain some years away from practical implementations. In the meantime unstructured and nested models afford a way to begin exploring the effects of mesoscale resolution on climate simulations. Nested ocean models may not be a general solution for global ocean and coupled climate configurations. It does, however, allow the identification and isolation of specific key processes and the quantification of their impact on the general large-scale circulation. In particular, with known biases in global models due to features such as overflows, inter-basin exchange and coastal boundary currents, nested models could guide the development and testing of parameterisations. The routine use of nested ocean models also provides an economic and flexible tool to set up and perform state-of-the-art ocean general circulation models for individual purposes, which makes the technique suitable even for a PhD level project.

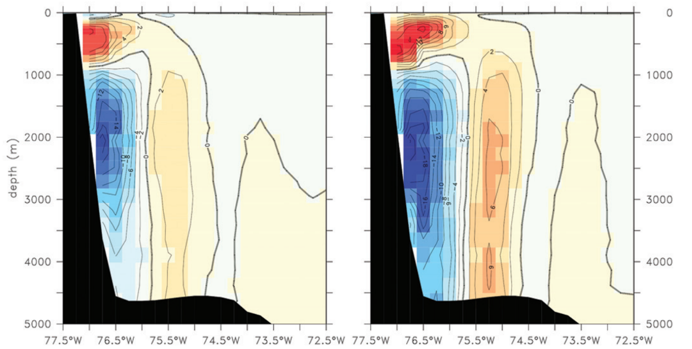

Figure 2: Western boundary current regime at 26°N in a 1/4° global ocean model (ORCA025, left) and in a base model at same resolution (right), but hosting a 1/20° nest north of 30°N (VIKING20). Shown is 10y-mean meridional velocity off Bahamas in cm s[-1].

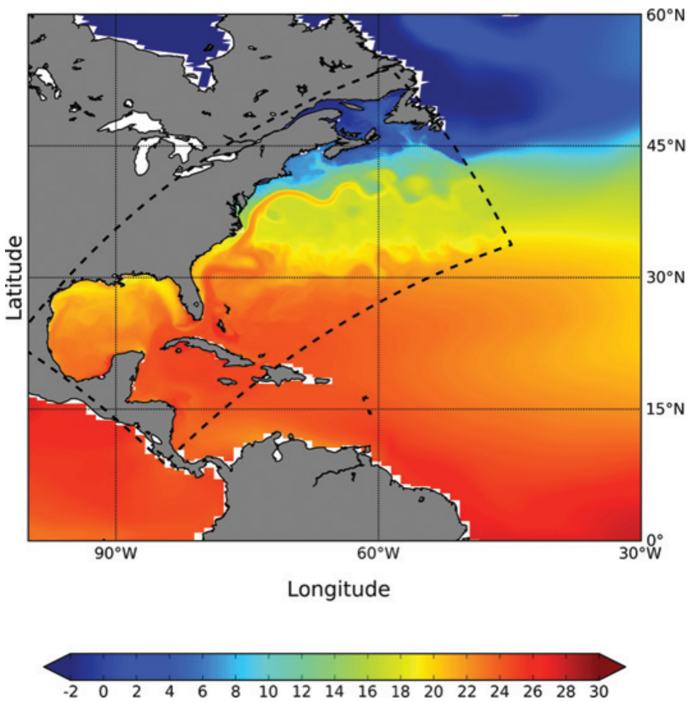

Figure 3: NCAR-CESM multi-scale configuration in the northwest Atlantic. The global ocean is POP at 1°; the regional model is ROMS at 7 km resolution. The ocean models are two-way coupled. A merged ocean SST is passed to the global atmosphere at each coupling timestep (R. Dussin, pers. comm.).

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Report on the 12th Session of the CLIVAR Working Group on Ocean Model Development (WGOMD) Kiel, Germany, 10-11 April 2014

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The 12th Session of the CLIVAR Working Group on Ocean Model Development (WGMOD) was held in Kiel, Germany on 10-11 April 2014 at GEOMAR, hosted by Dr. C. Böning. The meeting followed the WGOMD Workshop on High Resolution Ocean Climate Modelling. After a discussion of the main outcomes and recommendations of the workshop, the WGOMD meeting focused on reports on the status of the Coordinated Ocean-ice Reference Experiments phase II (CORE-II) analysis efforts, the CORE-II protocol, the WGOMD and CORE in the context of the Coupled Model Intercomparison Project (CMIP) process, future development of the Repository for Evaluating Ocean Simulations (REOS) and on the evolution of WGOMD as part of the changing structure and activities of the CLIVAR Project.

CORE-II

CORE-II represents an experimental protocol for ocean – sea-ice coupled simulations forced with interannually varying atmospheric data sets for the 1948-2007 period (Large and Yeager 2009). These hindcast simulations provide a venue for the following activities:

- To evaluate, understand, and improve ocean models
- To investigate mechanisms for seasonal, inter-annual, and decadal variability, and to evaluate the robustness of mechanisms across models
- To complement data assimilation by bridging observations and modelling;
- To provide ocean initial conditions for climate (decadal) predictions.

CORE-II simulations have garnered a tremendous interest from modellers and analysts with nearly 20 models participating. The simulations have fostered analysis efforts focused on the research areas listed below:

- Arctic Ocean (Wang et al., in preparation)
- Atlantic mean state (Danabasoglu et al., 2014)

• Atlantic variability (Danabasoglu et al., in preparation)
• Indian circulation (Ravichandran et al., in preparation)
• Pacific circulation (Tseng et al., in preparation)
• South Atlantic transport (Sitz et al., in preparation)
• Southern Ocean water masses and sea ice (Downes et al., in preparation)
• Southern Ocean eddy compensation and saturation (Farneti et al., in preparation)
• Sea level trends (Griffies et al., 2014)
• Water mass analysis (Zika et al., in preparation)

The plan is to publish each of these projects in a CORE-II special issue of the journal Ocean Modelling during 2014-2015. Indeed, we are very pleased to note that two manuscripts, documenting the mean states in the North Atlantic (Danabasoglu et al. 2014) and the global and regional sea level (Griffies et al. 2014) in the participating CORE-II simulations have been already published in the Ocean Modelling special issue. A broader and more unified scientific assessment of these CORE-II projects will be conducted upon completion of the above studies.
WGOMD discussed at length the CORE-II protocol and modelling group experiences in running the simulations. The protocol allows the use of surface salinity restoring, with participating groups employing a wide range of restoring time scales in their simulations, ranging from 50 days to 4 years (over 50 m thickness) to maintain stable simulations and obtain solutions in agreement with observations. In comparison to the impacts of such restoring, the deviations from the CORE-II protocol appear to be minor. However, more analysis is needed to understand the full impacts of the various modifications on the ocean simulations. The recommendations from the modelling groups could be the basis for developing a modified and improved CORE-II protocol.

An ocean biogeochemistry (BGC) and passive tracer protocol for CORE is being developed following the OCMIP-2 protocol - see the OCMIP-2 website: http://ocmip5.ipsl.jussieu.fr/OCMIP/phase2/simulations/.

WGOMD Contribution to CMIP

The WGOMD discussed the possible inclusion of the CORE-II experiments as a MIP within the CMIP framework in detail. WGOMD recognized that such inclusion would further increase the prominence of the CORE-II efforts and enforce the standardization of output data. However, there are some additional work and responsibilities that come with being a MIP. These include updates and improvements of the CORE-II protocol; continuing updates, corrections, and maintenance of the forcing data sets; and management of broader efforts, all requiring dedicated and not insignificant resources. After thoughtful deliberations, the WGOMD decided to i) to explore and seek funding opportunities to support the CORE-II efforts more reliably and ii) postpone participation in CMIP as a CORE-II MIP because we will have additional opportunities to do so in the near future.

WGOMD and Guests at the 12th Session, GEOMAR, Kiel, Germany

In preparation for CMIP6, WGOMD is requested by WGCM and the CMIP Panel to prepare the requirements for ocean model field output. The preference from the ocean modelling community is for the provision of data on native grids. However native grid output is complex for extracting latitude-longitude section from the ESGF. Steps are needed to help broaden the usage of native grid output, providing mapping scripts and guidelines on how to handle complex land-sea boundaries. Some basic variables, e.g. temperature, salinity, 2-dimensional state variables could be stored on a common, 1 degree latitude-longitude grid, thus requiring less work to use these data by the user community. The WGOMD will provide a list of ocean model output variables, based on the WGOMD document provided for CMIP5, prioritizing the output fields. We will also provide a short list of candidate variables for remapping onto a common regular latitude - longitude grid for easy analysis. A BGC output request, based on the CORE BGC protocol that is under development, will also be provided.

Repository for Evaluating Ocean Simulations (REOS)

WGOMD has developed the REOS web site, aimed at facilitating the research community's access to:
• basic datasets and analyses/syntheses products;
• metrics for evaluating variability and processes including input by the CLIVAR basin panels;
• guidance on ocean model validation;
• tools available for the community for ocean model data analysis;
• a comprehensive bibliography of papers, linked to the online articles where possible.

In order to make more use of this web site within CLIVAR, WGOMD recommends that it be incorporated into a broader pan-CLIVAR repository of sanctioned datasets, methods, projects, etc. Doing so will facilitate coordination across the observation-based and model panels. It will also leverage input from a broader group of CLIVAR scientists than available just within the WGOMD.

Links to CLIVAR SOP and GSOP

WGOMD is participating in discussions with P. Spence, M. England (and SOP/collaborators) on the design of coordinated experiments involving i) Southern Ocean wind perturbations and ii) Antarctic freshwater perturbation experiments.

With the maturation of the CORE-II effort and the availability of various reanalysis products, particularly via the GSOP/GODAE Ocean Reanalyses Intercomparison Project (ORA-IP), the WGOMD will collaborate with GSOP to start some preliminary comparisons of solutions from CORE-II experiments and those provided by reanalysis. The initial focus will be on Atlantic meridional overturning circulation (AMOC), Atlantic heat transport, and mixed layer depths. The analysis of overturning in temperature and salinity space being performed on CORE-II will also be extended to the ORA-IP simulations.

WGOMD evolving to CLIVAR Ocean Model Development Panel (OMDP)

The CLIVAR Scientific Steering Group (SSG) decided to instate the CLIVAR Ocean Model Development Panel (OMDP) at its 20th Session in July 2014, in accordance with the evolution of the overall direction of CLIVAR. OMDP will serve an advisory role in support of the CLIVAR ocean basin panels, in the development of the ocean observing system, and for the CLIVAR Research Foci. OMDP is a CLIVAR global modelling group and will continue in the role of providing leadership in the wider WCRP context on issues related to modelling the ocean as a component of the climate system, in partnership with the WCRP Working Group on Seasonal to Interannual Prediction (WGSIP) and Working Group on Coupled Modelling (WGCM) global modelling groups, under the coordination of the WCRP Modelling Advisory Council (WMAC).

The following areas have been discussed as future OMDP activities, beyond new activities as part of the CLIVAR Research Foci:

- CORE-III systematic Greenland ice melt experiment, forced by a realistic ice melt distribution
- AMOC MIP around Greenland in coupled models, comparing AMOC estimates with coupled climate model response
- Salinity restoring approaches
- High resolution CORE, development of high resolution CORE forcing dataset suitable for high resolution models, e.g., ERA-sat product (<T511 over satellite period, available in 2015)
- Climate response function CORE-type experiments
- CORE experiments with different wind products
- Partial coupling and other new forcing approaches for ocean only models
- Southern Ocean wind MIP, AABW cell intercomparison
- Exploring the feasibility and use of very high resolution 'truth' idealized experiments against which to compare models, parameterizations
- Community collaboration, CPT approach in parameterisation development tests, CVMix extension
- Coordinated activities that target systematic biases

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The International CLIVAR Climate of the 20th Century Plus (C20C+) Project: Report of the Sixth Workshop

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Introduction

The International CLIVAR Climate of the 20th Century Project (C20C; Folland et al., 2002) held its Sixth Workshop on 5-8 November 2013 at the University of Melbourne, Australia. C20C brings together climate modelling and data analysis groups to study climate variations and changes over periods up to the last 150 years using observational data and general circulation models (GCMs). There is an emphasis on atmospheric GCMs (AGCMs) forced with observed values of atmospheric composition (concentrations of greenhouse gases, aerosols, etc.) and surface conditions (SST, sea ice, land surface vegetation, etc.) as well as on natural variations alone. As agreed at the fifth Workshop in Beijing in 2010 (Kinter and Folland, 2011), the new C20C core project involves research in collaboration with the International Detection and Attribution Group and the international Attribution of Climate-related Events activity into the influence of anthropogenic forcing on climatic events, particularly extreme climate events. This is partly to support new research on quasi-operational attribution. The goal of the Sixth Workshop was to review early progress in the new core activity, observational data sets that will support C20C activities, and other key C20C projects. As in previous C20C meetings, the forcing data sets being used in a new set of coordinated model experiments, including the new Hadley Centre's SST and sea-ice analysis (HadISST2), were discussed including rapidly evolving arrangements for sharing the new experiments.

The 35 workshop participants from 18 institutions were welcomed by Prof. Janet Hergt, Dean of the Melbourne University Faculty of Science. A representative of the WCRP Working Group on Coupled Modelling, Dr Claudia Tebaldi, also attended. The Workshop enjoyed excellent hospitality from the University of Melbourne and the ARC Centre of Excellence for Climate System Science as well as the opportunity for a flutter on the prime local horse racing event of the year, the Melbourne Cup! The workshop web site (http://www.iges.org/c20c/dome.html) includes downloadable copies of most presentations and posters, and more detailed discussions of Workshop outcomes. A key decision was to rename the project C20C+ partly because of the new focus on research on operational attribution and because increasingly 21st century climate change is a crucial component of understanding variability and trend mechanisms. The projects and their progress are now described briefly.

The C20C+ Detection and Attribution Project

The aims are to characterise historical trends and variability in characteristics of damaging weather and short-term climate events as well as to determine the contribution of anthropogenic emissions to contemporary occurrence of these events. A key focus is on underlying uncertainties in these estimates. This will involve at least a dozen modelling groups around the world running atmospheric models in a semi-coordinated study of weather risk attribution. The experimental design can be summarised briefly as:

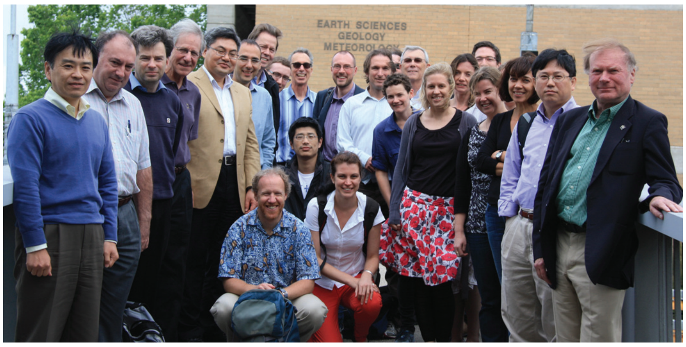
All-Hist: Simulations run under observed variations in radiative forcing and surface boundary conditions.

Nat-Hist: Simulations run under variations that might have occurred in radiative forcing and surface boundary conditions had anthropogenic emissions never interfered with the climate system.

In addition, it is planned that some coupled global models, regional models and selected impacts models will be used. Observations and models will also supply key climate indices. Output from the main simulations is being published on the Earth System Grid Federation (ESGF, http://esg.nersc.gov) under the project name "c20c". The data will also be used by other C20C+ projects. An important component of future experiments will be use of HadISST2 as the surface boundary condition to force the AGCMs.

HadISST

The Hadley Centre has developed an improved analysis of global SST and sea ice concentration so as to include



Participants in the 6th Workshop of the International CLIVAR Climate of the 20th Century Plus (C20C+) Project

more observations and attain greater accuracy and resolution. Titchner and Rayner (2014) describes the sea ice component. Key improvements of HadISST2 over HadISST1 are multiple (100) realizations (though only a small subset is likely to be used by C20C+), better resolution in time, new bias corrections to SST right up to the present, inclusion of AATSR satellite data and a considerably improved sea ice extent data set. Talks on the complex process used to create the SST and sea ice components of HadISST were presented at the Workshop.

Other Observational data sets

For the first time a C20C workshop highlighted key types of observational data sets of particular use to sub-projects other than HadISST2. The data sets included precipitation, reanalyses and a new extremes data set. The simulation of precipitation trends and events is a core C20C+ sub-project while the use of reanalyses to study changes in jet streams has already been important in atmospheric circulation studies. On the global scale there is much disagreement between precipitation data sets since 1979, and similarly there was much disagreement on the Indian Monsoon. Somewhat by contrast, six out of eight reanalysis data sets gave reasonably robust climatological characteristics of the main jet streams in the two hemispheres though, as might be expected, the 20th Century Reanalysis misses the existence of the near equatorial stratospheric jet associated with the QBO. HadISD is a new global sub daily data set underdevelopment that will be very useful for studying variations and trends studying daily and sub-daily extremes including storms (Dunn et al, 2012).

Weather Noise and performance of AGCMs Core Project

A key question for C20C+ is the extent to which AGCMs forced with observed SST and sea ice extents give reliable results when compared to their coupled model counterparts. This is also related to the problem of the relative contribution of SST forcing and weather noise to atmospheric variance. It is hypothesised that the information contained in detailed observational data sets like reanalyses can be best understood by identifying and comparing the weather noise and forced responses. The work has also been motivated by previously published results showing that AGCMs and CGCMs have different teleconnections. New results presented (e.g. Chen and Schneider, 2014) show that AGCMs can indeed be used to gain sights into teleconnections and SST forced responses as these are (almost) the same in the AGCMs and partner CGCMs. Some previous published results to the contrary appear to be strongly affected by model biases. However an exception is quantitative estimates of tropical precipitation where AGCMs clearly over-estimate responses to SST compared to CGCMs. The statistics of weather noise however appear to be similar in AGCMs and their coupled partners.

Precipitation variability and trends Core Project

This topic attracted a considerable number of presentations on observed and modelled trends and variations in precipitation and their forcing factors including the PDO. Included was progress in understanding the mechanisms of the spatial structure of Indian monsoon rainfall using the very high resolution (16km) ATHENA model forced with observed SST. For this workshop, relationships between ENSO and Australian rainfall were of considerable interest, particularly after the recent (2011-12) exceptionally wet period. One study concluded that ENSO remains dominant over anthropogenic effects for attributing such events. Results were shown from a very large ensemble of HadAM3P simulations obtained via the weather@home citizen science project. An interesting result is that HadAM3P can reproduce the asymmetric relationship between rainfall and ENSO over south east Australia in austral summer.

Atmospheric and Oceanic Variability and Atmospheric Predictability

A variety of talks were given including tropical cyclone trends and the reductions in South Australian rainfall in the last few decades which is also seen in many CMIP5 models under enhanced greenhouse gases. A complex mathematical method, developed partly under the auspices of C20C over the last decade, which distinguishes forced and internal atmospheric modes (Grainger et al. 2011) was presented to study the veracity of atmospheric circulation modes CMIP3 and CMIP5 coupled model simulations. CMIP5 models were shown to be an improvement. Among a number of other atmospheric studies, a presentation on the expansion of the tropics concluded that the modelled rate of tropical expansion in CMIP5 models is towards the low end of the range of measurements. A core atmospheric circulation project is that on the summer North Atlantic Oscillation Progress since the 5th Workshop includes two major papers (Linderholm et al, 2011, 2013) and increasing evidence was presented that boreal European summer climate has strong links to Southern Hemisphere winter climate as well as to the Atlantic Multidecadal Oscillation. The project now makes regular contributions to the annual Bulletin of the American Meteorological Society supplement on the State of the Climate.

New Early Career Scientist CLIVAR Initiative

Because of the number of young scientists present, Dr Sarah Perkins on behalf of CLIVAR presented a new initiative, the CLIVAR Early Career Scientist's Network. This is an international network involving website, conferences and social media that can include any climate scientist who considers themselves in the primary stages of their career, student, post-doc, or permanent research scientist. A full discussion of this evolving proposal is at: www.clivar.org/sites/default/files/ECS/Documents/CLIVAR_ECS_Survey_report.pdf.

Key Plans for the development of C20C+

A relatively full discussion is at the summaries of the two Workshop Breakout Groups at http://grads.iges.org/c20c/home.html Here we review the main plans not mentioned above.

Detection and Attribution Project

Project outputs will support the Bulletin of the American Meteorological Society's annual State of the Climate attribution supplement (Peterson et al, 2013). Before the next C20C+ Workshop, it is hoped to organise a journal special issue providing an overview of the first results from this subproject. More details can be found on the C20C+ web site http://grads.iges.org/c20c/home.html and at the subproject web site http://portal.nersc.gov/c20c/. Ideas for possible future experiments were discussed at the workshop, including a focus on the effects of specific forcings, a focus on projections for some future period, and a focus on geo-engineering problems.

Atmospheric circulation, rainfall and atmospheric noise

The Atmospheric Circulation and Rainfall Working Group discussed the plans for remaining C20C+ projects and their proposed participants. Following several recent published papers, some listed below, the atmospheric noise project will look at long time-scales of climate variability. These may be forced (partly at least) by atmospheric noise due to coupled atmosphere-ocean processes or ocean internal variability. With the help of AMIP-type ensembles to estimate weather noise, the noise component of key drivers of decadal variability in CGCM control experiments and decadal prediction experiments will be studied using diagnostic models.

There is increasing evidence that state of the art models represent processes affecting European climate considerably better than in the past. The role of C20C+ will be particularly in studying forcing mechanisms in the summer and for winter half year UK and European droughts. The latter project is studying the mechanisms of European summer climate variability following a number of recent papers on the summer North Atlantic Oscillation. Particular emphases are tropical rainfall forcing, SST forcing including the AMO, and the effect of the decline in recent Arctic sea ice. The project is also reconstructing the SNAO over the last millennium (Linderholm et al, 2013) and using CMIP5 models to try to simulate SNAO variations over this long period. Reanalyses are important to this project; the SNAO essentially involves changes in the tropospheric jet stream and the reanalysis project is expected to provide further advice on suitable reanalyses. There is considerable further potential to uncover the relative variance of potentially predictable modes and internal variability in both hemispheres using the Grainger et al. (2011) statistical method mentioned above.

In the context of forcing factors for atmospheric circulation modes and for studies of rainfall data in their own right, there is a considerable need for a review of, and advice on, rainfall data sets to support C20C+ activities. Many such global or quasi-global data sets now exist. It is expected that using the core C20C+ model data, a much more extensive analysis will be done on trends and variations in global precipitation; this activity will also cast light the strengths and weaknesses of the many observed data sets. Further investigation of monsoon rainfall mechanisms will be done using the IFS 16km – T1279 1o ocean resolution model. A 50-member ensemble is available for 1980-2010.

Finally as a background to these activities, it was agreed to add to the C20C+ project an ongoing study of the time-varying causes of the slowdown in global warming, and how this might end. It was also agreed that the next workshop would be in 2016 or 2017 at the Center for Ocean-Land-Atmosphere Studies, Maryland, USA.

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CONTENTS

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