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# Ocean Science - extended abstracts of the fourth CAWCR Workshop 9 November - 11 November 2010 Hobart, Tasmania, Australia

The Centre for Australian Weather and Climate Research A partnership between CSIRO and the Bureau of Meteorology

**CAWCR Technical Report No. 033** 

Andrew J. Hollis and Keith A. Day (editors) November 2010





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Centre for Australian Weather and Climate Research, GPO Box 1289, Melbourne, VIC 3001, Australia

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# FOREWORD

The Centre for Australian Weather and Climate Research (CAWCR) is a partnership between Australia's leading atmosphere and ocean research agencies – CSIRO Marine and Atmospheric Research (CMAR) and the Bureau of Meteorology. CAWCR, established in 2007, jointly manages the science capability within the Bureau and CMAR providing a single centre of research excellence. This year's CAWCR Workshop, '*CAWCR Ocean Science Workshop*' is the fourth Annual Workshop under the auspices of the Centre, continuing the series originating within the Bureau of Meteorology Research Centre.

The science undertaken within the Centre is applied in a variety of climate-related areas including climate change, physical oceanography, and seasonal climate prediction. There are many scientific questions to resolve. The key themes covered in this year's workshop are: (i) Global Ocean Climate Modelling; (ii) Ocean Processes; and (iii) Mesoscale Oceanography and Forecasting.

The workshop includes invited participants from leading research groups around Australia. We also welcome a number of contributors from overseas research institutes and national meteorological services to the workshop, with invited keynote presentations from Frank Bub (NAVOCEANO, USA), Charles Eriksen (University of Washington, USA), Stephen Griffies (Geophysical Fluid Dynamics Laboratory, USA), Nikolai Maximenko (International Pacific Research Centre, University of Hawaii, USA) and Motohiko Tsugawa (Japan Agency for Marine-Earth Science and Technology, Japan). We are grateful for these expert contributions and to all the participants' contributions to the debate and discussions.

We are particularly grateful to both Oracle Corporation Australia and the Integrated Marine Observing System (IMOS) for their generous support of this workshop.

Finally, I would like to thank the Local Organising Committee for the workshop, comprising Trevor McDougall (Chair), Gary Brassington, Keith Day, Andrew Hollis, Val Jemmeson, Simon Marsland and Susan Wijffels.

#### Tom Keenan

#### Director

Centre for Australian Weather and Climate Research: A partnership between the Australian Bureau of Meteorology and CSIRO

November 2010

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# **GLOBAL OCEAN MODELLING ACTIVITIES AT GFDL: 2004-2010**

Extended Abstract for a Science Workshop of the Centre for Australian Weather and Climate Research (CAWCR) Hobart, Australia, 9-11 November 2010

#### Stephen Griffies

#### NOAA/Geophysical Fluid Dynamics Laboratory Princeton, USA Thu Oct 14 10:18:14 2010

This presentation focuses on global ocean climate science and earth system modeling at NOAA's Geophysical Fluid Dynamics Laboratory (GFDL) during the years 2004-2010. Over this period, GFDL transitioned from its IPCC AR4 climate model, CM2.1, to a suite of AR5 climate and earth system models. Additionally, GFDL transitioned from the generalized level model, MOM (Modular Ocean Model), being the sole ocean code, to a partnership with the generalized layer model, GOLD (General Ocean Layer Dynamics). Elements of these two transitions will be described in this presentation.

# Climate and earth system modeling

The main focus of numerical ocean modeling research and development at GFDL concerns questions of climate science, including questions of earth system science in which biogeochemical processes are considered. During recent years, there has been the development of a suite of climate and earth system models, from intermediate complexity to fully realistic. The purpose of this section is to summarize the main features of these models and their scientific focus.

# CM2.1 and ESM2.1

The CM2.1 coupled climate model (Delworth et al. 2006; Griffies et al. 2005; Gnanadesikan et al. 2006; Wittenberg et al. 2006; Stouffer et al. 2006; Russell et al. 2006b) was completed in 2004 for IPCC AR4 science and projections. Although now outdated in some process physics and numerical algorithm aspects, a significant number of studies indicate that CM2.1 is the benchmark (see <a href="http://data1.gfdl.noaa.gov/CM2.X/">http://data1.gfdl.noaa.gov/CM2.X/</a> for a list of references) for future development of models at GFDL and elsewhere. The CM2.1 configuration is supported through the 18Dec2009 public release of MOM4p1, with this release the first for a GFDL coupled climate model configuration; see <a href="http://www.gfdl.noaa.gov/accessing-cm2-1p1">http://www.gfdl.noaa.gov/accessing-cm2-1p1</a> for details.

CM2.1 remains in use at GFDL for IPCC AR5 applications aimed at the decadal prediction component of CMIP5. In particular, a coupled ocean-atmosphere assimilation system has been constructed based on CM2.1 (Zhang et al. 2007), with an upgraded ocean model component using the MOM4p1 code of Griffies (2009). The assimilation system is used both for ocean reanalysis and the initialization of coupled climate predictions. See <a href="http://www.gfdl.noaa.gov/ocean-data-assimilation">http://www.gfdl.noaa.gov/ocean-data-assimilation</a> for details.

Plans are to conduct prediction experiments with this model based on 40 initial conditions (1970-2010), each run for 10 years using 10 ensemble members each (4000 years of simulation). Ensemble members use the same ocean initial conditions, with atmosphere initial conditions shifted by a few days. This approach helps to establish an upper limit on predictability of the variability simulated by this model.

CM2.1 forms the physical basis for one of GFDL's first earth system modeling efforts, ESM2.1. ESM2.1 uses early versions of the GFDL land and vegetation models, and it is a proto-type for the ESM2M and ESM2G models discussed later. ESM2.1 has been successfully run to millenial scale equilibrium using 1860 radiative forcing, and has been used for historical (1860-2000) and AR4 21st century warming scenarios. ESM2.1 produces a physical climate similar to CM2.1, but with some differences attributable to the interactive biogeochemistry in the ocean as represented by the GFDL TOPAZ model. There are additional plans to use ESM2.1 for millenial scale paleoclimate simulations focusing on the last glacial maximum.

### СМ3

Subsequent to the develoment of CM2.1, GFDL focused on two main development paths toward high-end models of use for AR5 and beyond. One avenue, leading to the climate model CM3, focused on atmospheric component priorities, including aerosol-cloud interactions, chemistry-climate interactions, and links between the troposphere and stratosphere. Updates to the land model used for ESM2M/G were also incorporated. To help achieve a state-of-the-science climate model tool using the new atmospheric model, in time for the AR5, we chose to keep the ocean and sea ice components of CM3 effectively the same as in CM2.1. Hence, CM3 uses the MOM4p1 code configured nearly as in CM2.1, with the single exception that the vertical grid uses the stretched geopotential coordinate  $z^*$  proposed by Stacey et al. (1995) and Adcroft and Campin (2004). Documentation of CM3 is given by Donner et al. (2010) (atmospheric component) and Griffies et al. (2010) (ocean and sea ice components).

# CM2M/ESM2M and CM2G/ESM2G

The second path toward AR5 models emphasized the needs of earth system modeling, in which interactive ocean biogeochemistry, land vegetation, and interactive carbon cycling are critical. This path used nearly the same atmospheric model as in CM2.1, and it led to two new earth system models, known as ESM2M and ESM2G, that differ only by their ocean components. ESM2M, based on MOM, uses the physical model CM2M, where CM2M has an updated suite of physical parameterizations and numerical methods beyond those used for CM2.1. The earth system model ESM2G is based on the CM2G physical model, in which the ocean component uses the GOLD isopycnal layer model.

ESM2M has been spun-up for ~2000y using pre-industrial radiative forcing. It has reached a stable climate equilibrium with a global mean ocean heat flux of less than  $0.01 W m^{-2}$ , and a corresponding stable carbon cycle. Consistent with the millenial scale CM2.1 and ESM2.1 simulations, ESM2M equilibrates to an ocean state somewhat warmer than present day observations. ESM2M is presently being integrated through the CMIP5 experiment suite. A similar plan was followed for ESM2G. However, ESM2G has yet to reach a stable climate equilibrium state, with roughly  $-0.2 W m^{-2}$  of heat leaving the ocean even after 2000y. It was decided that this particular version of ESM2G was not suitable for CMIP5. Consequently, further development is ongoing, with the working hypothesis being that the ocean model requires a more substantial level of interior mixing. A corresponding new setup for ESM2M is being considered as well, aiming to bring the interior mixing schemes for both models into a close agreement. It is unclear whether either of these two new configurations for ESM2M and ESM2G will contribute to CMIP5.

# CM2.5

In an attempt to address limitations of the one-degree class of ocean climate model (e.g., CM2.1 and ESM2M/ESM2G), GFDL has developed the CM2.5 climate model, which uses a  $1/4^{\circ}$  configuration of MOM4p1 coupled to a  $1/2^{\circ}$  cubed sphere atmosphere. Notably, this atmospheric configuration has shown some degree of realism in tropical cyclone simulations (Zhao et al. 2010). The ocean component of CM2.5 is similar to that used for the CM2.4 model of Farneti et al. (2010), yet CM2.4 used an atmosphere with 100km resolution. CM2.5 has been run for 200 years with 1990 radiative forcing, and it is presently being run in an idealized  $CO_2$  doubling experiment. Ongoing development with CM2.5 is focused on addressing sizable biases in the North Atlantic subpolar gyre. In particular, we are implementating the overflow scheme of Danabasoglu et al. (2011) in hopes of reducing biases on the Labrador Sea.

# CM2Mc and ICCM

In addition to the high-end "big science" efforts mentioned above, there are two notable model configurations developed through a far smaller effort. However, these configurations are proving to be very useful for many science questions, especially those posed by graduate students and post-docs. First, there is the CM2Mc configuration, developed by Eric Galbraith during his post-doc research with Jorge Sarmiento at Princeton University. CM2Mc is a coarsened version of both the atmosphere and ocean components of CM2M. It is focused on applications toward millennial scale earth system modeling and paleoclimate questions. This model is documented by Galbraith et al. (2011). Second, the Intermediate Coupled Climate Model (ICCM) of Riccardo Farneti was developed during his post-doc research period with Geoff Vallis at GFDL. This model uses a highly idealized land-sea geometry with a coarsened atmospheric resolution. It is documented in Farneti and Vallis (2009a), and its Atlantic variability is described by Farneti and Vallis (2009b). Both the CM2Mc and ICCM model configurations are available via the 18Dec2009 public release of MOM4p1.

# Future directions in GFDL ocean modeling

The purpose of this section is to highlight some of the code development aims and primary applications for numerical ocean models at GFDL over the next few years.

# Global eddy permitting simulations

GFDL ocean modeling will continue to stress the needs of global climate and earth system applications. The next few years will see particular emphasis on grid resolutions that admit a nontrivial portion of the ocean mesoscale eddy spectrum. Such global ocean eddy permitting models, as illustrated by CM2.5 described earlier, will in principle greatly assist in reducing uncertainties concerning the ocean's role in global climate variability and change. In addition to coupling such eddying ocean models to interactive atmosphere and land models as in CM2.5, we also plan a hierarchy of global ocean-ice models forced according to the protocol of the Coordinated Ocean-ice Reference Experiments (CORE) (Griffies et al. 2009). One application of the CORE hindcast simulations is to provide a mechanistic understanding of the instrumental record.

### Global and regional sea level

One of the most important research applications of global and regional ocean models at GFDL concerns the issues of sea level. Particular questions related to regional patterns of sea level (e.g., Yin et al. 2010), as well as effects on the gravity field due to major shifts in mass (Kopp et al. 2010), are two recent examples of how climate models at GFDL have been useful in deriving insights into sea level rise issues. Additional work

includes process studies that focus on interactions between the ocean and ice shelves, with such studies having direct implications for possible changes in continental ice sheets and thus for sea level.

### Southern Ocean response to climate change

The Antarctic Circumpolar Current (ACC) has spun-up in response to stronger and more poleward shifted southern westerlies since the 1950s. Changes in the westerlies have been attributed to CO2 induced warming and to depletion of ozone over Antarctica, both of which have increased the equator-to-pole temperature contrast in the middle atmosphere (Russell et al. 2006a). These changes are analogous to those as the earth warmed at the end of the ice age (Toggweiler and Russell 2008; Anderson et al. 2009). Theory and models suggest that stronger westerlies and a stronger ACC should induce a stronger AMOC and greater ventilation of the deep Southern Ocean (Russell et al. 2006a). However, the overturning is expected to weaken due to a stronger hydrological cycle. It is critical that this struggle between stronger westerlies and a stronger hydrological cycle be realistically simulated. Data analysis (Böning et al. 2008) and eddy permitting simulations (Hallberg and Gnanadesikan 2006; Farneti et al. 2010) indicate that climate models (Russell et al. 2006b) require refined resolution to accurately capture important physical processes (e.g., continental shelf processes, sea ice, mesoscale eddies) active in the Southern Ocean. Such forms a major focus both on the fine resolution eddy permitting simulations, and research into better eddy parameterizations for coarse models.

### Hurricanes and climate research

GFDL scientists have pioneered the coupling of hurricane models to ocean models, illustrating the importance of the negative feedback from cold waters upwelling under slow moving hurricanes (Bender et al. 2000). These early efforts used the Princeton Ocean Model. Recently, however, the hurricane research has incorporated MOM4p1 as its ocean component, with plans to couple a fine resolution MOM configuration (either 1/4° or 1/10°) to a 1/4° atmosphere for the study of global tropical cyclone and ocean interactions. This coupled model will require the use of an upper ocean wave model to distribute the momentum from the atmosphere to the ocean. The NCEP wave model has been ported to the GFDL Flexible Modeling System (FMS) for this purpose.

# A new paradigm for GFDL ocean code

As summarized by Griffies et al. (2000), the level and layered approaches to partitioning the vertical direction in an ocean model each have their respective strengths and weaknesses. The ocean community has yet to conclude which, if either, is optimal for global climate and earth system modeling. Given the intellects and passions at GFDL pursuing each path, GFDL has supported the two, with a notable increase in resources provided during years 2004-2010 to the layered approach. Goals of this two-pronged effort are to develop a rational platform to understand the ocean climate system, and to render better simulation tools. GFDL is at the cusp of exploiting this multi-year development effort, which in particular has resulted in the two earth system models ESM2M and ESM2G mentioned earlier. Over the course of the next few years, analysis efforts will be focused on both ESM2M and ESM2G, as well as other models constructed with MOM and GOLD as the ocean component. Ultimately, a path toward unification is envisioned, whereby there is just a single code base.

#### Process oriented research

Global modeling is the baseline for GFDL research and development. Nonetheless, idealized process oriented studies provide a key element in the development of new physical parameterizations and model algorithms. In particular, the development of a unified MOM/GOLD code will be fostered through a suite of

idealized process simulations where precise mechanisms, and potential code problems, are readily diagnosed. Additionally, the advent of a new Climate Process Team (CPT) focusing on internal gravity waves will motivate numerous idealized simulations that test our ability to understand and to parameterize ocean mixing.

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# HIGH RESOLUTION MODELLING WITH THE PARALLEL CUBIC GRID OCEAN GENERAL CIRCULATION MODEL

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# Introduction

The parallel cubic-grid ocean general circulation model (PCOM, Tsugawa et al 2008) is developed at the Japan Agency for Marine-Earth Science and Technology (JAMSTEC). The aim of the development is to make an ocean model which can be used for global high-resolution simulations, because the traditional longitude-latitude grid model is not suitable for the purpose.

The development starts from the research of the grid system suitable for global ocean modeling. Several grid systems including the tripolar grid, overset grids, icosahedral grids are investigated and a cubic grid is chosen as the grid system of the new ocean general circulation model (OGCM). There are several variants in the cubic grid, and the grid proposed by Purser and Rancic (1998, hereafter PR) is employed. The advantages of PR's cubic grid are as follows. The grid is quite homogeneous, so that the developed OGCM will not suffer the stringent time step limitation due to the CFL condition. In addition, the grid has quadrilateral grid cells so that the techniques used in traditional OGCMs can be implemented easier than the grid systems like the icosahedral grid. On the other hand, the disadvantage may be the lack of orthogonality and existence of the singular points.

In order to test the performance, PCOM is compared to the traditional latitude-longitude grid OGCM. The result of PCOM shows very good agreement with the latitude-longitude OGCM: it does not show any critical problem caused by the singular point and the non-orthogonality. PCOM is now available for scientific research, and is used for the simulation of the Agulhas Current. PCOM reproduces successfully the eddy activities around the Agulhas Current system, especially the characteristic cyclonic meanders observed in the current, the Natal Pulse.



*Fig. 1* The quasi-homogeneous cubic grid (left) and the stretched cubic grid used in the Agulhas Current simulation (right).

#### **Description of the Model**

The cubic grid is generated by mapping the grid cells on six faces of a cube to six tiles which cover the whole sphere. The mapping method proposed by PR is able to generate both a conformal grid and a quasi-homogeneous, non-orthogonal grid, and PCOM employs the quasi-homogeneous version (Fig 1).

The equations of the ocean model of the non-orthogonal grid are:

$$\begin{split} &\frac{\partial}{\partial t} u_{\alpha} = (f + \zeta) E_{\alpha\beta} u^{\beta} - K_{;\alpha} - w \frac{\partial u_{\alpha}}{\partial z} - \frac{1}{\rho_0} p_{\alpha} + F_{\alpha}, \\ &\frac{\partial \theta}{\partial t} = -(\theta u^{\alpha}) - \frac{\partial w \theta}{\partial z} + D_{\theta}, \\ &\frac{\partial s}{\partial t} = -(s u^{\alpha}) - \frac{\partial w s}{\partial z} + D_{s}, \\ &\frac{\partial w}{\partial z} = -u_{;\alpha}^{\alpha}, \\ &\frac{\partial p}{\partial z} = -g\rho, \end{split}$$

where <sup>u</sup> is the horizontal velocity vector, <sup>w</sup> is the vertical component of velocity, <sup>f</sup> is the Coriolis parameter,  $\zeta$  is the relative vorticity, K is the kinetic energy,  $\rho$  is the density,  $\theta$  is the potential temperature, s is the salinity, p is the pressure, Dh and Ds are the diffusion terms of potential temperature and salinity,  $\mathbf{F}_{\alpha}$  is the viscous force and  $\mathbf{E}_{\alpha\beta}$  is the anti-symmetric tensor. The Einstein summation convention is used here. The first equation is the vector-invariant-form momentum equation, second and third the temperature and salinity equations, forth the mass conservation equation and the last the hydrostatic equation. Greek sub- and superscripts represents the covariant and contravariant components, respectively. The covariant and contravariant components can be interchanged by using the metric tensor.<sup>g</sup> ab, as

#### $\mathbf{u}_{\alpha} = \mathbf{g}_{\alpha\beta}\mathbf{u}^{\beta}.$

In the traditional OGCMs, the orthogonal coordinate is used. The off-diagonal component of the metric tensor in the orthogonal coordinate vanishes, and that enable one to write the equations in a simpler and familiar form. On the other hand, PCOM uses the off-diagonal component, so it uses the more complex equations described above. However, the overhead of the complexity is not critical.

Although the grid system is different from traditional ocean models, the other features are quite similar to the traditional Bryan-Cox type OGCMs. The model employs the Boussinesq and hydrostatic approximations, the Arakawa B grid and vertical level coordinate. It employs an explicit free surface method for barotropic mode solver and partial bottom cell for bottom topography. PCOM is already vectorized and parallelized. The code of PCOM is written with Fortran90 and MPI.

The model is developed for global simulation, but PCOM can handle various geometries like f-plane, betaplane and latitude-longitude grid, by changing the geometric quantities such as the metric tensor. Especially, by use of a grid-stretching technique, PCOM can be used as a regional model, as shown later.

#### Model performance

PCOM is compared to a latitude-longitude grid model to examine its validity as a global ocean model. Longterm integrations are carried out and compared for the final state of the two models. The resolution of PCOM is C64, which denotes  $64 \times 64$  velocity points on each surface of the cube. The maximum grid size of C64 is about 160 km. PCOM is compared to  $1.5^{\circ}$  resolution latitude-longitude model (L1.5), hence the resolutions of C64 and L1.5 are similar. The initial condition is for motionless and homogeneous oceans. The integration period is 4000 years, which is long enough for models to reach their equilibrium state.



*Fig. 2* The sea surface height field (m) of PCOM (left) and the latitude-longitude model (right) after the 4000 years integration.

The sea surface height (SSH) field is shown in Fig. 1. As can be seen, major features corresponding to wind drive circulations such as western boundary currents and the Antarctic circumpolar current are reproduced. The SSH field of C64 shows very close resemblance to that of L1.5. The resemblance indicates that the dynamical balance in the wind driven circulation is correctly attained in the framework of the cubic grid system.



*Fig. 3* The zonal-mean meridional overturning circulation (Sv) of PCOM (left) and a latitude-longitude model (right) after 4000 years integration.

Figure 2 shows the meridional overturning circulations. The maximum transports in the Northern Hemisphere, which are contributed to mainly by the circulation of North Atlantic Deep Water (NADW), are about 16 Sv in both C64 and L1.5. Although the overall structure of PCOM is quite-similar to L1.5, the cells in high-latitudes show differences. It is speculated that such differences originate from the difference of representation of the coastline and bottom topography in high latitudes between the cubic grid and latitude-longitude grid. In the latitude-longitude grid, finer structure can be better represented and hence smaller scale flow is represented, compared to PCOM.



*Fig. 4* A snapshot of the sea surface velocity  $(m s^{-1})$  of the eddy-permitting model (C640) simulation.

In addition to the low-resolution test case, a global eddy-permitting simulation is carried out. The grid used in this model is C640, which corresponds to about 16 km resolution. Fig. 4 shows the sea surface velocity field after 20 years of integration. PCOM reproduces small scale features and eddy activities. For example, the strong western boundary currents, like the Kuroshio, the Gulf Stream and the Agulhas Current, are reproduced. In the equatorial Pacific Ocean, the tropical instability waves are reproduced. In the southern ocean, the Antarctic Circumpolar Current is reproduced with many meso-scale eddies.

# Simulation of the Agulhas Current

The Agulhas Current system is selected as the first scientific target of PCOM, because it shows various eddy activities and plays a crucial role in the global thermohaline circulation. Since PCOM is a flexible modeling system, it can be used like a regional OGCM. By use of a grid-stretching technique proposed by Murray (1996), which is the same as the Schmidt transformation, the resolution around the Agulhas Current is enhanced as shown schematically in Fig. 1. The C256 grid, which has nominal resolution of 40 km, is stretched and the resolution around the Agulhas Current system is about 10 km.

To use the model as a regional model, a target region is defined around the Agulhas Current System. The target region is the place where the distance from the center of the Agulhas Current System is less than 4000 km, hence it covers the eddy active region both upstream and downstream of the Agulhas Current. The temperature and salinity outside of the region is restored to the climatological value on all levels, and the sea surface wind stress is applied all over the world. These treatments supply realistic boundary conditions to the target region.

The stretched-grid PCOM reproduces the Agulhas Current realistically. The mean volume transport is about 80 Sv, which is near to the observed transport of 70 to 75 Sv. The retroflection loop usually lies west of 20°E, consistent with the observations, while in the coarser resolution models the Agulhas Current retroflects

at a more east longitude. PCOM also reproduces the major eddy activities in the Agulhas Current system. The Agulhas Rings exist around the southern tip of Africa. Also the Mozambique Eddy, which exists in the Mozambique Channel, upstream of the Agulhas Current, is also reproduced realistically.



Fig. 5 A snapshot of the sea surface velocity ( $m s^{-1}$ , shade) of the Agulhas Current simulation. The black arrow indicates a Natal Pulse.

Among the eddy activities, the Natal Pulse is focused in the following. The Natal Pulse is a cyclonic meander characteristic of the otherwise-stable northern Agulhas Current. The Natal Pulses are generated at the Natal Bight, when an anticyclonic eddy comes from upstream. The horizontal sizes of the Natal Pulses are several tens of kilometers when they are generated. The pulses move along the Agulhas Current at speeds about 20 km per day and grow as they move to at most 200km in horizontal size. The Natal Pulses play various roles in the Agulhas Current System. Especially, they relate the separation of the Agulhas Rings.

The generation and growth mechanism of the Natal Pulse is not understood completely. Until now, a mechanism proposed by de Ruijter et al (1999) has been considered as the most promising. He points out that the steep bottom slope along the northern Agulhas Current stabilizes the path of the Agulhas Current, but the bottom topography of the Natal Bight is not so steep. Hence the barotropic instability readily occurs there. It is difficult to reproduce such a small scale meander in the Ocean General Circulation Model until recently, so that the mechanism of the generation and growth has not been specified so far.

The Natal Pulse is reproduced in the simulation of the Agulhas Current system (Fig. 6). In the model, a Natal Pulse is generated when an anticyclonic eddy from the Mozambique Channel come to the Natal Bight. The Natal Pulse moves downstream with the anticyclonic eddy and grows as it moves. The features of the reproduced Natal Pulses are consistent with the observations. In addition, a supplemental simulation in which the bottom topography of the Natal Bight is manipulated to steep topography is carried out to validate the mechanism proposed by de Ruijter et al (1999). Though the generation should be suppressed according to the proposed mechanism, the Natal Pulses are reproduced. The features of reproduced Natal Pulses in the supplemental simulation are similar to those in the original simulation. Hence it can be concluded that the topography of the Natal Bight plays no role in the generation and growth of the Natal Pulses.



**Fig. 6** The velocity (vector,  $\mathbf{m} \mathbf{s}^{-1}$ ), depth (contour,  $\mathbf{m}$ ) and the potential vorticity (shade,  $10^{-6} \mathbf{m}^{-1} \mathbf{s}^{-1}$ ) of the layer between  ${}^{\sigma}\theta = 26.4$  and 24.6 around a Natal Pulse

To seek the generation and growth mechanisms, the eddy-mean barotropic and baroclinic energy conversion rates are calculated. The barotropic energy conversion is the main source of the growth of the Natal Pulse. The energy conversion rate analysis reveals that the interaction of the shear in the mean flow of the Agulhas Current and the Mozambique Eddy is the key factor of the generation and growth of the Natal Pulses. In addition, the analysis of the potential vorticity provides another view of the generation and growth mechanism. The high-potential vorticity between the coastline and the Agulhas Current is advected to offshore by the flow of the anticyclonic vorticity accompanying the pulse. Both the energy conversion and potential vorticity analyses show that the important component of the generation and growth of the Natal Pulse is the interaction between the anticyclonic vorticity and the mean flow of the Agulhas Current. The topography of the Natal Bight plays no role in the mechanism (Tsugawa and Hasumi 2010).

### Summary and future plan

A parallel cubic-grid ocean model (PCOM) has been developed. PCOM uses a homogeneous cubic grid in order to circumvent problems around the pole. It is tested and shows good performance of the global simulation. PCOM is designed for the global ocean circulation but it can also be applied to the regional research. It is applied to the Agulhas Current simulation by using a stretched grid and reproduces the various eddy activities of the Agulhas Current system.

The cubic grid PCOM is now available for ocean modeling research. A higher resolution regional model, which covers the region around the southern tip of Africa with about 5 km resolution, is being carried out to investigate the eddy activities and their role in the Indo-Atlantic interocean exchange. In addition, the development continues to make PCOM available for more realistic simulations. Several components, such as an ice model, isoneutral diffusion, thickness diffusion, mixed layer schemes and the higher-order advection schemes will be implemented.

A development project of a new coupled ocean-atmosphere model based on a cubic grid is starting under collaboration between CSIRO, CSIR in South Africa and JAMSTEC. The atmospheric component of the new coupled model is CCAM (McGregor 2005) and the ocean component is PCOM. The coupled model will be used for various climate change researches, and will include variable-resolution climate change simulations. The global simulations using the cubic grid coupled model will require less computational resource compared to the traditional coupled models. Hence the model is advantageous for challengingly high-resolution simulations on the fastest super computers, and also advantageous for medium resolutions on relatively cheap super computers. The coupled model will also be used for downscaling simulations. As shown in this paper, PCOM works well for the regional simulation with resolution enhancement by the grid

transformation. CCAM has also been applied successfully for various downscaling researches by using Schmidt transformation. Therefore, the new cubic-grid based coupled model will be a promising tool for the regional climate change researches with downscaling technique.

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# COORDINATED OCEAN-ICE REFERENCE EXPERIMENTS WITH THE AUSTRALIAN CLIMATE OCEAN MODEL

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# Abstract

The Australian Climate Ocean Model (AusCOM) is the ocean and sea-ice component of the Australian Community Climate and Earth System Simulator (ACCESS). AusCOM comprises the U.S. Geophysical Fluid Dynamics Laboratory Modular Ocean Model (MOM4p1), the U.S. Los Alamos National Laboratory Sea Ice Model (CICE4), and the French coupler system (OASIS3.2.5). Development of AusCOM is benefitting from participation in Coordinated Ocean-ice Reference Experiments (COREs). The CORE experiments conform to protocols adopted by the Climate Variability and Predictability (CLIVAR) Working Group on Ocean Model Development (WGOMD). The AusCOM model has been compared to the ocean and sea ice simulations of a number of IPCC-class models over the course of a 500 year benchmark simulation using a climatological atmospheric forcing dataset. Further experiments are in progress using interannually variable forcing over the period 1948-2007.

# Introduction

The Australian Climate Ocean Model (AusCOM; Bi and Marsland, 2010) is the ocean and sea-ice component of the Australian Community Climate and Earth System Simulator (ACCESS). AusCOM comprises the U.S. Geophysical Fluid Dynamics Laboratory Modular Ocean Model (MOM4p1), the U.S. Los Alamos National Laboratory Sea Ice Model (CICE4), and the French coupler system (OASIS3.2.5).

We conduct a Coordinated Ocean-ice Reference Experiment (CORE) and compare with previous efforts using six IPCC-class ocean and sea ice model components, as presented by Griffies et al. (2009). The CORE Normal Year Forcing (CORE-NYF) experiment is a multi-national effort endorsed by the World Climate Research Program Climate Variability and Predictability (CLIVAR) Working Group on Ocean Model Development (WGOMD). The CLIVAR/WGOMD CORE-NYF experiment uses an atmospheric forcing dataset that has been developed by scientists at the U.S. National Center for Atmospheric Research and is known as the CORE forcing (Large and Yeager, 2004; 2009).

The CORE-NYF is a composite of NCEP/NCAR Reanalysis data, bias-corrected reanalysis data, and observational data. It provides a climatological mean year forcing that incorporates realistic and self-consistent 6-hourly variability for air temperature, zonal and meridional wind speeds, specific humidity, and sea level pressure. The radiative heat fluxes (downward longwave, and shortwave fluxes) have daily temporal variability. The precipitative fluxes (rain and snow) have daily variability. Some further information on the CORE forcing can be found at the CLIVAR/WGOMD web site:

http://www.clivar.org/organization/wgomd/core/core.php

or the GFDL data distribution site:

http://data1.gfdl.noaa.gov/nomads/forms/mom4/COREv2.html

# Results



*Fig. 1* Adapted from Griffies et al. (2009), but including the AusCOM model. Time series of the globally averaged annual mean ocean temperature (left, degrees Celsius) and salinity (right, psu) from the CORE-I simulations.

Figure 1 shows the evolution of global mean potential temperature and salinity over the 500 years of the CORE-NYF experiment. The AusCOM model performance is similar to the other z-coordinate models used in Griffies et al. (2009). Further analysis and benchmarking considers the AusCOM simulation of the equatorial thermocline, the model's poleward heat transport and meridional overturning circulation, model drift of thermohaline fields at depth, mixed layer depths, and the seasonality of sea ice fields, amongst other diagnostics.

Overall, the model performs reasonably well when compared with the other 6 models that have participated in the CORE-NYF experiment. However, we find that the Antarctic Circumpolar Current is too strong, and that there is excessive open ocean convection in the Southern Ocean. These issues are being addressed through further experimentation related to tuning parameters associated with isopycnal mixing via a Gent and McWilliams (1990) style subgridscale eddy parameterisation. The lessons learnt from AusCOM participation in the CORE-NYF experiment are also further aiding development of the ACCESS modeling system.

AusCOM is also used in CORE Interannual Forcing experiments (CORE-IAF). CORE-IAF experiments are run with time varying atmospheric forcing over the period 1948-2007. Preliminary results from AusCOM CORE-IAF simulations will be presented at the workshop.

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## ASSESSMENT OF THE ACCESS COUPLED MODEL CLIMATE SIMULATIONS FOR CMIP5 AND IPCC AR5

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## Abstract

The Australian Community Climate and Earth System Simulator (ACCESS) will provide Australia with the next generation capability for weather prediction and climate/climate change simulation. The ACCESS coupled model was successfully assembled in mid-2009 and has been fully functional since early 2010. In the past 10 months or so, significant progress has been made so that the ACCESS model skill scores for the recent century-length runs surpass those for the preceding CSIRO Mk3 model delivered to IPCC AR4. A range of CAWCR scientists have been performing detailed testing to further improve the ACCESS model performance for the final AR5 version, and this paper provides preliminary analyses of the recent ACCESS control runs for assessing the suitability of this model to contribute to the Coupled Model Inter-comparison Project phase 5 (CMIP5) and IPCC AR5.

### Introduction

The ACCESS coupled model is built by coupling the UK Met Office atmospheric model UM (Unified Model, Dando 2004), and other sub-models as required, to the Australian Climate Ocean Model (AusCOM, Bi and Marsland 2010), an IPCC-class global coupled ocean and sea ice model consisting of the GFDL MOM4p1 ocean model (Griffies et al. 2004) and the Los Alamos CICE4.0 sea ice model (Hunke and Lipscomb 2008), under the PRISM 2-5 OASIS3 coupler (Valcke 2006) framework, as presented in Fig. 1. A set of century-scale control simulations have been performed with the current configuration of the ACCESS coupled model. These test simulations differ from one another in terms of ocean-atmosphere coupling frequency, number of vertical levels in the ocean, treatment of different types of sea ice albedo, treatment of certain cloud types and sub-grid scale cloud inhomogenity, and ocean model neutral physics parameterisations. This assessment will feature results from two recent (present day climate) simulations, which are judged overall to be superior to the previous experiments in terms of variety of criteria, especially the size of global average sea surface temperature (SST) bias and drift, and the overall skill in simulating a selection of key climatic fields (both globally and over Australia). For the purpose of this assessment, these two simulations are denoted "Base-12" and "Base-12c", which differ in the cloud inhomogeneous scaling factor (0.80 and 0.85, respectively, used to scale down the grid box mean cloud liquid water content to account for the effects of the sub-grid scale cloud inhomogenity) and sea ice albedo (Base-12 uses 0.65 for melting ice, 0.56 for bare ice, whereas Base-12c uses 0.57 and 0.61 for these two types of ice).

### **Global SST drifts and biases**

Figure 2 shows the global average annual mean SST evolution for Base-12 and Base-12c. In both experiments, the global SST exhibits a dramatic coupling shock (cooling at the beginning of the run, especially in the first year) and takes a few decades to gradually recover (and consequently shows

remarkable 'climate drift', i.e. a trend of 'global warming'), with Base-12 overshooting but stabilising to a level above the observed climate in the second half of the simulation, and Base-12c remaining under the observed level until nearing the end of the run (after the North Atlantic convection is re-activated around the Labrador Sea region, not shown here). Examination of the SST evolution from the two runs reveals the significant impact of the cloud scaling parameter on the global ocean surface heat budget.



Fig. 1 ACCESS coupled model components and coupling framework.



*Fig. 2 Time series of global annual mean SST (°C) in the two ACCESS runs.* 

Distributions of annual mean SST biases for the last decade of the runs are presented in the top panel of Fig. 3. Most regions of the world ocean show a deviation (from observation) of SST of less than 1 °C, with the largest biases found in the Atlantic Ocean where large areas show remarkable cooling (as commonly seen in other coupled models), especially in the Labrador Sea, one of the most climate sensitive regions. In the tropical Pacific, a region of major climate importance, there is only mild cooling in both runs, especially Base-12 because of its overall warmer status. This is a substantial improvement over previous runs which had a large 'equatorial cold tongue bias' with maximum cooling over 2 °C, as commonly seen in coupled models. The bottom panel of Fig. 3 shows time series of SST over the two targeted regions. While the Tropical Pacific gets through a relatively rapid warming adjustment and stabilises after year 40 in both runs, the northern North Atlantic experiences dramatic change associated with both the reduction of northward heat transport (as partly indicated by the surface cooling seen at the equatorial and low latitude Atlantic), and the shut-off and re-activation of local deep convection (not shown).



*Fig. 3* Top panel: Annual mean SST biases for the last decade of the two simulations; Bottom panel: Time series of annual mean SST over Tropical Pacific (100-220°W, 5°S-5°N) and northern North Atlantic (30-50°W, 45-65°N).

# Simulation of key climatic fields

Figure 4 shows the skill scores of the coupled model in simulating a range of key climatic fields globally. The degree of skill score for each field is indicated by the height of the lower bar. The skill for the present Base-12 is compared to that for the CSIRO Mk3.5 (the previous model that contributed to CMIP3 and IPCC AR4) and to an alternative observational data set, so as to get an indication of assessment limitations based on uncertainty in the observations. ACCESS is more skilful in simulating all the fields than Mk3.5. The mean skill score across the 9 fields is 0.695, 0.736 and 0.785, respectively, for CSIRO Mk3.5, ACCESS Base-12 and between observational sets. Over Australia (not shown here), the mean skill score across the 9 fields is 0.625, 0.685 and 0.814, respectively, for Mk3.5, ACCESS and observations. The improved performance of the ACCESS coupled model in simulating the selection of global (and regional) fields over previous model versions is a very important criterion for acceptance of ACCESS as being ready for CMIP5 and AR5.

# Conclusion

Multiple century-scale climate simulations have been conducted using the ACCESS coupled model in the past months. Recent experiments yield solutions featuring generally reasonable global climate with minimal drift and skill in simulating a range of key climate indices superior to previous Australian models used in CMIP3 and IPCC AR4. Most aspects of the climate solution fall within the range anticipated for CMIP5 simulations. However, there are a range of model biases which needs further attention before the model will likely be strong in the field of CMIP5 and AR5 contributors. Testing and tuning towards model improvement which may mitigate these biases has been ongoing. We conclude that the model is quite close to being ready for performing climate and climate change simulations contributing to CMIP5 and AR5, while it will still be able to benefit from further improvement.



**Fig. 4** Model skill measures M (plotted from the bottom), and M' (from the top) for 9 global quantities from simulated and observed climatologies. In each case, the average of the four seasonal values is shown, with the sign of the mean bias (when consistent over the four) indicated above. Fields include surface air temperature (Ta), sea-level pressure (SLP), downward shortwave radiation at the ground (SW<sub>g</sub>), upward SW and LW at TOA (top of atmosphere), precipitation (Pr), evaporation, total cloud radiative forcing (C<sub>f</sub>), and total cloud cover (Ct). The primary observed data are from the ERA-Interim reanalyses (1989-2008), except that TOA radiative fluxes and cloud forcing are from observations. The second observed data are from ERA40 reanalyses (1958-2002), except that the surface temperatures from CRU (1961-1990), the pr field from GPCPV2.1 (1979-2008), the radiation at ground from ISCCP (1983-2000), and the radiation and cloud fields from the ERA-Interim data.

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# SEA ICE IN THE ACCESS MODEL

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## **Model configuration**

The ocean model of the Australian Community Climate and Earth System Simulator (ACCESS) is the Australian Climate Ocean Model (AusCOM). AusCOM comprises the OASIS3 coupler, the MOM4 ocean model and the U.S. Los Alamos national Laboratory CICE4 sea ice model (Bi and Marsland 2010). The ACCESS model is built by coupling the UK Met Office Unified Model (UM7.3) atmosphere, and other sub-models as required, to AusCOM, via the OASIS3 coupling framework.

CICE4 has the following interacting components (Hunke and Libscomb 2008):

- A thermodynamic model that computes local growth rates of ice due to vertical conductive, radiative and turbulent fluxes, along with snowfall on the surface.
- A model of ice dynamics, which predicts the velocity field of ice based on a model of the ice material strength and elastic viscous plastic ice rheology for the strain rate tensor.
- A remapping transport scheme that describes advection of the areal concentration, ice volume and other state variables,
- a ridging parameterisation that redistributes ice among thickness categories (the ACCESS model has five categories), and
- routines that prepare and execute data exchanges with an external flux coupler, which then passes the data to other climate model components such as MOM4 and UM7.3.

The UM7.3 boundary layer has an implicit solver that treats all surface types in a passive manner by calculating the surface temperature in the UM code. The air temperature, net heat from the atmosphere, latent heat and conductive heat at the surface are computed within the atmospheric model and then passed to CICE4. The surface heat fluxes for each ice thickness category are passed through the coupler and to the thermodynamic ice model. This coupling approach requires a "Zero-Layer" thermodynamic configuration, i.e. the sea ice temperature changes linearly from surface to bottom. In the original coupling approach, implemented in the CICE4 model, the ice thermodynamics and surface fluxes are determined implicitly in the ice model itself, because some fluxes depend on the state of the ice. This original approach is not, however, compatible with the UM atmospheric boundary layer approach. We aim to upgrade the thermodynamic code to a multi-layer thermodynamic model jointly with our Hadley centre colleagues but we cannot realistically change the UM boundary layer approach of controlling the turbulent fluxes and surface temperatures and need to adapt around it.

CICE4 is configured to share the AusCOM orthogonal curvilinear horizontal grid with MOM4 (Bi and Marsland 2010). The horizontal grid has a nominal resolution of one degree in both zonal and meridional directions. A singularity at the North Pole is avoided by using a tripolar grid, which also provides reasonably fine resolution in the Arctic Ocean enhancing the computational efficiency and accuracy. Additionally, the horizontal grid is refined over the Southern Ocean. The coupling between CICE4 and MOM4 is relatively straightforward due to identical horizontal grids because exchanged data need no rotation, remapping or grid-

point shifting. In the ACCESS model all coupling fields are delivered by OASIS3 into or from CICE4 and there is no direct coupling between the UM7.3 and MOM4.

## **Current performance**

The ACCESS model simulates sea ice fairly realistically when compared to climatologies derived from simulations completed by other GCMs, such as HadGEM3 and CCSM4, and from observations (Fig. 1). Depending on the coupled model configuration affecting processes related to, for example, clouds, oceanic mixing and ice surface energy balance, the results vary over multi-decadal simulations. The ACCESS model has a cold SST bias in the Northern hemisphere being strongest in the North Atlantic region and affects the ice edge position. There have been simulations where albedos have been adjusted to reduce this bias and the ice extent is closer to the observed but then the Arctic ice is too thin. When albedos are more realistic the modelled ice is thicker in the Arctic basin but the ice extent covers too large an area. Additionally, some errors in the sea ice simulations based on ACCESS are regional and likely associated with errors in regional processes simulated by the atmosphere and ocean models. These errors include:

- Sea ice is too thin in the Central Arctic, while it is significantly and realistically thicker against Greenland and the Canadian Archipelago. This could be caused by unrealistically persistent winds pushing the ice toward Greenland and the Canadian Arctic Archipelago.
- The atmosphere is too cold in the Canadian Arctic Archipelago in summer, which reduces the sea ice melt and the ice becomes too thick over the years growing up to its equilibrium thickness of 20-30 metres. This may be associated with a cold SST bias of the Labrador Sea and a persistent high SLP bias in the Central Arctic. The cold SST cools down the atmosphere locally and the surface air temperatures don't warm above the melting point over adjacent areas in the Canadian Arctic in summer. The high SLP bias over the Arctic in summer directs geostrophic winds in the Central Arctic towards the western Arctic and transports cold air to the Canadian Arctic.
- The absence of tidal mixing impacts the ice thickness which becomes thicker in the coastal regions and narrow straits, but thinner over some areas such as the Central Arctic. For example, the difference in ice thickness between 25 year model runs with and without tides can be of 50 cm. In the Antarctic, as in the Arctic, sea ice becomes thinner close to the coast and thicker off the coastal regions due to tides. Tidal currents redistribute the sea ice and can increase the ice generation in the coastal regions up to 5 cm/day.
- The Antarctic sea ice is too thin. This could be due to too strong oceanic convection, where the upwelling warm water melts the sea ice and leaves too little surviving the summer. Additionally, too weak katabatic winds and cold air outbreaks might reduce the ice generation in the coastal regions. There also appears to be too little internal deformation and local ridging within the pack ice, possibly due to winds which may not have sufficient variability in direction.

It should be noted that other GCMs display errors of similar magnitude and some of the errors could be reduced by finding a set of optimal ice model parameter values and/or increasing the model resolution. However many of the errors are linked to errors in the climatology of the atmospheric and ocean models and considerable effort to reduce these biases may be required.


*Fig. 1* Sea ice concentration for (a) Arctic and (b) Antarctic from an ACCESS model run averaged over 10-60 model years. Thick black lines in (a) and (b) indicate the mean observed 10% ice concentration isoline from 1979-2000.

# **Future plans**

The ACCESS modelling team is preparing simulations to be included in the CMIP5 model archive by the mid-2011. Assessment and validation of the sea ice results is a significant part of this effort. The sea ice model development will continue in collaboration with the Hadley Centre at the UK Met. Office. This will include the implementation of improved schemes for the computation of surface albedos and shortwave radiation, wind stress, the CICE ice thermodynamics option with multiple vertical layers, and frazil ice formation in leads and polynyas. The coupling method that is intended for multi-layer ice poses a number of problems, partly due to a delay in the time step between the atmosphere and ice calculations, and also due to a stability issue with thin ice and snow layers.

A new version of CICE4 was released this year and the sea ice component of ACCESS will be upgraded to the new version. Significant improvements can be expected in future versions of CICE4 including, for example, a more advanced scheme for the computation of the impact of melt ponds by Flocco and Feltham (2010), internal ice salinity profiles and ice biogeochemistry.

The ACCESS developers aim to optimise parameters related to ice dynamics and thermodynamics. These include parameters associated with surface albedos, sea ice conductivity, mechanical redistribution and iceocean dynamic stress, primarily its turning angle. The sea ice thickness is quite sensitive to these, and potentially to the values of other parameters (Hunke 2010).

Modelling challenges will arise as Arctic sea ice decreases with an accelerating trend and its mechanical strength reduces resulting in increasing ice speeds (Rothrock et al. 2003; Rampal et al. 2009; Roberts et al. 2010). The accelerating trend is likely to be caused by the positive ice-albedo feedback, where decreasing sea ice cover leads to the warming of the upper ocean and lower atmosphere, reducing further sea ice. The increasing heat content of the upper ocean also reduces the volume of ice formed in subsequent seasons. The melting Arctic sea ice increases freshwater being transported from the Arctic to the North Atlantic, where the active convection driving the ocean thermohaline circulation takes place. Accordingly changes in the Arctic sea ice have global implications, which are yet to be fully understood and difficult to predict due to the complex processes involved, but some teleconnections across the Northern hemisphere have been observed post 2007 (Serreze 2009). Global climate models do not currently simulate realistically enough feedbacks

and interactions in the Arctic region. The failure of the CMIP3 models to simulate the dramatic decline in ice over the last few years since 2007 demonstrates this shortcoming (Wang and Overland 2009).

Continental shelves and coastal regions are important areas of atmosphere-ice-ocean exchanges, where the Arctic sea ice freezes and melts, changes the stratification of the upper ocean and impacts the overall circulation of the Arctic Ocean. Ocean currents form an essential part in this circulation by transporting heat and freshwater into and within the Arctic Ocean and have been shown to be difficult to model in low resolution coupled models (Uotila et al. 2006).

Changes related to sea ice, as already observed in the Arctic, can be expected to occur in the Antarctic region as the Southern Ocean warms in the coming decades (Liu and Curry 2010) increasing the meridional transport of heat and moisture (Uotila et al. 2007), where especially the role of mesoscale cyclones and their interaction with sea ice remains relatively unknown (Uotila et al. 2009). Observed regional and seasonal trends of the Antarctic sea ice can be attributed to the global change (Turner et al. 2009), although the surface warming comparable to the Arctic warming has been so far limited to the western Antarctic Peninsula.

The Australian Antarctic Division has published the *Australian Antarctic Science Strategic Plan 2011-12 to 2020-21*, which particularly addresses the need to improve the understanding of the role of Antarctica and the Southern Ocean in the global climate system. The plan introduces important pathways to enhance the performance of the sea ice component of ACCESS by targeted field campaigns within the sea ice zone, followed by process studies and improved numerical modelling of the atmosphere-ice-ocean system. In addition to the climate modelling the enhanced modelling capabilities will benefit the NWP service. Studies supported by the strategic plan will potentially include topics such as sea ice data assimilation in a NWP model, modelling of the fast ice and linking the sea ice physics with the research on Antarctic biological production. In summary, these activities promise an exciting decade for the sea ice modelling in Australia.

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## MODELLING THE INTERACTION BETWEEN ANTARCTICA AND THE SOUTHERN OCEAN

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#### Introduction

The increased loss of ice from the coastal margins of Antarctica (e.g. Wingham et al. 2006) is mostly due to the rapid thinning and retreat of ice streams, driven by the enhanced melting of the ice shelves that fringe the continent (e.g. Shepherd et al. 2004). The enhanced supply of cool and fresh glacial meltwater is the most likely cause of observed changes in the dense water that feeds the global overturning circulation (e.g. Rintoul, 2007). Understanding interactions between the ice sheet and oceans is essential for improved projections of the impacts of climate change on sea level rise and ocean heat and freshwater budgets.

Ice shelves form when large ice streams that reach the ocean become ungrounded and begin to float. Ice shelves are maintained in a dynamic equilibrium between the addition of ice from the inflowing glaciers and net snowfall on the upper surface and loss of ice due to melting that occurs at their base and iceberg calving from their front (Fig 1.b). The ice flux from Antarctica into the ocean is strongly influenced by the mass loss processes occuring beneath the ice shelves. The basal mass loss is determined by ocean processes that occur on the continental shelf (Fig 1.a). An inflow of Circumpolar Deep Water (CDW) can mix with the dense water product of sea-ice formation, such as High Salinity Shelf Water (HSSW) (solid curved lines), which can then sink down the continental shelf and melt the ice sheet at the grounding zone.



Fig. 1 (a) Important Antarctic coastal ocean processes. (b) Ice shelf mass balance.

Buoyant freshwater that is released during the melting process rises along the underside of the ice shelf as Ice Shelf Water (ISW) and can become locally supercooled at a shallower depth, leading to the formation of frazil ice (shown by the dots) and basal accretion of marine ice. The water created by the re-freezing process is analogous to that created by sea-ice formation (dashed curved lines). These processes are also important for deep water formation processes, which occur in the shelf seas around Antarctica and ventilate the abyssal oceans, such as Antarctic Bottom Water (AABW).

Currently we do not yet fully understand the processes that link ocean warming to ice-shelf melting and the retreat of ice streams and dense water formation. The most recent report from the *IPCC* (2007) AR4 highlights that the uncertainty in projections of future sea level rise are dominated by uncertainty concerning continental ice; many global models do not include glacial-ice/ocean interaction processes. Modelling studies have shown that ignoring ice shelves and associated freshwater fluxes leads underestimates sea ice thickness and increases bottom water formation and overturning circulation (*Hellmer*, 2004). Also, most global ocean models poorly resolve the continental shelf surrounding the Antarctic continent (such as polynyas and canyons). Recent modelling results highlight the importance of small scale features, such as grounded icebergs that aid the formation of polynyas, leading to enhanced dense water production (*Kusahara et al.* 2010). Quantifying the nature, rapidity and extent of the response of the ice sheets to climate change requires the development of our capacity to model Antarctic coastal processes.

# Current modelling at the ACE CRC

Research at the Antarctic Climate and Ecosystems Cooperative Research Centre (ACE CRC) will improve our understanding of the interaction between Antarctica and the ocean with the use of a high resolution model that incorporates the important ocean processes shown in Fig.1 a. The model horizontal grid has a resolution which varies from \_2.5 km at the south to \_7.5 km at the northern boundary located at 50<sub>o</sub>S (see Fig. 2) which will allow us to compare observations and predictions at decadal and seasonal scales. The grid was developed for future climate change projections that can examine the retreat of the Antarctic ice sheet.



*Fig. 2* Southern Ocean bathymetry and Antarctic bedrock topography (Timmermann et al. 2010) overlain with the models horizontal grid outlining groups of 20x20 cells (grey lines).



Fig. 3 Mid-winter snapshot of the surface ice growth rate in units of meters per year, where blue is ice melt and red is ice growth. The continental shelf break is indicated by the 1000 m bathymetry contour (thin black line). The ice shelf melt (blue) is calculated by the model and the sea ice growth (red) is prescribed – seen in small polynyas on the continental shelf and in the open ocean.

The model is based on the Regional Ocean Modeling System (ROMS; Shchepetkin andMcWilliams, 2005), which is able to simulate interactions between flows and topography, such as the important buoyant plumes of ISW that can rise under ice shelves and dense plumes of AABW that can flow down the continental shelf to the deep oceans. ROMS has been customised to include ice-shelf/ocean processes, frazil ice dynamics and tides. The model was initially developed for regional modelling studies and is able to simulate ice/ocean interactions at a high level of realism. The model is forced at the surface with winds and an imposed sea ice cover. The sea ice conditions are derived from Special Sensor Microwave Imager (SSM/I) observations (e.g. see the positive flux in Fig 3; Tamura et al. 2008) that include many important subgridscale features such as grounded icebergs and polynyas. Thermohaline and circulation fields are relaxed to smoothly varying reanalysis fields from ECCO2 at the northern boundary.

# **Future modelling**

Many of the problems facing the global ocean modelling community can be overcome by using results from the high resolution ocean model described here. The model has been developed to simulate the under ice shelf circulation and dynamics by parametrising the melting and freezing. The inclusion of frazil ice dynamics and glacial ice/ocean processes extends the capabilities of the model to correctly represent heat and mass transport into the ocean. However, the assumption made for these simulations is that the ice shelf is in steady state, reacting instantaneously to melting and freezing. For present day simulations, this is generally considered to be a reasonable assumption, since simplified studies suggest minimal change to the ice shelf draft in the short-term (Walker and Holland, 2007). The effects of basal melting and freezing on ice stream stability, over the longer timescales or under rapid retreat requires a different approach. Existing climate models suggest that climate warming would result in increased melting from coastal regions in Antarctica and an overall increase in snowfall. These models are also incapable of realistically simulating the outlet glaciers that discharge ice into the ocean and cannot predict the substantial acceleration already observed for some outlet glaciers.

Under the steady state assumption, present day and short term future changes to oceanic freshwater and heat budgets can be studied with relative ease. The results of the model shown here can be of benefit to both ice sheet and ocean modellers. Ice sheet models can use the basal fluxes beneath ice shelves (negative surface fluxes in Fig.3) and global ocean models can use vertical profiles of freshwater and heat fluxes near the Antarctic continent. Steps have already been made in this direction with the ACCESS ocean modelling group to use results from the model outlined here (*Marsland, pers comms.*). The modelling effort required to investigate rapid changes in ice streams and their impact on sea level is much larger. Several factors determine the position of the grounding line, and thus the stability of ice sheets. Of greatest concern are regions where the ice sheet is grounded below sea level (green areas inside the grounding line in Fig. 2). Future ocean models will need to respond to changes in processes that control the mass balance of ice shelves (arrows in Fig 1.b). To this end, work at the ACE CRC will begin by asynchronously coupling the ocean model shown here with an ice shelf model. This research aims to fill the gap in our current understanding of ice-ocean interactions and will help to provide realistic estimates of freshwater budgets and sea level rise in a global context.

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### **DEEP OVERFLOWS**

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## Introduction

Modification of dense, sinking waters in the oceanic bottom boundary layer plays an important role in setting the properties of many of the world's deep and bottom waters (Killworth 1983). Termed "overflows", the sinking waters initially form in reservoirs by brine rejection, evaporation or cooling. The reservoirs are generally marginal seas or coastal shelves. Dense water leaving the reservoir generally flows over some topographic feature such as a sill or strait, before going down the continental slope. Topographic steering can play a significant role in influencing properties, for instance a subsurface canyon can act as a conduit for dense water to rapidly descend. As waters descend they are often also significantly modified by entrainment of water from the ambient ocean (Smith 1975, Price and Baringer 1994). In recent years, there has been considerable effort put into improving our understanding of overflow processes (Legg et al. 2009).

From a utilitarian perspective, we are motivated to study overflows due to their climatic importance. Knowledge of the life cycle of deep and bottom waters is critical to answering many pertinent questions, such as ocean uptake of carbon dioxide, sea level rise and the stability of the meridional overturning circulation. The impact of overflows on these, and other oceanic properties provides strong motivation for the accurate representation of overflows in ocean climate models, particularly in the context of altered radiative forcing.



*Fig.* **1** A schematic of some of the physical processes affecting overflows around the Antarctic margins. From Baines and Condie (1998).

There are many examples of overflows throughout the world, including several in the Antarctic margins, the high latitude North Atlantic, the Mediterranean Sea, the Red Sea and numerous mid-ocean ridges, amongst others (Legg et al. 2009).

# The physics of overflows

The source waters of overflows are dense water in a marginal sea, on a shelf and/or behind a shallow sill. The dense water flows down the continental slope until it reaches its neutral depth or the bottom of the abyssal ocean. There are a great many processes that affect the pathway and properties of overflowing waters, including sill depth, Ekman drainage, rotation, baroclinic instability, bottom friction, shear instability and topographic steering (Smith 1975, Killworth 1977, Price and Baringer 1994, Özgökmen and Fischer 2008). A few of these processes are illustrated, for the Antarctic margin, in Fig. 1. There has also been recent process studies on the interaction of density currents with topographic features, such as ridges and canyons – processes which are particularly of interest around the Antarctic margins.

#### The parameterisation of overflows in ocean models

Typically, in global scale climate models, the sills are unresolved and the overflow waters are themselves are subgridscale. There have been a number of attempts to parameterise overflows in ocean climate models. Level coordinate models (i.e. models with a vertical coordinate system based on depth or pressure) have difficulty representing overflow processes until there are several grid points within the boundary layer and the horizontal resolution becomes  $\Delta x \approx \Delta z/|\partial_x H|$  where  $\Delta x$  is the horizontal grid spacing,  $\Delta z$  is the vertical grid spacing and  $\partial_x H$  is the gradient of the topography (Winton et al. 1998). The difficulty arises due to the steplike

nature of the topography. Shelf water moves horizontally to a deep ocean column, causing a vertical instability. The vertical convection parameterisation of the model then acts to reduce the instability via a vertical adjustment of tracer properties to reduce or eliminate the instability in the water column, as is shown diagrammatically in Fig. 2. As a result, source water properties are rapidly diluted.



Fig. 2 Overflows in level ocean models.

Campin and Goosse (1999) were mostly concerned with limiting excessive spurious mixing of overflow waters in level models, rather than being a physically based parameterisation. Efforts for more physically based parameterisations have been hampered by numerics. Slab bottom boundary layer schemes, such as that of Killworth and Edwards (1999), while being much more physically based have not been widely taken up by the ocean modelling community due to a number of significant drawbacks. Layer models, in which the vertical coordinate is based on an isopycnal coordinate, have much less difficulty representing overflow waters, as it is the diapycnal flux which is specified (Legg et al. 2006). A parameterisation for a layered model is a much more simple task, as it is essentially a parameterisation of entrainment or detrainment of the boundary layer. This natural way of representing overflows in ocean climate models has been a one of the motivating factors in the development of layered and hybrid coordinate ocean models.

# Overflows in the present generation of ocean climate models

Up until recently, at least in level ocean models, there has been little improvement in the overflow parameterisations in the last decade. As a result, most ocean climate models either do not have a parameterisation for overflows, or they use one of the parameterisations mentioned above. Some recent developments are being incorporated in the present generation of ocean climate models. Modifications to boundary conditions to represent subgridscale straits and channels (e.g. the Strait of Gibraltar) include cross land mixing, as well as thin and partial barriers. Having straits that are artificially widened, or completely closed spuriously affects the properties of the source waters. Methods for improving the topographic represent undersea canyons and other subsurface topographic features, such as sea mounts. Another approach which has been implemented in the CCSM4 for the Denmark Strait (Danabasoglu, et al 2010), the Mediterranean and the Weddell and Ross Seas (Briegleb et al 2010). The approach is to define a subsurface source region and a number of predefined subsurface injection locations. Water is taken from the source region and instantaneously transported to the injection location. The water may be modified based on the

properties of a predefined input region and entrainment region, which are intermediate regions between the source region and the injection locations. The amount of source water transported from the source region, how much the water is modified and the choice of injection location, where the product water is placed, are all based on a physical model and calculated. The procedure requires the modification (raising) of topography in order to stop the grid scale model from carrying out the procedure shown in Fig 2.



*Fig. 3* Cross section of mean zonal velocity (cm/s) along the continental slope across 69W in an uncoupled OGCM. Left: without the Nordic Seas overflows parameterised, Right: with the Nordic Seas overflows parameterised. From Danabasoglu et al. (2010).

The implementation of the above parameterisation in the Nordic Sea overflows has had important non-local effects when implemented in the CCSM. The largest effect has been the removal of a spurious poleward deep western boundary current below about 2600m (Fig. 3), which almost universally improves the properties of the North Atlantic (Danabasoglu et al 2010).

It is hypothesised that significant improvements in the overflows of the Antarctic margins should also improve Antarctic bottom water (AABW) properties. Unrealistic AABW formation rates and incorrect AABW properties are ubiquitous in the present generation of ocean and coupled climate models. As an example, Downes et al (2010) examines the AABW production in different configurations of the GFDL coupled models, and ECCO products. They show for transport across 30°S, the highest transport of the models examined is less than half the lowest AABW transport estimates from observations.

# The future of overflows in ocean climate models

Recent studies provide ample motivation for the community to invest in the improvement of the representation of overflows in ocean climate models. Layered models provide one avenue with which to overcome the shortcomings of the representation in the traditional level models. The maturation of layered models means that it is no longer impractical to use them in realistic climate simulations. While there will probably be more modelling groups using layered models for realistic climate simulations over the next 5-10 years, there remain many advantages to using level coordinate models. As such, a significant proportion of the models used for realistic climate simulations will still be using level coordinate models.

The overflow parameterisation of Briegleb et al (2010) has the potential for the inclusion of more detailed physics. Its dramatic improvement in the properties of the North Atlantic has already motivated other model development groups to implement equivalent schemes. A fundamental drawback of their approach is, however, that each overflow must be prescribed and tuned. With the inclusion of better physics there is scope to further improve simulations of the present climate, however, it stands to reason that one must be very cautious when drawing conclusions about the response of overflows to altered forcing.

Another approach, which is being developed at UNSW and GFDL, is the embedding of a Lagrangian model to represent overflows in level models. The Lagrangian "blobs" transport material properties, are formed and destroyed dynamically and obey a different, more appropriate set of physics, to the processes being represented. The exchange of properties between the Eulerian and Lagrangian models (to representent entrainment and detrainment) is given by some parameterisation. The Lagrangian model theoretically admits a finer resolution topography than that resolved by the Eulerian model, so, it offers the possibility of being able to have gravity currents "feeling" finer scale topography than that being used by the Eulerian model. While the embedded Lagrangian model is still highly developmental, its first principles formulation gives much promise of being able to provide a more realistic response to altered forcing than what is presently available in level models.

The embedded Lagrangian model presents many technical challenges to ocean model developers, however, it is anticipated that the prototype algorithms being developed at present will, over the course of the next 5-10 years, be improved in terms of the realism of entrainment/detrainment parameterisations as well as the efficiency, robustness and other performance issues.

As the overflow parameterisation and the embedded Lagrangian model mature, these and perhaps other approaches to handling overflows in level ocean climate models will become virtually mandatory over the next 5-10 years for any serious large scale model study of the climate system.

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# THE IMPACT OF OCEANIC VERTICAL MIXING SCHEMES ON ACCESS CLIMATE SIMULATIONS

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A partnership between CSIRO and the Bureau of Meteorology

Oceanic mixing plays a critical role in controlling the exchanges of energy and momentum between the ocean and atmosphere. Here we apply three vertical mixing schemes and investigate how and where they impact the mean climate state and interannual variability in the Australian Community Climate and Earth-System Simulator (ACCESS) - a state-of-the-art atmosphere-ocean-land coupled system. Non-local Kprofile parameterization (KPP; Large et al., 1994), Chen (Chen et al. 1994) and integer power (IP; Wilson 2000) mixing schemes (the latter being local gradient Richardson number based) are investigated in long-term climate runs of the ACCESS model. Our results show that the model sea surface temperatures (SSTs) are closely associated with the choice of mixing schemes. IP makes the ACCESS model produce too high global average SST compared to observations. KPP and Chen schemes, in contrast, produce more realistic globally averaged SST. Compared to the more widely used KPP scheme, the Chen scheme has a smaller cold tongue bias in the equatorial Pacific as well as better temperature simulations in the middle and high-latitude North Atlantic regions. The differences of SST distributions between KPP and Chen also influence the model's precipitation and sea surface salinity (SSS). The equatorial Undercurrents (EUCs) in both KPP and Chen schemes are weaker than observations whose maximum amplitude is about 1.0m/s. However, Chen's EUC whose core is about 0.7m/s is 0.1m/s higher than KPP's. Both KPP and Chen schemes produce too strong interannual variability in the tropical Pacific. However, the Chen scheme's climate variability is a little weaker than KPP's and thus closer to observed.

# Introduction

ACCESS is a coupled climate and earth system simulator that will deliver a new generation of numerical models to improve weather and climate research in Australia. Its climate simulation results will contribute to the Intergovernmental Panel on Climate Change (IPCC) 5th Assessment Report. ACCESS will be served as seamless multi-time scale predictions of the climate system ranging from intraseasonal, seasonal to decadal predictions. So far most atmosphere-ocean-land coupling work for the ACCESS model has been done. At the present stage, one key issue is how to improve ACCESS performance in climate simulations by tuning of various physical parameterisations in both the atmosphere and ocean models. Here we report how the oceanic vertical mixing schemes impact ACCESS model simulations.

The oceanic vertical mixing plays a significant role in climate model simulations, but it cannot be explicitly resolved in models and has to be parameterized due to small-scale turbulent process. To date many mixing schemes have been proposed (Bryan and Lewis 1979; Mellor and Yamada 1982; Large et al. 1994; Chen et al. 1994, Wilson 2000 etc). Here we only explore KPP, Chen and IP mixing schemes and investigate their performance in the ACCESS model. The KPP scheme explicitly predicts an ocean boundary layer depth and

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the turbulent mixing in this boundary layer is parameterised using a nonlocal bulk Richardson number and the similarity theory of turbulence. The Chen scheme is a hybrid scheme that combines a homogeneous mixed layer model with Price's dynamical instability model (Price et al. 1986). The IP scheme uses integer power approximation to hourly observations and is a function of the gradient Richardson number.

# **Model description**

ACCESS comprises three components. They are the atmospheric component UK Unified Model (UM 7.3), the Australian Climate Ocean Model (AusCOM; Bi and Marsland 2010), and the land component CABLE 1.4 (Kowalczyk et al. 2006). Presently, we still use the land model MOSES that is built into the UM model while CABLE coupling nears completion. The coupler is PRISM-2-5 OASIS3 (OASIS3.25) (Valcke 2006). The ocean-atmosphere coupling interval is 3 hours.

The UM7.3 atmospheric model applies HadGEM3 (Hadley Centre Global Environment Model version 3) configuration (<u>http://www.metoffice.gov.uk/research/modelling-systems/unified-model/climate-models/hadgem3</u>). Its resolution is N96L38. The cloud scheme is the PC2 scheme (prognostic cloud fraction and prognostic condensate). The cloud inhomogeneous scaling parameter of 0.8 is applied to stratiform water clouds everywhere except 0.5 in the equatorial Pacific region bounded by 5°S to 5°N from 130°E to 120°W<sup>†</sup>

AusCOM is a coupled ocean-sea ice system. The ocean component is MOM4p1 (Griffies 2009) from the U.S. Geophysical Fluid Dynamics Laboratory (GFDL) and the sea ice model is CICE4.0 from the U.S. Los Alamos National Laboratory (Hunke and Lipscomb 2008). The horizontal mesh in the ocean model is 360x300. The model is nominally one degree resolution, with latitudinal enhancement to 1/3° from 10°S to 10°N. A Mercator grid is used in the Southern Ocean with 1° latitudinal resolution at 30°S reducing to 0.25° at 78°S. In the Arctic a tripolar grid is used. The vertical discretisation has 50 levels covering 0-6000 m with a 10m-resolution from the surface to 200m.

## Results

## Annual mean state

The ACCESS model has run for 85 years with KPP and Chen vertical mixing schemes and 73 years with IP schemes. Here we investigate the model long-term mean states and compare them with observations. Figure 1 shows the time evolutions of global average sea surface temperature (SST) with different vertical mixing schemes. All schemes have SST drifts over the first 30 years and then seem stable after that. IP scheme (green line) produces much higher global average SST than KPP (red line) and Chen (black line) schemes, and also much higher than the model initial global average SST (18.28°C) after the first 10-years of model integration. So we will not analyse IP results further in this study. The global average SSTs in KPP and Chen schemes are approximate 0.3°C higher than the initial value after 30-year integration, but they are still acceptable. The KPP scheme has larger variations of global average SST than the Chen scheme.

<sup>&</sup>lt;sup>†</sup> The inhomogeneous parameter is defined to account for the subgrid-scale cloud variability. The cloud scheme in coupled model provides grid mean values of cloud liquid/ice water content. The radiative impact of clouds, however, is sensitive to the variance of the cloud liquid/ice content rather than the mean values. Cahalan et al. (1994) have approved the following relationship,

 $<sup>\</sup>overline{\log(W)} = \log(s\overline{W})$ 

where  $0 \le s \le 1$  is the scaling parameter, which has both diurnal and spatial variations. Shok et al. (2010) study shows that *S* tends to be smaller for convective clouds than for stratiform clouds. Since the convective clouds are dominant over the equatorial Pacific region, a small value of s may be necessary to adequately account for the inhomogeneous effect. Please note that the equatorial Pacific inhomogeneous setting in this study is only for scientific research and will not be used in simulations for IPCC AR5.



Fig.1 The time evolution of global average SST with IP (green), KPP (red) and Chen (black) schemes

Figure 2 displays the annual mean SSTs and their errors with Chen Scheme, SST errors with KPP scheme and the SST differences between Chen and KPP schemes. A problem that the cold tongue in the equatorial Pacific extends too far west in the ACCESS model is considerably improved by the choice of smaller inhomogeneous parameter (not shown). The Chen scheme looks a little better than the KPP scheme with regards to the equatorial cold tongue bias. Large SST differences between Chen and KPP occur in middle and high latitudes of the northern hemisphere. In particular, the Chen scheme produces a smaller cold bias in the North Atlantic.



*Fig.2* The long-term annual mean SST with Chen scheme (a), SST errors with Chen scheme (b), SST errors with KPP scheme (c) and the SST difference between Chen and KPP scheme (d).



*Fig.3* The SSTA standard deviation with Chen scheme (a) and the difference between Chen and KPP (b).



Fig.4 The time series of Nino3 SSTA with Chen (black) and KPP (red) schemes.

## Interannual variability

The first 20-years of the model results are not used for the analysis of model climate variability due to spin up. Figure 3 shows that both Chen and KPP produce realistic interannual variability in the tropical Pacific. However, the standard deviations of sea surface anomalies (SSTAs) are a little larger than observed in this region. The SSTA standard deviations in the Chen scheme are weaker than those in KPP scheme (Fig.3b), and thus they are closer to observations. The time series of Nino3 (5°N-5°S, 90°W-150°W) index (Fig.4) in both schemes looks similar, but the KPP scheme (red line) has a larger amplitude for La Nina events (e.g., sometime it even reaches -4°C and that has never happened in the observational records) than Chen scheme (black line).

# **Discussion and summary**

We have investigated three oceanic vertical mixing schemes in the ACCESS model and find that the IP scheme produces too high global average SST. The Chen scheme is better than the KPP scheme in two important aspects: a smaller cold tongue bias and more realistic SSTA standard deviation in the tropical Pacific: and a smaller cold bias in the middle and high-latitude North Atlantic. We will tune the Chen scheme further as a next step. In addition, tidal mixing is not accounted for in this study yet. It should have a positive impact on the coastal regions and sea ice simulations in high-latitude regions.

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# REGIONAL INTERANNUAL-TO-DECADAL SEA LEVEL VARITIONS IN THE WESTERN TROPICAL PACIFIC REVEALED BY ADJOINT SENSITIVITY ANALYSIS

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The global mean sea level (SL) has been rising for the past several decades mainly due to global ocean thermal expansion (e.g. Bindoff et al. 2007). However SL variations are not globally uniform and there are also significant regional SL variations relative to the global mean. Some coastal regions and islands may be especially vulnerable to regional SL changes.

In this study, the regional SL variations in the western tropical Pacific on interannual to decadal time scales are examined using the MIT General Circulation Model (MITgcm, Marshall et al. 1997) and its adjoint model. Generated by automatic differentiation technique (Heimbach et al. 2005), the adjoint model calculates sensitivity (i.e. partial derivative) of the cost function J with respect to control variable  $v (\partial J / \partial v)$ . The model dynamics is contained in the chain rule of partial derivatives. The control variables can be model state, forcing fields, model parameters, initial and open boundary conditions. Here the cost function J is chosen as regional SL in a defined region in the western tropical Pacific. The adjoint model calculates the sensitivities of above-defined SL to atmospheric forcing fields and ocean state going backwards in time.

The adjoint-derived sensitivity of regional SL to surface wind stress is dominated by wind stress patterns propagating eastward with increasing time lags. The propagation speed is faster than the theoretical first baroclinic mode's phase speed, but in good agreement with altimeter observation (Chelton and Schlax 1996). This temporal evolution of the sensitivity of regional SL to wind stress resembles wind-forced Rossby wave dynamics, suggesting it as main underlying mechanism. Averaging the monthly sensitivity to wind stress over a year shows a broad anticyclonic wind stress curl pattern east of cost function region.

The adjoint sensitivity can be compared to a response function or regression coefficient. Consequently the derived adjoint sensitivity can also be applied in the hindcasting scenario. The seasonal-to-decadal variations of regional SL in the western tropical Pacific can be hindcasted by multiplying adjoint sensitivity and time-lagged wind stress over the whole model domain and summing over time lags. The success of this hindcast experiment implies that surface wind stress is the determining factor for regional SL variations on seasonal to interannual time scales in the western tropical Pacific.

Similarly, effects of surface heat and freshwater fluxes can also be examined using the same adjoint sensitivity analysis, though their roles are found to be secondary to that of wind stress. This adjoint sensitivity analysis provides a method to separate SL variations due to various physical processes (such as wind-driven). Its potential application for both historical SL reconstruction and future SL projection will also be discussed.

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# **GLIDERS AS OBSERVING TOOLS**

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# Abstract

Autonomous underwater gliders have proliferated as observational tools over the past decade. To understand they can be used effectively to study climate and climate processes, it is important to know how they function and what limitations they face. Ultimately, gliders will be judged by how cost effective they are in observing the ocean. This review describes how gliders can be used to observe processes in different ocean environments distinguished by their eddy energy. New developments that extend potential glider operating regions to nearly the entire world ocean are described.

# Introduction

It is helpful to review the rapid technological changes that have led to glider development. Affordable access to the sea continues to be the principal impediment to making ocean observations. Historically, most of what oceanographers know about the ocean has come from ship-based observations. Collecting samples from instruments on winch-lowered cables has been the principal means of probing the ocean since the beginning of oceanography. A number of automated means to observe the ocean began development in the last half of the 20<sup>th</sup> century. Three waves of instrumentation development have changed the nature of ocean observation immensely. First, moored platforms made possible the collection of time series, a breakthrough that enabled estimation of frequency spectra of oceanic variables. While still reliant on ships to install and recover them, moorings considerably reduced the labor of data collection by employing the developments of low power electronics. The development of satellite remote sensing produced the second huge advance in ocean observation. Despite being limited to measuring the very near surface ocean, the spatial character of oceanic processes illuminated by satellite observations was a considerable breakthrough. However, the need to observe the ocean interior with finer lateral resolution than is feasible with moorings remained. Autonomous platforms are the third wave of ocean technology that is transforming how the ocean is observed. The development of floats, initially intended to provide a Lagrangian description of ocean processes, has enabled observation of the upper ocean interior regularly and continuously over nearly the entire world ocean at scales comparable to the separation between satellite orbit tracks. The hybridization of platforms that drift along a depth or density horizon with vertical profilers floats led to the profiling float, the centerpiece of the global Argo array. A principal limitation of floats, though, is their inability to navigate deliberately.

Autonomous underwater gliders are steerable profiling floats, the use of which is our current subject. Two key technical developments from outside oceanography are key to their development and continued use: GPS navigation and low-power satellite communications. Without them, accurate location and near real-time retrieval of data and the ability to globally control gliders would not be possible. Iridium has emerged as the standard system for glider communication.

# Functionality, performance, and cost

Underwater gliders combine the desire to collect oceanic profiles with that of controlling their geographic location. Gliding efficiently accomplishes traverse both vertically and laterally through the ocean. That is, vehicle buoyancy is used both to supply the potential energy necessary to climb across the pycnocline and to balance vehicle hydrodynamic drag and lift, converting a purely vertical force to forces with both vertical

and horizontal components. Generally, a glider must supply energy to overcome both the potential density gradient of oceanic stratification and any mismatch in compressibility between its hull and seawater. Most gliders operate with a fixed energy supply for a mission in the form of batteries. The amount of energy and its usage determine mission endurance and range as well as the amount of data recorded and transmitted. While conventional autonomous underwater vehicles typically have endurance and range measured in hours and tens of kilometers, underwater glider endurance and range is measured in months and thousands of kilometers.

Underwater gliders are attractive for their high endurance and long range, but these attributes come via limitation to low speed. Energy supplied for propulsion ultimately is dissipated through hydrodynamic drag. Since drag laws are typically quadratic, halving speed quadruples endurance while doubling range. Gliders intended to maximize endurance and range operate only fast enough to navigate through ocean currents. Today's operational gliders are unable to match the speed of the fastest ocean currents, so alternate strategies are necessary. Since currents are typically surface intensified, the strategy of diving deeply is often effective. The average current over the depth range of a glider dive is the relevant metric. Typical gliders are designed to operate at ~0.5 kt ~ 0.5 m/s.

Low drag is important, since for a given speed and glide slope, the buoyancy force supplied for glider thrust is proportional to the drag coefficient. Appendages for instrumentation can account for appreciable fractions of the total drag, directly reducing endurance and range. The lower the drag coefficient, the higher the possible lift-to-drag ratio, hence gentler the glide path slope can be. Operational gliders typically attain glide slopes between about 1:1 and 1:5. While steep compared to glide slopes of sail planes, underwater glider glide slopes are much steeper than ocean property surfaces on the separation scale of successive dive cycles. Except very close to the stall limit, horizontal speed varies proportionally to glide slope for a given supply of propulsion power. For example, a glide slope of 1:3 produces 50% more horizontal speed, hence range in still water, than one of only 1:2. Fig. 1 shows the range of buoyancy, power consumption by drag and attack angle for a Seaglider as a function of vertical and horizontal speed.



Fig. 1 Seaglider flight performance inferred from regression of the difference between observed and modeled vertical speed against lift, drag, and induced drag coefficients. Curves contour attack angle, buoyancy, and delivered power. Colored dots indicate vertical speed model misfit at a variety of speed component combinations of vehicle pitch and buoyancy in an ensemble of field dive cycles.

Dive depth affects glider performance beyond setting the depth over which average current compares to vehicle horizontal speed. Glider navigation in shallow waters is more difficult because barotropic flows tend to be more pronounced. Since glider heading at the start of each dive is determined by the direction of the wave field, turn rate must be faster in shallower dives in order to navigate effectively. While weaker stratification reduces the demand on a buoyancy system to traverse it, shear tends to weaken concomitantly, reducing the effectiveness of diving deeply to reduce depth-averaged current. Another factor is that pumps used in gliders tend to be more efficient at higher pressure. For example a pump may draw only double the power at 1 km depth than it does at 100 m depth, hence it is five times more efficient on the deeper dive. Finally, biofouling occurs roughly in proportion to cumulative light exposure in much of the ocean, so deeper dive depth reduces the duty cycle of exposure to the euphotic zone and extends mission endurance in roughly inverse proportion.

Since work performed by a glider buoyancy engine provides not only propulsion but compensation for volume differences between the vehicle and an equal mass of seawater in its environment, reduction of the latter extends endurance and range. Volume compensation for the environment could be complete if the thermal expansion, haline contraction, and compressibility of a glider could be matched to the seawater equation of state. Both metal and composite hulls can both be engineered to compress nearly the same as seawater, but not over the complete pressure range of the ocean. Typical cylindrical pressure hulls made of metal are about half as compressible as seawater, so relative volume changes vary by  $\sim \frac{1}{4}$  % over 1000 m depth in the ocean. Relative volume changes due to temperature and salinity may amount to 1% or more over the same depth range in the upper ocean or may be negligible (e.g. in the case of a deep mixed layer). By comparison, sufficient buoyancy for propulsion is produced by volume differences of  $\sim \frac{1}{4}$  % or less. Appendage of a highly compressible and expansible material is effective in reducing the thermal and pressure induced volume differences to  $\sim 0.1$ %, less than half what is required for propulsion, over the full deep ocean water column. The relative amount of this added material must be tuned to the regional ocean environment, with the result that a vehicle can be made nearly neutrally buoyant from the sea surface to deep ocean floor. Figure 2 shows an example for conditions in the tropical western Atlantic.



*Fig. 2* Relative volume changes of a fixed mass of seawater, a Deepglider hull, and compressee / expansee amounting to 12% of the total vehicle mass, along with volume changes necessary to propel the vehicle at 0.25 m/s. Net buoyancy, the difference between seawater and vehicle volume, varies by at most ~0.1% of the vehicle volume from its average over the water column.

Many of the underwater gliders now in use employ an un-pumped electrode cell to measure electrical conductivity, used together with temperature to estimate salinity and density. While a pumped system can provide steady flow through the cell, for which correction for thermal inertia effects is a standard procedure for data from conventional ship-lowered conductivity-temperature-depth (CTD) instruments, the power used for the pump is an appreciable fraction of the total glider power budget. Gliders using a pumped CTD often collect measurements only on half of the dive-climb cycle to save energy, effectively reducing the horizontal resolution of the measurements. Correction for thermal inertia effects can be accomplished with an unpumped glider CTD if care is taken to model glider speed as a function of buoyancy and pitch attitude. At typical glider speeds (~0.2-0.4 m/s), induced flow through an un-pumped conductivity cell crosses the transition from viscous pipe flow to flow barely slowed by a frictional boundary layer. The thermal inertia correction to conductivity within the un-pumped cell is forced by a temperature perturbation corresponding to the oceanic temperature change over ~0.2 m depth (for a pumped cell it is about 1/3 as big). To adequately estimate the response requires a sampling interval short compared to the thermal relaxation time of the cell (~15-25 s for typical glider speeds). To attain a given salinity accuracy standard, sampling must be proportionately more frequent where the temperature gradient is higher.

The cost-effectiveness of using gliders to observe the ocean is largely determined by their endurance and range. The major costs of glider use are associated with preparation, launch and recovery. These include the costs of refurbishment, testing, shipping, travel, and vessel use. Costs that scale with mission duration are relatively minor, such as data communication and piloting costs. Because of the preponderance of fixed costs for each mission, the higher the endurance and range, the lower the cost per unit distance and time of survey. Due to relative inefficiency of shallow dives, continuous glider operations in coastal waters limited to a month or two duration tend to be several times more expensive than with missions with deeper dives which can extend to many months and are projected to exceed one year. For perspective, the cost of a single mission is comparable to the daily cost of operating a research vessel. Usage costs can be reduced significantly by using small boats from near shore for launch and recovery, thanks to the small size of operational gliders, which can be handled manually by a team of two persons. Glider acquisition cost is only a few times that of mission cost, so investment in glider technology is relatively quickly dominated by usage costs. Figure 3 shows that dramatically larger range and endurance are achieved by applying modest buoyancy forcing on deep dives.



*Fig. 3* A projection of Seaglider endurance, range, dive cycle time, effective horizontal speed (including time pumping and at the surface), and number of dives in a mission for a Seaglider, plotted as a function of dive depth and buoyancy used for propulsion.

# Surveys in regions of weak, moderate, and strong current

The principal utility of gliders is to repeatedly survey a region, a mission that is prohibitively expensive to do with a ship for very long. The ability to persistently resolve spatial structure on time scales relevant to ocean processes is their greatest attribute. But as a consequence of the need to go slowly in order to achieve high endurance and range, glider surveys are less synoptic as the survey extent grows. The mesoscale eddy field largely sets not only the required spatial and temporal resolution, but also the extent over which a single glider can survey effectively.

Examples of surveys in three different eddy environments illustrate how gliders can be used: 1) near Ocean Station P in the northeast Pacific, a low eddy energy region, 2) over the northeast Pacific continental slope off Washington state, an eastern boundary current region with moderate eddy energy, and 3) over the Iceland-Faroe Ridge, an eddy rich region of the northeast Atlantic Ocean. Repeated surveys were carried out from 1.5 to just over 5 years duration in these regions using typically a single Seaglider.

#### Repeat surveys near Ocean Station P

Ocean Station P (145°W, 50°N) has been visited at various intervals over the past half century, first by a Canadian weather ship and now thrice annually by CCGS Tully (in mid-winter and at the beginning and end of summer). Although the eddy field is weak in the vicinity of Ocean Station P, evolution of upper ocean structure on seasonal time scales cannot be recovered from only 3 samples per year.



*Fig. 4* Track of three successive Seaglider missions over 19 months. The bow-tie pattern was repeated roughly fortnightly proceeding in sequence to the labeled targets NW, SW, NE, SE, and NW again. Circles show surface positions between dive cycles.

In cooperation with a program to collect moored time series at the site, a sequence of three Seaglider deployments was made to jointly survey spatial structure in the vicinity and its evolution through two summers and the intervening winter. An initial 3-month deployment was followed by two deployments each over 9 months in duration, the two longest missions to date by any autonomous underwater vehicle. They operated for 279 & 292 days at sea, and glided 4911 & 5528 km through the water, respectively. Each mission began by launch from a ship near Ocean Station P, but since endurance was not sufficient to last a full year, the last mission *included transit to offshore Washington, where the Seaglider was recovered*.



Fig. 5 Track and estimated depth-averaged current over 0-1 km depth on the third Seaglider mission started at Ocean Station P. After 7 months executing the bow-tie survey, the vehicle transited ~1500 km toward the coast for recovery by a small sport-fishing boat chartered on a day trip from the Washington coast.

During these missions, the Seagliders dove to 1000 m depth continuously, reaching the surface every  $\sim$ 9 hours where profiles of temperature, conductivity, dissolved oxygen, fluorescence, and optical backscatter from both dive and climb portions of the cycle were communicated via the Iridium satellite network to our laboratory. Despite considerable productivity in the region, biological fouling of the vehicles was negligible.

These gliders executed a bow-tie pattern inscribed within a 50 km by 50 km box centered on a mooring, passing it weekly, taking two weeks to repeat the survey. Currents in the region were weak, typically ~0.05 m/s depth-averaged over the top 1 km. Currents did evolve, but on seasonal time scales. Water properties tended to have correlation scales of 2-3 weeks and ~20 km. Estimates of absolute geostrophic current (referenced to glider-inferred depth-averaged current) and horizontal gradients are sufficient to estimate horizontal advection of density. Because the Rossby number of the flow was only ~0.01, vertical stretching can plausibly be inferred from meridional flow, enabling an estimate of vertical advection as well. The vertical mixing inferred from the sum of time rate of change and horizontal and vertical advection implies vertical diffusivity of  $O(10^{-4})$  m<sup>2</sup>/s in the upper pycnocline averaged over a season.

While the bow-tie track was easy to repeat, eddy activity at the end of the transit to the coast interfered with glider progress and ability to follow the desired track. Depth-averaged current exceeded glider speed at the end of the transit.

#### Cascadia – Repeat surveys over the continental slope offshore Washington

Eastern boundary current systems are attractive sites to employ gliders to resolve the eddy field and detect seasonal and interannual water property and circulation changes. In a sequence of Seaglider missions off the Washington coast lasting more than 5 years, a single vehicle at a time was sent along a V-shaped transect seaward 220-240 km from the continental shelf edge (Fig. 6). Each leg of the surveys was completed roughly

fortnightly. Gliders were launched at the shelf edge in order to extend endurance and to avoid exposure to barnacle larvae that populate continental shelf waters. Small boats on day trips from shore were used for launch and recovery, except when a vehicle had to be rescued beyond their range. Full missions typically lasted 5 months, since Seagliders with the original (smaller) complement of batteries were used in this project.

While eddies are a common feature of circulation over the continental slope, gliders were generally able to stem their flow and reach the targets at the ends of the transects that defined the survey. An onboard Kalman filter was used to correct course in response to current. This was effective despite the simplistic model state used by the filter: that drift was due to two tidal components (semidiurnal and diurnal) and a spatially uniform current. The widest deviations from the intended track were likely due to improper correction for hard iron contribution to the vehicle compass on one of the missions.



Seagliders 22 August 2003 – 7 January 2009

Fig. 6 Seaglider tracks offshore Washington, USA. Circles indicate surface positions between dive cycles. Gliders were launched and recovered from small boats transiting 30-40 nautical miles from Grays Harbor, La Push, and Neah Bay. Dives were to near the sea floor or 1 km depth, whichever was shallower.

#### Iceland-Faroe Ridge repeat surveys

The Iceland-Faroe Ridge is a barrier between relatively warm, saline, Atlantic waters and cool, fresh, Nordic Seas waters. The ridge extends from Iceland to the Faroe Islands. Nordic Seas waters overflow through the narrow gap between the Faroe Islands and Faroe Bank, the Faroe Bank Channel to supply what evolves, with mixing of the overlying Atlantic water, into North Atlantic Deep Water, one of the major constituent water masses of the world ocean. From November 2006 to November 2009, Seagliders were launched and recovered from F/V Magnus Heinason, the Faroese Fisheries Laboratory research vessel, on its quarterly hydrographic survey and mooring service cruises. Because of frequent severe weather and sea conditions and a relatively wide continental shelf, using small boats to service gliders was impractical, although on two occasions during summer months a small passenger ferry was chartered to recover a glider. Despite heavy seas, gliders were routinely recovered manually from a working deck with 7 m freeboard.

Initially, the intent was to send gliders along the Ridge toward Iceland along a sequence of pre-defined sawtooth transects and periodically repeat them. Unlike other missions, dives in this project were deliberately to within a few meters of the sea floor, normally in water less than 1 km (the operating depth limit for Seaglider). Because dive depths typically were in the 400-1000 m range, endurance was roughly 4 months, but due to the ship schedule, missions lasted no more than 3 months. Although dives were through the entire water column, stratification was relatively modest and depth averaged currents were frequently swifter than Seagliders were able to glide. Instead of a regular sequence of transects, gliders were pushed randomly up and down the slope as they tried to follow a particular isobath along the Ridge's Atlantic flank. On several occasions, cross-Ridge transects were carried out.

The result of trying largely to sample along a given isobath on the Ridge flank was a broad swath of dive locations. Figure 7 shows the location of Seaglider dives as well as bottom temperature. Unlike most ocean regions where sampling frequency can be controlled via depth bins, the thermocline hugs the bottom on the Atlantic side of the Ridge. At the thermocline base, a bottom mixed layer was typically found. The temperature distribution in Fig. 7 clearly shows the cold overflow plume exiting the Faroe Bank Channel, with part of it spreading along the Ridge toward Iceland, warming as it goes. Some additional overflow crosses the Ridge directly.

Despite nearly random sampling locations within a swath controlled by general glider progress through the eddy field, patterns of the average flow are apparent.



Fig. 7 Bottom temperature (or at 1000 m, whichever is shallower) over the Iceland-Faroe Ridge in the northern North Atlantic, sampled by Seagliders in quarterly surveys over 3 years (November 2006 – November 2009). The eastern edge of Iceland appears in the upper left of the frame, with the Faroe Islands in the lower right. The Faroe Bank Channel is to the southwest of the Faroe Islands. Vigorous eddies and tidal currents prevented surveys along well-defined tracks to be carried out and repeated. Heavy depth contours are drawn at 500 & 1000 m.

# **New developments**

Seaglider technology has been extended to enable operations under ice-covered waters and to the sea floor in the open deep ocean. Under-ice navigation on Seagliders is carried out using RAFOS beacons. The new Seaglider derivative called Deepglider has recently made dives to nearly 6000 m depth. By compensating for compressibility and thermal expansion differences, Deepglider energy consumption is reduced sufficiently to enable missions up to 18 months duration and 10,000 km range. Modifications to Seaglider suggest endurance of over 1 year is feasible.

# Conclusions

Repeat surveys can be carried out by gliders in a variety of ocean environments. Both vehicle efficiency and efficient usage strategy are key to their successful, cost effective use in the open ocean. Continued access to GPS navigation and low power satellite communication are essential to their viability as ocean research tools.

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# **LEEUWIN CURRENT - DYNAMICS, VARIABILITY & IMPACTS**

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The Leeuwin Current is an anomalous poleward-flowing eastern boundary current, carrying warm, surface ocean waters from northwest Australia southward along the west coast of Australia. The meridional pressure gradient in the southeast Indian Ocean, set up by the Indonesian Throughflow in the tropics and by latent heat fluxes (cooling) in the mid-latitude, accounts for the existence of the Leeuwin Current. The Leeuwin Current is shallow in the north (about 150 m) and deepens towards the south (about 300 m), with an average volume transport of 3-4 Sv. The flow rate of the Leeuwin Current is typically one knot and the core of the current is anchored at the upper continental slope. Mesoscale meandering structures occur ubiquitously along the pathway of the Leeuwin Current and some of the meanders form long-lived anti-cyclonic (warm-core) eddies that detach from the current and propagate offshore. The Leeuwin Current has the strongest eddy energy among the mid-latitude eastern boundary current systems. Mesoscale eddies draw their energy from the Leeuwin Current due to mixed barotropic and baroclinic instability and play important roles in the momentum and heat balances of the Leeuwin Current.

Off the west coast of Australia, the Leeuwin Current is stronger in austral autumn and winter and is weaker in austral summer, mostly due to seasonal variations of alongshore winds. Strong northward winds between November and March slow the Leeuwin Current and drives episodically northward inshore currents and localised upwelling. The strongest southward surface current occurs in April-May when the opposing winds weaken. Eddy energetics in the Leeuwin Current experiences a strong winter enhancement, lagging the seasonal variation of the mean current by about 2 months. Due to the existence of equatorial and coastal waveguides, the interannual variations of the Leeuwin Current and its eddy field respond to the El Niño/Southern Oscillation (ENSO) variability in the Pacific, being stronger during a La Niña year and weaker during an El Niño year. Seasonal and interannual variability of sea level and thermocline off the west coast have been found to have downstream impacts along the south coast of Australia.

Under the influence of human-induced climate change, climate models suggest that the mean state of the tropical Pacific will likely shift towards an El Niño-like condition, which would induce shallow thermocline depth anomalies in the equatorial western Pacific and the southeast Indian Ocean and a weakening trend of the Leeuwin Current. There have been observed shallow thermocline anomalies in the region during mid-1970's to early 1990's, however, this trend may have reversed in the last one and half decades due to enhanced decadal climate variability in the Pacific.

The existence of the Leeuwin Current induces deep thermocline depth, suppresses wind-driven coastal upwelling, and causes the oligotrophic marine environment off the west coast of Australia. Pelagic production off the west coast is low however there is relatively large benthic community, supporting the iconic western rock lobster fishery. The recruitment processes of the western rock lobster and some other fisheries off the west coast are susceptible to ENSO-related interannual variations of the marine environment. Understanding the future climate in the Indo-Pacific Ocean and the impacts of climate change and variability on the marine environment are crucial for the long term management of the marine resources in the region.

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# OCEAN VARIABILITY OFF EASTERN AUSTRALIA AND FUTURE OBSERVING STRATEGIES

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#### Introduction

The ocean regions around Australia include many dynamically interesting and unique features. Here we focus on the waters to the east of Australia. This region forms a complex intersection of the Pacific and Indian Oceans. includes a major current system, the East Australian Current, complex topography, and large ENSO related interannual variability in the southwest Pacific. A monitoring system capable of addressing the broad requirements of this region needs to draw upon a diverse range of observation strategies

#### East Australian current and its eddies

The ocean is also a very turbulent environment - over periods of days to weeks the variability is dominated by mesoscale eddies, which have a strong signature in surface pressure and thus sea level. There are very energetic regions of mesoscale eddies associated with the major current systems: the EAC (Fig. 1).



*Fig. 1* The RMS variability of the sea surface height as determined from satellite altimetry data from 1992-2010.

These eddies play an important role in the dynamical and heat balances of the major current systems – acting to modulate the strength of the mean currents and their regional temperature footprint. They also flux heat and nutrients across current systems and isobaths. Along the Australian shelf break, the eddy field likely mediates the transport of these quantities between the offshore and shelf environments. The resolution and prediction of the eddy field and its impacts on the structure of the mean ocean flow is a key research challenge.

In the EAC system, about 2-3 large eddies grow and are pinched off every year (~90 day period) – believed to derived from a mixed barotropic/baroclinic instability of the mean flow (Mata et al, 2007). These eddies exhibit a westward vertical tilt with increasing depth indicating that they are actively fluxing heat poleward (Oke and Griffin, submitted). These eddies frequently move onto the continental shelf and close inshore and influence the local circulation patterns. At prominent coastal features the EAC moves away from the coast,

driving upwelling which draws nutrient-rich water from a depth of 200-m or more. However, while the EAC may drive nutrient-rich water onto the shelf, upwelling-favourable winds (northerly) bring the water to the surface (Rochford, 1984; Cresswell, 1994; Church and Craig, 1998).

# **Observation strategies**

In general, the meso-scale features such as the EAC present a major challenge to observing systems. Presentgeneration systems simply cannot sample the world's ocean frequently enough with high resolution, except in the case of sea surface temperature and colour, and even those variables are under-sampled wherever cloud cover is at all dense. These shortcomings are compounded for these boundary currents since they are energetic and narrow and cannot be adequately sampled with the broadscale networks (Argo floats and satellite altimetry) that we rely on in the interior. They pose particular observational challenges. We need to resolve these systems at very fine scales to measure heat and freshwater transport. Boundary currents produce turbulence on multiple scales in response to coastline, bathymetric irregularities, and flow instabilities. The high current speeds within the EAC make them inherently nonlinear, producing internallygenerated variability that can be the dominant term in the momentum and vorticity balances, and which demands sustained sampling. The resulting meanders and eddies can produce a bolus flux that in some cases appears to accomplish most of the transport (e.g. the EAC south of 30°S; Mata et al. 2000). High flow speeds through thick layers can make operation of various platforms untenable in some regions.

The general circulation structure has been obtained from traditional research cruise measurements – now greatly expanded in time and space by Argo network. This is designed to capture the broadscale structure. Including satellite measurements allows the finescale features to be resolved.



Fig. 2 The major current systems off eastern Australia. Potential sites for observing the major boundary currents systems. Full-depth mooring arrays at the Red sections – EAC (26°S, off Brisbane), Tasman Outflow (southern Tasmania). Glider across current deployments shown by yellow tracks, Hiri Current (north Queensland, 14°S), EAC Extension (eastern Tasmania) and along the core of the EAC.

#### Fixed mooring arrays

These diverse challenges will not be met be any particular tool, and a boundary current strategy will demand multiple observational techniques tuned to the particular conditions of each current system. A more realistic approach is to choose a limited number of strategic locations or 'chokepoints' where representative

measurements of the currents may be obtained. Off eastern Australia potential chokepoints that include the Tasman Outflow (south of Tasmania, across the South Tasman Saddle), the EAC off Brisbane ( $26^{\circ}$ S) and the straits and channels east of PNG ( $14^{\circ}$ S).

# **Deepwater gliders**

Deepwater gliders provide an alternative technology for monitoring the boundary currents. They potentially provide repeated measurements of mean velocity, T, S,  $O_2$  and a suite of biological properties. The glider deployments need to be fully integrated into the strategy for monitoring boundary currents. The deployments need to move from single glider process-study proposal-driven mode into a more sustained pattern of repeat tracks with 2-or more gliders at each location preferably integrated with other infrastructure components (eg moorings). In order to resolve the time-varying flow, deployments should allow immediate replacement and where possible multiple gliders deployed. Deployments should be focused on regions of highest signal to noise conditions. The glider deployments should be guided by both experience gained in the completed deployments and by simulation studies based on Bluelink model results. In more benign current environments repeated transects may be feasible (Hiri Current) whereas in energetic systems such as the EAC, glider tracks may follow zig-zag paths down the current.

# Resolving the basin-scale flow

Here we maintain and extend hi-density XBT sections and Argo deployments. These data are pivotal to providing the interior basin circulation - they extend the observing system beyond the endpoint of the boundary arrays and largescale context of the boundary flow. Being high-resolution and exactly repeating, these transects when combined with altimetry observations resolve the temporal variability of the EAC eddy fields

For many years the backbone of ocean monitoring has been the XBT network. Since 1991 3 high-res XBT transects enclosing the Tasman Sea (Tasman Box) has been the central component of EAC observation. Methods have been developed to merge these data with satellite observations to resolve, seasonal, interannnual, decadal variability of the main current features.

# Satellite observations

A further approach is to utilize satellite observations of SST and ocean colour which resolve fine spatial scales. The next most densely-sampling observing system is satellite altimetry. This density of altimetry coverage, along with the fact that sea level is a key measure of ocean circulation, is the reason that altimetry is the single most important data source for mesoscale-resolving models such as Bluelink. The Jason-2 altimeter samples with a repeat cycle of 10 days and spacing of 315km. This is a sufficient density for large-scale phenomena such as El Nino but it is insufficient for rapidly-changing and/or smaller-scale phenomena such as extra-tropical eddies and boundary currents. In addition, within the coastal region there is a growth in the altimeter error which quickly approaches the magnitude of the signals of interest. One way of using this data is to exploit high resolution models/data assimilation systems such as BlueLink to provide full spatial and temporal coverage through the water column. Here we need to provide appropriate data streams such as Argo, SST and altimetry for assimilation into the models and for subsequent validation.

One way of using this data is to exploit high resolution models/data assimilation systems such as BlueLink to provide full spatial and temporal coverage through the water column. Here we need to provide appropriate data streams such as Argo, SST and altimetry for assimilation into the models and for subsequent validation.

# Monitoring the circulation on the continental shelf

The above approaches are focused on the boundary currents which tend to be located on the continental slope and into the adjacent deep ocean regions. The circulation on the continental shelf regions is also of major importance. These systems tend to be weaker but more complex with smaller time and space scales. A systematic effort to monitor the shelf region is essential. Again this may be addressed by strategic location of mooring platforms (inshore of the deepwater arrays), coastal gliders, coastal radar and satellite altimetry.

The 'coastal altimetry' community are starting to produce datasets for use as close as 5 -10km from the coast, subject to the availability of a local tidal correction. Track-normal geostrophic velocity estimates could be accurate at ~3km along-track resolution. Validation of this method at a few places therefore has the potential to unlock a 17-year (and growing), Australia-wide record of boundary current variability.

Altimetry, SST and ocean colour reveal the location of eddies and fronts but are insufficiently accurate, or fine-grained in space or time, to measure the details of what is happening in the surface layers, let alone the sub-surface layers. This information is required in order to make progress on understanding the links between the boundary currents, eddies and biogeochemical processes. Gliders are a cost-effective way to measure the subsurface water column properties of eddies, and the structure of fronts, with high spatial and temporal resolution. In shallow shelf regions where currents are relatively weak, gliders are ideally suited to collect sustained long-term observations on repeated tracks. Coastal radars measure the detailed structure of surface currents, in some cases with sufficient accuracy to infer the vertical velocity from the divergence of the field. This sort of information is required to explain the as-yet inexplicable features we see in ocean colour imagery.

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#### DEEP OCEAN CHANGE IN THE WESTERN PACIFIC OCEAN

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# Abstract

Repeated occupations of three hydrographic sections in the western Pacific Ocean from the 1990's to 2000's show the spatial distribution of Antarctic Bottom Water change in the Southwest Pacific Basin. The largest property change – temperature, salinity, total carbon and oxygen – are found in the deep western boundary current from 50°S to the equator. The magnitude of the property change decreases with increasing distance from the western boundary. The abyssal warming and freshening has resulted in a local decrease in density of the deep Southwest Pacific basin, and hence a reduction in the volume of Antarctic Bottom Water within the Basin.

# Introduction

Predicting climate change and sea level rise depends critically on knowledge of the changing budgets of greenhouse gases, energy and water, and the processes determining them. The upper ocean (above 2000 m) contribution to sea level and the Earth's energy budget is calculated from in situ ocean observations (Domingues et al. 2008). However to accurately monitor and simulate the Earth's energy budget and provide estimates of the global and regional sea level budget all components of the earth climate system, including the uncertain deep ocean, must be included. While the sparse data makes quantification of deep ocean thermal changes difficult, they do appear large enough to be significant contributors to the Earth's energy (some tens of percent of the upper ocean heat storage) and sea level budgets. In addition, estimates of the impact of aerosols on the Earth's energy budget (Murphy et al. 2009) is provided as a residual of ocean, land and atmosphere heating and the known forcing agents (solar, carbon dioxide and other greenhouse gasses). Improved estimates of the deep ocean heat contribution to the energy budget will provide more robust projection of sea-level rise and better constrain poorly known estimates of aerosol impacts on climate.

The Global Ocean Ship-based Hydrographic Investigations Program (GO-SHIP, <u>www.go-ship.org</u>) is coordinating international efforts for sustained observations of the full-depth ocean. This program is providing highly accurate deep ocean observations at specific sites in the global ocean. Analyses of repeat sections show how variability in the North Atlantic overflows (e.g. Yashayaev, 2007), spreads throughout the North Atlantic. Decadal variations in Labrador Sea Water properties can be traced to 20°W (Johnson et al. 2005), and North Atlantic Deep Water property variations to 24°N (Cunningham and Alderson, 2007). Similar to the North Atlantic, warming deep ocean temperatures linked to Antarctic Bottom Water (AABW) sources appear to be widespread in the South Atlantic, Indian, and Pacific Oceans. Deep and bottom warming has been detected over the past several decades in the Weddell Sea (Fahrbach et al. 2004). Downstream of this region, statistically significant warming has been detected in the bottom waters of all the western basins of the South Atlantic (Johnson and Doney, 2006), and warming signals appear to extend into the western North Atlantic (Johnson et al. 2008). Deep ocean warming, emanating from bottom water source

regions of eastern Antarctica and the Ross Sea, is apparent throughout the Pacific Ocean (Kawano et al. 2006; Johnson et al. 2007), even to the rather surprising location of 47°N in the Pacific Ocean (Fukasawa et al. 2004).



Fig. 1 Location of repeat hydrographic sections (P06, P21 and P15) occupied in the western Pacific Ocean between 1990 and 2009/2010. Bathymetry (gray shading) at 1000 m intervals shows the extent of the Southwest Pacific Basin (> 4000 m).

Here eight realizations of three sections occupied in the western Pacific are analysed with respect to AABW changes (Fig. 1). Two repeat zonal sections at approximately 17°S (P21, occupied in 1994 and 2009) and 32°S (P06, occupied in 1992, 2003 and 2009/2010) and a meridional section 170°W (P15, occupied in1990, 2001 and 2009) reveal the spatial distribution of the deep ocean change in the Southwest Pacific Basin.

#### Results

The combined analysis of all sections consistently shows that the largest property changes are found near the western boundary of the Southwest Pacific Basin. The meridional section (P15 - 170°W) shows that warming of the abyssal ocean below 4000 m extends from 50°S to the equator (Fig. 2, and Fig. 3). The magnitude of the warming decreases with increasing distance from the western boundary. The largest warming signal is associated with topographic constraints on the western boundary at 45-35°S, 20°S and 10°S (Fig. 3). These topographic features strongly constrain the northward deep western boundary current. Similar temperature change along 17°S (P21) are associated with the topography as the AABW flows northward along the deep western boundary.

Analysis of salinity and total carbon changes (not shown) indicate a freshening and slight increase of total carbon concentration in the deep basin that is consistent in all sections. Decreased oxygen concentrations are due to the change in solubility driven by the temperature increase.

The observed changes in the abyssal Pacific Ocean are significant and impact the global climate. Given the signals found in the expanding deep observational record, we need to assess the simulation of these changes in global climate models. Such studies will highlight deficiencies in model simulations of the deep ocean. Focusing on these deficiencies will provide the impetus to fix model physics in the next generation models.



*Fig. 2* (Upper panel) Decadal potential temperature change ( $^{\circ}C/decade$ ) along  $170^{\circ}W$  near the western boundary of the Southwest Pacific Basin. (Lower panel) Mean vertical potential temperature change (x  $10^{-2} \, ^{\circ}C \, decade^{-1}$ ) for the southern, middle, and low latitude regions of the section.



*Fig. 3* Spatial distribution of decadal potential temperature change (°C/decade) below 4000m in the Southwest Pacific basin.

#### Conclusion

Outlining the need and strategy for monitoring AABW formation sites, deep passages, and overturning circulation in the southern hemisphere oceans is a priority. A deep observing system needs to monitor:

- The global ocean temperature and salinity below 2 km by building a broad-scale full-depth array of floats/gliders/moorings;
- Temperature, salinity, velocity, and carbon near formation regions including AABW locations in Ross Sea and Adelie Land to Weddell Sea;

- Temperature, salinity, and velocity downstream from these formation sites in Indian and Pacific Oceans and the Southern Ocean Choke points and;
- The meridional overturning circulation in South Atlantic, Indian and Pacific and Southern Ocean.

The major observed change of the deep ocean is in the Southern Ocean and therefore the implementation of a deep ocean observing system must firstly target the Southern Ocean and then expand to other regions.

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### **EDDIES AND CLIMATE**

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#### Abstract

The mean circulation of the global ocean is contaminated by a strong eddy field, which in many cases dominates the local ocean state. Such eddies are (by computational necessity) parameterised in climate models. However, eddies may play a role in contributing to low-frequency climate variability and modulating the ocean's response to climate change.

#### Introduction

Ocean eddies are a ubiquitous feature of the circulation. However, their small scales (50 - 200 km) imply that ocean or climate models require  $1/6^{\circ}$  resolution (or better) to explicitly represent eddies in numerical simulations. Limits on computational resources mean that few climate models can reach this resolution; thus, eddies are parameterised in all such models. Evaluation of the role of eddies in global climate is therefore restricted to idealised or regional high-resolution model simulations.

Here I will review two cases in which the time-dependent behaviour of the mid-latitude ocean circulation is dependent upon explicit resolution of eddies. The first case concerns eddying models of the double-gyre circulation, representing areas such as the North Pacific or North Atlantic Oceans. Here, modes of low-frequency (decadal) variability arise through interactions between the eddy field and the mean flow, while coupled simulations indicate the potential for such behaviour to feed back on the atmospheric variability. The second case is specific to the Southern Ocean, where eddy-resolving models produce results counter to those of coarse-resolution models. These two cases are outlined in brief in the following sections.

# The double gyre Turbulent Oscillator

It is well-known that the mean circulation in the mid-latitude ocean basins is comprised of gyres driven by prevailing westerly winds (e.g. Munk, 1950). If we consider a Northern Hemisphere ocean basin (such as the North Atlantic or North Pacific Oceans), positive wind stress curl south of the maximum wind stress generates southward flow in the interior, returning as a narrow boundary current along the western edge of the domain. Thus, a westward intensified subtropical gyre forms in these regions. Conversely, a subpolar gyre exists in the northern part of the basin, with northward interior flow and a southward western boundary current (WBC). This mean circulation is shown by the thermocline depth (denoted HO1) in an idealised quasigeostrophic model driven by wind stress with a simple maximum at y=2400 km in Fig. 1(c) (from Hogg et al. 2006). Here, solid contours indicate an elevated thermocline and anticlockwise circulation, while dashed contours indicated a depressed thermocline and clockwise circulation. The maximum velocities (steepest thermocline gradient) are along the WBC, and the coalesce into a jet (the WBC extension) which penetrates the ocean interior and marks the division between the two gyres.

An important characteristic of this flow is that the instantaneous flow differs substantially from the mean at all times, principally because of strong geostrophic turbulence. This field is shown in Fig. 1(a), where a strong, meandering jet and associated eddy field dominates the flow.



*Fig. 1* (a) Instantaneous thermocline depth; (b) instantaneous SST field; (c) mean thermocline depth and (d) mean SST field from a quasigeostrophic model of an idealised double-gyre ocean. From Hogg et al. (2006).

The Sea Surface Temperature (SST) fields in Fig. 1 (b,d) reflects the gyre circulation, with sharp SST gradients in the WBC extension region, indicating a likely influence of this region upon the atmosphere above.

Eddy-resolving simulations have been used to investigate the fundamental dynamics of the double gyre circulation. Berloff and McWilliams (1999) demonstrated the emergence of modes of low-frequency variability when eddies are resolved in such flows. This result was confirmed by Hogg et al. (2005) and Berloff et al. (2007) for higher dimensional systems. The characteristics of the variability is shown in Fig. 2 (reproduced from Hogg et al. 2006) using an Empirical Orthogonal Function (EOF) analysis. Figure 2(a) shows the spatial patterns from the first mode Hilbert (or complex) EOF of the thermocline depth; this mode accounts for 43% of the low frequency variance in the circulation. Hilbert EOF analysis is similar to a standard EOF, except that it picks up spatial patterns from two different phases and so is ideal for identification of travelling waves. The two patterns in Fig. 2(a) describe a strengthening (left) and a shifting (right) of the jet dividing the two gyres. This mode of variability affects SST in a similar fashion (Fig. 2b), while the power spectrum in both fields (Fig. 2c) is dominated by a strong interdecadal peak (20 years in this case). This interdecadal mode of variability was dubbed the "Turbulent Oscillator" by Berloff et al. (2007).

Such low-frequency variability is absent from double-gyre simulations in which the eddy field is parameterised, indicating the likelihood that the observed mode owes its existence to the presence of resolved eddies.



*Fig. 2* The first Hilbert EOF of (a) thermocline depth and (b) SST. (c) Spectra of the Principal Components showing interdecadal variability in both modes. From Hogg et al. (2006).

Berloff et al. (2007) investigated the fundamental dynamics driving the Turbulent Oscillator, and found that the phenomenon depended principally upon nonlinear interactions between the inertial mean flow and the eddy field. The WBC extension jet which divides the two gyres, provides a barrier to eddies, and thereby inhibits eddy-driven flux of potential vorticity (PV) between the gyres. In such a state, the PV contrast between the gyres strengthens, leading to a stronger jet and a stronger barrier. In its strongest state, the jet is attracted south, towards the strongest gyre, until such time as WBC instabilities act to destabilise the jet. In this unstable state, the jet is permeable to eddy PV flux between the gyres, weakening both gyres and resulting in a return of the jet to its northward position. Thus, the mechanism depends upon the ability of the nonlinear jet to oscillate between a barrier and a conduit to the exchange of eddies between the gyres.

The above results established the principle that the ocean could generate intrinsic low-frequency variability in the large-scale circulation. In addition, Hogg et al. (2006) used a couple quasigeostrophic model to demonstrate that the Turbulent Oscillator could drive low-frequency variability in the atmosphere. This study used a combination of uncoupled, partially coupled and fully coupled experiments to show that atmospheric variability at low-frequency was enhanced by the addition of a dynamic, eddy-resolving ocean. Importantly, the coupled system did not generate new modes of atmospheric variability, but instead projected onto existing modes, enhancing their low-frequency components. The result was that fully coupled simulations resulted in greater oceanic and atmospheric variability, implying that the ocean-generated variability could be amplified by coupling to the atmosphere.



*Fig. 3* Spinup of an idealised quasigeostrophic Southern Ocean simulation. Snapshots are shown from (a) 5 years; (b) 10 years, indicating the presence of baroclinic instability; (c) 15 years and (d) a fully developed state at 25 years. From Hogg & Blundell (2006).

#### Southern Ocean eddies

The Southern Ocean has an even stronger eddy field, as well as a stronger mean flow, than other midlatitude regions of the ocean. Thus, it is likely that eddy-mean flow interaction plays a strong role in setting the dynamics of the circulation in this region. The Southern Ocean can be modelled with very similar physics to the double-gyre problem, except that periodic boundaries in the zonal direction mean that reentrant flow alters the flow state to a series of jets, rather than gyres. The development of such jets in a quasigeostrophic eddy-resolving model is shown in Fig. 3 (reproduced from Hogg & Blundell, 2006). Here we again show thermocline depth, which initially tilts (upwards in the south, downwards in the north) to balance the imposed westerly wind stress forcing. After approximately 10 years, the flow becomes baroclinically unstable. As the mean flow builds, so does the eddy field and a sequence of multiple jets, which resembles the flow state of the Southern Ocean. Each of the jets shown in Fig. 3(d) describes eastward flow, summing up to substantial eastward transport known as the Antarctic Circumpolar Current (ACC) in the real ocean.



*Fig. 4* Diagnostics from the ensemble mean of 12 simulations for three different wind stress forcings. (a) Wind stress; (b) ACC Transport; (c) Potential Energy; (d) Kinetic Energy and (e) Temperature in a region south of the ACC.

A number of Southern Ocean studies have focussed on the total volume transport of the ACC (e.g., through Drake Passage). There is some consensus that the ACC transport is set by wind stress forcing, and coarse-resolution models appear to show a linear dependence of transport upon the wind (e.g. Allison et al, 2010) in keeping with some theoretical predictions (e.g. Marshall & Radko, 2003). However, it is also possible that Straub's (1993) prediction that transport may be independent of wind stress above a certain threshold is close to reality. This prediction is consistent with eddy-permitting model results (e.g. Hallberg & Gnanadesikan, 2006) and has lead to the term "eddy saturation" to describe this effect.

Figure 4 shows results from a quasigeostrophic model for the Southern Ocean, which is close to the eddy saturation limit. Timeseries of ensemble mean quantities (using 12 individual simulations) from 3 different types of experiment are shown; in two of the experiments, wind stress is increased for a 2-year period as

shown in Fig. 4(a). However, the ACC transport (Fig. 4b) responds only weakly to the wind stress changes, despite a coherent amplification of potential energy (Fig. 4c). Potential energy peaks soon after the maximum wind stress, and results in a cascade of energy into the eddy kinetic energy field (EKE; approximated by the kinetic energy timeseries in Fig. 4d). EKE peaks 1-2 years after the maximum wind stress, a result which is consistent with satellite observations of the ACC's response to interannual wind stress variations (Meredith & Hogg, 2006). This result is the first observational result to support the predicted eddy-saturated state. More recent eddy-resolving simulations using a range of models (Screen et al. 2009, Farneti et al. 2010, Hogg 2010) support the notion of an ACC which is close to the eddy-saturation limit, indicating that global climate models may omit the fundamental dynamics of the response to change in the Southern Ocean.

One consequence of an eddy-saturated ACC which was spelled out by Hogg et al. (2008) and confirmed by Screen et al. (2009) is that enhanced EKE following strong Southern Ocean wind events is likely to result in an increase in southwards eddy heat flux. This effect is shown in Fig. 4(e), where a delayed warming of the Southern Ocean south of the ACC is generated by the enhanced eddy field. This result implies that increasing wind stress over the last 50 years may have indirectly generated enhanced warming of the Southern Ocean by activating eddy heat transport.

#### Discussion

The above two examples serve to indicate that ocean eddies may play a more important role in climate and its variability than previously thought. In both cases, eddy resolving models produce different results from coarse-resolution models in which eddies are parameterised, and there seems little prospect of adjusting existing parameterisations to account for observed effects. Thus, if climate models are to improve their ability in predicting regional climate variability, resolving ocean eddies may be one of the critical steps.

The Turbulent Oscillator in the double-gyre system is an interdecadal mode of variability which has, so far, only been catalogued in idealised simulations. However, it may have links with existing known modes of variability in the coupled system, such as the Pacific Decadal Oscillation (PDO) and the North Atlantic Oscillation (NAO). Recent improvements in observations of the PDO (Qiu & Chen, 2010) indicate some similar characteristics to the proposed Turbulent Oscillator; in particular, the quadrature between jet strength and position. However, the observational record remains short and resolving this question will most likely depend upon global eddy-resolving climate simulations.

The role of Southern Ocean eddies in the climate system is an active area of current research. Results described here, indicating the possibility of an eddy-saturated ACC, have spurred further questions about whether eddies may also modulate the wind-driven upwelling and overturning circulation in the Southern Ocean (e.g. Boening et al. 2008). Changes in this overturning circulation have major implications for the carbon cycle and heat uptake of the ocean; however there is no real consensus in future predictions. The dynamics underpinning wind-driven upwelling in the Southern Ocean needs further study to resolve this question.

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#### HEAT IN THE OCEAN MODELS

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#### Abstract

Global climate models forced with all radiative forcing factors (both natural and anthropogenic) underestimate global trends in oceanic heat content (OHC) over the late 20th century. This is possibly due to the models overestimating the oceanic cooling influence of the indirect aerosol effect. Volcanic forcing alone cannot account for the weaker modelled OHC trends. The models results support the hypothesis that anthropogenic aerosols are critically important for simulating the structure of the observed temperature change in the Southern Hemisphere; this change is strongest in the midlatitude band 35°-50°S, and are associated with a Sverdrup-like response to poleward strengthening winds and a poleward shift of the Southern Hemisphere supergyre of about 1°. The heat required for this midlatitude warming is mostly derived from surface heat fluxes from south of 50°S, which is advected northward by enhanced Ekman transport induced by the poleward-strengthening winds. These results may, however, be significantly altered in eddy-resolving models.

#### Introduction

Significant warming has occurred across many ocean basins over the past 50-years, measured as an increase in the oceanic heat content (OHC) or depth integrated temperature change.

Globally, the OHC (surface-3000m) increased over the 1961-2003 period by  $14.2 \pm 2.4 \times 10^{22}$  J (or an equivalent heating rate of  $0.21 \pm 0.04$  W m<sup>-2</sup>) (Bindoff et al. 2007), with slightly longer-term measurements over 1960-2007 (Wijffels et al. 2008) suggesting a warming (surface-2000m) of close to  $19 \times 10^{22}$  J (Cai et al. 2010a). Differences in OHC datasets, which are mostly due to quality control techniques employed and actual measurements used, are small, whilst long-term trends (on multi-decadal time scales) in the data show good agreement (Bindoff et al. 2010, Cai et al. 2010a). There is a growing consensus that the warming is likely due to anthropogenic forcing, however the potential cooling offset of aerosols from volcanos and anthropogenic sources (e.g., sulfates) may have masked an even stronger warming signal, as shown in recent climate modelling studies (Church et al. 2005, Cai and Cowan, 2007).

The strongest oceanic warming has occurred in Southern Hemisphere (SH) midlatitude band of 35°-50°S, with a temperature increase of between 0.2-0.4°C from 1960-2007 extending down to a depth of 1000m (see Fig. 4c in Cai et al. 2010b). The relative importance of wind and surface heat flux changes in driving this regional warming pattern are the subject of much research. For example, a poleward shift in the midlatitude gyre circulation is associated with an upward trend of the Southern Annular Mode (SAM), partially forced by an increase in greenhouse gases and stratospheric ozone depletion (Thompson and Solomon, 2002). Furthermore, anthropogenic aerosols have also been shown to intensify the SAM trend through an air-sea feedback process (Cai and Cowan 2007). However, recent Argo float measurements infer a downward and southward displacement of isopycnals, which suggests that wind forcing alone may not be the only driver of the warming (Roemmich and Gilson, 2009).

Studies therefore have to rely upon ocean-only and fully coupled climate models to evaluate the role of wind and surface heat fluxes in driving the warming trends. This is achieved using ocean temperature, surface wind, and surface heat flux components from climate model experiments submitted as part of the Third Coupled Model Intercomparison Project (CMIP3) for the Fourth Assessment Report of the Intergovernmental Panel on Climate Change. Surface heat flux outputs allow the calculation of a simple heat budget in the SH midlatitude oceans to ascertain the origin of the heat that has led to a warming of the 35°-50°S latitude band.

Furthermore, multiple model ensembles and targeted modelling experiments allow us to isolate (and attribute) the proportionate impact of external radiative forcings such as volcanic and anthropogenic aerosols on global and SH OHC trends, and spatial temperature patterns.

In the following sections we discuss the role of wind and surface heat fluxes in driving the fast warming of the SH midlatitude band through multiple CMIP3 model simulations, and investigate the potential reasons that CMIP3 models simulate an underestimation of recent global OHC trends.

#### **Results**

Outputs of full-depth temperature, surface heat fluxes, and wind stress from one simulation of the late 20th century climate from 17 CMIP3 models were utilized (see Table S1 of auxiliary material section of Cai et al. (2010a)). The 17 models were selected for this study due to their availability of a control run to remove spurious (subsurface) drift. Monthly anomalies and trends of full-depth temperature, OHC (defined as the integrated temperature over the upper 2000m, as so to compare with observations from Wijffels et al. 2008 and Levitus et al. 2009), surface wind stress, wind stress curl, and surface heat fluxes were constructed for the late 20th century (1951-1999).

Depth accumulative global OHC trends are calculated for the two observational datasets over 1955-2009 (Fig. 1a diamonds) and 1960-2007 (Fig. 1a, square), and for two CMIP3 model ensembles over 1951-1999: 7 models with all radiative forcings (ALL, Fig. 1a, up-triangle) and 10 models with only anthropogenic forcings (ANT, Fig. 1a, down-triangle).



*Fig. 1* (a) Depth-accumulative global OHC trends over 1951–1999, averaged over CMIP3 model experiments with all forcings (ALL, up-triangles), and with anthropogenic forcings (ANT, down-triangles) only (i.e. without natural forcings). Individual model results are shown in Fig. 1a and 1b of Cai et al. (2010a). Shown also are observed OHC trend estimates over 1955–2009 (diamonds) from Levitus et al. (2009), and an estimate over 1960-2007 (square) from Wijffels et al. (2008). (b) Depth-accumulative trends from targeted modelling runs isolating natural and aerosol forcing components.

The results of Fig. 1a show that models without natural forcings (volcanic aerosols and solar fluctuations) are around 70% larger than observed OHC trend (at 2000m). Furthermore the ALL ensemble is around 40% smaller. The difference in OHC trends between the two groups cannot be assumed to be due to the inclusion natural forcings, as 5/7 models in ALL ensemble contain the cooling indirect aerosols effect, whereas 8/10 models in the ANT ensemble do not (see Cai et al. 2010a for more details). To isolate the individual impacts of natural forcings (predominantly volcanic aerosols) and anthropogenic aerosols on simulated global OHC trends, model ensemble experiments from the CSIRO Mk3A and NCAR PCM1 coupled climate models were also analysed, and are shown in Fig. 1b. Results show that natural forcings alone (Fig. 1b, dotted lines) are unable to account for the difference in the ALL and ANT CMIP3 ensembles (Fig. 1b, thick line). It is likely that anthropogenic aerosols, particularly the addition of the indirect aerosol effect, account for much of the difference; however as the CMIP3 ALL trend is 40% smaller than observed OHC trends, one such reason could be that CMIP3 models overestimate cooling of the indirect aerosol effect (Cai et al. 2010a).

CMIP3 models that include ALL forcings also better simulate the SH warming of the 35°-50°S midlatitude band, and the subsurface cooling in the SH subtropics (Fig. 2d of Cai et al. 2010a). An explanation for this warming is a result of two process: firstly, poleward strengthening winds, as driven by changes in the SAM (as a result of increased greenhouse gases, ozone depletion and anthropogenic aerosols), have shifted the subtropical gyres south by about 1° (Cai et al. 2010b). Wind changes impact the subsurface ocean via Rossby wave propagation, where heat is distributed slowly (rapidly) to a shallower (deeper) depth off the western (eastern) side of the South American continent. This eventually leads to a steady state

Sverdrup balance. Steric height calculations on the 50-yr CMIP3 model zonal wind stress trends (using a wind-only driven ocean model) confirm this, with wind changes effectively resulting in heat being stored in the 35°-50°S latitude band. However, this process is not adiabatic and requires a heat source. The second process is where this heat is coming from, and for this surface heat fluxes are used to calculate a simple heat budget over the SH oceans. Heat transport is calculated as the difference between actual OHC changes and the heat gain from the atmosphere (surface heat fluxes).

The results show that about 50% of heat accumulated south of 50°S (from the atmosphere) is advected northwards with very little heat transfer occurring north of the fast warming band at 35°S (~1%). The process responsible for this northward heat transfer is the wind-induced Ekman transport, allowing upwelled water at the surface to gain heat before sinking (and moving equatorward) along the steep isopycnal surfaces to form Antarctic Intermediate Mode Water (Cai et al. 2010b). CMIP3 models simulate this process quite well, however the ensemble OHC change is weaker north of 35°S, perhaps reflecting that models underestimate how much heat is derived from north of 35°S. These processes, driven by increases in greenhouse gases, anthropogenic aerosols and stratospheric ozone depletion have all contributed to the fast warming of the SH midlatitude band, however the extent to which each play their part is still unresolved. Adding to this, while none of the 'current-crop' of CMIP3 models resolve eddies there remains some uncertainty of the role of mean eddy fluxes in offsetting the increase in the northward Ekman transport.

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#### **OCEAN MIXING IN THE ACC**

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#### Introduction

#### *To Mix = To Put Together*

One's idea of mixing depends on the scales of one's interest. According to the American Meteorological Society *Glossary of Meteorology*, ocean mixing is "any process or series of processes by which parcels of ocean water with different properties are brought into intimate small-scale contact, so that molecular diffusion erases the differences between them". They draw a distinction between stirring, which moves the water parcels into intimate contact, and mixing, the final process of molecular diffusion that blends the water parcels together. However, they acknowledge that the term "mixing" is currently used to describe all of the processes, including molecular diffusion.

In practical terms, mixing is the combined effect of all processes acting to homogenize the ocean that are not resolved in observations or in a numerical model. For a climate system model of the ocean with 1° resolution, mesoscale eddies are not resolved and must be parameterized as mixing. For an eddy-resolving regional model, what is termed mixing might be any process acting at scales of 10 kilometres and smaller. Presumably, if we had perfect observations and perfect models, mixing would only be molecular diffusion.

#### The Southern Ocean View

In the Southern Ocean, the meridional circulation is thought to be the result of two competing pathways: a wind-driven circulation that acts to tilt isopycnals, and an eddy-driven pathway that acts to flatten isopycnals (e.g. Speer et al. 2000, Marshall and Radko, 2003) The result (ignoring Antarctic Bottom Water Formation and other things to the south) is upwelling of Circumpolar Deep Water south of the Polar Front, northward Ekman transport, and subduction of Antarctic Intermediate Water and Subantarctic Mode Water north of the Polar and Subantarctic Fronts, respectively (e.g. Speer et al. 2000). Mixing processes in the Antarctic Circumpolar Current are key to this circulation, controlling the rate at which water sinking at high latitudes in the North Atlantic returns to the surface in the Southern Ocean (Naveira Garabato et al. 2007).

Mixing process in the interior of the ocean can be divided into those occurring along density surfaces (isopycnal) and those mixing across density surfaces (diapycnal). The traditional view was that isopycnal mixing dominated in the ocean interior with diapycnal mixing present near the sea surface in response to air-sea-ice interactions. Estimates of interior isopycnal diffusivity ( $\kappa_{H}\sim 10^3 \text{ m}^2\text{s}^{-1}$ ) are many orders of magnitude larger than background diapycnal diffusivity observed outside the ACC ( $\kappa_{V}\sim 10^{-5} \text{ m}^2\text{s}^{-1}$ , Ledwell et al. 1993). However, recent observations of mixing in the Southern Ocean using physical and chemical tracers have shown that diapycnal diffusivity is of order  $\kappa_{V}=10^{-4} \text{ m}^2\text{s}^{-1}$  in the vicinity of rough topography, and can be up to  $\kappa_{V}=10^{-3} \text{ m}^2\text{s}^{-1}$  (Naveira Garabato et al. 2004, Naveira Garabato et al. 2007, Amelie Meyer PhD results). Estimates of the associated diapycnal upwelling indicate that it is only 1-2 orders of magnitude weaker than regional mean isopycnal upwelling, and comparable to the zonal average isopycnal upwelling (Naveira Garabato et al. 2007). Thus, mixing in the Southern Ocean is a juxtaposition of two players of comparable strength.

In the Southern Ocean, where direct measurements are feats of endurance for instruments and people, advances in our understanding of the circulation have been led by theoretical studies, numerical modeling and remote sensing. In this age of Argo, the large-scale circulation above 2000 m in ice-free regions is now relatively well observed. However, observations outside this zone, and observations of small-scale, rapid variability such as mixing processes are still very rare and costly. A UK-Australia collaboration called SOFINE (Southern Ocean Fine Structure Project) combined a series of traditional and novel instruments to measure mixing in the ACC at the northern Kerguelen Plateau and figure out the momentum balance in this region where the ACC collides with rough bathymetry. The impressive list of acronyms that describe the types of instruments used are: CTD, LADCP, SADCP, TSG, VMP, ISW, APEX, SOLO and EM-APEX. It is the EM-APEX observations that will be discussed in this talk.

#### New observations of mixing and other things

The EM-APEX is an enhancement to Webb Research Corporations' version of the Argo float, APEX (Autonomous Profiling EXplorer), developed by Tom Sanford's team at the University of Washington, USA. An electromagnetic subsystem is added to the Argo float, that includes a package with compass, accelerometers and electrodes to measure motionally-induced electric fields generated by currents moving through the vertical component of the Earth's magnetic field (Sanford, 1971). The electrical potential measurements are processed on board the float, and subsequently converted to horizontal velocities. Individual velocity profiles are relative to an unknown constant offset. By pairing profiles, and obtaining a GPS position for each, we can determine a full depth profile of absolute velocity.

Eight EM-APEX profilers were deployed in the SOFINE experiment, distributed at various dynamic heights across the ACC at the western edge of the northern Kerguelen Plateau. The EM-APEX profiled 4 times per day, resolving the inertial frequency, and returned approximately 1600 profiles each of temperature, salinity and horizontal velocity to 1600m depth, north of Kerguelen Island as they drifted eastward in the ACC. The vertical spacing of samples was approximately 3 dbar; horizontal spacing of profiles was between 2 and 10 km, depending on the speed of the float's drift during profiling and over a 4-hour rest at 1000m every 4<sup>th</sup> profile. The time spent at the sea surface was between 15 and 40 minutes.

The beauty of these data is that they give us coincident watermass property and velocity information and we can diagnose the dynamical processes leading to the observed watermass distributions. The data returned capture processes at many scales including:

- 1 The large-scale distribution of temperature, salinity and velocity along and across the ACC, upstream and downstream of the Kerguelen Plateau, from the surface to 1600 m.
- 2 The vertical structure of a mesoscale eddy and the implications of its tilt and vertical overturning to the transfer of heat and salt across the ACC.
- 3 Lateral stirring of density-compensated temperature-salinity filaments that we use to estimate isopycnal mixing rates across the ACC.
- 4 The identification and characterization of internal waves in the velocity profiles, and an estimation of the diapycnal diffusivity associated with the breaking of these waves.
- 5 The tracing of the internal wave rays back in time through a high-quality, time-varying climatology to identify their generation site at the surface or floor of the ocean.

- 6 The penetration of wind momentum into the surface boundary layer through Ekman dynamics. Ekman spirals are prevalent in the profiles and it is clear that steady state Ekman theory is insufficient to capture the variability of the spirals.
- 7 The penetration of surface heat and freshwater fluxes into the surface mixed layer, and potentially through the pycnocline and into the ocean interior.
- 8 Surface wave characteristics.

Investigation of 1-7 is in progress; 8 will come later. In the talk, Helen will present the most recent results from each of these analyses. We provide a preview of one area below.

#### **Ekman spirals**

The data reveal Ekman spirals (Fig. 1) - well-known in theory but rarely observed. We find equal numbers of anticlockwise-rotating (consistent with steady-state Ekman theory) and clockwise-rotating spirals. We are investigating whether the counter-rotation is due to wind forcing at frequencies higher than inertial (periods shorter than16 hours at 45°S) as found by Rudnick and Weller (1993). The Ekman decay scale in the steady state theory is the depth at which velocity decays by 1/e and rotates by 1 radian relative to the surface velocity. Fitting amplitude decay and rotation rates to each velocity profile, we find that these depths are not the same: a mean of ~36m for decay-based estimates from anticlockwise spirals, and ~57m for the rotational estimate from the same profiles. These incompatible depth estimates imply a compression of the observed Ekman spiral relative to theoretical predictions, also observed in Drake Passage by Lenn et al. (2009), and suggest that stratification of the water column must be given consideration.



*Fig. 1* Vertical profile of absolute horizontal velocity in the surface boundary layer observed with EM-APEX 3760. An estimate of the geostrophic velocity has been subtracted from this profile, equal to the vertical average of velocity components between 100 and 200 dbar.

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### SEASONAL PREDICTION OF THE TROPICAL INDO-PACIFIC: SUCCESSES AND CHALLENGES

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#### Introduction

Prediction of seasonal climate derives primarily from the ability to predict slow variations of the tropical oceans and, in particular, ocean surface temperatures associated with the occurrence of El Niño/La Niña. Although successful prediction of El Niño was demonstrated 25 years ago, many challenges remain in order to deliver useful and reliable seasonal climate forecasts. These include the ability to predict different varieties of El Niño and the surface temperature variations in the equatorial Indian Ocean associated with the Indian Ocean Dipole, for which Australian (and global) climate is sensitive. Here we assess our current capability to predict El Niño, including its different flavours (or different patterns of surface temperature variations) and its oceanic teleconnections, and to predict the Indian Ocean Dipole. We highlight common success and challenges by comparing results from our own forecast system (POAMA) to three other international forecast systems.

#### **Forecast models**

The primary results are derived from the seasonal forecasts produced the BoM POAMA seasonal forecast system (Hudson et al. 2010). Two versions of the system are studied here: POAMA1.5b and POAMA2.4c. The ocean model (based on MOM2 with 2° zonal by 0.5° meridional grid spacing) and atmosphere model (T47L17) components are identical in these two versions of POAMA. The only difference between the two versions is that POAMA24.c uses improved ocean initial conditions produced by the POAMA Ensemble Ocean Data Assimilation System (PEODAS; Yin et al. 2010). PEODAS assimilates available subsurface temperatures and salinity using an ensemble approach, where as the ocean assimilation used in POAMA1.5 is univariate in temperature with fixed temperature error covariances. In POAMA1.5 salinity is not adjusted after temperature is incremented, hence salinity is not in balance with temperature. Based on comparison to independent data, the initial ocean state depicted by PEODAS has reduced error, is more dynamically balanced, and results in more skilful forecasts compared to POAMA1.5b.

Assessment of forecast skill is based on a 10-member hindcast set that is initialized on the first of each month 1982-2008. Forecasts are initialized from observed ocean and atmospheric states (Hudson et al. 2010). Forecasts are 9 months in duration. The ensemble generation strategy is slightly different in the two versions of POAMA. POAMA1.5b uses a single ocean state and 6 hour lagged atmospheric states. POAMA24.c uses a single atmospheric state and perturbed ocean states as provided by the PEODAS assimilation. These different approaches result in initially different ensemble behavior (e.g. atmospheric spread is much less in 24.c), but these differences disappear after about 1 month of the forecasts.

We compare the forecasts from the POAMA system to those produced by the ECMWF ECSy3 (Molteni et al. 2007), the US NCEP CFS (Saha et al. 2006), and the University of Tokyo SINTEX-F (Luo et al. 2005). These forecasts were provided to us, respectively, by M. Balmaseda, S. Saha, and J.-J. Luo. These models

represent a cross section of the best practice in coupled model seasonal climate forecasting and, as for POAMA, all of then are used to make seasonal climate predictions in real time. Hence, they are indicative of the current capability to make seasonal climate forecasts. One fundamental difference in these systems is that POAMA, ECMWF, and NCEP all are initialized with observed atmospheric/oceanic initial states, where as SINTEX-F is initialized through a coupled relaxation to observed ocean surface temperature. Although the SINTEX-F model uses much less observed initial information (none in the atmosphere and none in the ocean subsurface), the coupled relaxation has the benefit of reducing shock. As will be discussed below, impacts of not using observed information at the initial time are evident early on in the SINTEX-F forecasts. However, some apparent benefit of reduced shock is evident at longer lead times.

The available hindcasts from each of these forecast models is of similar length and numbers of ensembles (typically 1982-2008 with at least 8 members). However, we only have available from ECSys3, CFS, and SINTEX-F the predictions of some standard indices of ocean surface temperature variability, including the Nino3, Nino4, Nino3.4 SST indices in the Pacific and the Dipole Model Index and its east and west components in the Indian Ocean. In all cases, anomalies are computed relative to the respective model climate, which is a function of start month and lead time.

# Predicting ENSO, Its flavours and oceanic teleconnections

We focus on some key results for the current capability to predict ENSO and its two distinctive flavours: El Niño events that have greatest amplitude in the eastern Pacific, referred to as cold tongue El Niño, and El Niño events that are more concentrated in the central Pacific, referred to as warm pool El Niño. Australian rainfall is more sensitive to WP El Niños (Wang and Hendon 2007; Hendon et al. 2009). We also assess the prediction of the ENSO oceanic teleconnection into the Indian Ocean (via the Indonesian throughflow region and onto the West Australian coast) and its implications for predicting variations in the Leeuwin Current. Key results are summarized as:

- 1 Improved ocean initial conditions in POAMA2.4c have increased the skill in predicting El Niño over that from POAMA1.5b by at least 1 month lead time. The skill for predicting El Niño using POAMA2.4c is now comparable to the best systems;
- 2 The "spring" predictability barrier is largely overcome through use of dynamical models that make use of initial subsurface ocean conditions;
- 3 Predicting key differences in the patterns of CT and WP El Niño are limited to lead times of 1-2 months (Hendon et al. 2009); beyond this, the patterns are indiscernible, reflecting drift of the CT ENSO mode. It is not yet clear whether the forecast models faithfully simulate the mechanisms of WP El Niño. Regional climate prediction across Australia at longer lead times is negatively impacted by this inability to discern two types of El Niño (Lim et al. 2009);
- 4 Simulating ENSO with realistic amplitude in the forecast models remains a challenge, which limits the ability to make regional climate predictions that are strongly reliant on ENSO teleconnections;
- 5 The MJO acts to limit the accuracy with which the strength of El Niño can be predicted because the details of the evolution of individual MJO events matter (Li et al. 2009). To represent realistic uncertainty, couped models need to faithfully represent the MJO so that adequate forecast spread is generated;
- 6 Although the MJO is not predictable beyond a few weeks, forecasts of El Niño are sensitive to the initial state of the MJO for a lead time of a few months because a coupled response is instigated by

7 The oceanic teleconnection of ENSO into the Indian Ocean is faithfully represented in the POAMA model, which means that prediction of the variations of volume transport of the Leeuwin Current offshore of Fremantle is skilful to a lead time of 9 months (Hendon and Wang 2009).

#### The Indian Ocean

It is now well understood that surface temperature variations in the tropical Indian Ocean play a key role in driving global climate variability, and, in particular, cool seasonal rainfall variations across southern Australia (e.g., Cai et al 2010). A primary driver of these variations is the Indian Ocean Dipole. Predicting the IOD has proven to be a much more challenging task than predicting El Niño (e.g., Zhao and Hendon 2009). Some key results of comparing POAMA's predictions to other forecast systems are:

- 1 The hit rate for predicting the occurrence of a IOD event in springtime (when the IOD has greatest amplitude and impact) drops to less than 50% after ~2 month lead time (3 months for the best system and 1 month for the worst system). In contrast, hit rates exceed 50% for the prediction of El Niño to beyond 9 months;
- 2 Improved ocean initial conditions result in negligible improvement in forecast skill of the IOD;
- 3 Models with overactive IODs (i.e. the standard deviation of the model's simulated IOD is greater than observed) tend to have lower skill;
- 4 Model's using observed initial states have higher skill initially;
- 5 There is little indication of any predictive capability of the IOD beyond that being driven by El Niño;
- 6 Improving the simulated relationship beyond the IOD and El Niño should lead to improved prediction of the IOD.

# Some key challenges

Model error is the key limitation now for prediction of the main driver of climate: ocean temperature variations in the Indo-Pacific. In the Pacific, an improved understanding of the mechanisms of warm pool (WP) and cold tongue (CT) El Niños is required in order to advance our modelling capability: unless we understand the different mechanisms, we won't be able to improve the models. The ability to discern WP and CT El Niños appears to be tightly tied to the simulation of the mean state. Hence, it is a major challenge to improve the simulation of the mean state of the Pacific (including the seasonal cycle and the salinity field) so that El Niño operates with realistic amplitude and locality. At a fundamental level, an improved understanding of the role of salinity variations for the evolution of El Niño is also required. We know that salinity in the warm pool varies systematically during El Niño, but we don't yet know its impact. Improved assimilation of salinity clearly impacts the mean state of the coupled model many months into the forecast, so we know the effects of salinity are not trivial. However, we don't yet know how vital the proper variation of salinity through the evolution of El Niño.

Simulating the mean state of the tropical Indian Ocean is as big a challenge as in the Pacific. The east-west slope of the thermocline along the equator is a key determinant of the strength and seasonality of the IOD, but this slope is notoriously hard to get right because it depends critically on the mean surface westerlies

along the equator, which themselves depend on a the distribution of near-equatorial atmospheric convection. Hence, improved simulation and prediction of the IOD will hinge on improved simulation of tropical convection. Simulation and prediction of the IOD also requires faithful depiction of coastal upwelling along Java-Sumatra. The sensitivity of this upwelling to model resolution needs to be established. As much of the predictive capability of the IOD stems from its connection to El Niño, an improved simulation of the teleconnection of El Niño into the Indian Ocean is also required, which will require improved simulation of the location and seasonality of El Niño in the Pacific.

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# TRACER-CONTOUR INVERSION: PATHWAY TO GLOBAL MIXING FIELDS?

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The main aim of physical oceanography, both observational and theoretical, since it began as a discipline in about 1900 was to estimate the ocean circulation.

In the last 25 years we have been so bold as to also ask about the strength of ocean mixing processes, mostly because our climate models need to know.

Starting in 1978 three inverse methods have been invented to use hydrographic data in order to estimate the ocean circulation and mixing processes; one was by Carl Wunsch of MIT, one by Hank Stommel of Woods Hole, and one by Peter Killworth of the UK. These inverse methods have been the basis for many (perhaps 100) PhD thesis and they have had some skill in pinning down the ocean circulation, but they have had limited success in estimating the mixing strengths or in providing estimates for the rates of formation of subducted waters (e.g in the Southern Ocean).

The Tracer-Contour Inverse Method builds on both Killworth's Bernoulli method and the box model method to provide a very clean set of equations in which the advection/diffusion balance is represented; there is no longer a need to assume a level of no motion. Also, the method is ideally set up to estimate the rate of subduction of water masses from the sea surface into the ocean interior.

The first three papers (all with Jan Zika as lead author) have proven the value of the method in regions of the ocean that are approximately ten degrees of latitude and longitude on a side. The method has proven able to deduce the strength of diapycnal mixing even when it is very small. The fact that the diapycnal diffusivity increases with depth and the lateral diffusivity decreases with depth comes clearly out of the model results.

On-going work in Hobart with Andrew Meijers and Bernadette Sloyan is aimed at applying the Tracer-Contour Inverse Method to a global hydrographic atlas in order to find maps of diapycnal and lateral diffusivities that will be useful to forward ocean models.

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# THE OCEAN AND PROGRESS IN SEASONAL FORECASTING

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# Abstract

Dynamical seasonal prediction has grown rapidly over the last decade or so. At present, a number of operational centres issue routine seasonal forecasts produced with coupled ocean-atmosphere models. These require real-time knowledge of the state of the global ocean since the potential for climate predictability at seasonal time scales resides mostly in information provided by the ocean initial conditions, in particular the upper thermal structure.

Assimilation of observations into an ocean model forced by prescribed atmospheric fluxes is the most common practice for initialisation of the ocean component of a coupled model. Assimilation of ocean data reduces the uncertainty in the ocean estimation arising from the uncertainty in the forcing fluxes and from model errors. Latest improvements in ocean data assimilation will be described, including the impact of the assimilation on the seasonal forecasts. Focus will be on the POAMA system from CAWCR, but will also include an inter-comparison with other international assimilation systems.

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# GOOS AND GODAE THE EARLY VISION AND REFLECTIONS ON UNFINISHED BUSINESS

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# Introduction

The idea for a Global Ocean Observing System (GOOS) was conceived in the late 1980's and formalised in 1991. The concept was built on an analogy with meteorological systems and the foundations emerging from the Tropical Oceans-Global Atmosphere Experiment and the World Ocean Circulation Experiment, much as the First GARP Global Experiment laid foundations for the modern global atmospheric observing system. Experimentation and scientific research were recognized as fundamental underpinning for GOOS but the mechanisms for translating from a research management and governance to sustained (operational) support was left open.

The Global Ocean Data Assimilation Experiment (GODAE) emerged in the late 1990's and, though it had common research roots with GOOS, it was conceived as a very different type of experiment – it was to demonstrate the feasibility and practicality of global ocean prediction, including within the context of GOOS (Smith and Lefebvre, 2008). Science was not the primary goal or outcome, but rather prototypes and operational systems supporting ocean prediction.

This paper briefly looks at the history of GOOS and GODAE, their connections to ocean research, and finishes with some thoughts on the future.

# The Global Ocean Observing System

Smith (2010) provides a number of perspectives on ocean science within the context of the 50<sup>th</sup> anniversary of the Intergovernmental Oceanographic Commission, including transformational events during the last 50 years and expectations of science and technology for the future. Trends in observing infrastructure emerge from that analysis, among other things.

As introduced above, GOOS was seen as a natural legacy from the large global-scale experiments TOGA and WOCE, but the emergence of a stable and effective system has proved rather more challenging than those early conceptual discussions envisaged. Considerable effort was devoted to the design but it was only the open ocean/climate aspects that gained any real momentum in the early years.

The Ship-of-Opportunity eXpendable BathyThermograph (XBT) network was regarded as one of the earliest candidates for transition and, indeed, by the late 1990's, a number of lines had been transitioned to a quasi-operational footing. The sea level network, established to monitor inter-annual and longer-term change, was also an early element of GOOS. Sea surface temperature observing networks were also claimed as an element of GOOS though it could be argued they owed much to numerical weather prediction and climate change for their existence.

The Tropical Atmosphere-Ocean array was one of the most significant legacies from TOGA, with around 60 upper ocean moorings stretched over the entire equatorial Pacific in a 16° band straddling the equator. Though they were logistically challenging to service, the array was superbly designed for tropical

phenomena, providing real time data of upper ocean variability and winds, the key element of the El Nino/Southern oscillation phenomenon.

To a large extent, these elements provided independent information on the ocean but two technologies from WOCE began yielding whole-of-ocean, integrated views.

The TOPEX/Poseidon altimeter was designed to measure variations in sea level across the globe, on a roughly 10 day repeat cycle. It was unable to measure absolute sea level but through calibration and an ingenious blending with long-term sea level data the altimeter has been able to provide global measures of sea level variability and change, a legacy that has been continued through the Jason series. The success of the altimeter owes little to GOOS – research use has been the main driver – but at some stage the community will have to settle on a means for long-term support.

Profiling float technology also emerged from WOCE. The *Argo* program has revolutionised the way we go about observing the ocean and provides one of the few examples where the design of GOOS was used to lower the priority of one technology (low density XBT lines) and favour of another (Argo profiles). The scientific strategy for observing the upper ocean has fundamentally changed because of Argo and its ability to deliver cost-effective, high-quality profiles of both temperature and salinity down to 2000m.

Recent meetings have invariably pointed to the partial completion of the climate components of GOOS (around 62% completed relative to the 1998 design) and the still strong dependence on research funding as major reasons for concern. To those who have been working on GOOS for over two decades, the picture looks amazingly bright compared to where we were.

Progress with the coastal and non-physical elements of GOOS has been less spectacular, this despite the fact that the coastal community probably arrived at a better strategic plan compared with climate. There are many reasons for this, one of which concerns the maturity of the science, the other with its local and regional nature which does not lend itself well to a global initiative, at least as it has been tried with GOOS.

The Integrated Marine Observing System developed in Australia is probably the stand-out exception for coastal GOOS. Through a mixture of good fortune and very wise planning and implementation, the Australian community has landed on a model that is delivering substantial value in the coastal region and in a framework that honours free and open exchange of data. The data management systems are world-leading and it looks like a system compared with an *ad hoc* collection of individual scientific pursuits.

This positive experience has not been replicated through the rest of the world, with the possible exception of Europe and parts of the US. It might just be a matter of time, but there may also be fundamental flaws in the approach. Smith (2010) discusses some potential alternative approaches to local and regional science issues where it is not global phenomena that are the binding force (cf El Nino or global ocean circulation). In this case GOOS would focus on aspects where it could add real value and give less attention to the top-down to the coastal GOOS concept.

# The scientific foundation of GODAE

As introduced above, the opportunity to develop GODAE arose from a number of scientific and technical factors:

- the critical advances provided by research programs like TOGA and WOCE;
- the development and maturity of remote and direct observing systems, making global real-time observation feasible;
- the steady advances in scientific knowledge and our ability to model the global ocean and assimilate data at fine space and time scales; and

• the genuine enthusiasm of the community and, in particular, the remote sensing community, to promote and implement *integrated* global observing systems.

The World Ocean Circulation Experiment had championed altimeter and scatterometer measurements, both of which would be critical for GODAE. It had extended the TOGA observing system globally through the ship-of-opportunity program among other initiatives, and had promoted new technologies such as surface drifters and subsurface floats.

GODAE was conceived and implemented as a finite period "experiment". The strategy for the development of GODAE products was built on the concept of a GODAE "Common" shared by and accessible to all GODAE Partners contributing to the goals and objectives of GODAE. Interestingly, this concept has a lot of features similar to the Creative Commons concept that is being used increasingly for licensing of data and information access.

The dependence on ocean data streams was recognized from the start, but also constituted a significant risk. GODAE believed altimetry was crucial, both in high-precision low resolution and low-precision high-resolution modes, the latter for resolving and initialising eddies. The fact that significant uncertainty still surrounds future altimetric missions suggests we have not yet provided the unequivocal evidence, or the appropriate mode, for transitioning these systems to a sustained footing. GODAE also championed high-resolution sea surface temperature products, both for its own purposes and for use by the numerical weather prediction community. The GODAE (now Global) High-Resolution SST project has developed a new generation of SST products, exploiting the full potential of the various sources of data.

When the first "gap analysis" was done for GODAE, the glaring weakness was *in situ* profiles of the ocean. Profiling floats developed in WOCE, combined with ship-based techniques, appeared to offer some potential but at that time it was just that: potential. *Argo* emerged from discussions in GODAE in early 1998, though thereafter it controlled and determined its own destiny. The governance model was similar to GODAE in that it was semi-autonomous and self-sufficient. In many ways *Argo* has become the flagship for ocean observing systems. *Argo* is providing data that is unparalleled in terms of quality and extent, particularly in relation to salinity.

The role of models and estimation tools was emphasized from the outset, drawing heavily on the experiences of WOCE, particularly in the approach to ocean state estimation and error characterization. The generation of globally consistent fields of ocean temperature, salinity and velocity components through the synthesis of multivariate satellite and in situ data streams into ocean models was a central theme of GODAE. From the outset it was recognized that quantifying the error and utility of GODAE products would be important and led to the development and establishment of metrics upon which ocean prediction systems could be evaluated and intercompared. Ultimately the integrity of operational systems rests upon the degree to which the user has confidence in the quality and assurance that the products are fit for the intended purpose.

Global high-resolution ocean model assimilation systems were the main focus of GODAE, initially eddypermitting, but later eddy resolving. Regional prototypes proved critical for development and for regional applications. Sector-specific systems (e.g., for global climate estimates) were also an important aspect, particularly in the context of ocean re-analyses. Along the way, many scientific questions arose, some of which remain valid post-GODAE.

- To what extent is the ocean predictable, and what are the scale dependencies?
- Does this predictability vary, for example in boundary currents, and does it vary with time (for example, in the presence of significant interannual variability)?
- What are the dependencies and limitations of this predictability for practical purposes?

GODAE has now moved onto its next phase but the scientific challenges are as daunting now as they were in 1998. More attention is being focused on coastal regions and biophysical models, and on climate change adaptation. Some useful design guidance for ocean observing systems is emerging and the ocean prediction community now provides powerful advocacy for ocean observations.

BLUElink, the Australian component of GODAE, began a few years after GODAE but in the last 5 years has been recognized as one of the leading players. A number of other presentations in this Workshop will show just how far it has come.

# Conclusion

GOOS and GODAE, though related, have taken rather different pathways to where we are today. The latter is similar in concept to the process taken with our ACCESS modelling system and its translation into an operational numerical weather prediction. An exciting mix of high-end science and technology, with a focus on the end application rather than the delivery of scientific papers. By most standards, GODAE will be judged a success, delivering both on its primary goal and opening up an exciting scientific field of research.

For GOOS, the legacy is less clear, despite the many fine contributions on its behalf. GOOS is not driving scientific agenda and is not providing the international advocacy for ocean observations that was envisaged 20 years ago. It is high profile in intergovernmental circles such as within the Intergovernmental Oceanographic Commission but, with the exception of the Joint Technical Commission for Oceanography and Marine Meteorology, this has not translated into an effective implementation mechanism. There is much good science that is relevant to GOOS but little that owes its existence to GOOS.

There is little doubt in my own mind that oceanography related to observations and ocean prediction is in a more exciting and healthier state now than at any time in its history. It is comforting to know Australian science, and particularly that within CAWCR, is at the leading edge.

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# HOW TO BUILD AN OCEAN FORCASTING CENTRE FROM SCRATCH

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I begin by noting that this is not exactly how the Naval Oceanographic Office (NAVOCEANO) has gone about developing our modeling capability! But if we could start over, I can only hope that we'd consider our experiences and follow some of the recommendations presented below. For those of us that already have an operational center, I expect that none of this is new but hope that I can provide some useful tidbits. For those of us thinking about it, perhaps I'll be able to provide some useful guidance. Modeling has been equated to making sausage, so let's see how we might do it right.

# You need a reason to do ocean modeling.

Specific tasking must be backed up by continued support. There must be a willingness to spend the money necessary to develop an operational ocean modeling system that works, both initially and in the future. This shouldn't be undertaken lightly as everything about this business is expensive, takes more time than expected, and isn't always done correctly the first time. There is a vast difference between ocean modeling in an academic environment and the actual operational daily delivery of complete, accurate, and timely products. The commitment needed to transition an ocean model from Research and Development (R&D) to full operations is tremendous. For the US Navy, the commitment is the result of our need to understand, work with, and exploit our operating environment. Some examples include safety of navigation; energy efficiency; underwater sound propagation; currents; water clarity; density structure; waves and surf; and so on. I wouldn't suggest someone take on this sort of a project without a lot of money and dedication—you should consider having some real requirements and finding a few partners if you really want to build an operational ocean modeling center.

## We will start with hardware requirements.

Large supercomputers are vital. To resolve mesoscale ocean processes, we work with horizontal resolutions that are at least one-fourth of those of our atmospheric partners, and, in some cases, we must get this down to tens of meters. We need enough vertical levels to define mixed layer thickness, the thermocline, and the deep sound channel. The current Global Navy Coastal Ocean Model (G-NCOM) covers the ocean pole-to-pole at 1/8 degree with 40 levels. For much of the ocean, this is not adequate for Navy operations, so we are implementing a 1/12 degree Global Hybrid Coordinate Ocean Model (G-HYCOM) and plan to halve this to 1/25 when we have enough computer power (late 2104?). Even this does not meet some requirements, so we have set up a number of nested 3km (1/36 degree) regional NCOM and even 300m (1/360 degree) coastal NCOM domains.

Thus, the computers need to be big and powerful so that the daily model runs complete quickly enough to be useful for Navy operations. Input-output and internal memory both have to handle massive amounts of data efficiently—an issue that has led to a lot of creative work by our R&D partners. We are fortunate at NAVOCEANO that our mainframes are refreshed every three to five years with new state-of-the-art systems.

The downside of this is that each new system requires a painful transition of our models and scripts. It is apparent that we have not achieved the concept of "transportable" ocean models. Obviously just having the computer power is only part of the story and we need the expertise to keep the systems up and running. We rent our major modeling capacity from the Department of Defense Supercomputing Resource Center (DSRC) —a national R&D system—so it provides these operational needs. The DSRC currently has IBM and Cray mainframe computers with over 240 teraflops of computing power (<u>http://www.navo.hpc.mil</u>). NAVOCEANO has operational access to about 15% of this with 1,380 cores on IBM Power 5+ and 6 systems and 1,530 cores on a Linux-based Cray XT5.

We will also need plenty of storage, including both rapid access and deep archiving capabilities. Just our daily G-NCOM NetCDF files are about 25 gigabytes, but there are about 600 gigabytes of data that float around during its production. Today's HYCOM quadruples these values. What to keep and what to discard is a constant dilemma. A properly constructed, reasonably accurate ocean model can be the source for an ocean climatology that should be better than anything that the analysis of historical data could produce, but we'd need to store years of data to do this right (atmosphere people tell us 30 years). An operational center like NAVOCEANO is unable to go back and rerun the models, so we are trying to save at least the daily analysis fields for such a climatology as well as exercise reconstructions and responses to planning questions such as "What will the ocean be like out there next February?" The Naval Research Laboratory (NRL) is preparing a 15-year retrospective analysis of the ocean using NOAA Global Forecast System (GFS) atmospheric forcing and G-HYCOM (see below). This will go a long way toward developing a model-based climatology and setting up scenarios for training, but the storage requirements will be massive.

Just having the data somewhere is not enough, however. We have over half a petabyte of data in storage, but getting the information we need out and analyzing it in a timely manner is not so easy. For our "new system" I'd design the storage directories as well as naming and content conventions up front so they were carefully regulated and consistent. This would include a ready-access to deep archiving system that was automatic, a plan to routinely purge expired data and just keep the useful stuff, and a quick and easy extraction system that would allow us to find, subset, stage, and deliver only what is needed. Good, fast data compression and retrieval capabilities are essential. This means we should develop a "system specification" before the modeling even starts (more on this later).

In addition to the "big iron," we need an infrastructure that collects, stores, and moves observation data, atmospheric forcing fields, and boundary conditions into the modeling environment. It must also move finished products back out. At NAVOCEANO, this hardware and software are maintained by our engineering department (separate from DSRC). This means we work with two separate providers and must adhere to the rules and functions of both while we transfer data back and forth. Security and information assurance are always issues. Thus, a modeling system that will depend on externally managed computational power will need to ensure that the surrounding infrastructure is robust, compliant, well protected, and closely coordinated with the provider.

# An equally important resource is people.

We have a modeling operations team that works with the developers, computer providers, data managers, and customers to monitor and manage our modeling system. The team's areas of expertise include system engineering, data management, software design, ocean modeling, and system troubleshooting. Each person has a few production responsibilities although we are developing standard operating procedures (SOPs) that document our main functions to get us away from a traditionally "one deep" situation. We have established watch-standing duties to monitor the daily progress of our main model production schedule. This is not a full-time job as we have built the model runs to be automated, so the watch consists of following a check sheet, reviewing a series of monitoring web pages, and recording problems and resolutions. Most of the

watch functions are this review and reporting, troubleshooting using the SOP, or calling in an expert for the really difficult problems. We will talk more about the automated monitoring system later.

Of course, the list of people who contribute goes far beyond the immediate model operations team. Computer operators for the DSRC and NAVOCEANO hardware keep the computers humming and information flowing. This distribution network extends beyond our walls and into the Navy's data delivery systems and, thanks to our NOAA partners, to external users via their data servers (see below). The R&D scientists at NRL and other government or private labs develop ocean models, deliver upgrades, and help us solve problems beyond our capabilities. We have skilled oceanographers on staff with extensive modeling backgrounds (many at doctorate level) that work with NRL during the transition of a system and become the system "owners" at NAVOCEANO.

As providers of our atmospheric forcing data (see below), the Fleet Numerical Meteorology and Oceanography Center (FNMOC) in Monterey, California has an equally large production staff for delivering weather data to the Navy. NAVOCEANO is well known for our bottom data collection and analysis skills, and the Bathymetry Department is the first tasked to provide an accurate "ocean bottom" when we implement a new model domain. Observations for the correction of our models by assimilation (also discussed later) are collected by a vast number of *in situ* and remote systems and collated by the NAVOCEANO Data Collection Department. At the end of the production cycle, someone has to figure out what all these numbers are telling our users. We are developing a team of "ocean forecasters" that interprets the model fields in the context of Navy operations.

Leadership and oversight are necessary to keep "the vision" intact, translate requirements into capabilities, weigh these against what is available, plan for and articulate future needs, keep management informed on what can and can't be done, maintain schedules, and ensure that the products get delivered. Bottom line, if we were to build an ocean forecasting organization that was dedicated only to operational model production, it would require nearly 100 people with advanced skills in a wide range of categories. And, even then, we would be dependent on a lot of outside help to get the needed information into our oceanographic center and the products back out to the users.

## Next we need some operational models.

The models need to produce information that matches the sponsor's requirements—and this is where pure scientific development and operational needs often diverge. That is, if your main concern is safety for ships at sea, weather and waves will be most important, and there is little need to resolve deep ocean processes. For this application, the locations and intensities of ocean fronts and eddies really don't matter much. If you are conducting Anti-Submarine Warfare (ASW) and want to find a submarine and hold contact, you need to know about the mixed layer and upper thermocline which, unfortunately, are changing constantly due to internal tides. Below the axis of the deep sound channel (about 1000m), the ocean doesn't change enough to warrant high vertical resolution, so most of our modeled layers are near the surface. If you want to hunt for mines placed in harbors, you need to know currents from both tides and river runoff. Here, wind effects are generally minimal (until a storm event stirs things up and sends a lot of water downstream!). If you are planning an amphibious landing, waves and surf, coastal and longshore currents, and the shape and composition of the beach and offshore bar become important, so a high-resolution, integrate-process modeling system like Delft3D is needed.

These are Navy-centric examples. The recent Gulf of Mexico oil spill demonstrated a need for a threedimensional current model that can reliably forecast pollutant trajectories and behavior far into the future. This includes density structure, currents, and cross-slope processes. Coastal engineers need models that show the long-term and/or catastrophic effects of currents on their designed structures. Biologists what to know where the ocean is favorable for spawning lobsters and where the little creatures will go as they grow. Aquaculturists need to know if their cages will be lost due to storm currents or whether the fish detritus will be swept away by the tides. And so on... User applications will dictate how you will concentrate your modeling efforts.

An ocean model is based on a few simple rules of physics. Newton's second law, force equals mass times acceleration, is translated into the description of environmental processes by the Navier-Stokes equation. This, along with conservation equations and water density's relationship with temperature, salinity, and pressure, makes up the so-called "primitive equations." Transcribing these into a numerical form is the art of modeling. In Physical Oceanography 101, we are told that the only place that you can dynamically influence ocean properties is at the surface, where the atmosphere provided momentum and heat fluxes. Everywhere else, only mechanical mixing can change temperature and salinity. Thus the atmosphere and ocean are coupled, and weather modelers that support the oceanographers need to pay attention to these air-sea exchanges. Unfortunately, they usually ask us for only sea surface temperature (we're working on improving this coupling!). Since we never have enough resolution to describe all processes and we need to make computational compromises, the relevant sub-grid scale processes must be estimated or parameterized. The idea that the ocean is a relatively benign three-layer environment that is taught in this same course is far from the real world. The mixed layer isn't all that well-mixed, and internal waves move the "stratified" thermocline vertically at tidal time scales. Only the deep ocean is relatively constant (unless you are concerned with its capacity for heat storage as it relates to global change) but most of our 40 layers are near the surface, so we aren't properly modeling deep water features like Antarctic Bottom Water. The challenge of ocean modeling is to depict this dynamic environment in a manner that meets the customer's requirements, and unfortunately there is not a universal version that answers everyone's needs. As a result, we end up running a suite of models. The objective should be to minimize the number we have in order to reduce operational complications.

NAVOCEANO runs a set of circulation and wave models that update ocean conditions daily. Today's primary ocean model is the G-NCOM with a nominal horizontal resolution of 1/8 degrees (14km/7.5nm) and 40 vertical layers. It is based on the Princeton Ocean Model (POM) with numerous modifications by the developers at NRL Stennis. NCOM is forced by FNMOC wind stress and heat flux forecasts from the half-degree Navy Operational Global Atmospheric Prediction System (NOGAPS). Observations from satellites (remotely sensed sea surface temperature and elevation) and surface and/or profile temperature and salinity data from a number of sources are assimilated via the Navy Coupled Ocean Data Assimilation (NCODA) system. Tidal forcing is not included in G-NCOM although barotropic tides from the Oregon State University Tidal Inversion Software (OTIS) are linearly added afterwards. Daily forecasts of temperature, salinity, eastward and northward currents, and surface elevation are delivered as graphics and data fields at 3-hour intervals to 96 hours, with work underway to extend this forecast period to eight days.

The 1/12 degree G-HYCOM is running in a pre-operational mode at NAVOCEANO and is expected to replace G-NCOM after an operational test in 2011 (see below). It is also based on the POM. Atmospheric forcing and data assimilation are the same as NCOM. Our initial version of HYCOM will have tides from OTIS added linearly although work is underway to integrate them into the model forcing fields.

Nested within the global model are higher resolution regional and coastal NCOM domains using the same algorithms as G-NCOM. The regional domains are normally 1/36 degree (3km/1.7nm) grids with 55 layers. The inner nests of coastal domains are as high as 1/360 degree (300m) resolution with the number of vertical layers based on water depth. Depending on requirements, forecasts to 96 hours at 1 or 3 hours are provided. These are forced by 15km Coupled Ocean/Atmosphere Mesoscale Prediction System (COAMPS) fields from FNMOC.

Work is underway to add Delft3D to the production suite for coastal and estuarine support. A Polar Ice Prediction System (PIPS) forecasts ice properties in the Arctic using the Los Alamos Sea Ice (CICE) algorithms. An Arctic ice cap forecast system coupling PIPS and HYCOM via the Earth Systems Modeling Framework (ESMF) will become operational in 2011, with the global HYCOM and PIPS coupled by 2012. Truly integrated environmental modeling systems are emerging today and the transition of an ESMF coupled ocean-atmosphere-wave modeling system, entitled COAMPS-OS, is planned for 2011. The ocean, atmosphere, and wave models are NCOM, COAMPS, and Simulating Waves Nearshore (SWAN), respectively.

The NAVOCEANO Wave Model (WAM) provides twice daily wave forecasts at global and regional levels to 48 hours, and the SWAN model forecasts high resolution wave and surf properties in the nearshore. We anticipate that WAM will be replaced with the latest version of Wave Watch III in 2012.

Now, if we were to design our modeling system from scratch, we would have created a set of specifications that told the R&D people what we needed to accomplish and would have described in detail the operating environment into which it would need to fit. Obviously, this is nearly impossible as, for example, it has taken ten years to bring HYCOM from conception to reality (and it was heavily based on the experiences gained from the NCOM development). Often, the reality of available computational power dictates progress. For example, the operational implementation of NCOM had to wait until a 2006 DSRC upgrade was in place. The concept of model portability is a myth—you can anticipate better performance when new computers arrive but you had better plan to take time to translate the algorithms into the new (or "improved") operational environment.

# These models rely on data.

As mentioned, atmospheric forcing by FNMOC's global NOGAPS and regional COAMPS provides fluxes of momentum (wind stress) and heat (solar, infrared, latent, sensible) to the ocean. We rely on FNMOC as it pays special attention to the air-sea interface to ensure that our ocean models receive this information properly. As with the ocean models, computational assets are always a problem, and we generally use a 5:1 step-down from the atmosphere to ocean.

Ocean observations are routinely collected and collated for our modeling effort. These data serve two purposes: (1) correction of the initial model analysis fields by assimilation and (2) metrics for the assessment of the models' forecast skills. Data include remotely sensed (satellite) sea surface temperature and altimetry, surface temperature, and *in situ* subsurface temperature and salinity profiles from a variety of platforms. Current and drift observations are collected for assessment but not yet assimilated. NCOM and HYCOM are initiated with quality-controlled observations via NCODA using a Multi-Variant Optimal Interpolation (MVOI) scheme. Plans for replacing MVOI with the variational scheme 3DVAR are well underway and 4DVAR is awaiting further development, as well as a lot more computer power. With only surface forcing and without the corrections provided by data assimilation, our models would not be very good as they quickly reverted to climatology.

# More discussion on infrastructure is needed.

At NAVOCEANO, we have spent a lot of time and effort trying to refine the management of data flow through our system, but it is still very complicated. We've prepared an extensive "wiring diagram" to document this. Because of its complexity, coupled with the fact that we don't have the people (or need) to oversee it 24/7, model production has been extensively automated. A Linux-based collection of scripts called the "Traffic Cop" ensures the data and products get where they need to be on time. Traffic Cop is linked to a group of production instructions called the Modeling Output Graphics System (MOGS). Here, new product

requirements are entered via web-based tables and translated into scripts that then direct their production. All modeling processes must produce log files whose entries are collected and evaluated by a MySQL database called ROAMER. Web-based tables that monitor production are maintained by ROAMER with green entries indicating all is well and various other colors alerting the model operations watch that processes are late, deliveries are incomplete, computer systems are down, and so forth. ROAMER can also either activate fallback or recovery routines or suggest how to troubleshoot and resolve issues. All this is to demonstrate that to be truly operational, you will need to design and maintain robust modeling systems that keep to schedules, can "self-heal" and will only require human intervention when things go wrong.

To quote Hendrik Tolman of NOAA, models just provide guidance for a forecaster. As noted, we are developing a team of ocean forecasters who are translating the numeric outputs into "actionable" information. The ocean forecaster's role is to tell the customer what the ocean looks like, how it relates to their requirements, how he/she can best use the information, and where the strengths and weaknesses of the models can be found. To do this, we have developed a number of viewing and analysis environments and tools enable a forecaster to view and manipulate graphics, test ideas, answer questions, find alternatives, and prepare briefing material that will tell the operators what they need to know.

#### We complete the system with assessment, transition, and oversight.

Once our ocean modeling system is up and running, we need to demonstrate that it is functional, reasonably accurate, robust, and ready to meet the sponsors' requirements. Internally, the first step is to show that the delivery can run in the operational environment-properly using the available assets and meeting the necessary schedules. This is no mean feat as you really need to work with R&D to ensure that the model is efficient and that the output gets to the customer before it is useless. The Navy has developed a three-level development and transition process that is governed by a headquarters-based Administrative Modeling Oversight Panel (AMOP). Milestone I is met when the R&D center (i.e. NRL) demonstrates the scientific veracity of the new system, generally through the acceptance of peer-reviewed publications. A Validation Test Panel (VTP) is established with members of the R&D and production centers. Guided by the VTP, the R&D center prepares a series of Navy-relevant tests to demonstrate that the transition will meet operational needs, resulting in a scientifically-oriented Validation Test Report (VTR). Milestone II is passed with AMOP approval of the VTR. During this period, the pre-operational system is installed in our production center job stream. After we start to run the system on our own, an Operational Test (OPTEST) is completed to show that NAVOCEANO (or FNMOC) has fully integrated the delivery into its production cycle and found it meets Navy requirements. AMOP Milestone III is met with the acceptance of the OPTEST report, indicating the new model or upgrade is approved for operational use.

Even after Milestone III, we need to keep our eye on the models' strengths and weakness and inform the users accordingly. The NCODA system includes a number of model assessment tools that routinely document model skills and depict ocean property parameters such as regions of higher error. NAVOCEANO and NRL have developed a program (entitled Auto-Metrics) that routinely collects and compares temperature and salinity profiles from models that are concurrent with shipboard XBT, CTD, profiling float (e.g., ARGO), or glider observations. While this program is relatively new, we hope to gather enough information to determine reliable error statistics as well as spatially and seasonally changing model skills in order to advise customers on the levels of model uncertainty. We are using limited ensemble approaches to guide sampling and further assess model variability and skill. How to pass on this uncertainty as part of the model product suite remains a challenge. Our Navy users are slowly being weaned from relying on out-of-date analyses and learning that our two- to five-day forecasts are accurate and useful for real-time planning and operations.

### In summary.

First, you need a purpose to model the ocean and a sponsor who will support your efforts. You need lots of reliable computer power, including processors, speed, memory, storage, and transfer capabilities. You need people who can develop the models, transition them into operations, keep the data flowing, monitor the process, fix problems, and translate the results into useful information. You need some models that run in a reasonable amount of time and consistently produce results that are acceptable to your customers. You need observations and an assimilation system that keeps the model honest. You need someone to oversee the system that can maintain the vision, keep the plans, manage its various parts, and tell leadership what can and can't be done. Finally, you need customers that tell you what they need and who actually rely on the products.

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http://edac.northerngulfinstitute.org/opendap\_index.html.

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# **BLUElink OceanMAPS**

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## Introduction

The BLUElink-2 project (2008-2010) provided an opportunity to consolidate the Ocean Model, Analysis and Prediction System (OceanMAPSv1.0, Brassington et al. 2008), analyse the mesoscale ocean dynamics represented by the model-data state estimations, identify and optimise weaknesses in the system and implement a major upgrade. Throughout this same period the routine delivery of ocean products have enabled events to be tracked, users to be identified and in some cases see genuine impact. In this short presentation we will focus on a number of highlights that have guided the BLUElink-3 three year workplans and the future for ocean services.

## What ocean does OceanMAPS represent?

At its core the prediction system is based on an Ocean General Circulation Model (OGCM) which is a numerical representation of the governing equations of ocean circulation. In this case, the software are various versions of the Modular Ocean Model version 4 (Griffies et al. 2004) from Geophysical Fluid Dynamics Laboratory (GFDL) consistent with that of ACCESS In this case the ocean is assumed to be incompressible (i.e. conserves volume) relative to a reference compression otherwise known as potential temperature/density, hydrostatic (i.e. vertical momentum is negligible or is otherwise parameterised to both represent the physics and manage numerical stability). In the OFAM model (Schiller et al. 2008) the horizontal gridscale ( $\Delta x \ge 10$  km) greatly exceeds the vertical gridcale ( $\Delta z \ge 5$ m) and in general this assumption remains valid. Vertical convection processes such as occur in the mixed layer are parameterised used a turbulence scheme (Chen et al. 1994) or deep convection parameterised using instantaneous cell to cell mixing.

The ocean exhibits temporal scales from seconds through millennia and spatial scales from mm to basin scales. The specific target of OFAM is the so-called geostrophic turbulence of eddies and fronts which ranges from O(10-100) km. MOM4 uses second order numerical methods which resolve length scales  $>4\Delta$ . The horizontal resolution of  $1/10^{\circ}$  can resolve fronts exceeding 40 kms or eddies of diameters >80 kms. The OFAM therefore only partially resolves the spectrum of geostrophic turbulence. A resolution of  $1/30^{\circ}$  therefore approaches fully resolving geostrophic turbulence.

An important physical process that is resolvable by an ocean model using 1/10° resolution is tides. Tides in a global model can be explicitly modelled using a body forcing whilst for a regional model both a boundary condition and synchronous body force is require (although for sufficiently small domains the body forcing contribution becomes negligible). At present, OFAM and therefore OceanMAPS does not explicitly represent tides.

Estimating the ocean state requires a background field, an assimilation method and an observing system. The BLUElink Ocean Data Assimilation System (BODAS) is based on an EnOI scheme (Oke et al. 2008) where

the background error covariances are diagnosed from the spatial covariance of the dynamics resolved by the ocean model. The in situ observing system is sparse with the Argo array targeting one float for every  $3^{\circ} \times 3^{\circ}$ in the open ocean complemented by moorings and XBT's. Sea surface temperature (SST) is observed by multiple satellites providing high temporal and spatial coverage. Sea surface height anomalies (SSHA) are observed by multiple narrow-swath altimeters currently based on two Jason-class 10 day (256 passes) repeat orbits and track separation at the equator of 156 kms and Envisat 35 day (1002 passes) repeat orbit and an equator track separation of 39.9 kms. The time-space resolution of altimetry is therefore a balance between the time-space scales where the background errors decorrelate. EnOI (and other 3D DA methods) perform an analysis for a specific central time at each analysis which does not formally account for the temporal decorrelation of background error which is the motivation for 4D schemes. However, whilst ocean modelling continues to pursue a higher resolution pathway, the use of 4D schemes remain beyond the resources of computational resources. OceanMAPS currently uses a time-window for altimetry of 10 days which provides a complete orbit of Jason1 and Jason2 and a minimum equatorial spatial resolution of 78 km. The same 10 day window is applied to Envisat which provides an incomplete orbit and therefore provides higher spatial resolution in different regions with each analysis. The data volumes of remotely sensed SST and SSHA have been shown to have a dominant influence over ocean analysis as compared with in situ profiles. The limitations of the background error covariance (used in BODAS) into the vertical limit the constraint of errors with depth.

In this presentation, we will address specific questions relating to the fitness for purpose of the present system and the motivations for current development pathways. These will include:

## Where does the physical ocean matter?

In developing any systems including an ocean prediction system it is sufficient to have one really important application (e.g., Tsunami's) or in the case of BLUElink Australia's defence. The Royal Australian Navy have been intent on supporting BLUElink to develop a tactical advantage to the defence forces. However, at the inception the expectation from the Bureau of Meteorology was that ocean forecasting would not only support defence but a raft of other applications and thereby rank alongside Numerical Weather Prediction as a core public service. We will present an overview of some of the applications that have emerged and the impact.

#### What have we learnt about the system?

A growing activity within BLUElink, nationally and internationally is the development of performance monitoring and intercomparisons (Hernandez et al. 2009). We will review what has been learnt about OceanMAPS and outline some of the next steps toward extending the suite of metrics, establishing an international common for an operational monitoring for routine intercomparisons and the first steps toward consensus forecasting.

## How do we extend the performance?

Having identified a number of high value variables (e.g., SST, heat content surface currents, sonic layer and coastal sea level) what is the current performance, are there risks to that performance and what is planned to extend that performance.

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## DEVELOPMENTS OF BLUELINK OCEAN DATA ASSIMILATION SYSTEM

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#### Introduction

Ocean data assimilation traditionally involves the assimilation of ocean observations to initialise an ocean model. Ocean observations that are assimilated are traditionally sea surface temperature (SST) and sea-level anomalies (SLA) from satellites, and in situ temperature and salinity (T/S) from a range of sources that include moorings, ship-borne transects, and profiling floats. Ocean variables that are typically updated include T, S, velocities, and sea-level. The ocean data assimilation capability developed under the Bluelink project employs an ensemble-based method to combine ocean observations and model fields (e.g. Oke et al. 2008). This capability has been used operationally for several years (Brassington et al. 2007), for both high-resolution short-range applications (e.g. Schiller et al. 2008), and seasonal applications (e.g. Yin et al. 2010). The applications to high-resolution short-range applications, under Bluelink, have benefited by recent improvements in the ocean model, and in the initialisation. These improvements are described in this article, along with a preliminary demonstration of the improvements.

The problem of data assimilation can be thought of as a projection of ocean observations onto ocean variables. However, the employment of an ensemble-based method for data assimilation readily facilitates the extension of the ocean data assimilation system to coupled applications. This is achieved by simply augmenting the ensemble with other variables. This could include atmospheric variables, biogeochemical variables, sea-ice variables, and so on. This capability has been developed in the Bluelink system by generalizing the code, so that the variables to be analysed are no longer prescriptive. Instead, the variables to be analysed are simply an array of two-dimensional and three-dimensional. Provided an ensemble of a variable can be constructed, the variable can be analysed. In time, this development could be used for coupled data assimilation.

As a step towards coupled data assimilation, we first consider the simpler problem of coupled initialisation. Consider an application to either a coupled ocean-atmosphere model (two-way coupling), or an ocean model that is forced by the atmosphere (sometimes *deceptively* called one-way coupling). Coupled initialisation is different from coupled data assimilation. Within this article, we differentiate these methods as follows. We refer to coupled data assimilation as the case where the state of both the ocean and atmosphere are updated together, using observations from either the atmosphere or the ocean, or both. In this case, the atmospheric state and the oceanic states are adjusted and both components of the coupled model is reinitialised. By contrast, we regard coupled initialisation as the process where the ocean observations are projected onto atmospheric variables that are immediately relevant to the ocean (i.e., surface fluxes). In this case, the ocean "feels" the adjusted fields, but the atmospheric model remains unchanged.

With this perspective in mind, coupled initialisation answers the question: "Based on the model-data misfits, how should we adjust the surface fluxes, so that the ocean model produces an ocean state that is more consistent with observations".

# Data assimilation system

The data assimilation system developed under Bluelink is based on ensemble optimal interpolation (EnOI; Oke et al. 2002; Evensen 2003). The details of the Bluelink system are described by Oke et al. (2008). Fundamental to EnOI is the employment of a time-invariant (Oke et al. 2005; Fu et al. 2008; Counillon and Bertino 2009), or seasonally varying (Brasseur et al. 2005), ensemble to approximate the system's background error covariance. The background error covariance of the system typically includes variables that are directly observed (e.g. T/S, SST, SLA) and variables that are not observed (e.g., velocities). The adjustments to the unobserved variables are based on the covariances that are implicit to the ensemble. If the errors of the unobserved variables are correlated with the errors of the observed variables, according to the ensemble, then the adjustments to the unobserved variables will be non-zero. By simply augmenting the ensemble with additional variables, the EnOI system can readily project the observations onto those additional variables. If the errors of the additional variables are fully project the observations onto those additional variables. If the errors of the additional variables will be non-zero. As a first test of this, we have augmented the ensemble with surface flux fields including wind stress, precipitation, evaporation, and the various components of the surface heat flux. This includes short-wave radiation, long-wave radiation, latent heat flux, and the sensible heat flux.

For the experiments presented below, the ensemble is constructed by computing intraseasonal anomalies from a spin-up run of the Bluelink ocean model. The surface forcing for this spin-up run is 6 hourly fields from the European Center for Medium Range Weather Forecasting (Kallberg et al. 2004). The specific product used here is ERA-Interim. The ocean model is forced by these fields and is free to evolve, without data assimilation. This yields a time-varying picture of the ocean circulation that is consistent with the prescribed surface fluxes. The ensemble captures the covariability of the ocean variables and the surface fluxes for zero time-lag. The next step that we haven't addressed yet will include a time lag. This may be important, because the response of the ocean to the atmosphere often has a time lag of a few hours to a few days.

# Ocean data assimilation

Recent developments of the Bluelink ocean data assimilation system and the under-pinning model – without coupled initialisation – have yielded good improvements in the system performance. Tested through a 5-year ocean reanalysis, spanning the period 2005-2009, the quality of the reanalyzed state of the ocean in the latest reanalysis is an improvement over previous versions. To demonstrate this, we compare the standard deviation of SLA with the root-mean-squared error (RMSE) of SLA in Fig. 1. This figure uses along-track SLA (atSLA) from altimetry. For each grid point, atSLA data from within 100 km are grouped and either the standard deviation is computed (Fig. 1a), or the RMSE is computed (Fig. 1b,c). To produce Fig. 1b,c, it is implicitly assumed that the atSLA data are error free. This is clearly untrue, but for the purposes of evaluation, the atSLA data provides the best available measure of the true state of the ocean. In reality, the expected error of the atSLA data is 3-5 cm.

The fields presented in Fig. 1 include results from version 2p1 of the Bluelink ReANalysis (BRAN) that is described by Schiller (2008), and version 3p0 of BRAN that has only recently been performed. The main differences between BRAN2p1 and BRAN3p0 are the source of the ensemble, the data assimilated, and the initialisation scheme. Briefly, BRAN3p0 uses fields from a more recent version of the Bluelink model that includes a parameterisation of vertical mixing and bottom stress that accounts for tidal mixing, improved topography, and improved surface fluxes. BRAN3p0 assimilates AVHRR and AMSR-E SST observations, while BRAN2p1 only assimilates AMSR-E observations. Finally, BRAN3p0 uses an adaptive nudging scheme (Sandery et al. 2010) for initialisation, while BRAN2p1 uses Newtonian nudging.

The improvements in BRAN3p0 over BRAN2p1 are most obvious in the Tasman Sea, between about 30-40°S, and along the path of the Antarctic Circumpolar Current, between around 50-60°S. In these regions the RMSE for SLA is measurably less in BRAN3p0 than it is in BRAN2p1. This reflects improvements in both the ocean model and the initialisation scheme that is used. Analyses of BRAN3p0 are ongoing.



*Fig.* 1 (a) Standard deviation of SLA, derived from along-track SLA during 2005; and RMSE of SLA, assuming that along-track SLA is error free, during 2005 from the (b) 2008-version and the (c) 2010-version of the Bluelink Reanalysis.

# **Coupled initialisation**

By augmenting the ensemble with surface flux fields, the ocean observations are readily projected onto the atmospheric fields. An example of this projection, off South Australia is presented in Fig. 2, showing analysis increments to the surface heat flux and wind stress, based on the assimilation of ocean observations. For this example, the adjustment to the ocean state includes a decrease in surface temperature over much of the region. For the western-most part of the domain, this is consistent with a negative increment to the surface heat flux. Along the coast, the negative surface temperature increment is consistent with the upwelling favourable adjustments to the surface wind stress (winds from the south-east). This example demonstrates that, like 4d-var systems, ensemble-based data assimilation systems can be used to optimise surface fluxes. This research is ongoing.



*Fig. 2* Example of ensemble-based increments to the ocean surface heat flux and wind stress (1<sup>st</sup> layer), and the ocean temperature and velocities at different depths (2<sup>nd</sup>-4<sup>th</sup> layer) for a region of southern Australia.

# Conclusions

The success of any data assimilation system is closely related to the skill of the under-pinning model. Recent improvements to the Bluelink model have yielded measurable improvements in the reanalysis system. Improvements to the Bluelink model include changes to parameterisations that now better represent tidal mixing, changes to the surface fluxes (now using ERA-Interim), and changes to the bottom topography. The performance of the data assimilation system also depends on the initialisation of the model. If the model is poorly initialised, the subsequent model integration is often degraded by an initialisation shock. Recent developments in this area (Sandery et al. 2010) have yielded improvements to the Bluelink reanalysis system.

Development of an ensemble-based ocean data assimilation system continues under Bluelink, and related projects. Recent developments of the Bluelink system include a generalization of the model code so that analyses for any model variable can be readily computed. This generalisation has facilitated the augmentation of the ensemble with atmospheric variables. In the first instance, this new capability will be evaluated through a series of experiments using coupled initialisation, where the surface fluxes are modified in an attempt to improve the simulated ocean circulation. Extensions of this coupled assimilation capability

to include other oceanic variables, like mixed-layer depth, depth of isopycnals, and biogeochemical properties are planned.

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### ENSEMBLE PREDICTION AND THE MESOSCALE OCEAN: FORECASTING THE EAC

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#### Abstract

We describe the application of an ensemble prediction methodology to forecasting the mesoscale ocean. In particular we focus on the East Australian Current (EAC) and the role of dynamical instabilities and flow dependent errors of the day. The region where the EAC separates from the coast, is characterized by significant mesoscale eddy variability, meandering and is dominated by nonlinear dynamics thereby representing a severe challenge for operational forecasting. Using analyses from OceanMAPS, the Australian operational ocean forecast system, we explore the structures of flow dependent forecast errors over timescales from days to weeks. Forecast ensemble perturbations are generated using the method of bred vectors allowing the identification of those perturbations to a given initial state that grow most rapidly. We consider a 6 month period spanning the Austral summer that corresponds to the season of maximum eddy variability. We find that the bred vector (BV) structures align and anti-correlate with the forecast errors and that these structures typically occur in regions of instability and in particular where the EAC boundary current separates. We also find that very few BVs are required to identify regions of large forecast error and on that basis we expect that even a small BV ensemble would prove useful for adaptive sampling and targeted observations. The results presented also suggest that it may be beneficial to augment the static background error covariances typically used in operational ocean data assimilation systems with flow dependent background errors calculated using a relatively cheap ensemble prediction system (EPS).

Ocean forecasting has seen major advances in the past decade. Many countries now perform operational forecasts of the mesoscale ocean circulation. Many of the advances in ocean forecasting are on the back of numerical weather prediction (NWP), particularly those in data assimilation. Ocean forecasting is underpinned by satellite observations of sea-level anomalies (SLA) and sea-surface temperature (SST), and in situ observations from Argo floats and a sparse array of tropical moorings. The most energetic scales of the oceans are in the mesoscale, which is characterized by eddies and meanders, and occurs particularly in western boundary current (WBC) regions. These scales are only marginally resolved by the above-mentioned components of the global ocean observing system. As a result, we expect the errors of operational ocean forecasts to be variable in time. The errors of the day are likely to depend on the coverage of assimilated observations and the stability of the ocean's circulation. In this presentation we develop a computationally efficient method for representing flow dependent forecast errors associated with dynamic instabilities and examine their role in determining the predictability of the East Australian Current (EAC).

The reasons that deterministic forecasts fail over reasonable prediction periods has been found to be in large part due to the inherent non-linearity in the system (deterministic chaos; see Lorenz (1963)), errors in the

initial conditions, model deficiencies and forcing errors. In comparison to a single control forecast, ensemble forecasts can provide not only improved estimates of the forecast (ensemble mean), but also estimates of the forecast error covariance and possibly the higher-order moments. Rather than try to sample the entire probability distribution function one approach is to seek to identify the spatial structures of a system's fastest growing modes. This may be achieved through a periodic rescaling of the evolved perturbation vectors and judicious choice of initial perturbation amplitude, evolution period and rescaling norm. By this "breeding" process information about the fast growing errors may be incorporated into the initial ensemble of forecast perturbations. For particularly dynamic flows, such as when emergent coherent structures are developing (for example atmospheric high-low blocking dipoles or eddy shedding associated with oceanic boundary current separation), errors arise due to fast-growing large-scale instabilities. For such systems bred vectors are finite amplitude, finite time (local) stochastically and non-linearly modified versions of the leading Lyapunov vectors (global).

O'Kane et al 2010, [1] demonstrated how one may use bred vectors to identify when and where instabilities are likely to occur in the ocean and so-doing identify when and where the forecast skill of an operational ocean forecast system is likely to be low. They developed a bred vector EPS for a regional ocean model of the Tasman Sea. The main features of the Tasman Sea are the EAC and its eddies. At any point in time, the EAC is typically characterized by a narrow, strong southward flow adjacent to the continental shelf between about  $15 \circ S$  and  $32 \circ S$ . The EAC typically separates from the coast near Sugarloaf Point, forming a complicated field of warm- and cold-core eddies. Warm-core eddies are typically large, with diameters of several hundred kilometers, forming every 90 days or so. Cold-core eddies are smaller, perhaps 50-100 km across, and often form at the point where the EAC separates from the coast, or on the peripheries of warm-core eddies. While warm-core eddies are usually well-resolved by altimetry, cold-core eddies are often missed.

The results presented in Figs 1 & 2 extend the study of O'Kane et al 2010 [1] from forecast time scales of 1 week, relevant to covariance estimation in ocean data assimilation to 1 month. In Fig. 1 we compare successive 1 month unperturbed or control forecasts (e, f & g) to verifying analyses (a, b, & c) from the Australian Bureau of Meteorology Ocean Model Analysis and Prediction System (OceanMAPS). Individual bred vectors are then calculated from an ensemble of perturbed forecasts (m, n & o) and averaged to determine a low-dimensional subspace where they align and hence identify regions of instability (i, j & k contours). The ensemble averaged bred vector contours are then shown to broadly correspond to regions of large control forecast error (i, j & k shaded).

In Fig. 2 we examine T250 and through the water column in vertical sections along regions of coincidence between the evolved bred vector averages and the control forecast errors on the 8*th* April 2008. We find that not only do the bred vectors project onto the control forecast error throughout the vertical column but that the ensemble forecast (4 members) RMS errors are substantially reduced over the deterministic control forecast. Our results clearly show that even over a 1 month period forecast errors arise largely due to dynamic instabilities and that these forecast errors can often be expected to dominate analysis errors. Moreover a very small ensemble of bred perturbation vectors may provide an effective, computationally inexpensive means to calculate flow dependent background information or to identify regions where additional observations may be targeted.

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Fig. 1 Surface height on T cells. Columns 1, 2 & 3 depict results valid for the 12<sup>th</sup> February, 11<sup>th</sup> March & 8<sup>th</sup> April 2008 respectively. Row 1 figs a, b, c; Analysis fields, Row 2, figs e, f, g; 28 day control forecasts. Row 3, figs i, j, k; Comparison of ensemble averaged (4 members) bred vectors (±0:4m, ±0:25 & ±0:15 contours) and day 28 forecast error (shaded). Row 4, figs m, n, o; ±0:4m, ±0:25 & ±0:15 contours for each of the 4 individual bred vectors.



*Fig. 2* Comparison of day 28 forecast error (left), ensemble averaged bred vectors (middle) & ensemble averaged forecast error (right) (4 members) valid on the 8<sup>th</sup> April; Showing T250 (a-c) and vertical sections along 30:05°S (d-f) and 38:85°S (g-i)

#### **OCEAN STRIATIONS**

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This paper overviews the pattern of mesoscale, quasi-zonal, jet-like features (striations), recently detected in satellite and in situ data (Fig.1). Various aspects of the dynamics of striations, such as different kinds of forcing and interaction with mean meridional flow and eddies, are discussed. Properties of a broad variety of striations can be generalized within the concept of non-linear beta-plume. We analyze the roles that in situ observations and remote sensing played in revealing the striations. Striations may impact the climate system in at least two ways: (i) by increasing the anisotropy of horizontal mixing in the geophysical turbulence and (ii) via the signature that striations produce in atmosphere through the effect of the sea surface temperature on stability of the planetary boundary layer. Numerical models can benefit from dynamical comparison with the dense global grid, provided by striations. General theory of ocean circulation can be improved by better understanding the mechanisms, regularizing the process of baroclinic instability.



*Fig. 1 Mean zonal geostrophic velocity at the ocean surface, high-pass filtered horizontally with twodimensional 4° filter. Gray colors vary from -1 cm/s (white) to +1 cm/s (black).* 

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## MESOSCALE ZONAL FEATURES IN THE SOUTHEAST INDIAN OCEAN AS SEEN IN THE BLUELINK REANALYSIS 2.1 (BRAN2.1)

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# Introduction

It has been presumed that the geostrophic turbulence in the ocean is unorganised such that any mean patterns are a result of under sampling due to the finite averaging period and Schlax and Chelton (2008) highlighted the challenge to detect potential organization. The recent discovery of multiple long term open ocean zonal striations from high resolution numerical models (Nakamo and Hasumi 2005; Richards et al. 2006) and satellite observation analysis products (Maximenko et al. 2005; Maximenko et al. 2008) have drawn much attention and invoked debate on their origin. Geostrophic turbulence theory has been shown to form zonal jets on a beta plane (Rhines 1975) and the resemblance of these mean ocean zonal striations to those visible on giant planets (Galperin et al. 2004) has been noted. Studies of eddy induced zonal velocity bandings have shown features with comparable width and magnitude to striation structures found from satellite altimetry (Schlax and Chelton 2008). In this study we analyze current patterns of the southeast Indian Ocean from an eddy-resolving 14-year ocean reanalysis. We examine the reanalysis annual and seasonal mean currents to reveal the surface and sub-surface structure of these alternating quasi-zonal bandings.

## Data

The results presented are drawn from a 14 year ocean reanalysis (BLUElink>ReANalysis 2.1; BRAN2.1) providing a "best" ocean state estimate from 1993 to 2006. BRAN2.1 combines the Ocean Forecasting Australia Model (OFAM) with ocean observations using the BLUELink Ocean Data Assimilation System (BODAS) to provide the first high resolution ocean reanalysis for the Asian-Australian region (Schiller et al. 2008). OFAM is a global model based on MOM4p0d (Griffies et al. 2004), with 1/10° horizontal resolution around Australia (90-180°E, south of 16°N) and 47 vertical levels. The model was initialised by a climatological ocean state and forced by ERA40 (1993 to mid 2002) and ECMWF (mid-2002 to 2006) surface fluxes. BODAS uses an ensemble optimal interpolation scheme based on a stationary ensemble of modelled anomalies from the forced model integration without data assimilation (Oke et al. 2008). BODAS employ a seven day analysis cycle assimilating all available satellite altimeters (ERS, GFO, Topex/Poseidon, Envisat and Jason), SST observations from Pathfinder and AMSR-E missions and in situ profiles from e.g., Argo floats and XBT. The quality of BRAN2.1 has been objectively compared with other ocean state estimations (Hernandez et al. 2009) demonstrating comparable or superior performance in the southeast Indian Ocean. In general, the performance is expected to decline below the thermocline as the background error covariability of remotely sensed observations reduces, although this is partly compensated by a reduction in ocean variability. The BRAN2.1 Mean Dynamic Topography (MDT) surface geostrophic currents are compared with geostrophic surface currents derived from a MDT by AVISO (Rio et al. 2009) with horizontal resolution of 0.25°; and a MDT supplied by Asia-Pacific Data Research Center (APDRC) (Maximenko and Niller 2005) with horizontal resolution of 0.5°. In this study we analyze data for the southeast Indian Ocean, in the latitude range of 50S to the equator and between longitudes of 90E and 120E.

#### Results

The zonal component-only representation of annual mean ocean current for depth averaged and meridional depth section along 105°E are shown in Fig1. Figure 1a shows quasi-zonal patterns of alternating positive and negative velocities in mean ocean current from BRAN2.1. These alternating mean currents show slight inclination towards equator especially in the subtropics and have meridional widths of 2-3°. Meridional depth section (fig. 1b) reveals that many of these mean features extend coherently to the abyssal ocean with a decline in meridional width and velocity magnitude as seen in other high resolution model studies without data assimilation (Richards et al. 2006). The eastward flows are largely confined to the upper 250m of the depth section with smaller abyssal extensions. Westward flow dominates at mid-depth (250-1000 m) and shows more interconnections with adjacent flows of similar direction. Even though zonal mean features (otherwise called "striations") are easily distinguishable in zonal component-only representation, we find such representations are more misleading as opposing meridional anisotropy (resulting from mean meridional velocity) can play a crucial role in total mean current. Thus we adopt a colorwheel to better represent total current; with different colours depicts resultant total current direction with a gradient in whiteness handle current magnitude (as extended abstracts are strictly in black and white format such colorwheel figures are not represented here but results are discussed briefly below).



*Fig. 1* Contour plots of (a) Depth averaged (0-3500 m) 14-year mean zonal velocity (b) 14-year mean zonal velocity meridional section along 105°E from BRAN2,1; thick and dotted lines represent 0.01ms-<sup>1</sup> and -0.01ms-<sup>1</sup> respectively.

By applying a colorwheel tool to surface geostrophic currents derived from MDTs (AVISO, APDRC and BRAN2.1) we have shown that between the South Equatorial Current and 30S latitude in the southeast Indian Ocean the circulation is dominated by a broad eastward flow regime towards the west-Australian coast regularly punctuated by narrow fingers of near westward flow. These broad eastward flows are in general consistent between the three MDT's. We identify a series of five eastward jets in this broad eastward regime. The eastward jets in the tropics propagate into the coast and Leeuwin current system though Exmouth Plateau and northwest shelf region. We also note a well developed mean westward flow in MDT surface geostrophic velocities connecting both Naturaliste and Broken plateau. Meridional depth sections represented using the colorwheel also shows coherent mean patterns as shown in Fig. 1b. But these vertically coherent mean features are similarly quasi-zonal especially poleward of 15°S. The north faces of bathymetry features consistently show a westward component of flow; similarly eastward flows consistently align with the southern face of the bathymetry. 14-year mean depth averaged currents represented using colorwheel

reveals a more complex view of the current systems than flow structures in the geostrophic surface currents from MDT's or the simpler alternating flow patterns seen in other high resolution models (Richards et al. 2006). This sort of flow structure, which we refer to as the "Arteries" of the ocean circulation, shows a variety of pathways with a range of directions with meridional connections and meandering. The complexity, in other words "Arterial" nature of mean flow patterns increases with depth. We note increased eastward transports in the upper 250 m during seasons that correspond to stronger Leeuwin transports. Conversely increased westward flows align with weaker Leeuwin current periods. Such seasonal patterns are minimal at mid-depth (250-1000 m), where westward mean jets dominate with weaker eastward mean flows. At the deeper layer (1000-3500 m) alternating eastward and westward features are comparable in both magnitude and width with greater "Arterial" nature. We also note a deep poleward flowing shelf transport in the deeper layer. This mean pole-ward flow starts from the southern side of the Exmouth Plateau and closely follows the shelf contours until separating westward along the north face of the Naturaliste Plateau. From the Naturaliste Plateau this deep coastal shelf current joins the westward flowing zonal flow between Naturaliste and Broken Plateaus.

# Conclusion

The study shows relatively broad geostrophic eastward surface flows embedded with a series of eastward jets and punctuated by narrow fingers of westward flow of the southeast Indian Ocean using an ocean reanalysis and available high resolution MDT datasets. By applying a colorwheel visualization for total current, we have shown that the current patterns in the high resolution model and MDT are much more complex that are otherwise masked by the conventional visualization of zonal current in the eddy resolving datasets. These circulation patterns in the southeast Indian Ocean are more arterial with considerable meridional variations. The prominent eastward flow patterns at the surface and westward flow patterns at mid-depth give some promising insights into the recent discovery of Leeuwin Current source flow regions in the southeast Indian Ocean previously mentioned by Domingues et al. 2007. Increase and decline of these eastward flow structures closely coincides with the phases of the Leeuwin Current, making an interesting link with the Leeuwin Current dynamics. There is a specific orientation of reanalysis currents near bathymetry indicating they are on average influenced by and interacting with the bathymetry.

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# MESOSCALE COASTAL SEA LEVEL FORECASTING: AUSTRALIAN CONTEXT, RECENT FINDINGS AND DIRECTIONS

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# Abstract

This presentation will overview coastal sea level forecasting in the Australian context and highlight recent findings and directions. Attention will focus on the forecasting of mesoscale ocean dynamics.



*Fig. 1* Coastal sea surface from the perspective of a real time tide gauge.

Coastal sea level is an aspect of ocean forecasting of special and direct relevance for a range of human activities. The variety of end-uses is reflected in the qualitatively different forecasting approaches in use around the globe. These forecasting approaches can be roughly categorized on the basis of forecast time windows and forecasting schedules.

Compared to other regions of the globe, the Australian coastline is particularly vast and exposed to open oceans but with highly localized population centers and areas of commercial interest. In Australia, coastal sea level data and forecasting is not centralized under any single organization and both public and private interests are involved.

The Bureau of Meteorology's National Tidal Centre does however play a central role for certain activities and insitu observations at tide gauges continue in general to be of great importance to forecasting. Recent upgrade works to the Bureau's real-time sea level observation network increasingly open doors to operational data assimilation and validation.

<sup>&</sup>lt;sup>‡</sup> National Tidal Centre, Bureau of Meteorology

<sup>&</sup>lt;sup>§</sup> OMC International. WWW: <u>http://www.omc-international.com</u> Email: <u>giles@omc-international.com</u>

Forecasts in the form of harmonic tide predictions are well established and omnipresent. However, astronomical tide predictions alone are not suitable for many end-uses and the recent research highlighted here approaches the role of non-tidal mesoscale dynamics in qualitatively different ways.

Explicitly considering the combined dynamic interaction of tidal and non-tidal forcings in systems such as OceanMAPS is one research direction that offers new perspectives to coastal sea level forecasting. This is early in the research phase but results are expected to appear as part of the Bluelink 3 project.

Recent findings forecasting storm surge hazard around Tasmania (McInnes) are based on extreme value statistical methods of combining historical records, shallow water surge modeling, tide predictions and high resolution topography. This methodology allows the net effect of mesoscale interactions to be projected well beyond the time scales of the mesoscale itself and is of value for coastal planning and engineering end-uses.



*Fig. 2* Hobart storm tide inundation excedance probabilities.

Extreme value statistics are also the focus of ongoing commercial research by Under Keel Clearance service provider OMC International, though here the forecast window can be as short as a few hours and the attention is on highly localized reductions in sea level. This Australian company reports of research into the use of data assimilation techniques and the determination of site specific empirical relationships to blend between real-time observations and numerical model forecasts.

Recently published results detailing the observed forecast skill of OceanMAPS at coastal tide gauge locations across Australia has lead to the development of trial routine 'total sea level' forecast products. This approach to forecasting coastal sea level combines the output of existing operational forecast systems maintained by the Bureau of Meterology in order to estimate the net sea level observed at real-time tide gauges within a 1 week forecast window.



Fig. 3 Routine sea level forecast overlaid with real-time observations.

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# **OPERATIONAL WAVE FORECASTING**

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# Introduction

Wind generated ocean waves are almost always present at sea. These waves are generated by winds somewhere on the ocean, be it locally, so called wind sea, or thousands of kilometres away, referred to as swell. These waves affect a wide range of activities such as shipping, fishing, recreation, coastal and offshore industry, coastal management and pollution control. They are also very important in climate processes as they play a large role in exchanges of heat, energy, gases and particles between the ocean and the atmosphere. The following paper presents an overview of wave modelling and its development at the Bureau of Meteorology (the Bureau), followed by a short discussion of future directions.

# **Describing wind waves**

Wind waves are generally described with an energy density spectrum. An example of such a spectrum in given in Fig. 1. On the left, polar direction shows the direction in which the waves propagate, with frequency indicated by radial distance from the centre. This example shows a spectrum with two clearly identifiable peaks, indicating two separate wave systems, the broad wind sea peak travelling to the NW and a narrow swell peak travelling to the NE. Integrating over directions gives the energy density as a function of frequency only, presented on the right.



*Fig. 1 Example spectrum, represented as a polar plot showing spectral energy density in both frequency and directional space on the left, and integrated over direction on the right.* 

In addition to the spectral representation above, the wave field can be described with mean wave parameters. The most common parameter is the significant wave height  $H_s$ . This wave height is defined as the mean height of the 1/3 of the highest waves in a sample, and can be computed from the energy spectrum by:

$$H_s = 4E, \qquad E = F f, \theta \, df d\theta \tag{1}$$

where E is the total energy in the wave energy spectrum F , expressed in terms of the wave frequency f and direction  $\theta$ .

#### Wave modelling

Modern wave models employ a spectral description of waves, and are based on a form of the spectral wave energy or action equation.

$$\frac{DF \ f,\theta}{Dt} = S \ f,\theta + S_{nl} \ f,\theta + S_{ds} \ f,\theta \quad \dots$$
(2)

The left hand side of this equation represent the spatial and temporal evolution of the wave spectrum, and the right hand side represents sources and sinks of wave energy: the input of wave energy by wind  $(S_{in})$ , nonlinear interactions between waves  $(S_{nl})$  and dissipation due to wave breaking or 'whitecapping'  $(S_{ds})$ . The dots in Eq. (2) represent additional (usually shallow water) processes such as bottom friction and other wave-bottom interactions.

The evolution of wave models can broadly be categorised by the treatment of the nonlinear interaction term, representing the process that transfers energy to longer waves. Exact computation of this terms is prohibitively expensive for operational use, and early wave models employed rudimentary parameterisations of these effects (termed 1<sup>st</sup> and 2<sup>nd</sup> generation models). The shortcomings of these models led to the establishment of the WAM Group, whose mission was to develop a model that would explicitly resolve the energy balance equation. This was achieved with the development of the discrete interaction approximation (DIA: WAMDI Group, 1988). These models are denoted as third generation (3G) models. The WAM model (Komen, 1994) became the first operationally feasible 3G wave model. Since the development of WAM, a limited number of other 3G wave models have been developed, WAVEWATCH III<sup>®</sup> (WW3) (Tolman & D Chalikov, 1996; Tolman et al. 2002) and SWAN (Booij, Ris, & Holthuijsen, 1999) being the most well known of these.

The original WAM the input and dissipation terms, often referred to as source terms, were based on Snyder et. al. (1981) and Komen and Hasselmann (1984, hereafter WAM3). During the development of these source terms, testing was performed for wind sea growth in the absence of swell, which was later found to have unnatural effects on the corresponding model results (Tolman and Chalikov, 1996). Modifications by Janssen (1991, hereafter WAM4) and (Bidlot, Janssen, & Abdalla 2008, hereafter BAJ) gave more emphasis on the high-frequency part of the wave spectrum resulting in a more realistic interaction between windsea and swell.

The original release of WW3 employed the source term package of Tolman and Chalikov (1996) (hereafter TC96) consisting of the input source term of Chalikov and Belevich (1993) and Chalikov (1995) and two dissipation constituents. The previous release (version 2.22) of WW3 maintained the option to use WAM3 source terms within the WW3 model framework. In the most recent release (version 3.14), WAM4 source terms variants have also been added (Ardhuin et al. 2007).

Further details on the development of third generation models can be found in (Komen, 1994) and (Group et al. 1988), with more up to date reviews found in (Tolman et al. 2002) and (Janssen 2007).

# Operational wave modelling at the Bureau of Meteorology

The history of numerical wave prediction at the Bureau began with a parametric wave model in 1983, followed by a first generation spectral model in 1986 (see Greenslade 2004). Since June 1994, the Bureau has run a version of the WAM model known as AUSWAM (Bender and Lesley, 1994), which has remained in operational use for the last 15 years. This model was replaced on 23 August 2010 with AUSWAVE, based on the WW3 model. This follows the implementation of the ACCESS (Australian Community Climate and Earth-System Simulator) Numerical Weather Prediction (NWP) systems which formally replaced the GASP, LAPS and related NWP systems on Tuesday 17 August 2010.

The initial implementation of AUSWAVE was intended to replicate much of the AUSWAM system. Grid resolutions remain the same, as do spectral resolutions. Evaluation against observations demonstrates that BAJ source terms provide the best overall forecast. This is a notable difference to the systems run by NOAA and the UK MetOffice, both of which employ the default TC96 terms. One significant difference between the operational implementations of AUSWAVE and AUSWAM is that AUSWAVE does not incorporate the assimilation of satellite altimeter  $H_s$  data.

The new ACCESS forced AUSWAVE shows significant improvement of the GASP forced AUSWAM. Despite the lack of data assimilation, AUSWAM shows lower RMS error at both the short and longer forecast periods when verified against both buoy and altimeter data. A slight negative bias is present in the modelled H<sub>s</sub>, due primarily to a negative bias in the ACCESS winds, as verified against scatterometer data.

# Ongoing issues and future directions

The overall accuracy of a wave model forecast depends on the accuracy two models, the wave model itself, and that of the atmospheric model providing the forcing winds. 10-15 years ago, wave model development had reached a point where a significant proportion of the wave forecast error could be attributable to errors in the forcing winds, limiting gains that could be realised by improvements in the wave models themselves. The intervening period has seen great improvement in the atmospheric models, resulting in a larger proportion of the error being attributable to the wave model, stimulating renewed vigour in wave model source term research.

Recent advances in theoretical knowledge, as well as increasing computer resources have lead to the development and testing of more physically based source terms. The dissipation source term has long been used as the 'closure term' in the balance equation, used mainly to tune models. Substantial progress has been made in the understanding of wave breaking, which in turn has lead to more physics-based source term formulations for wave models (a good review can be found in Babanin & van der Westhuysen (2008)).

The introduction of new source term formulation is hindered by the fact that each term cannot be considered in isolation. The DIA, while providing the ability to compute  $S_{nl}$  at operationally feasible cost, is an approximation that introduces error in the evolution of the spectrum, error for which the input and dissipation terms have been tuned to compensate. While the option is available in most 3G models to compute this term explicitly, typical computational increases by a factor of 300 keep this beyond operational feasibility. Several techniques have been developed which strike a compromise between these two extremes (e.g. Cavaleri et al. 2007).

Other processes that play an important role in operational wave modelling are two-way coupling, data assimilation and ensemble forecasting. Historically, environmental modelling has focused on isolated problems, like weather, ocean, storm surge or waves. Increasingly, it is understood that progress in many of these fields will depend on properly accounting for the interactions between such systems, and hence on the

coupling of these models. For waves, prime benefits are expected from coupling to current models and storm surges. Waves in turn can provide direct forcing for inundation (storm surge) models, and more physical parameterisations of boundary layer processes in both the atmosphere and the ocean. Whereas some of these coupling activities have been addressed for several decades, systematic coupling between systems in an operational environment has only recently started to be addressed.

Wave models do not represent an initial value problem, and data assimilation is not critical in order to produce a good wave forecast. Nevertheless, data assimilation can improve short term forecasts (12 to 24h) in general, and potentially can improve swell prediction within ocean basins up to two weeks into a forecast. It should be noted that, compared to best practices in atmospheric data assimilation, data assimilation in wave modelling is rather primitive. This implies that much work can be done in wave data assimilation, particularly by using specific wave physics in new data assimilation methods (Greenslade & Young, 2004).

Finally, probabilistic methods are becoming more and more important in weather forecasting. In such an approach, multiple model runs are made with different initial conditions to address the uncertainty in the model guidance, and to get the best consensus. Wave model ensembles based on such weather model ensembles have been produced at meteorological centres for up to a decade, but the forecast benefits and probabilistic reliability of the ensembles has not yet received much attention.

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## **TSUNAMI FORECASTING AND WARNING**

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#### Introduction

Following the Indian Ocean Tsunami of 26 December 2004, there was a significant increase in the effort put towards developing tsunami warning systems around the world. The Australian Government committed \$68.9M to the development of a comprehensive tsunami warning system for Australia. This was a four-year project and was completed in June 2009. One of the major outcomes of this project was the establishment of the Joint Australian Tsunami Warning Centre (JATWC), operated jointly by the Bureau of Meteorology (the Bureau) and Geoscience Australia (GA). Another component of the project was the development of an operational model-based tsunami prediction system. This extended abstract describes the basic elements of this tsunami prediction system.

## Tsunami warning systems

Prior to 2004, there were few operational tsunami warning systems around the world, and none within the Indian Ocean. Australia relied on the Australian Tsunami Alert Service (ATAS) which provided a limited notification and alerting capability, but not a comprehensive monitoring and warning system.

Historically, tsunami warning systems operated on the basis that warnings were required for regions near the earthquake, with the hazard reducing as the distance from the earthquake increased. This is a reasonable assumption as a first approximation, but there are a number of factors that are not taken into account. These include the complex dynamics of tsunami propagation resulting from interaction with bathymetric features and the fact that earthquake ruptures are typically not point sources. Given the increased capability in a number of areas relating to tsunami forecasting over the past few years, current systems are now significantly more sophisticated, and most are based on numerical models.

# **Tsunami Models**

There are three main components to a tsunami model: the determination of the initial condition due to the earthquake; the propagation of the tsunami across the deep ocean; and inundation onto land. Typically, these components are modelled separately.

The model used within the JATWC is the Method Of Splitting Tsunamis model (MOST; Titov and Synolakis, 1998). The initial condition component of this assumes that the tsunami has been generated by a submarine earthquake. The model determines the sea floor deformation assuming a double-couple source model (Okada, 1985). The required inputs for this are the details of the earthquake rupture. These are: the location of the rupture (latitude and longitude), depth of the rupture (below the sea-floor), the size (length, L, and width, W) of the rupture, the

amount of slip for the rupture,  $u_0$ , and the elasticity of the oceanic crust,  $\mu$ , usually taken to be a constant. The fault orientation is specified by the strike, i.e. the geographical orientation of the subduction zone, and the dip (measured down from the horizontal). The direction of the slip also needs to be specified. See Fig. 1 for how these are specified for MOST. Once the sea-floor deformation has been determined, this is imposed instantaneously as the initial condition on the ocean surface, under the assumption of incompressibility.



**Fig. 1** Schematic of the parameters describing a rupture for the MOST model. L is the length, W is the width,  $u_o$  is the slip, D marks the location where the depth is defined and X marks the location where the latitude and longitude are defined.

The seismic moment  $(M_0)$  of the earthquake is related to the rupture characteristics via:

#### $M_o = \mu L W u_o$

And the moment magnitude is related to seismic moment as:

$$M_{w} = \frac{2}{3} (\log_{10} M_{0} - 9.1)$$

Once the initial displacement has been determined, the propagation across the deep ocean is modelled. MOST solves the depth-averaged non-linear shallow-water equations without bottom friction factors, artificial viscosity, or an explicit dispersion term. Numerical dispersion is exploited to approximate physical wave dispersion.

The implementation of MOST for the inundation aspect of tsunami modelling is not currently used within the JATWC so this is not described here. Further details on MOST can be found in Titov and Synolakis (1998).

Numerical tsunami models are fairly well developed, but they present a challenge to run in realtime; partly due to computational limitations (in many situations, the model runs cannot be completed quickly enough to forecast impacts), but also due to a lack of knowledge of the details of the initial condition. In particular, the specific details of the earthquake and resulting seafloor displacement needed to initialize the tsunami model are not known until well after an earthquake event, if at all. For these reasons, typical tsunami forecast systems, including the system implemented within the JATWC are based on pre-computed tsunami "scenarios". A "scenario" is a single tsunami model run that is calculated ahead of time with the initial conditions carefully selected so that they are likely to represent an actual tsunamigenic earthquake.

## T2 Scenario database

The scenario database currently implemented within the JATWC is the T2 scenario database (Greenslade et al. 2009; Greenslade et al. accepted). T2 is described in detail in Greenslade et al. (2009) so only a few of the major features are presented here.

The basis for the source locations within T2 is all known and potential subduction zones within the Indian and Pacific Oceans. The South Sandwich subduction zone in the South Atlantic Ocean is also included. Earthquake sources are spaced at 100 km intervals along the strike of each subduction zone. This results in a total of 521 source locations for T2. These are all shown in Fig. 2.



Fig. 2 Source locations included in the T2 scenario database.

The T2 scenario database includes 4 earthquake magnitudes of  $M_w = 7.5$ , 8.0, 8.5 and 9.0 at each source location. These have pre-defined rupture dimensions for which details can be found in Greenslade et al. (2009). This results in a total of 1,866 individual scenarios in the T2 scenario database.

Intermediate magnitude earthquakes, i.e. those with magnitudes other than 7.5, 8.0, 8.5 and 9.0 are derived from the pre-computed scenarios by applying a scaling factor to the sea-level elevation (Greenslade et al. 2009; Simanjuntak and Greenslade, submitted). This provides guidance for earthquakes with magnitudes ranging from 7.3 to 9.2 at intervals of 0.1. Some verification of the forecast sea-level from T2 for a number of previous tsunami events can be found in Greenslade et al. (accepted).

# **JATWC Tsunami warnings**

Tsunami warnings issued by the JATWC are based on the T2 scenario database and use a threshold technique. When a tsunami event occurs, the closest scenario is selected from the T2 scenario database according to the seismic parameters that are available in real-time, i.e. location and magnitude. Tsunami amplitudes for the scenario are assessed within coastal regions around the Australian coastline and offshore territories to determine the level of threat. Tsunami

warnings are issued according to whether the maximum amplitudes exceed pre-determined threshold values. There are three tiers of warning:

- *1* No threat,
- 2 *Marine threat*: warning of potential danger in the marine environment and immediate foreshore,
- 3 Land threat: warning of major land inundation in low-lying coastal areas.

The threshold values for each level of warning have been derived empirically by consideration of coastal observations and impacts of past events in Australian coastal waters. Further details of this technique can be found in Allen and Greenslade (submitted).

## **Future directions**

There are a number of short-term and longer-term improvements that are planned for the modelbased components of the JATWC warning system. One of the first priorities is to extend range of earthquake magnitudes for which numerical guidance is provided. This will involve development of a series of  $M_w = 7.0$  scenarios at each source location, as an extension to the T2 scenario database. The importance of this is not so much to provide forecast guidance for Australia (there are very few instances in which warnings would be expected for events with  $M_w < 7.3$ ) but rather to address the JATWC's responsibilities as a Regional Tsunami Watch Provider within the Indian Ocean Tsunami Warning System. The possibility of developing a set of larger magnitude scenarios, i.e.  $M_w > 9.0$ , is also being considered.

One of the issues with verification of tsunami forecasts is the limited amount of data available. The most common data sources are sea-level observations from coastal tide gauges and deepocean tsunameters. Furthermore, observable tsunami events are relatively rare. This means that every tsunami that occurs and that is well-observed provides an extremely valuable new dataset. As future tsunamis occur, further verifications of T2 will be possible and perhaps more importantly, as impacts are observed (or not observed) in the Australian region, assessment and tuning of the warning thresholds can be performed.

In addition to model verification, sea-level observations can also be used within the forecast system. Currently, sea-level observations are used mainly to confirm whether or not a tsunami has been generated, and there is also some limited subjective use within the JATWC relating to validation of the scenario selection. A high priority is the need to develop an objective method to optimise the use of the sea-level observations. This is currently an area of active research. The initial aim of such a scheme is to use observations of sea-level from tsunameters to modify the T2 scenarios during an event and provide improved estimates of coastal sea-level. Preliminary results of this research have been encouraging.

As mentioned previously, tsunami forecasts for Australia do not currently involve any inundation modelling. Considerable effort has gone towards the development of a warning scheme based on the T2 scenarios, which currently are limited to greater than 20m depth. Ideally, warning guidance should be based on forecasts of coastal flooding, but computational and data constraints have limited the development of this capability. As computational capacity

increases, this can be reconsidered. Eventually, it is envisaged that when a potentially tsunamigenic earthquake occurs, the deep-water component of the tsunami model will be run in real-time on the Bureau's supercomputing facilities, initialised using event-specific rupture details obtained from seismic data and incorporating a sophisticated data assimilation scheme. Detailed inundation models could then be run for areas identified to be at risk, using boundary conditions from the real-time deep-water forecast. Tsunami warnings could be refined based on the output of the inundation models. In addition to the need for high resolution coastal bathymetry and topography data, significant computational resources would be required for this sort of system, and research into optimising MOST, or possibly another tsunami model code for real-time integration would be necessary.

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# LITTORAL ZONE FORECASTING

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#### Introduction

Simulating and forecasting oceanic variables in the nearshore or littoral zone, is presented here in the context of the Bluelink Project (http://www.bom.gov.au/bluelink), a collaborative effort involving CSIRO, the Bureau of Meteorology and the Royal Australian Navy (RAN). The overall aim is to provide ocean forecasts across a range of space and time scales from ocean basin scale to the littoral zone. A component of this project is to develop LOMS (Littoral Ocean Modelling System), a relocatable hydrodynamic modelling suite configured to provide threeday predictions of waves, currents and morphological response on an arbitrary beach. As waves break on approaching a beach they create radiation stress gradients that can drive local circulation causing rip currents and longshore currents. The pattern of wave-breaking and circulation is strongly influenced by the bathymetry. The requirements for LOMS, and the challenges facing its implementation, are described here focussing on the implementation and validation of the hydrodynamic model XBeach. A field experiment at Secret Harbour, WA, provided a dataset that was used to validate XBeach. Model simulations are found to be sensitive to long-terms changes in the bathymetry, which points to future research in this area.

# Requirements

The aim is to forecast wave (Height, Breaking, Direction) and current (U, V) conditions in the surf zone. This necessitates using a state-of-the-art hydrodynamic model with: high resolution (5-10m), depths less than twenty metres, and easily relocatable (various shoreline profiles). Forcing is supplied by: local wind speed and direction, tide level, and integral wave parameters at some offshore location. Also, the three-day forecast needs to be able to be done in reasonable time.

Two main features of littoral-zone modelling present major challenges. The model domain will generally be a rectangular region with one side over land. The other three sides over the ocean will require the treatment of open boundary conditions for the current and wave equations in the hydrodynamic model. Available models vary considerably in their treatment of open boundary conditions. The second challenge is the paucity of detailed knowledge of the bathymetry and the shoreline location that need to be supplied to the model. Observations are often from quite diverse datasets. The model can be validated for case studies where we do have adequate knowledge, but for the model to be easily relocatable the level of knowledge of bathymetry is an issue. Furthermore, even with adequate knowledge of the bathymetry and shoreline, the action of waves and wave-driven currents can significantly change the bathymetry and shoreline by transporting sediment.

#### Modelling

Various hydrodynamic models were considered for inclusion in LOMS. After considerable testing, it was found that the main challenge to successful implementation was the numerical treatment of wave and flow conditions at the open boundaries. The model XBeach (Roelvink et al. 2009) proved to be effective in its treatment of boundary conditions, compared to other models. XBeach was developed at TU Delft and Deltares and consists of a wave module coupled to a circulation module. The wave module uses conservation of wave action (energy/frequency) with a separate roller equation. The circulation model consists of the vertically-averaged conservation of momentum and volume. XBeach also includes a morphology module that allows evolution of the seabed in response to water movement. This is an area of future research. A separate consideration is considering hydrodynamic models that allow three-dimensional current velocity fields (see e.g. Warner et al. 2008). Such models are of necessity much more expensive to run than two-dimensional models, and moreover have not been yet been widely deployed in the surf-zone. However they will be addressed in future research.



*Fig.* 1 Observed (solid) and modelled (dotted) time series of alongshore current speed during 17-19 February 2009, at the locations of VEC1, VEC2 and VEC3.

# Validation

As part of Bluelink, a month-long series of observations of wave and current conditions was measured at Secret Harbour WA during the Austral summer. This field experiment was designed to measure wave and current conditions on a longshore-uniform barred beach over several sea-breeze cycles, and hence provide observations for the validation of a littoral-zone model. XBeach was configured for this location and forced by hourly integral wave parameters measured at a Nortek AWAC situated at seven metres depth outside the surf zone. XBeach was also implemented in other locations such as Palm Beach, NSW, a curved beach with rip channels, and at Marmion Lagoon, WA, which has a complex reef structure.

Figure 1 shows the time series of observed and modelled alongshore current speeds at three locations: VEC1 (closest to the beach), VEC2 (shoreward of the bar crest) and VEC3 (seaward of the bar crest). Although the alongshore current is somewhat underestimated at VEC3, in general the agreement is good. Over the course of the field experiment described above, the shape of the sandbar changed significantly. XBeach was run over a high-resolution bathymetry surveyed at the end of the observing period. The model/data misfit seen in fig. 1 is most likely due to the bathymetry changing over the duration of the experiment. The model was also run with perturbed bathymetries to investigate the range of response it produces in the velocity field.

A natural dynamical process we expect to be present in the observations and the model output is the presence of edge waves and shear waves. These are infragravity motions, with periods much longer than surf waves. Another means of validating a littoral-zone model against observed time-series is to compute power cross-spectra and investigate the existence of infragravity motions. Some evidence of infragravity waves was found in the phase cross-spectrum between sea-surface elevation and cross-shore velocity, both in XBeach-generated time-series and observed time-series for Secret Harbour.

# Conclusions

Preliminary results were presented illustrating the performance of the hydrodynamic model XBeach against observations from the Secret Harbour field experiment. The agreement is quite good at the vector current meter locations. Differences are primarily due to morphology changes of the bar position and depth. Such changes cause the location of wave shoaling and breaking to vary, and hence drive currents at different locations. There is evidence of the generation of infragravity motions observed in the data and the model output. The XBeach model displays sufficient ability to capture the wave-driven currents during the sea-breeze cycle. The model also is able to simulate realistic wave and current conditions at Marmion Lagoon. Not only does XBeach meet the requirements for littoral-zone forecasting, it also shows good potential to study morphological changes in future research. Methodologies from the field of ocean data assimilation may be applied to problem of varying bathymetry, treating the bathymetry as control variable that influences the model/data misfit through the model. The major component of future model development with be the inclusion of a morphology module coupled to the wave and current modules. The challenge will be the large difference in time-scale between short-wave forcing (seconds) and morphological changes (hours-days).

#### References

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