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Sampling Physical Ocean Field in WCRP CMIP5 Simulations: CLIVAR Working Group on Ocean Model Development (WGOMD) Committee on CMIP5 Ocean Model Output

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Sampling Physical Ocean Fields in WCRP CMIP5 Simulations

CLIVAR Working Group for Ocean Model Development (WGOMD)
Committee on CMIP5 Ocean Model Output

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Abstract

This document presents recommendations for sampling physical ocean fields for the World Climate Research Program (WCRP) Coupled Model Intercomparison Project #5 (CMIP5) numerical experiments. These experiments are of particular interest for the 5th IPCC assessment (IPCC-AR5). We include guidelines for space and time sampling, and rationalizations for a list of fields to be archived. The perspective taken here is that of physical ocean scientists aiming to enhance the scientific utility of model simulations contributing to CMIP5. We focus on the liquid ocean in this document. The audience for this document includes the CLIVAR Working Group for Coupled Modeling (WGCM), ocean modelers contributing results to CMIP5, and scientists aiming to analyze CMIP5 simulations.

Note about updates to this document

This version of the report was prepared on February 12, 2009. The most updated version is available at the following two websites:

- The CMIP5 location on the PCMDI website
  `www-pcmdi.llnl.gov/`
- The CLIVAR Working Group for Ocean Model Development website
  `www.clivar.org/organization/wgomd/wgomd.php`

Addendums and clarifications to this report will also be provided at these websites. Please refer to either website prior to finalizing your ocean variables submitted to the CMIP5 repository.

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Chapter 1

Executive Summary

This document presents recommendations for sampling physical ocean fields for the World Climate Research Program (WCRP) Climate Model Intercomparison Project #5 (CMIP5) numerical experiments. These experiments are of particular interest for the 5th IPCC assessment (IPCC-AR5). We include guidelines for space and time sampling, and rationalizations for a list of fields to be archived. The perspective taken here is that of physical ocean scientists aiming to enhance the scientific utility of model simulations contributing to CMIP5. We focus on the liquid ocean in this document. The audience for this document includes the CLIVAR Working Group for Coupled Modeling (WGCM), ocean modelers contributing results to CMIP5, and scientists aiming to analyze CMIP5 simulations.

1.1 Space and time sampling

The following presents requirements for space and time sampling of ocean model fields (refer to Section 3.1 for more details and discussion).

- **Time averages**: Time averaging should include all model time steps over the given range of the average. Products of time dependent fields should be time averaged as a product, using all model time steps to build the average.
- **Space averages**: Space averaging should include all model grid points within the relevant domain (e.g., Atlantic-Arctic, Indian-Pacific, Global). For two dimensional fields, the average is over the horizontal area of the field. For three-dimensional fields, the averages include all points in the horizontal domain within a column.
- **Horizontal grid (assuming grid description standards)**: Simulation results must be provided on the model's native grid (including grid staggering). Full specification of this grid (i.e., all grid factors) must be provided in a standardized grid file. A Fortran remapping subroutine must be provided by each modeling center for the purpose of remapping scalar fields onto an associated spherical latitude-longitude grid, whose resolution is comparable to the native grid (i.e., roughly the same number of grid points). Vector fields (e.g., velocity, fluxes, mass and tracer transports) should remain on the model's native grid (including staggering) and should not be remapped to a non-native spherical grid. Transports across a section (e.g. meridional overturning at a given latitude, transport through a passage, or vertically integrated poleward heat transport) should be computed consistent with the native grid by finding a nearly equivalent path to the section that has been "snapped" to the native grid (often resulting in a "zig-zag" path).[1]
- **Horizontal grid (assuming no grid description standards)**: Simulation results must be provided on the model's native grid (including grid staggering). Absent a standardized grid specification file and remapping tools, each modeling center must provide remapped scalar fields onto a spherical grid, with comparable resolution to the native grid (i.e., roughly the same number of grid points), in addition to the native grid scalar fields. Vector fields (e.g., velocity, fluxes, transports) should remain on the model's native grid (including staggering) and should not be remapped to a spherical grid. Transports across a section (e.g. meridional overturning at a given latitude, transport through a passage, or vertically integrated poleward heat transport) should be computed consistent with the native grid by finding a nearly equivalent path to the section that has been "snapped" to the native grid (often resulting in a "zig-zag" path).
- **Vertical grid**: All model output should be reported on the model's native vertical grid. For isopycnal or terrain following models, output should also be recorded on an associated rescaled geopotential or rescaled pressure

grid, with resolution directly comparable to that of the native grid. Vertical remapping to the non-native grid should be performed every model time step and in a conservative manner. Remapping subsampled fields can lead to erroneous analysis, so must be avoided.

- **Flux Units**: Fluid transport should be recorded as a mass transport (kg/s) rather than volume transport (m^3/s).

1.2 Ranking the fields

WGCM (2007) recommends a list of fields for the CMIP3 archive. They propose two levels of priority: *high* and *low*. Given the more extensive list of variables presented here, with more anticipated for future CMIPs, we propose the following characterization that allows more fine grain specifications:

- Level-0 fields provide:
  - A rudimentary characterization of the model configuration;
  - A broad assessment of the simulated climate state;
  - A broad assessment of the simulated climate change.

  These fields serve as a baseline for the CMIP5 repository.
- Level-1 fields are characterized by one or more of the following qualities:
  - fields that render a measure of the mass and tracer transports over the globe, within semi-enclosed basins, or across sections;
  - fields that allow for a quantification of mass, heat, salt, and momentum budgets on a global or regional scale;
  - subsets of Level-0 fields that facilitate a more convenient means of performing a quick assessment.
- Level-2 fields are characterized by one or more of the following qualities:
  - auxiliary fields that render a more complete characterization of the simulation;
  - fields that allow for process level budget analysis;
  - fields that provide for a more detailed study of transients and/or variability;
  - fields that characterize subgrid scale parameterizations.

Many CMIP5 participants will have limitations in resources that preclude them from submitting some of the fields, even some of the baseline Level-0 fields. Incomplete submissions also occurred for CMIP3. As the CMIP process is voluntary, there is no *all or nothing* criterion imposed on contributors. So the prioritization terms *high* and *low* have an uncertain meaning. In effect, prioritization occurs within each model group based on their desire to have their model output included in the many comparison studies spawned by the archived results. We thus prefer the proposed *level categorization* defined here, as it allows for a reasonably coherent partitioning of the fields suggested for CMIP5 by identifying common scientific uses of the fields.

1.3 Size of requested output

Relative to CMIP3, we are proposing a rather large number of new fields. It is thus appropriate to consider whether a modeling center should aim to archive all fields for all experiments. We anticipate that doing so is unreasonable, both for the modeling centers and for the archive repository. It is particularly difficult to archive new three dimensional fields. We are thus led to consider how to garner a nontrivial level of scientific use from a field without overwhelming the archives. We address this issue in the following manners:

- Some of the newly requested fields will be of scientific use even if they are saved *only* for the historical experiment. The CFC-11 tracer suggested in Section 4.2 is an obvious candidate for saving just in the historical experiment.
- Fields listed in Tables 2.9 and 2.10 (see Sections 4.6 and 4.7), which support a characterization of the vertical and lateral subgrid scale (SGS) parameterizations, should be saved as climatological monthly means computed from 01JAN1986 to 31DEC2005 for the CMIP5 “historical” experiment (rather than saving the fields for every month simulated and every experiment).

Each of these recommendations leads to a tremendous reduction in the model output, yet without sacrificing much of the scientific utility of these new fields.

1.4 Names of fields

To facilitate archiving fields, it is useful to have an associated standard name. WGCM (2007) presents standard names for the fields in CMIP3. We recommend using the same names for CMIP5 fields that are the same as in CMIP3. The CF metadata conventions "standard names" listed here are in some cases provisional. A final approved list will appear on the PCMDI website alter in 2009.

Notes

[1]The zig-zag approach yields conservative transport *if* the model formulation is based on finite volumes, as is the case for many ocean climate models. Finite element models are a notable exception, where different techniques have been proposed to compute transports (Cotter and Gorman, 2008; Sidorenko et al., 2008).

Chapter 2

Ocean Fields for CMIP5

This chapter presents a tabulated list of ocean model fields recommended for the CMIP5 archive. In each table, we present the field name; its relation to CMIP3 as detailed in WGCM (2007) (same field, modified field, or new field); physical units; time sampling for output (mean over day or month, or max over a day or month); spatial shape of the field; level characterization; and for what experiment the fields should be saved (all experiments, just the historical experiment, or only one entry as a representation for all experiments). Unless otherwise specified, results should be submitted for the full length of each IPCC scenario experiment. All fields should be reported as "missing" over grid cells that are entirely land. The spatial shape of a field that has no horizontal dimension(s) is indicated by a 0, whereas horizontal two dimensional fields are denoted XY, vertical two dimensional fields are denoted YZ or Yρ, and three-dimensional fields are denoted XYZ.[2]

STANDARD NAME	CMIP3	UNITS	TIME	SHAPE	LEVEL	EXPT
sea_water_equation_of_state	new	fcn of θ, S, p	N/A	Fortran	0	once
sea_water_freezing_temperature_equation	new	fcn of S, p	N/A	Fortran	0	once
reference_sea_water_density_for_boussinesq_approximation	new	kg/m^3	static	0	0	once
sea_floor_depth_below_geoid	same	m	static	XY	0	once
grid specification (lengths, areas)	new	m and m^2	static	XY	0	once
region	new	dimensionless	static	XY	1	once

Table 2.1: This table presents static fields and functions to be saved for the ocean model component in CMIP5. These fields provide basic information about the model configuration, and need only be archived once for all the model experiments in the CMIP5 repository (hence the "once" entry for the experiment column). Listed are the field name, its relation to recommendations for CMIP3 from WGCM (2007) (same, new, modified), units, time average for output, shape of the field, and rank level. The CF metadata conventions "standard names" listed here are in some cases provisional. A final approved list will appear on the PCMDI website later in 2009. Note that the equation of state and the seawater freezing temperature should be archived as Fortran subroutines. Refer to Section 4.1 for more details.

Notes

[2]XYZ is a shorthand for the more detailed prescription of both horizontal and vertical grids given in Sections 3.1.2 and 3.1.3. For example, isopycnal coordinate ocean models must submit three-dimensional fields on both XYZ and their native XYρ. In contrast, pressure coordinate models need not remap their output to geopotential, so that XYZ represents a shorthand for XYp.

STANDARD NAME	CMIP3	UNITS	TIME	SHAPE	LEVEL	EXPT
sea_water_mass	new	kg	month	0	0	all
sea_water_pressure_at_sea_floor	new	dbar	month	XY	0	all
sea_water_pressure_at_sea_water_surface	new	dbar	month	XY	1	all
sea_water_volume	new	m^3	month	0	0	all
sea_surface_height_above_geoid	same	m	month	XY	0	all
square_of_sea_surface_height_above_geoid	new	m^2	month	XY	2	all
global_average_sea_level_change	same	m	month	0	0	all
global_average_steric_sea_level_change	new	m	month	0	0	all
global_average_thermosteric_sea_level_change	same	m	month	0	0	all
sea_water_mass_per_unit_area	new	kg/m^2	month	XYZ	0	all
cell_thickness	new	m	month	XYZ	0	all
sea_water_potential_temperature	same	K	month	XYZ	0	all
sea_water_potential_temperature	same	K	month	0	0	all
sea_surface_temperature	same	K	month	XY	1	all
sea_surface_temperature	new	K	day	XY	2	all
square_of_sea_surface_temperature	new	K^2	month	XY	2	all
square_of_sea_surface_temperature	new	K^2	day	XY	2	all
sea_water_salinity	same	psu	month	XYZ	0	all
sea_water_salinity	new	psu	month	0	0	all
sea_surface_salinity	new	psu	month	XY	1	all
sea_water_potential_density	same	kg/m^3	month	XYZ	2	all
sea_water_age_since_surface_contact	new	year	month	XYZ	2	all
moles_per_unit_mass_of_cfc11_in_sea_water	new	mol/kg	month	XYZ	2	hist
ocean_barotropic_mass_streamfunction	mod	kg/s	month	XY	2	all
ocean_mixed_layer_thickness_defined_by_sigma_t	mod	m	month	XY	2	all
square_of_ocean_mixed_layer_thickness_defined_by_sigma_t	new	m^2	month	XY	2	all
ocean_mixed_layer_thickness_defined_by_mixing_scheme	new	m	daily max	XY	2	all
ocean_mixed_layer_thickness_defined_by_mixing_scheme	new	m	monthly max	XY	2	all

Table 2.2: This table summarizes the scalar fields that should be saved from the ocean component in CMIP5 ocean model simulations. Listed are the field name, its relation to recommendations for CMIP3 from WGCM (2007) (same, new, modified), units, time average for output, shape of the field, and rank level. The CF metadata conventions "standard names" listed here are in some cases provisional. A final approved list will appear on the PCMDI website later in 2009. Note that CFC-11 should be archived *only* for the historical experiment. See Section 4.2 for more details.

STANDARD NAME	CMIP3	UNITS	TIME	SHAPE	LEVEL	EXPT
sea_water_x_velocity	same	m/s	month	XYZ	0	all
sea_water_y_velocity	same	m/s	month	XYZ	0	all
upward_ocean_mass_transport	mod	kg/s	month	XYZ	0	all
square_of_upward_ocean_mass_transport	new	$(kg/s)^2$	month	XYZ	0	all
ocean_mass_x_transport	new	kg/s	month	XYZ	1	all
ocean_mass_y_transport	new	kg/s	month	XYZ	1	all
ocean_meridional_overturning_mass_streamfunction	mod	kg/s	month	YZ-basin	1	all
ocean_meridional_overturning_mass_streamfunction	new	kg/s	month	$Y\rho$-basin	1	all
ocean_y_overturning_mass_streamfunction	new	kg/s	month	YZ-basin	1	all
ocean_y_overturning_mass_streamfunction	new	kg/s	month	$Y\rho$-basin	1	all
ocean_meridional_overturning_mass_streamfunction_due_to_bolus_advection	new	kg/s	month	YZ-basin	2	all
ocean_meridional_overturning_mass_streamfunction_due_to_bolus_advection	new	kg/s	month	$Y\rho$-basin	2	all
ocean_y_overturning_mass_streamfunction_due_to_bolus_advection	new	kg/s	month	YZ-basin	2	all
ocean_y_overturning_mass_streamfunction_due_to_bolus_advection	new	kg/s	month	YZ-basin	2	all
northward_ocean_heat_transport	mod	W	month	XY	1	all
northward_ocean_heat_transport_due_to_bolus_advection	new	W	month	XY	2	all
northward_ocean_heat_transport_due_to_diffusion	new	W	month	XY	2	all
ocean_heat_x_transport	new	W	month	XY	1	all
ocean_heat_y_transport	same	W	month	XY	1	all
ocean_heat_x_transport_due_to_bolus_advection	new	W	month	XY	2	all
ocean_heat_x_transport_due_to_diffusion	new	W	month	XY	2	all
ocean_heat_y_transport_due_to_bolus_advection	new	W	month	XY	2	all
ocean_heat_y_transport_due_to_diffusion	new	W	month	XY	2	all
northward_ocean_heat_transport_due_to_gyre	same	W	month	Y-basin	1	all
northward_ocean_heat_transport_due_to_overturning	same	W	month	Y-basin	1	all
northward_ocean_salt_transport_due_to_gyre	same	kg/s	month	Y-basin	1	all
northward_ocean_salt_transport_due_to_overturning	same	kg/s	month	Y-basin	1	all

Table 2.3: This table summarizes the vector fields, or components of vector fields, that should be saved from the ocean component in CMIP5 simulations. Listed are the field name, its relation to recommendations for CMIP3 from WGCM (2007) (same, new, modified), units, time average for output, shape of the field, and rank level. The CF metadata conventions "standard names" listed here are in some cases provisional. A final approved list will appear on the PCMDI website later in 2009. Certain of the fields should be partitioned into Atlantic-Arctic, Indian-Pacific, and Global regions. See Section 4.3 for further details.

STANDARD NAME	CMIP3	UNITS	TIME	SHAPE	LEVEL	EXPT
barents_opening	new	kg/s	month	0	1	all
bering_strait	new	kg/s	month	0	1	all
canadian_archipelago	new	kg/s	month	0	1	all
denmark_strait	new	kg/s	month	0	1	all
drake_passage	new	kg/s	month	0	1	all
english_channel	new	kg/s	month	0	1	all
equatorial_undercurrent	new	kg/s	month	0	1	all
faroe_scotland_channel	new	kg/s	month	0	1	all
florida_bahamas	new	kg/s	month	0	1	all
fram_strait	new	kg/s	month	0	1	all
iceland_faroe_channel	new	kg/s	month	0	1	all
indonesian_throughflow	new	kg/s	month	0	1	all
mozambique_channel	new	kg/s	month	0	1	all
taiwan_and_luzon_straits	new	kg/s	month	0	1	all
windward_passage	new	kg/s	month	0	1	all

Table 2.4: This table summarizes the sections for archiving the vertically integrated mass transport from the ocean component in CMIP5 simulations, to be identified by the CF standard name sea_water_transport_across_line. Each geographical region has an associated string-valued coordinate given by the name in this table (see Section 4.4 for further details). Listed are the field name, its relation to recommendations for CMIP3 from WGCM (2007) (same, new, modified), units, time average for output, shape of the field, and rank level. The CF metadata conventions "standard names" listed here are in some cases provisional. A final approved list will appear on the PCMDI website later in 2009.

STANDARD NAME	CMIP3	UNITS	TIME	SHAPE	LEVEL	EXPT
rainfall_flux	mod	kg/(m^2 s)	month	XY	1	all
snowfall_flux	mod	kg/(m^2 s)	month	XY	1	all
water_evaporation_flux	new	kg/(m^2 s)	month	XY	1	all
water_flux_into_sea_water_from_rivers	new	kg/(m^2 s)	month	XY	1	all
water_flux_into_sea_water_from_icebergs	same	kg/(m^2 s)	month	XYZ	1	all
water_flux_into_sea_water_due_to_sea_ice_thermodynamics	new	kg/(m^2 s)	month	XY	1	all
water_flux_into_sea_water	new	kg/(m^2 s)	month	XY	1	all
water_flux_into_sea_water_without_flux_correction	new	kg/(m^2 s)	month	XY	1	all
water_flux_correction	same	kg/(m^2 s)	month	XY	1	all

Table 2.5: This table provides a summary of the air-sea and ice-sea boundary fluxes of water mass that should be saved from the ocean model component in CMIP5 simulations. Positive fluxes are into the ocean, with the single exception of evaporation, which is positive for water leaving the liquid ocean. Listed are the field name, its relation to recommendations for CMIP3 from WGCM (2007) (same, new, modified), units, time average for output, shape of the field, and rank level. The CF metadata conventions “standard names” listed here are in some cases provisional. A final approved list will appear on the PCMDI website later in 2009. See Section 4.5.1 for details.

STANDARD NAME	CMIP3	UNITS	TIME	SHAPE	LEVEL	EXPT
virtual_salt_flux_into_sea_water_due_to_rainfall	new	kg/(m^2 s)	month	XY	1	all
virtual_salt_flux_into_sea_water_due_to_evaporation	new	kg/(m^2 s)	month	XY	1	all
virtual_salt_flux_into_sea_water_from_rivers	new	kg/(m^2 s)	month	XY	1	all
virtual_salt_flux_into_sea_water_due_to_sea_ice_thermodynamics	new	kg/(m^2 s)	month	XY	1	all
virtual_salt_flux_into_sea_water	new	kg/(m^2 s)	month	XY	1	all
virtual_salt_flux_correction	same	kg/(m^2 s)	month	XY	1	all
downward_sea_ice_basal_salt_flux	new	kg/(m^2 s)	month	XY	1	all
salt_flux_into_sea_water_from_rivers	new	kg/(m^2 s)	month	XY	1	all

Table 2.6: This table provides a summary of the air-sea and ice-sea boundary fluxes of salt mass that should be saved from the ocean model component in CMIP5 simulations. Positive fluxes are into the ocean. Listed are the field name, its relation to recommendations for CMIP3 from WGCM (2007) (same, new, modified), units, time average for output, shape of the field, and rank level. The CF metadata conventions “standard names” listed here are in some cases provisional. A final approved list will appear on the PCMDI website later in 2009. See Section 4.5.2 for details.

STANDARD NAME	CMIP3	UNITS	TIME	SHAPE	LEVEL	EXPT
upward_geothermal_heat_flux_at_sea_floor	new	W/m^2	month	XY	1	all
temperature_flux_due_to_rainfall_expressed_as_heat_flux_into_sea_water	new	W/m^2	month	XY	1	all
temperature_flux_due_to_evaporation_expressed_as_heat_flux_out_of_sea_water	new	W/m^2	month	XY	1	all
temperature_flux_due_to_runoff_expressed_as_heat_flux_into_sea_water	new	W/m^2	month	XYZ	1	all
heat_flux_into_sea_water_due_to_snow_thermodynamics	new	W/m^2	month	XYZ	1	all
heat_flux_into_sea_water_due_to_sea_ice_thermodynamics	mod	W/m^2	month	XYZ	1	all
heat_flux_into_sea_water_due_to_iceberg_thermodynamics	new	W/m^2	month	XYZ	1	all
surface_net_downward_longwave_flux	mod	W/m^2	month	XY	1	all
surface_downward_latent_heat_flux	mod	W/m^2	month	XY	1	all
surface_downward_sensible_heat_flux	mod	W/m^2	month	XY	1	all
surface_net_downward_shortwave_flux	mod	W/m^2	month	XY	1	all
downwelling_shortwave_flux_in_sea_water	new	W/m^2	month	XYZ	1	all
heat_flux_correction	same	W/m^2	month	XY	1	all

Table 2.7: This table provides a summary of the air-sea and ice-sea boundary fluxes of heat that should be saved from the ocean component from the ocean component in CMIP5 simulations. Positive fluxes are into the ocean. Listed are the field name, its relation to recommendations for CMIP3 from WGCM (2007) (same, new, modified), units, time average for output, shape of the field, and rank level. The CF metadata conventions "standard names" listed here are in some cases provisional. A final approved list will appear on the PCMDI website later in 2009. See Section 4.5.3 for details. For the geothermal heating, most models use a static geothermal heating, in which case only one time step need be archived. If time dependent, then monthly fields are requested. Note that many climate models place boundary fluxes at the ocean surface. However, more general couplings are being considered (e.g., a sea ice model that interacts with more than the surface ocean cell). To allow for such generality, we note that many of the fluxes can be three-dimensional.

STANDARD NAME	CMIP3	UNITS	TIME	SHAPE	LEVEL	EXPT
surface_downward_x_stress	mod	N/m^2	month	XY	1	all
surface_downward_y_stress	mod	N/m^2	month	XY	1	all
surface_downward_x_stress_correction	mod	N/m^2	month	XY	1	all
surface_downward_y_stress_correction	mod	N/m^2	month	XY	1	all

Table 2.8: This table presents the net surface stress applied at the liquid ocean surface by air-sea plus ice-sea interactions that should saved from the ocean model component in CMIP5 simulations. Positive fluxes accelerate the ocean. Listed are the field name, its relation to recommendations for CMIP3 from WGCM (2007) (same, new, modified), units, time average for output, shape of the field, and rank level. The CF metadata conventions "standard names" listed here are in some cases provisional. A final approved list will appear on the PCMDI website later in 2009. See Section 4.5.4 for details. Note that the units N/m^2 are identical to Pa.

STANDARD NAME	CMIP3	UNITS	TIME	SHAPE	LEVEL	EXPT
ocean_vertical_heat_diffusivity	new	m^2/s	$\langle mon\rangle^{20}$	XYZ	2	hist
ocean_vertical_salt_diffusivity	new	m^2/s	$\langle mon\rangle^{20}$	XYZ	2	hist
ocean_vertical_tracer_diffusivity_due_to_background	new	m^2/s	$\langle mon\rangle^{20}$	XYZ	2	hist
ocean_vertical_tracer_diffusivity_due_to_tides	new	m^2/s	$\langle mon\rangle^{20}$	XYZ	2	hist
tendency_of_ocean_potential_energy_content	new	W/m^2	$\langle mon\rangle^{20}$	XYZ	2	hist
tendency_of_ocean_potential_energy_content_due_to_tides	new	W/m^2	$\langle mon\rangle^{20}$	XYZ	2	hist
tendency_of_ocean_potential_energy_content_due_to_background	new	W/m^2	$\langle mon\rangle^{20}$	XYZ	2	hist
ocean_vertical_momentum_diffusivity	new	m^2/s	$\langle mon\rangle^{20}$	XYZ	2	hist
ocean_vertical_momentum_diffusivity_due_to_background	new	m^2/s	$\langle mon\rangle^{20}$	XYZ	2	hist
ocean_vertical_momentum_diffusivity_due_to_tides	new	m^2/s	$\langle mon\rangle^{20}$	XYZ	2	hist
ocean_vertical_momentum_diffusivity_due_to_form_drag	new	m^2/s	$\langle mon\rangle^{20}$	XYZ	2	hist
ocean_kinetic_energy_dissipation_per_unit_area_due_to_vertical_friction	new	W/m^2	$\langle mon\rangle^{20}$	XYZ	2	hist

Table 2.9: This table summarizes some fields that support the study of vertical/dianeutral SGS parameterizations. These fields should be saved from the ocean model component in CMIP5 simulations. Listed are the field name, its relation to recommendations for CMIP3 from WGCM (2007) (same, new, modified), units, time average for output, shape of the field, and rank level. The CF metadata conventions "standard names" listed here are in some cases provisional. A final approved list will appear on the PCMDI website later in 2009. See Section 4.6 for details. In particular, the dominant scientific use of the fields in this table are realized by archiving *just* the climatological monthly means computed from 01JAN1986 to 31DEC2005 for the CMIP5 "historical" experiment; hence the notation $\langle mon\rangle^{20}$ for the time.

STANDARD NAME	CMIP3	UNITS	TIME	SHAPE	LEVEL	EXPT
ocean_tracer_bolus_laplacian_diffusivity	new	m^2/s	$\langle mon\rangle^{20}$	XYZ	2	hist
ocean_tracer_bolus_biharmonic_diffusivity	new	m^4/s	$\langle mon\rangle^{20}$	XYZ	2	hist
ocean_tracer_epineutral_laplacian_diffusivity	new	m^2/s	$\langle mon\rangle^{20}$	XYZ	2	hist
ocean_tracer_epineutral_biharmonic_diffusivity	new	m^2/s	$\langle mon\rangle^{20}$	XYZ	2	hist
ocean_tracer_xy_laplacian_diffusivity	new	m^2/s	$\langle mon\rangle^{20}$	XYZ	2	hist
ocean_tracer_xy_biharmonic_diffusivity	new	m^4/s	$\langle mon\rangle^{20}$	XYZ	2	hist
tendency_of_ocean_eddy_kinetic_energy_content_due_to_bolus_transport	new	W/m^2	$\langle mon\rangle^{20}$	XYZ	2	hist
ocean_momentum_xy_laplacian_diffusivity	new	m^2/s	$\langle mon\rangle^{20}$	XYZ	2	hist
ocean_momentum_xy_biharmonic_diffusivity	new	m^4/s	$\langle mon\rangle^{20}$	XYZ	2	hist
ocean_kinetic_energy_dissipation_per_unit_area_due_to_xy_friction	new	W/m^2	$\langle mon\rangle^{20}$	XYZ	2	hist

Table 2.10: This table summarizes some fields that support the study of lateral SGS parameterizations. These fields should be saved from the ocean model component in CMIP5 simulations. Listed are the field name, its relation to recommendations for CMIP3 from WGCM (2007) (same, new, modified), units, time average for output, shape of the field, and rank level. The CF metadata conventions "standard names" listed here are in some cases provisional. A final approved list will appear on the PCMDI website later in 2009. See Section 4.7 for details. In particular, the dominant scientific use of the fields in this table are realized by archiving *just* the climatological monthly means computed from 01JAN1986 to 31DEC2005 for the CMIP5 "historical" experiment; hence the notation $\langle mon\rangle^{20}$ for the time.

Chapter 3

Purposes and Perspective

This document serves the following purposes:

- To rationalize a list of physical ocean model fields to be archived for the Coupled Model Intercomparison Project (CMIP5) supporting the IPCC-AR5.
- To offer guidance to ocean climate modelers for enhancing the scientific relevance of sampled model output.
- To articulate certain needs of ocean scientists aiming to analyze CMIP5 model output, and whose research directly supports Working Group 1 (WG1) goals.

This document was solicited by the CLIVAR Working Group for Coupled Modeling (WGCM), with the list of fields provided by WGCM (2007) used as a starting point. The perspective taken in this document is that of physical ocean scientists aiming to enhance the scientific potential available from simulations planned for CMIP5. Included here are discussions of space and time sampling, preferred units for transport, and rationale for archiving certain model fields. We include technical notes at the end of the document to further support the recommendations presented in the main text.

Before addressing the main topics of this document, we give some perspective on the goals of the very successful CMIP3 project maintained by PCMDI supporing IPCC-AR4, and how we hope to further enhance the success of the ocean model output archived for CMIP5. The following is stated on the PCMDI website

`www – pcmdi.llnl.gov/ipcc/about_ipcc.php`

> This unprecedented collection of recent model output is officially known as the "WCRP CMIP3 multi-model dataset." It is meant to serve IPCC's Working Group 1, which focuses on the physical climate system – atmosphere, land surface, ocean and sea ice – and the choice of fields archived at the PCMDI reflects this focus.

The perspective forming the basis for the present document is fully aligned with this quote from PCMDI. Most notably, we aim to support WG1 research by enhancing the ocean model output archived as part of CMIP5.

As noted in various places in the following, we believe there are specific shortcomings in the ocean model output contained in CMIP3, and these shortcomings compromise the output's utility for ocean scientists aiming to support WG1 science. Broadly, the following outlines the major shortcomings with the ocean model output in CMIP3.

- There is insufficient model output for constructing budgets of mass, heat, and salt. In particular, there is incomplete information regarding the surface boundary fluxes, and those boundary fluxes that are present are generally *not* archived on the ocean model native grid.
- Vector fields (e.g., velocity, transport, fluxes) are remapped to a spherical grid from the commonly used non-spherical native grids of global ocean models. This remapping occurred despite the absence of a generally agreed upon remapping algorithm. The result is incomplete ocean model contributions, and/or output with remapped vector fields that are generally untrustworthy.
- There is a paucity of fields of use for studying the impact of subgrid scale (SGS) parameterizations used by the models. SGS parameterizations are of leading order importance for most IPCC class ocean models. Numerous IPCC class ocean models targeted for CMIP5 will incorporate a broad suite of SGS methods. The CMIP5 archive thus provides the ocean research community with a tremendous opportunity to evaluate the integrity and impact of the various SGS schemes.

Many CMIP5 models will include some form of an ocean biogeochemistry model. "Physics first" is the *mantra* from biogeochemists aiming to reduce the huge parameter phase space in their models. Absent a more robust understanding of ocean physics, the research community is hard pressed to rationalize the new tracers and tracer methods being employed to simulate ocean biogeochemical cycles. As evidenced by the wide range in physical model behaviour in a recent comparison of global ocean-ice models (Griffies et al., 2009), it is unwise to assume the physical problem is "solved." We therefore must continue to address the physical problem head-on, with the CMIP5 simulations providing a promising opportunity.

By articulating shortcomings with ocean model output in CMIP3, and in turn by identifying certain needs of ocean scientists aiming to analyse model output, this document aims to inform those charged with establishing protocols for CMIP5. Entraining more ocean science into the WG1 process will enhance the robustness of climate model projections, which is of fundamental importance for the IPCC process.

3.1 Sampling Ocean Model Fields

There are many uncertainties associated with climate modeling, with some uncertainty related to poor sampling methods. Poor sampling compromises on the goal of providing an honest record of a simulation. The result is a lack of trust that analysts place on a particular model's output. This section offers basic principles for sampling an ocean model simulation to help maintain a high degree of integrity in the model output. We also comment on the relevant physical dimensions for use in measuring fluid transport.

3.1.1 Temporal sampling

A climate model simulation exhibits multiple regimes of temporal behaviour. Furthermore, the refined resolution models now envisioned being integrated in support of IPCC-AR5, and beyond, possess an ever growing spectrum of fluctuations. Correspondingly, analysis methods require multiple time averages depending on the phenomena to be studied (e.g., daily, monthly, seasonal, annual, decadal, centennial, millenial). We are thus motivated to suggest the following, very basic, requirements for temporal sampling of ocean model fields.

- **Time averages**: Time averaging should include all model time steps over the given range of the average. Products of time dependent fields should be time averaged as a product, using all model time steps to build the average.

Sub-sampling generally arises from the desire to reduce input/output volume or intermediate storage requirements. However, time averages based on sub-sampling introduce the potential for aliasing, and reporting a time averaged product from the product of the time averages ignores possible correlations.[3]

An example helps to illustrate the above considerations. A commonly saved diagnostic requiring nonlinear products is the mass transport passing through the face of a grid cell: $\mathcal{V}^{(\hat{\mathbf{n}})} = \rho\,\mathbf{v}\cdot\hat{\mathbf{n}}\,\mathrm{d}A$, where ρ is the density, $\mathbf{v}$ is the velocity, and $\hat{\mathbf{n}}\,\mathrm{d}A$ is the area of the cell face through which the flow is passing, with outward normal $\hat{\mathbf{n}}$. In particular, the $\hat{\mathbf{y}}$ component of this field is used to construct the overturning streamfunction discussed in Section 4.3. For a rigid lid[4] Boussinesq geopotential ocean model, the only time-dependent field appearing in the transport $\mathcal{V}^{(\hat{\mathbf{n}})}$ is the velocity, since ρ is set to the constant reference density ρ_o, and cells have a time independent volume. However, for a free surface model, the top cell has a time dependent thickness, and for isopycnal and sigma models, all thicknesses are time dependent. Finally, for non-Boussinesq models, the density factor ρ becomes time dependent, thus making $\mathcal{V}^{(\hat{\mathbf{n}})}$ the product of three time dependent fields.[5]

3.1.2 Spatial sampling in the horizontal

Given the rather complicated grids used by some ocean models, in both the horizontal and vertical, it is important to be clear that spatial averaging must include the appropriate grid factors (e.g., grid cell volume, horizontal area, density factors, etc.).

- **Space averages**: Space averaging should include all model grid points within the relevant domain (e.g., Atlantic-Arctic, Indian-Pacific, Global). For two dimensional fields, the average is over the horizontal area of the field. For three-dimensional fields, the averages include all points in the horizontal domain within a column.

An especially controversial aspect of CMIP3 is the requirement for model output to be reported on a spherical latitude-longitude grid. That is, from WGCM (2007):

> Latitude-longitude grids must be rectilinear; i.e., have a unique set of longitudes that applies to all latitudes.

Reporting fields on a spherical grid simplifies the following aspects of model analysis:

- Visualization
- Analysis and comparison between models and observational data
- Calculations of multi-model statistics.

However, an increasing number of modeling centers use generalized locally orthogonal grids (i.e., *curvilinear grids*) to avoid the spherical coordinate singularity at the North Pole. The evolution away from the latitude-longitude ocean models, which often employ unphysical methods to stabilize the simulations, represents an important step toward enhancing the scientific integrity of Arctic simulations in global climate models.

For ocean models employing non-latitude-longitude grids, the spherical requirement in CMIP3 necessitates a remapping procedure to move model output from the native grid to a spherical grid. The following represents the scientific reasons to abandon horizonally remapped output:

- Since coastlines cannot be uniquely remapped to a non-native grid, conservation of internal flux integrals over any area containing a coastline will be lost regardless the "conservative" properties of the remapping algorithm.
- The remapped fields are of little use for analysis requiring further manipulations, such as for approximating transports or forming dot products. Relatedly, budgets computed with remapped output are spurious.
- There are unresolvable ambiguities involved with mapping algorithms moving discrete vector fields from the native grid to the spherical grid. For example, the "eastward velocity component," which is a required field for CMIP3, is undefined at the North Pole. Consequently, there is no universally agreed upon method to remap discrete vector fields from their native grid to a spherical grid.[6]

For these reasons, some modeling groups failed to submit ocean model results for CMIP3 in a timely manner. The nontrivial chance that a scientist would produce misleading results analyzing the remapped output outweighed benefits accrued by having the output in the CMIP3 archive housed at PCMDI. This argument held the day in many groups for some of the vector fields, whereas in other groups it resulted in nontrivial delays. From the user perspective, many ocean scientists associated with WG1 research avoided accessing remapped ocean output, especially vector fields, since the output was not trusted.

A workable middle ground is desired to maintain unity amongst the various IPCC working group needs, and integrity of the model results. The middle ground we propose recognizes the scientific relevance of native grid results, the utility of distributing scalar fields on a spherical grid, and the dis-utility of remapping vector fields. We thus propose the following specification of the horizontal grid for ocean model fields.

- **Horizontal grid (assuming standards)**: Simulation results must be provided on the model's native grid (including grid staggering). Full specification of this grid (i.e., all grid factors) must be provided in a standardized grid file. A Fortran remapping subroutine must be provided by each modeling center for the purpose of remapping scalar fields onto an associated spherical latitude-longitude grid, whose resolution is comparable to the native grid (i.e., roughly the same number of grid points). Vector fields (e.g., velocity, fluxes, mass and tracer transports) should remain on the model's native grid (including staggering) and should not be remapped to a non-native spherical grid. Transports across a section (e.g. meridional overturning at a given latitude, transport through a passage, or vertically integrated poleward heat transport) should be computed consistent with the native grid by finding a nearly equivalent path to the section that has been "snapped" to the native grid (often resulting in a "zig-zag" path).

The feasibility of this recommendation depends on the availability of a standard for the model's grid specification file, and associated standard utilities to remap scalar fields from native to spherical grids. We have more to say on this point in Section 4.1, where we mention the existence of a proto-type for a standard grid file. However, given that the community has yet to fully converge on a standard, we present the following modified recommendation to allow for the potential lack of convergence in time for AR5.

- **Horizontal grid (assuming no standards)**: Simulation results must be provided on the model's native grid (including grid staggering). Absent a standardized grid specification file and remapping tools, each modeling center must provide remapped scalar fields onto a spherical grid, with comparable resolution to the native grid (i.e., roughly the same number of grid points), in addition to the native grid scalar fields. Vector fields (e.g., velocity, fluxes, transports) should remain on the model's native grid (including staggering) and should not be remapped to a spherical grid. Transports across a section (e.g. meridional overturning at a given latitude, transport through a passage, or vertically integrated poleward heat transport) should be computed consistent with the native grid by finding a nearly equivalent path to the section that has been "snapped" to the native grid (often resulting in a "zig-zag" path).

3.1.3 Spatial sampling in the vertical

There are two questions to answer regarding the vertical coordinate:

- Should model output be remapped in the vertical to a common vertical coordinate?
- If remapped, then what is a scientifically relevant vertical coordinate?

There is no ambiguity regarding the vertical grid when working with Boussinesq rigid lid geopotential-coordinate ocean models, as each grid has a fixed vertical position. It was thus sensible for WGCM (2007) to recommend that output in the vertical be on a geopotential grid, preferably remapped to the 33 depth levels used by Levitus (1982).[7] The more recent trend towards free surface geopotential models raises only trivial issues with the surface grid cell, and these issues can be ignored without much loss of accuracy.[8] However, the move towards pressure, isopycnal, and terrain following models increases the relevance of addressing the vertical coordinate questions.

We make the following recommendations regarding the vertical grid.

- **Vertical grid**: All output should be reported on the model's native vertical grid. For isopycnal or terrain following models, output should also be recorded on an associated rescaled geopotential or rescaled pressure grid, with resolution directly comparable to that in the native grid. Vertical remapping to the non-native grid should be performed every model time step and in a conservative manner.

We make the following observations and clarifications regarding these recommendations:

- For the isopycnal and terrain following models, we have given an affirmative answer to the question: "Should model output be remapped in the vertical to a common vertical coordinate?" This answer reflects a recognition that the majority of ocean climate models used for IPCC assessments remain level coordinate geopotential models. Hence, the application of analysis methods is prejudiced towards these models. However, the answer does *not* reflect our belief that geopotential (or pressure) based analysis is the physically most relevant for all purposes. Indeed, many purposes are best served by analyzing output on alternative vertical coordinates, especially on isopycnal coordinates for characterizing water masses. To support the evolution towards a more balanced analysis methodology, in Section 4 we recommend certain fields (e.g., overturning streamfunction) be archived on both geopotential/pressure and density surfaces.
- Vertical remapping with straightforward linear interpolation is reasonably accurate so long as the remapping is done every model time step. Remapping subsampled fields can lead to erroneous analysis, especially with isopycnal models.
- Contrary to the situation in the horizontal, separate vector components can be treated as scalars for the purpose of remapping in the vertical.
- There are two useful choices for remapping isopycnal and terrain following models. For Boussinesq isopycnal and terrain following models, it is natural to consider remapping to the *rescaled geopotential* coordinate (Stacey et al., 1995; Adcroft and Campin, 2004)

$$z^* = H\left(\frac{z-\eta}{H+\eta}\right). \tag{3.1}$$

  In this equation, z is the geopotential, $z = -H(x, y)$ is the ocean bottom, and $z = \eta(x, y, t)$ is the deviation of the free surface from a resting ocean at $z = 0$. To better understand the ratio, note that $z - \eta$ is the thickness of seawater above a particular geopotential, and $H + \eta$ is the total thickness of seawater in the fluid column. Surfaces of constant z^* correspond to geopotentials when $\eta = 0$. For most practical applications of global ocean modeling, z^* surfaces only slightly deviate from constant geopotential surfaces even with nonzero η fluctuations. The advantage of z^* over geopotential is that it has a time independent range $-H \le z^* \le 0$, thus allowing for a more straightforward mapping from a free surface isopycnal or terrain following model.
- The *rescaled pressure* coordinate is defined as

$$p^* = p_{\mathrm{b}}^{o}\left(\frac{p-p_{\mathrm{a}}}{p_{\mathrm{b}}-p_{\mathrm{a}}}\right), \tag{3.2}$$

  where p is the pressure at a grid point; $p_{\mathrm{a}}(x, y, t)$ is the pressure applied at the ocean surface due to overlying atmosphere, sea ice, and/or ice shelves; $p_{\mathrm{b}}(x, y, t)$ is the pressure at the ocean bottom; and $p_{\mathrm{b}}^{o}(x, y)$ is a static reference bottom pressure, such as the initial bottom pressure. To better understand the ratio, note that in a hydrostatic ocean, $g^{-1}(p - p_{\mathrm{a}})$ is the mass per horizontal area of seawater situated above a pressure level p, and $g^{-1}(p_{\mathrm{b}} - p_{\mathrm{a}})$ is the total mass per horizontal area of seawater in the fluid column. For most practical applications of global modeling, constant p^* surfaces only slightly deviate from constant pressure surfaces, even with nonzero fluctuations of p_b. The advantage of p^* over pressure is that p^* has a time independent range $0 \le p^* \le p_b^o$, thus allowing for a more straightforward mapping from a non-Boussinesq isopycnal or terrain following model.

- For visualization purposes, the distinction between geopotential (or rescaled geopotential) and pressure (or rescaled pressure) can be ignored to within a few percentage accuracy, so long as geopotential is measured in metres and pressure is measured in decibars.
- For analysis purposes, the distinction between geopotential (or rescaled geopotential) and pressure (or rescaled pressure) can be ignored when working with model native scalars and fluxes. The differences *cannot* be ignored when performing off-line integration of velocity components to approximate fluxes. This is a central reason that we recommend archiving mass fluxes in addition to velocity components (Section 4.3).

3.2 Physical dimensions for fluid transport

An increasing number of ocean models have removed the Boussinesq approximation, and so are now mass conserving non-Boussinesq models. One benefit of non-Boussinesq models is that the sea level height is more accurate, since these models include steric effects within the prognostic equations (Greatbatch, 1994). We detail these points in Section 4.2.

A notable consequence of moving to a non-Boussinesq model involves the physical dimensions of transport fields. Namely, fluid transport is measured as a mass flux rather than a volume flux. For example, as discussed for temporal sampling in Section 3.1.1, the mass transport $\mathcal{V}^{(\hat{\mathbf{n}})} = \rho\,\mathbf{v}\cdot\hat{\mathbf{n}}\,\mathrm{d}A$ is the mass per time passing through the $\hat{\mathbf{n}}$ face of a grid cell. This transport is naturally represented using the mass Sverdrup 10^9kg/s rather than the volume Sverdrup $10^6\mathrm{m}^3/\mathrm{s}$ commonly used in a Boussinesq model. For a Boussinesq model, the density factor ρ becomes a constant reference density ρ_o, which trivially allows for use of the mass Sverdrup as the unit of transport as well.

In summary, we recommend the following physical dimensions of transport fields:

- **Flux Units**: Fluid transport should be recorded as a mass transport (kg/s) rather than a volume transport (m^3/s).

Notes

[3]Fluctuating fields generally exhibit correlations: $\langle A\,B\rangle = \langle A\rangle\langle B\rangle + \langle A'\,B'\rangle$. Only in the rare case of vanishing correlations, $\langle A'\,B'\rangle = 0$, can we set $\langle A\,B\rangle = \langle A\rangle\langle B\rangle$.

[4]Nearly all IPCC-class ocean models constructed during the past decade have eliminated the rigid lid method, which contrasts to the dominance of rigid lid methods until the mid-1990s. Reasons to jettison the rigid lid include its inability to straightforwardly incorporate real water fluxes, its inability to simulate an ocean with astronomically forced tides, and inefficiencies of the rigid lid solver in a parallel computer environment, (Huang, 1993; Dukowicz et al., 1993; Griffies et al., 2001). Fields saved in ocean models thus need to reflect the scientifically motivated trend away from the rigid lid class of models.

[5]For the special case of a hydrostatic pressure coordinate ocean model, $\rho\,\mathrm{d}z = -g^{-1}\,\mathrm{d}p$, with $\mathrm{d}p$ the pressure increment for the grid cell. In this case, the non-Boussinesq pressure model becomes isomorphic to the Boussinesq free surface geopotential model (Huang et al., 2001; DeSzoeke and Samelson, 2002; Marshall et al., 2004).

[6]The mathematical remapping problem in the continuum is well defined. But fields on a discrete lattice are generally displaced relative to one another, and this displacement presents new problems that do not arise in the continuum. For example, on the Arakawa C-grid, the horizontal components of velocity are not co-located. The same staggering holds for fluxes leaving tracer cells on either the Arakawa B or C grids. Furthermore, the presence of topography greatly complicates the remapping problem.

[7]This recommendation was not a requirement, so many groups participating in CMIP3 chose to report their output on the model's native vertical grid.

[8]We know of no group that considers the question of remapping model fields in the top model cell of a free surface geopotential model to a pre-defined geopotential level. Indeed, there is little reason to do so, as the top cell, whether it has a center at $z = -1$m or $z = 1$m, for example, still represents the model's version of the sea surface.

Chapter 4

Details of CMIP5 Ocean Fields

In this chapter, we present recommendations for ocean model fields to be archived as part of CMIP5. Tables summarizing the fields are presented in the Executive Summary in Chapter 1. Where warranted, we offer formulational details and rationalizations that help to motivate the inclusion of a field on the list. In the section headings for each field, we indicate the ranking of the field according to the system described in Section 1.2.

4.1 Some static fields and functions

This section presents our recommendations for certain static fields and functions that are needed to describe elements of the ocean model. A summary of these fields and functions is given in Table 2.1.

4.1.1 Equation of state (0)

- sea_water_equation_of_state

For certain analyses, it is useful to have access to the model's equation of state. For example, it might be sufficiently accurate to compute the density offline using the model's time mean temperature, salinity, and pressure. To facilitate this recommendation, we recommend archiving a standard Fortran subroutine that allows for the computation of *in situ* density (kg/m^3) as a function of the model's prognostic temperature (typically potential temperature, but possibly conservative temperature), salinity, and pressure. By setting pressure to a constant, one can then trivially compute the potential density referenced to that pressure. Literature reference to the equation of state should also be part of the Fortran code.

4.1.2 Temperature for seawater freezing (0)

- sea_water_freezing_temperature_equation

Ocean models use a variety of equations to determine when liquid seawater freezes to form frazil and then sea ice. It is very useful for studies of high latitude processes to have an archive of the equation used to compute the freezing point (in degrees C) of seawater, as a function of salinity and pressure.

4.1.3 Boussinesq reference density (0)

- reference_sea_water_density_for_boussinesq_approximation

Many ocean climate models employ the Boussinesq approximation, in which there appears a constant reference density ρ_o within budgets for tracer and momentum, and volume is conserved rather than mass. It is useful to have an archive of this constant for CMIP5.[9]

4.1.4 Bathymetry (0)

- sea_floor_depth_below_geoid
- sea_floor_depth_below_geoid_velocity

For global primitive equation ocean models, the geoid is assumed to correspond to the geopotential surface $z = 0$. The distance from $z = 0$ to the ocean bottom defines the ocean depth field $H(x, y)$, or the ocean *bathymetry*. This solid earth boundary used by the model should be archived. Precisely, the bathymetry representing the ocean bottom from the perspective of the model's tracer fields defines the field sea_floor_depth_below_geoid. For non-isopycnal models, it is also necessary to provide the bathymetry sea_floor_depth_below_geoid_velocity as seen by the velocity fields, as this depth generally differs from that seen by tracers. For isopycnal models, the bottom for the velocity field is a time and flow dependent function, and so need not be archived.

If the lateral area for exchange of fluid between columns (e.g., mass transport) is anything other than a simple function of the tracer column depths, then the modulated areas effecting the exchange should be provided. For example, this additional information is necessary for models that allow a strait to be more narrow than the nominal width of the cell.

4.1.5 Grid specification (0)

As discussed in Section 3.1.2, a grid specification file is required to fully describe the model's horizontal grid. A common aspect of the grid file of use for analysis is the horizontal areas of each grid cell, including any staggering that may account for differences in tracer and velocity cell areas. We acknowledge that at present, there is no community-wide standard for the grid specfication file. This situation is unfortunate, as it adds a layer of complexity to the description of an archived field in CMIP5.

A prototype for a standard description of grids can be found in the *Mosaic Gridspec* at

`http://www.gfdl.noaa.gov/˜vb/gridstd/gridstd.html`

This grid file contains a comprehensive description of grids for the purposes of interpolation to other grids, and for vector and tensor operations on gridded fields. A set of tools for creating gridspec files and using them for regridding and plotting gridded fields is described in `http://www.gfdl.noaa.gov/˜vb/grids/`. Regardless the level of standardization between the model grid files, it is critical to have grid information in the CMIP5 archive.

4.1.6 Tracer and velocity cell region masks (1)

- region

Analysis of budgets and properties over ocean basins is commonly performed for the purpose of assessing the integrity of simulations. This analysis generally involves the use of a mask that partitions the model grid into ocean basins (some enclosed seas may be missing from the model). We recommend the following mapping between ocean regions and integer, with the names corresponding to standard CF basin names found at

`http://cf-pcmdi.llnl.gov/documents/cf-standard-names`.

1. southern_ocean
2. atlantic_ocean
3. pacific_ocean
4. arctic_ocean
5. indian_ocean
6. mediterranean_sea
7. black_sea
8. hudson_bay
9. baltic_sea
10. red_sea

These region masks are set according to the following flag_values and flag_meanings, which should be recorded as attributes of the variable:

- flag_values=1,2,3,4,5,6,7,8,9,10
- flag_meanings="southern_ocean, atlantic_ocean, pacific_ocean, arctic_ocean, indian_ocean, mediterranean_sea, black_sea, hudson_bay, baltic_sea, red_sea"

It is very useful for the analyst to have access to this mask, with many analyses requiring the tracer mask. Additionaly, for some grid staggering (such as B-grid), the tracer mask will differ from velocity mask, in which case a mask for the velocity cells should be provided to the CMIP5 archive as a distinct output variable, with the same standard name of region. The two variables are distinguished in netCDF by their coordinates, one being on the tracer grid and the other on the velocity grid.

4.2 Scalar fields

This section presents recommendations for scalar fields to be archived as part of CMIP5. These fields are summarized in Table 2.2.

4.2.1 Total mass of liquid seawater (0)

- sea_water_mass

For the purpose of global budgets in non-Boussinesq models, it is essential to have the total mass of liquid seawater in the ocean domain. This scalar field includes all seawater contained in the liquid ocean, including any enclosed seas that are part of the ocean model integration.[10] As a discrete sum of the three-dimensional grid cells, sea_water_mass is given by

$$\mathcal{M} = \sum_{i,j,k} \rho \, \mathrm{d}x \, \mathrm{d}y \, \mathrm{d}z, \tag{4.1}$$

with ρ the *in situ* density, $\mathrm{d}x \, \mathrm{d}y$ the horizontal area of a grid cell, and $\mathrm{d}z$ the vertical thickness. For time dependent cell thicknesses, the mass is evaluated as the product of time dependent terms. Accuracy thus requires computing this product each time step in order to include temporal correlations. If there is no net boundary flux of mass, then a non-Boussinesq model ideally should retain a constant total mass to within numerical roundoff. In contrast, a Boussinesq model will generally alter its mass even without boundary fluxes, since Boussinesq fluids conserve volume rather than mass.

4.2.2 Pressure at ocean bottom (0)

- sea_water_pressure_at_sea_floor

The bottom pressure in a hydrostatic ocean is given by the gravitational acceleration acting on the mass per area of a fluid column, plus any pressure applied at the ocean surface from the overlying atmosphere or ice. In a discrete model, sea_water_pressure_at_sea_floor is given by the vertical sum over the $k-levels$ in the column

$$p_{\mathrm{b}} = p_{\mathrm{a}} + g \sum_{k} \rho \, \mathrm{d}z \tag{4.2}$$

where p_{a} is the pressure applied at the ocean surface, and we assumed a constant gravitational acceleration (presumably assumed for all CMIP5 simulations). Hence, $g^{-1}(p_{\mathrm{b}} - p_{\mathrm{a}})$ is the mass per horizontal area of a fluid column. The bottom pressure is a prognostic field in non-Boussinesq models, whereas it is diagnosed in Boussinesq models. Anomalies of bottom pressure with respect to a suitable reference value, such as $\rho_o \, g \, H$, provide a means for measuring mass adjustments throughout the water column.

If the model is non-hydrostatic (very uncommon for global climate models), the bottom pressure is affected by the mass per area of the ocean fluid, plus non-hydrostatic fluctuations in the pressure field.

4.2.3 Net pressure from atmosphere, sea ice, ice shelves, etc (1)

- sea_water_pressure_at_sea_water_surface

The pressure applied to the ocean surface from the overlying atmosphere is often neglected in climate simulations. However, for those models that incorporate this effect, they provide a means to simulate the inverse barometer, which presents a rapid barotropic forcing to the ocean. In these cases, it is important to have a map of the applied pressure from the atmosphere acting on the ocean.

In addition to atmospheric mass impacting on the ocean, there is mass from overlying sea ice, ice shelves, icebergs, etc. This mass should also be included in this applied pressure field. Note that solid runoff is defined as all frozen water that enters the ocean from land, such as from snow and land-, lake- and river ice. For example, snow may enter in its frozen state when a land model has a buffer layer of a certain thickness, with all snow exceeding this buffer conveyed to the ocean. Land-ice may enter the ocean as icebergs that may result from an ice sheet/shelf model, or be formed from snow excess. There is no increase is liquid ocean water until the solid runoff melts. However, the presence of solid ice affects the pressure felt within the liquid ocean column.

Please note that rigid lid ocean models, the term "surface pressure" refers to the hydrostatic pressure at $z = 0$ associated with the layer of liquid water between $z = 0$ and $z = \eta$. This pressure is also sometimes referred to as the "lid pressure." It can be positive or negative. This "surface pressure" field is distinctly *not* what we refer to here by sea_water_pressure_at_sea_water_surface. Instead, the field sea_water_pressure_at_sea_water_surface records the non-negative pressure applied at $z = \eta$ due to media above the ocean surface interface. Such pressure may be set to zero in some approximate model formulations, such as the rigid lid, in which case the ocean dynamics is not influenced by movement of overlying media.

4.2.4 Total volume of liquid seawater (0)

- sea_water_volume

For the purpose of global budgets in Boussinesq models, it is essential to have the total volume of liquid seawater in the ocean domain.[11] As a discrete sum of the three-dimensional grid cells, sea_water_total_volume is given by

$$\mathcal{V} = \sum_{i,j,k} \mathrm{d}x \, \mathrm{d}y \, \mathrm{d}z. \tag{4.3}$$

If there are no net boundary fluxes of volume, then a conservative Boussinesq model will retain a constant total volume to within numerical roundoff. In contrast, a non-Boussinesq model will generally alter its volume in cases where the ocean density changes (i.e., via *steric effects*). We detail steric effects in our discussion of sea level in Sections 4.2.5 and 4.2.7.

4.2.5 Sea level (0)

- sea_surface_height_above_geoid

There are various technical points regarding the sea level that are important to understand in order to properly interpret the simulations.

- The ocean surface height is a prognostic field in free surface Boussinesq models, and it can be trivially diagnosed in non-Boussinesq hydrostatic pressure models (which generally compute the bottom pressure prognostically, rather than the surface height). For rigid lid Boussinesq models, the surface height must be computed through a diagnostic elliptic method (Pinardi et al., 1995). In general, surface height is a proxy for the sea level (Wunsch and Stammer, 1998), with the surface height from a non-Boussinesq model most closely reflecting ocean processes responsible for affecting sea level.
- It should be noted in the "comment" attribute whether sea_surface_height_above_geoid is obtained directly, as in a free-surface model, or has been derived, for example, from geostrophy using diagnosed velocities at some level or from geostrophy relative to an assumed level of quiescence, or some other technique. Various possible methods of estimating sea-level in rigid-lid models are described in the Appendix of Gregory et al. (2001).
- Non-Boussinesq models incorporate those ocean effects contributing to sea level, including *steric effects* associated with changes in the seawater density. In contrast, the Boussinesq approximation precludes steric effects, thus making its free surface an inaccurate approximation to sea level, especially for global warming, where steric

effects are expected to increase. In this endnote, we detail the kinematics that identifies these differences.[12] It is thus important that the "comments" attribute for this variable contain information regarding whether the field was obtained from a Boussinesq or non-Boussinesq ocean model simulation.

- In some coupled models, sea ice at a grid cell depresses the liquid seawater through its mass loading (appearing as an applied surface pressure on the ocean model). This depression occurs independent of the subgrid scale distribution of sea ice, as it is a result of the mass of sea ice in a grid cell acting on the liquid ocean. There is, however, no dynamical effect associated with these depressions in the liquid ocean sea level, so that there are no associated ocean currents. Hence, when reporting sea level, modelers may choose to report the effective sea level as if sea ice (and snow) at a grid cell were converted to liquid seawater, as this is the dynamically relevant sea level (Campin et al., 2008). It should be noted in the comments for this variable whether it represents the ocean model sea level or the effective sea level from the sea ice model.
- If the ocean model feels the effects from the applied atmospheric forcing (the *inverse barometer effect*), then include this fact in the "comments" section for the field sea_surface_height_above_geoid.
- There are a couple of acceptable options for reporting this field: 1) if the geoid is defined to relate to the instantaneous mass of the ocean, the global mean of sea_surface_height_above_geoid will always be zero, and 2) if the geoid is defined relative to a time-mean sea level over some period, then the global mean of sea_surface_height_above_geoid will be time-dependent. Analysis of global mean sea level changes will not rely on this field. Rather, it will rely on the global mean fields global_average_steric_sea_level_change and global_average_sea_level_change.

4.2.6 Squared sea level (2)

- square_of_sea_surface_height_above_geoid

The field square_of_sea_surface_height_above_geoid is requested to help measure the variability simulated in the sea level, by computing the variance

$$\langle(\eta - \langle\eta\rangle)^2\rangle = \langle\eta^2\rangle - \langle\eta\rangle^2, \tag{4.4}$$

with η a shorthand for sea_surface_height_above_geoid, and the angle brackets represent monthly time means.

4.2.7 Changes in sea level (0)

- global_average_sea_level_change
- global_average_steric_sea_level_change
- global_average_thermosteric_sea_level_change

The potential for increased sea level due to global warming presents some of the most pressing issues for adaptation to a warmer world. Sea level changes also provide a baseline assessment of the changing ocean climate in the simulations. It is thus of primary importance to consider the effects from sea level rise as simulated in the CMIP5 models. Results from model simulations should be carefully documented in order to properly interpret the CMIP5 archive. We offer the following technical points regarding the evolution of sea level.

- The field global_average_sea_level_change is the finite time increment $\Delta\overline{\eta}$ in global mean of the sea level

$$\overline{\eta} = \sum \eta \, dx \, dy, \tag{4.5}$$

  including changes in ocean density and boundary fluxes of mass. The reference for $\overline{\eta}$ should be taken with respect the model at the start of the control run, in which case the ocean surface height may not generally vanish. Importantly, the reference state must be the same for all scenario experiments within the CMIP5 ensemble.
- Non-Boussinesq models contain all ocean effects (including steric) within the ocean acting on the sea level. Hence, the field global_average_sea_level_change in a non-Boussinesq model is determined from its global averaged sea level increment $\Delta\overline{\eta}$. In contrast, for a Boussinesq model, the steric effect must be diagnosed and then added to the model's global mean sea level. Please note in the "comment" attribute any assumptions or methodological details related to calculation of this time-series. We offer the following endnote with kinematic details to better understand how global mean sea level evolves.[13]

- The field global_average_thermosteric_sea_level_change represents that part of the global mean sea level change due to changes in ocean density arising just from changes in temperature. In the following endnote, we offer definition of this field.[14] For many purposes the sea level rise under a warming planet is dominated by thermal effects, in which case the steric and thermosteric contributions are nearly the same. However, saline effects become nontrivial as increasing fresh water melt enters the ocean, thus making it important to distinguish all three contributions.

4.2.8 Mass per area or thickness of grid cell (1)

- sea_water_mass_per_unit_area
- cell_thickness

To compute tracer budgets in non-Boussinesq models, we require the mass of seawater in the cell, per horizontal area of the cell

$$\text{sea\_water\_mass\_per\_unit\_area} = \rho\, dz, \tag{4.6}$$

with units of kg/m^2. For a Boussinesq model, the *in situ* density factor ρ is set to the constant reference density ρ_o, in which case the mass per area is equivalent to the scaled cell thickness. For non-Boussinesq models, we request a separate archiving of the cell thickness

$$\text{cell\_thickness} = dz \tag{4.7}$$

in order to measure the distance (in metres) between surfaces of constant vertical coordinate. This information is useful, in particular, for measuring changes in thickness between pressure surfaces in a pressure model exposed to increasing anthropogenic radiation. We note that temporal dependence of the mass per area and the cell thickness is a function of the vertical coordinate used in the ocean model, and how the sea surface is treated.

4.2.9 Potential temperature in liquid seawater (0)

- sea_water_potential_temperature

WGCM (2007) recommends the three dimensional monthly mean potential temperature be archived, where the reference pressure is at the ocean surface. This field is the most common prognostic tracer in ocean models used to measure heat,[15] and so should be included in the archive.

In addition to the three-dimensional field of potential temperature, it is important to archive the global mean potential temperature. This field has the same standard name as the three-dimensional potential temperature, but is distinguished by the cell methods attribute (area and depth mean). The calculation of global mean potential temperature differs depending on the use of Boussinesq or non-Boussinesq ocean equations. In a non-Boussinesq model, the mean is given by the mass weighted mean

$$\overline{\theta}_{\text{non-Bouss}} = \frac{\sum_{i,j,k} \rho\, \theta\, dx\, dy\, dz}{\sum_{i,j,k} \rho\, dx\, dy\, dz}, \tag{4.8}$$

where θ is the model's potential temperature field. In a Boussinesq model, the mean is computed as the volume weighted mean

$$\overline{\theta}_{\text{Bouss}} = \frac{\sum_{i,j,k} \theta\, dx\, dy\, dz}{\sum_{i,j,k} dx\, dy\, dz}. \tag{4.9}$$

The distinction between non-Boussinesq and Boussinesq models arises from the differences in the underlying conserved fields in the two model formulations. For both cases, it is necessary to accumulate each model time step when producing the time mean, since the mean is built from the product of time dependent terms (e.g., density and grid cell thicknesses are generally time dependent).

The global mean temperature presents the analyst with a very convenient measure of drift in the model, and a measure of the model's deviation from estimates of the observed global mean temperature. Furthermore, when combined with the boundary fluxes and total mass/volume, one can diagnose the degree to which the ocean model conserves heat.[16]

4.2.10 Monthly mean SST of liquid water (1)

- sea_surface_temperature

In the CMIP archive, it is quite valuable to have the full three-dimensional fields, such as potential temperature and salinity. However, for many purposes, just the surface fields are sufficient. The sea surface temperature (SST) and sea surface salinity (SSS) are two such fields we recommend, as well as the daily mixing layer depth. Given that the surface fields are far smaller in size than the full three-dimensional fields, archiving a selection of surface ocean fields will facilitate a much simpler analysis of the results.[17] For this purpose, we request the monthly mean SST.

4.2.11 Daily mean SST of liquid water (2)

- sea_surface_temperature

We recommend that daily mean SST be saved for the purpose of computing space-time diagrams to diagnose propagating signals, such as Tropical Instability Waves. The daily mean SST is also of use for understanding the potential for enhanced coral bleaching in a warming world. Coral bleaching is one of the major potential environmental consequences resulting from global warming. Remotely sensed estimates of coral bleaching have converged on a measure based on degree-heating weeks (Strong et al., 2004). Quantifying this meausure in models requires an archive of daily mean sea surface temperatures.

4.2.12 Daily and monthly mean squared SST of liquid water (2)

The field

- square_of_sea_surface_temperature

is requested to help measure the variability simulated in the sea surface temperature, so that one may compute the variance

$$\langle(\theta - \langle\theta\rangle)^2\rangle = \langle\theta^2\rangle - \langle\theta\rangle^2, \tag{4.10}$$

with θ a shorthand for the field sea_surface_temperature, and the angle brackets represent either daily or monthly time means.

4.2.13 Salinity of liquid water (0)

- sea_water_salinity

We recommend archiving the three dimensional monthly mean ocean salinity field.[18] In addition, as for potential temperature, we recommend saving the global mean salinity of liquid seawater. This mean is computed in a non-Boussinesq model by the mass weighted mean

$$\overline{S}_{\text{non-Bouss}} = \frac{\sum_{i,j,k} \rho\, S\, \mathrm{d}x\, \mathrm{d}y\, \mathrm{d}z}{\sum_{i,j,k} \rho\, \mathrm{d}x\, \mathrm{d}y\, \mathrm{d}z}, \tag{4.11}$$

whereas for a Boussinesq model it is the volume weighted mean

$$\overline{S}_{\text{Bouss}} = \frac{\sum_{i,j,k}\ S\, \mathrm{d}x\, \mathrm{d}y\, \mathrm{d}z}{\sum_{i,j,k}\ \mathrm{d}x\, \mathrm{d}y\, \mathrm{d}z}. \tag{4.12}$$

In either case, the global mean salinity presents the analyst with a very useful measure of the drift in the model, a measure of the model's deviation from the estimates of the observed global mean salinity, and a means for checking for conservation of total salt. As for the global mean temperature, it is generally necessary to compute each of the terms in the average on each time step, since the average is generally built from the product of time dependent terms.

4.2.14 Sea surface salinity (1)

- sea_surface_salinity

The sea surface salinity (SSS) provides a useful means for detecting changes in the high latitude thermohaline forcing, which can present the analyst with a quick diagnosis of whether a simulation is more or less prone to modification of the overturning circulation. For example, fresh water capping can be seen by diagnosis of the SSS. In this case, signals in SSS may motivate more detailed analysis of the three-dimensional fields. Absent the SSS field in the archive, the analyst is burdened with unpacking the full three-dimensional fields to make even the most rudimentary analysis. Given the far more convenient nature of the smaller two-dimensional SSS relative to the full three-dimensional salinity, we recommend that monthly mean SSS field be archived for CMIP5.

4.2.15 Potential density (2)

- sea_water_potential_density

Potential density referenced to the ocean surface (σ_0) is a useful auxiliary field for garnering a measure of the stability of the upper ocean. WGCM (2007) recommended that the monthly mean potential density be archived for CMIP3, and we concur. Note that for some purposes, it may be sufficient to compute offline the potential density given the time mean temperature, salinity, and pressure. However, for careful analysis of water column stability, especially in high latitudes, it is more accurate to save the model's potential density online during the simulation.

4.2.16 Ideal age tracer (2)

- sea_water_age_since_surface_contact

A number of groups participating in CMIP3 included ideal age tracer (Bryan et al., 2006; Gnanadesikan et al., 2007). This tracer (Thiele and Sarmiento, 1990; England, 1995) is set to zero in the model surface level/layer at each time step, and ages at 1 yr/yr below. Ideal age is particularly useful for revealing surface-to-deep connections in regions such as the Southern Ocean where these connections have spatiotemporal variability. It can also be used to estimate uptake of anthropogenic tracers such as carbon dioxide (Russell et al., 2006).

There are differences in the methods used to implement the ideal age in ocean models. These differences relate to (A) the surface boundary condition, (B) the initial condition. For the surface boundary condition, we recommend that the ideal age be set to zero when the field enters the top model grid cell, or upper ocean boundary layer. An alternative is to provide a surface restoring of age towards zero, with the limit of infinite restoring strength leading to the recommended approach. So long as the restoring strength is sufficiently strong, there should be trivial distinctions between the two approaches.

To facilitate direct comparison of ideal age in the different model simulations, we recommend initializing age globally to zero at 01Jan1901 in the various scenario experiments. Measuring age in years, rather than seconds, is the traditional approach in ocean modeling, and is the recommended units for ideal age in CMIP5.

4.2.17 CFC-11 (2)

- moles_per_unit_mass_of_cfc11_in_sea_water

Chlorofluorocarbons (CFCs) have been increasingly utilized in evaluating Ocean General Circulation Models, largely due to a good observational data base (the World Ocean Circulation Experiment, WOCE, upon which Global Ocean Data Analysis Project, GLODAP, Key et al. (2004) is largely based) and their well-known atmospheric concentrations. These passive tracers are particularly useful in assessing model mixing processes, ventilation rates, deep water formation, and circulation characteristics. The surface CFC fluxes are calculated for the 1931-2000 period following the Ocean Carbon Model Intercomparison Project (OCMIP-2) protocols (Dutay et al., 2002). However, instead of the protocol specified atmospheric fields, CMIP5 models should use their own atmospheric data sets in these flux equations. Because CFC-11 and CFC-12 exhibit similar behavior in model simulations, we request monthly-mean distributions only for CFC-11 and *only* from the historical simulations.

The units of CFC-11 should be reported as the moles of CFC-11 per kilogram of seawater.

4.2.18 Barotropic or quasi-barotropic streamfunction (2)

- ocean_barotropic_mass_streamfunction

The barotropic streamfunction is a useful field for mapping the vertically integrated fluid transport. However, many ocean models have jettisoned the rigid lid assumption of Bryan (1969) for both computational and physical reasons. Absent a rigid lid assumption, the vertically integrated mass transport[19] $\mathbf{U}^{\rho} = \int_{-H}^{\eta} \rho\,\mathbf{u}\,dz$ generally has a non-zero divergence,[20] thus precluding it from being fully specified by a single scalar field. Instead, both a streamfunction and velocity potential are needed to specify the transport. For those models that do not compute a barotropic streamfunction, we introduce the notion of a *quasi-barotropic streamfunction* ψ^{U} in this endnote,[21] with this field serving as a useful approximate alternative to the barotropic streamfunction.

In summary, we recommend either of the following scalar fields be archived for purposes of mapping the vertically integrated mass transport:

- Barotropic streamfunction for those models that compute this function using an elliptic solver.
- The quasi-barotropic streamfunction ψ^{U} for cases when the model does not distinguish the streamfunction from the velocity potential.

Consistent with our discussion in Section 3.2, we recommend that the dimensions of the streamfunction be mass transport (kg/s), rather than volume transport (m^3/s).

4.2.19 Mixed layer depth (2)

- ocean_mixed_layer_thickness_defined_by_sigma_t
- square_of_ocean_mixed_layer_thickness_defined_by_sigma_t

An assessment of model mixed layer depth (MLD) is useful for understanding how water-mass formation in the simulations is regulated by upper ocean stratification and surface water overturn. In addition, with the squared MLD one may deduce a measure for the variability of the simulated MLD.

Unfortunately, there is no universally agreed upon criterion for defining the mixed layer depth. For the purpose of fostering a consistent comparison of simulated mixed layers from ocean model components in CMIP5, we recommend that the "sigma-t" criterion introduced by Levitus (1982) be followed. To furthermore support the direct comparison of simulated MLDs from those presented in observational analyses, we recommend that the simulated MLDs be diagnosed from the monthly mean temperature and salinity, rather than accumulated over each model time step.[22]

4.2.20 Mixing layer depth (2)

- ocean_mixed_layer_thickness_defined_by_mixing_scheme

One of the key issues for ocean biological cycling is the length of the growing season. In high latitudes this length is controlled by the shallowing mixed layers, allowing the surface ocean to receive sufficient light. Since plankton respond to light on time scales of days, capturing the growing season requires relatively high temporal resolution, as well as an ability to distinguish the layer that is actively mixing from the layer that has low stratification. To support the study of these issues in the ocean model components of CMIP5, we recommend archiving the daily maximum values of *mixing layer depth*, with monthly maximums also submitted to allow for easier study of longer term trends.

The mixing layer depth is distinct from the mixed layer depth discussed above. The mixing layer depth is determined by the model's planetary boundary layer parameterization, and is the depth over which upper ocean mixing is active. In isopycnal models with a bulk mixed layer, the mixing layer depth is the depth of the bulk mixed layer. In models which use a boundary layer scheme such as Large et al. (1994), the mixing layer depth is the planetary boundary layer depth used to calculate the bulk Richardson number. In models with a turbulence closure scheme of the Mellor and Yamada (1982) class, the mixing layer depth is defined as the depth over which diffusivities are significantly higher than the interior values.

4.3 Vectors or components to vectors

We now consider vector fields, or components to vector fields, suggested for the ocean model components to CMIP5. Refer to Table 2.3 for a summary of the fields.

To reemphasize the discussion in Section 3.1.2, we recommend *against* remapping vector fields (e.g., horizontal velocity, mass and tracer transports, fluxes) from the model's native grid to a spherical grid. However, we do recognize that in order to easily and clearly compare models, it is useful to calculate and to make available the poleward transports of mass and tracers, integrated along latitude circles in each ocean basin. This mandate was taken by many modelling groups in CMIP3 to be synonymous with requiring that non-spherical grid models remap transports in the computation of poleward transports. The remapping step is problematic since it cannot ensure flux conservation. This remapping was in fact not explicitly requested in CMIP3. An alternative approach that is conservative is recommended.

We recommend that each group using non-spherical grids develop a native-grid algorithm that computes the closest native grid approximation to the basin integrated poleward transports. That is, transports across a section (e.g. meridional overturning at a given latitude, transport through a passage, or vertically integrated poleward heat transport) should be computed consistent with the native grid by finding a nearly equivalent path to the section that has been "snapped" to the native grid (often resulting in a "zig-zag" path). This approach retains the native grid variables, and so allows for conservation of transports. It also avoids ambiguities associated with defining a remapped land/sea mask. The resulting transports should be made available as a function of latitude (even though the integrations are not exactly along latitude circles). The latitude spacing should be comparable to that of the model's grid spacing.

4.3.1 Horizontal velocity field (0)

- sea_water_x_velocity
- sea_water_y_velocity

WGCM (2007) recommend archiving the monthly mean horizontal velocity field, and we concur with this recommendation. This field should be archived on the native model grid, including staggering, and should not have any rotation or remapping applied. If needed, a remapping can be performed off-line to compute the spherical components of the velocity.

4.3.2 Vertical mass transport (0)

- upward_ocean_mass_transport

In a hydrostatic ocean model, Adcroft and Hallberg (2006) identify two general means to compute the mass flux of seawater that crosses surfaces of vertical coordinate (s =constant)

$$\mathcal{V}^{(s)} = \rho\, w^{(s)}\, \mathrm{d}x\, \mathrm{d}y, \tag{4.13}$$

where $w^{(s)} = (\partial z/\partial s)\, \mathrm{d}s/\mathrm{d}t$ is the dia-surface flux across vertical coordinate surfaces, $\mathrm{d}s/\mathrm{d}t$ the material time derivative of the vertical coordinate, and $(\partial z/\partial s)$ measures the inverse stratification of the s-surfaces. The *in situ* density ρ is replaced by the constant reference density ρ_o in the case of a Boussinesq fluid. Quasi-Eulerian models (e.g., conventional geopotential, pressure, and terrain following models) diagnose $\mathcal{V}^{(s)}$ through mass or volume conservation, whereas quasi-Lagrangian models (conventional isopycnal models) specify $\mathcal{V}^{(s)}$ based on physics of cross coordinate surface flow (e.g., diapycnal transport in isopycnal models). For notational simplicity, the term "cross-coordinate mass transport" is abbreviated as "vertical mass transport" in this document. To touch bases with the familiar Boussinesq geopotential case, note that division of the vertical mass transport $\mathcal{V}^{(s)}$ by the reference density ρ_o and the horizontal area $\mathrm{d}x\,\mathrm{d}y$ of a grid cell (available in the model's grid specification file as discussed in Section 4.1) yields the vertical velocity component.

The vertical seawater mass transport $\mathcal{V}^{(s)}$ (measured in kg/s) is more valuable for analysis than the vertical velocity component. Indeed, one of the most common steps in forming water mass analyses given the vertical velocity component, density, and grid areas, is to approximate the seawater mass transport. Conversely, *if* there is reason to determine the vertical velocity component, it can be trivially diagnosed from the vertical mass transport in a Boussinesq model, and approximated in a non-Boussinesq model. Hence, rather than archiving the vertical velocity component, we recommend archiving the vertical mass transport of seawater.[23]

4.3.3 Horizontal mass transport (1)

- ocean_mass_x_transport
- ocean_mass_y_transport

For the same reasons as the vertical mass transport is useful, the horizontal mass transport of seawater is also of use for studying water mass processes. We therefore recommend that the horizontal components to the mass transport

$$\mathcal{V}^{(x)} = \rho\, u\, \mathrm{d}y\, \mathrm{d}z \tag{4.14}$$

$$\mathcal{V}^{(y)} = \rho\, v\, \mathrm{d}x\, \mathrm{d}z \tag{4.15}$$

be archived at each model grid cell. These are three-dimensional fields on the model's native grid, staggered as used in the ocean model's prognostic equations, and result from the product of time dependent fields. Note that for a Boussinesq model, the *in situ* density factor ρ is set to the constant reference density ρ_o.

4.3.4 Squared vertical mass transport (2)

- square_of_upward_ocean_mass_transport

This field is requested to help measure the variability simulated in the vertical transport, so that the variance of this transport may be computed via

$$\langle (w - \langle w \rangle)^2 \rangle = \langle w^2 \rangle - \langle w \rangle^2, \tag{4.16}$$

with w a shorthand for sea_water_mass_transport_vertical, and the angle brackets representing monthly time means. If only a single depth is saved, then 2000m, or thereabouts, is recommended.

By using a coupled atmosphere-ocean-sea-ice model with an ocean of 1/4° resolution, Komori et al. (2008) found high-frequency internal gravity waves propagating equatorward from midlatitude stormtrack regions. A particularly effective means of diagnosing these waves is to compute the variance of the vertical transport. The waves appear to be ubiquitous in the Pacific, Atlantic and Indian Oceans, and they have been found only in coupled simulations with high frequency forcing. Further research into the importance of these waves is an active research area, thus motivating that this field be saved especially in the eddying model simulations submitted for CMIP5. To reduce output load, only results from the historical experiment are recommended.

4.3.5 Poleward and $\hat{\mathbf{y}}$-ward overturning streamfunction from all transport processes (1)

- ocean_meridional_overturning_mass_streamfunction
- ocean_y_overturning_mass_streamfunction

The transport of fluid northward in each of the basins Atlantic-Arctic, Indian-Pacific, and World Ocean, as a function of depth/pressure and density, is of interest for many purposes of ocean climate dynamics.[24] This transport is of particular interest for the study of tracers, such as heat and salt. Consequently, we are interested in transport arising from the model's resolved velocity field, as well as transport arising from all SGS processes such as Gent and McWilliams (1990); Gent et al. (1995) and the submesoscale mixed layer transport scheme from Fox-Kemper et al. (2008), amongst others.

The issue of generalized horizontal coordinates adds complexity to the diagnosis of the northward mass transport when using non-spherical grids. As stated at the start of Section 4.3, instead of remapping mass fluxes to a spherical grid, and then computing the basin transports, we recommend computing the transports across native grid lines that approximate latitude circles, and reporting these as a function of latitude. Such algorithms can be implemented in a conservative manner for finite volume based models, even those with complex grids. Finite element models, in contrast, require extra care. Indeed, the diagnosis of transport for these models just now being considered in the literature (Sidorenko et al., 2008).

For those models using a non-spherical coordinate horizontal grid, in addition to archiving the meridional overturning streamfunction, we recommend archiving the model's native grid $\hat{\mathbf{y}}$-ward overturning streamfunction, where $(\hat{\mathbf{x}}, \hat{\mathbf{y}})$ are directions defined according to the model's native grid. We also use the synonyms (*iward*, *jward*), using the familiar (i, j) notation for horizontal grid indices. For many purposes and for many of the most commonly used non-spherical grids, the $\hat{\mathbf{y}}$-ward native grid streamfunction is sufficient. The following reasons can be given for those cases where the $\hat{\mathbf{y}}$-ward native grid streamfunction is sufficient:

- The two commonly used generalized horizontal coordinates for ocean models in CMIP5 include the tripolar grid (see, for example, Griffies et al., 2005), where all latitudes south of roughly $65°N$ remain spherical, or the displaced pole grid (see, for example, Smith and Gent, 2004), where the coordinate North Pole is moved over a land region in the northern hemisphere. In either case, northward transport, at least for regions south of the

Arctic Circle, is readily approximated as the $\hat{\mathbf{y}}$-ward transport, as defined along the model's native grid lines. These streamfunctions are sensibly compared between models with varying grid choices, again since regions where the grid lines are most highly distorted from the sphere are precisely those regions where the flow is very weak and thus of less interest for scientific purposes.

- Transport in regions north of the Arctic Circle is very weak relative to transport in the south, and poorly sampled from observations. Hence, its diagnosis is of secondary concern for comparison to observations.
- The overturning streamfunction is not a directly observed field. Instead, it is partialy inferred through selected transport measurements at very few ocean sections. To constrain the recording of the simulated streamfunction to be oriented precisely along a line parallel to geographical longitude is not warranted for reasons of comparing to observations.

A general expression for the ocean mass transport overturning streamfunction is given by

$$\Psi(y, s, t) = -\int_{x_a}^{x_b} \mathrm{d}x \int_{-H}^{z(s)} \rho\, v^{\text{total}}\, \mathrm{d}z, \tag{4.17}$$

where v^{total} is the native grid approximation to the poleward transport arising from the resolved velocity component, plus all SGS processes contributing to mass transport. Note that the zonal integral is computed along surfaces of constant s, where s is either a geopotential/pressure surface, or a potential density surface. That is, we recommend that the following versions of the overturning streamfunction be archived at monthly time averages in the CMIP5 repository, with results for the Atlantic-Arctic, Indian-Pacific, and Global Oceans:

- **poleward-depth overturning streamfunction** and **$\hat{\mathbf{y}}$-ward-depth overturning streamfunction**: The depth $z(s)$ corresponds to either the depth of a geopotential or the depth of a pressure surface, depending on whether the model is Boussinesq or non-Boussinesq, respectively.
- **poleward-density overturning streamfunction** and **$\hat{\mathbf{y}}$-ward-density overturning streamfunction**: The depth $z(s)$ corresponds to the depth of a predefined set of σ_{2000} isopycnals, with the definition of these isopycnals at the modeler's discretion. This field presents complementary information relative to the $\hat{\mathbf{y}}$-ward-depth overturning streamfunction, and is very useful particularly for diagnosing water mass transformation processes.[25]
- Consistent with the discussion in Section 3.1.1, it is critical that the time average of the streamfunction be accumulated using each model time step, in order to avoid problems with aliasing and problems ignoring correlations.

4.3.6 Poleward and $\hat{\mathbf{y}}$-ward overturning streamfunction from SGS transport (2)

- ocean_meridional_overturning_mass_streamfunction_due_to_bolus_advection
- ocean_y_overturning_mass_streamfunction_due_to_bolus_advection

We follow the same philosophy as above for diagnosing the poleward and $\hat{\mathbf{y}}$-ward overturning streamfunction arising just from SGS transport. There are two notable SGS methods whose overturning we recommend archiving: Gent et al. (1995) and the newer mixed layer transport scheme from Fox-Kemper et al. (2008). Other schemes may also be included, so long as they impact on the overturning mass streamfunction. That is, both the Gent et al. (1995) and Fox-Kemper et al. (2008) introduce a new streamfunction, with the name "bolus" generically ascribed to this streamfunction based on historical reasons. Note that since the Fox-Kemper et al. (2008) scheme applies only in the mixed layer, only its poleward-depth and $\hat{\mathbf{y}}$-depth version is relevant. For the Gent et al. (1995) streamfunction, it is useful to map this in both depth and density space.[26]

4.3.7 Vertically integrated heat transport from all processes (1)

- northward_ocean_heat_transport
- ocean_heat_x_transport
- ocean_heat_y_transport

There are many processes in the ocean that affect the heat transport: resolved advective transport, diffusion, skew diffusion, overflow parameterizations, etc. In the analysis of ocean model simulations, it is very useful to have a measure of each component of the heat transport. We suggest the vertically integrated $\hat{\mathbf{x}}$-ward and $\hat{\mathbf{y}}$-ward heat transport from all processes, archived as monthly means for the Atlantic-Arctic, Indian-Pacific, and World Ocean, and maintained on the model's native grid. In addition, following from the approach taken for the meridional overturning streamfunction, each ocean model using non-spherical coordinate horizontal grids should compute the poleward heat transport in each of the basins, approximated using the model's native grid fields without remapping. The approximated poleward transport will generally consist of transports crossing a "zig-zag" path. The resulting poleward heat transport should be reported as a function of latitude, with latitudinal resolution comparable to the model's native grid resolution. This recommendation follows that from WGCM (2007), with the method to calculate the transports now explicitly specified.

4.3.8 Vertically integrated heat transport from SGS processes (2)

- northward_ocean_heat_transport_due_to_bolus_advection
- northward_ocean_heat_transport_due_to_diffusion
- ocean_heat_x_transport_due_to_bolus_advection
- ocean_heat_x_transport_due_to_diffusion
- ocean_heat_y_transport_due_to_bolus_advection
- ocean_heat_y_transport_due_to_diffusion

In support of understanding the importance of various SGS physical parameterizations, we recommend that separate heat transports should be archived as follows.

- Parameterized SGS advection, as from Gent and McWilliams (1990); Gent et al. (1995) and the scheme from Fox-Kemper et al. (2008), should be included in the fields "due_to_bolus_advection", even if the implementation of the schemes appears as a skew diffusion. In addition, include the effects from neutral/isopycnal diffusion (see comments below).
- Any remaining SGS processes, such as overflow parameterizations, etc., should be included in the "due to diffusion" fields.

These standard names are imprecise. They are largely based on historical reasons that are somewhat obsolete. The main reason to combine the SGS advection processes, such as Gent et al. (1995), with neutral diffusion is that many model implementations follow the approach of Griffies (1998). Here, neutral diffusion and Gent et al. (1995) are combined into a single operator, in which case the "bolus advection" from Gent et al. (1995) is transformed to a skew diffusion. The combined neutral diffusion and skew diffusion is thus better considered a single "neutral physics" operator, with it impractical algorithmically to then split this operator into two pieces for purposes of diagnostics. We therefore request models to include all of these "neutral physics" processes within the fields "due_to_bolus", even if the models do not implement Gent et al. (1995) according to a skew diffusion approach. Information regarding these details should be clearly stated in the meta-information for the fields.

We recommend archiving these heat transports both on the model's native grid, as well as the poleward transport approximated using the a "zig-zag" path method discussed previously. The components should be archived as monthly means for the Atlantic-Arctic, Indian-Pacific, and World Ocean. The transports should be reported as a function of latitude, with the latitudinal spacing comparable to the model's native grid spacing.

4.3.9 Vertically integrated heat & salt transport from gyre and overturning components (1)

- northward_ocean_heat_transport_due_to_gyre
- northward_ocean_heat_transport_due_to_overturning
- northward_ocean_salt_transport_due_to_gyre
- northward_ocean_salt_transport_due_to_overturning

The $\hat{\mathbf{y}}$-ward advective transport of a tracer within a particular ocean basin is given by the integral

$$\mathcal{H}^{\hat{y}}(y,t) = \int_{x_1}^{x_2} \mathrm{d}x \int_{-H}^{\eta} \mathrm{d}z\, \rho\, C\, v, \tag{4.18}$$

where C is the tracer concentration, $z = -H(x, y)$ is the ocean bottom, $z = \eta(x, y, t)$ is the ocean surface, and x_1 and x_2 are the boundaries of the basin or global ocean. It is useful for some analysis to decompose the transport (4.18) into "gyre" and "overturning" components, with these terms defined in the following endnote.[27] We recommend that the monthly means for the components to heat and salt transport, partitioned according to Atlantic-Arctic, Indian-Pacific, and Global Oceans, be archived in the following manner:

- For all processes affecting the tracer, including resolved and SGS processes.
- For just the resolved scale flow.

We recommend archiving these transports both on the model's just as the poleward transport approximated using the a "zig-zag" path method discussed previously. The components should be archived as monthly means for the Atlantic-Arctic, Indian-Pacific, and World Ocean. The transports should be reported as a function of latitude, with the latitudinal spacing comparable to the model's native grid spacing.

4.4 Integrated mass transports

- sea_water_transport_across_line

1. barents_opening
2. bering_strait
3. canadian_archipelago
4. denmark_strait
5. drake_passage
6. english_channel
7. equatorial_undercurrent
8. faroe_scotland_channel
9. florida_bahamas
10. fram_strait
11. iceland_faroe_channel
12. indonesian_throughflow
13. mozambique_channel
14. taiwan_and_luzon_straits
15. windward_passage

There are a number of climatologically important straits, throughflows, and current systems whose integrated mass transport is measured observationally (though some have wide uncertainties). These mass transports provide a useful means to characterize the simulation. Transports were not archived in CMIP3, thus necessitating a diagnostic calculation from the archived velocity field and/or the barotropic streamfunction. Such diagnoses, however, can be subject to uncertainty, especially for models with complex horizontal and vertical grids. It is thus more direct and accurate to have these transports computed by each participating model group, and archived as part of CMIP5. Table 2.4 provides a list of recommended transports for CMIP5. Each geographical region has an associated string valued coordinate given by the name.

We make the following recommendations regarding the recording of integrated mass transports.

- In the following, we note the approximate geographical longitude and latitude coordinates of the straits and currents. Given considerations of model grid resolution and grid orientation, precise values for the coordinates may differ for any particular model. *In general, we recommend computing the simulated transport where the strait is narrowest and shallowest in the model configuration, and where the model grid is closely aligned with the section.*
- For most ocean model grids, the requested transports can be diagnosed by aligning the section along a model grid axis. In this case, it is straightforward to assign a positive sign to transports going in a pseudo-north or pseudo-east direction, and negative signs for the opposite direction. We use the term *pseudo* here as it refers to an orientation according to the model grid lines, which in general may not agree with geographical longitude and latitude lines. In general, the sign convention chosen for the recorded transport should be clearly indicated in the metadata information for the transport field.

- Some models may have a strait artificially closed, due to inadequate grid resolution. In this case, a zero transport should be recorded for this strait.
- We present references to observational estimates for the mass transports. These references will be kept updated on the WGOMD web page `www.clivar.org/organization/wgomd/reos/reos.php` as new results become available. Notably, there are some straits with large uncertainties. Even so, recording transport results from the various models will present the community with a valuable means to characterize the model flow fields.
- Some of the following transports are defined in accordance with the Global Ocean Data Assimilation Experiment (GODAE), as detailed in the report by the MERSEA project (MERCATOR, 2006).

The following provides details for the various regions where integrated mass transport is requested for CMIP5.

1. Barents Opening: The Barents Opening separates Spitsbergen from Norway. Vertically integrated transport through the Barents Opening occurs geographically roughly between the points

   - Barents opening = $(16.8°E, 76.5°N)$ to $(19.2°E, 70.2°N)$.

   Observational estimates range from 1.5-2.0 Sv northwards, with large variability, thus necessitating longer time series to get a zero order estimate.

2. Bering Strait: The Bering Strait separates Alaska from Siberia. Vertically integrated transport through the Bering Strait provides the only exchange between Pacific and Arctic waters. It is defined geographically from

   - Bering Strait = $(171°W, 66.2°N)$ to $(166°W, 65°N)$.

   An observational estimate from Roach et al. (1995) is 0.8Sv northward from the Pacific into the Arctic Ocean.

3. Canadian Archipelago: The Canadian Archipelago refers to the wide range of Arctic islands in northern Canada. The transport through these islands connects waters of the open Arctic to the North Atlantic through the Davis Strait and into the Labrador Sea. Vertically integrated transport through the Canadian Archipelego can be defined according to the following geographic region

   - Canadian Archipelego = $(128.2°W, 70.6°N)$ to $(59.3°W, 82.1°N)$.

   Observational estimates range from 0.7 to 2.0Sv southward (Sadler, 1976; Fissel et al., 1998; Melling, 2000).

4. Denmark Strait: The Denmark Strait separates Greenland from Iceland. Vertically integrated transport between Iceland and Greenland occurs over the following geographical region

   - Denmark Strait = $(37°W, 66.1°N)$ to $(22.5°W, 66°N)$.

   Observational estimates are 0.8Sv for the net northward transport (Osterhus et al., 2005) and 3Sv for the net southward transport (Olsen et al., 2008).

5. Drake Passage: The Drake Passage separates South America from Antarctica. It presents the narrowest constriction for the Antarctic Circumpolar Current. Vertically integrated transport in the Southern Ocean through the Drake Passage is determined by flow through the region

   - Drake Passage = $(68°W, 54°S)$ to $(60°W, 64.7°S)$.

   An observational estimate from Cunningham et al. (2003) is an eastward transport of 135Sv.

6. English Channel: The English Channel separates Britian from the European continent. Vertically integrated transport in the English Channel occurs geographically through the region

   - English Channel = $(1.5°E, 51.1°N$ to $(1.7°E, 51.0°N)$.

   Observational estimates from Otto et al. (1990) are roughly $0.1 - 0.2$Sv northward.

7. Equatorial Undercurrent: A commonly used region to measure transport in the equatorial undercurrent is given by the region

   - Equatorial Undercurrent = $(155°W, 3°S)$ to $(155°W, 3°N)$ over the depth range 0-350m.

   Observational estimates range between 24Sv-36Sv in an eastward direction (Lukas and Firing, 1984; Sloyan et al., 2003).

8. Faroe-Scotland Channel: The Faroe-Scotland Channel separates the Faroe Islands from Scotland. Vertically integrated transport between the Faroe Islands and Scotland occurs geographically through the region between

- Faroe-Scotland Channel = ($6.9°W, 62°N$) to ($5°W, 58.7°N$)

Observational estimates are 3.8Sv for the net northward transport (Osterhus et al., 2005) and 2.1Sv for the net southward transport (Olsen et al., 2008).

9. Florida-Bahamas Strait: Since 1982 cables have been used to measure the transport of the Florida Current between Florida and the Bahamas near 27N. We thus define this transport according to the following geographical locations

   - Florida-Bahamas Strait = ($78.5°W, 26°N$) to ($80.5°W, 27°N$).

   Observational estimates range from 29Sv-35Sv (Leaman et al., 1987). Updated information is available from AOML at `www.aoml.noaa.gov/phod/floridacurrent/`. See also Figure 2-6 from the MERSEA project (MERCATOR, 2006).

10. Fram Strait: The Fram Strait separates Spitsbergen from Greenland. Vertically integrated transport in the Fram Strait occurs geographically through the region

    - Fram Strait = ($11.5°W, 81.3°N$ to ($10.5°E, 79.6°N$).

    Observational estimates from Schauer et al. (2004) are 4 ± 2Sv southwards.

11. Iceland Faroe Channel: The Iceland Faroe Channel separates Iceland from the Faroe Islands. Vertically integrated transport between Iceland and the Faroe Islands occurs geographically through the region between

    - Iceland-Faroe Channel = ($13.6°W, 64.9°N$) to ($7.4°W, 62.2°N$)

    Observational estimates are 3.8Sv for the net northward transport (Osterhus et al., 2005) and 1Sv for the net southward transport (Olsen et al., 2008).

12. Indonesian Throughflow: Vertically integrated transport through the Indonesian Archipelogo is defined approximately by

    - Indonesian Throughflow = ($100°E, 6°S$) to ($140°E, 6°S$).

    An observational estimate from Gordon et al. (2003) is roughly 10Sv from the Pacific to the Indian Oceans.

13. Mozambique Channel: The Mozambique Channel separates Madagascar from the African continent. Vertically integrated transport through the Mozambique channel separating Madagascar from Southeast Africa is defined approximately by

    - Mozambique Channel = ($39°E, 16°S$) to ($45°E, 18°S$).

14. Taiwan-Luzon Straits: We ask here for the vertically integrated transport giving the combined inflow to the South China Sea through the Taiwan and Luzon straits. The value from observations is positive when entering the South China Sea, and Yaremchuk et al. (2008) present a review of observed values.

15. Windward Passage: The Windward Passage lies between the easternmost region of Cuba and the northwest of Haiti, and is defined approximately by

    - Windward Passage = ($75°W, 20.2°N$) to ($72.6°W, 19.7°N$).

4.5 Boundary fluxes

The ocean is a forced-dissipative system, with forcing largely at its boundaries. To develop a mechanistic understanding of ocean simulations, it is critical to have a clear sampling of the many forcing fields. Some of the following fields can be found in other parts of the CMIP5 archive as part of the sea ice or atmosphere components. However, these fields are typically on grids distinct from the ocean model grid. Fluxes on grids distinct from the ocean make accurate budget analyses difficult to perform, and such was a major shortcoming of the CMIP3 archive (WGCM, 2007). For CMIP5, we wish to resolve this problem by requesting an archive of the precise boundary fluxes used to force the ocean model.

We offer the following general comments regarding the boundary flux fields. Details of the requested fluxes are given in the subsequent subsections.

- All fluxes (water mass, salt mass, heat, momentum) are normalized according to the horizontal area of the ocean model grid cell. In some cases (e.g., rainfall), the flux computation requires integrating the rainfall over the ice-free sea (to get a mass per time of rainfall) and then dividing by the ocean grid cell area (to get mass per time per area). For these fluxes, according to the CF metadata conventions, the cell_methods attribute for the fields should read

  – area: mean where ice_free_sea over all_area_types.

  In other cases (e.g., melting sea ice) the flux computation requires integrating the sea ice melt over the sea ice covered portion of the ocean grid cell, and then dividing by the ocean grid cell area. For these fluxes, according to the CF metadata conventions, the cell_methods attribute for the fields should read

  – area: mean where sea_ice over all_area_types.

- Multiplication of a boundary flux by the ocean model grid cell area allows for the mass, heat, or momentum passed to the ocean, per time, to be computed.
- Many climate models place boundary fluxes just at the ocean surface. However, more general couplings are being considered (e.g., penetrative shortwave heating; sea ice models that interact with more than the surface ocean cell). To allow for such generality, we ask that those fluxes that are three-dimensional be archived with their full three dimensional structure.
- To remphasize a point raised throughout this document, the wind stresses and ice stresses should be reported on the ocean model native grid, with no remapping or rotation.
- The term "flux correction" in Tables 2.5, 2.6, 2.7, and 2.8 refers to the imposition of a prescribed flux that has at most a monthly variability (sometimes only an annual mean adjustment is imposed). Flux corrections (also called flux adjustments) have no interannual variability. They are added to some climate models for the purpose of reducing model drift. However, flux corrections are becoming less common as models are improved, in which case they are zero (see, for example, Section 8.4.2 of McAvaney et al., 2001).
- Some ocean models do not allow for the passage of water mass across the liquid ocean boundaries. Virtual salt fluxes are instead formulated to parameterize the effects of changes in salinity on the density field (Huang, 1993; Griffies et al., 2001). The models that use virtual fluxes do not have a physically correct water cycle, as there is zero exchange of water between the ocean and other components of the climate system. Correspondingly, they do not have a physically correct salt budget, since the real ocean system has a trivial net flux of salt across the ocean surface boundary, contrasting with the nontrivial virtual salt fluxes.
- The passage of water across an ocean boundary (via precipitation, evaporation, and runoff) corresponds to a transfer of tracer across the boundary, since the water generally carries tracer (e.g., carbon, heat). Heat from this water transport (relative to $0°C$) is requested in the Table 2.7. Notably, this heat transport is distinct from the heat transport associated with phase changes. Instead, the heat transport is due solely to the nontrivial heat present in water that moves between the ocean and other components of the climate system. Models that artificially preclude water to cross the ocean boundary (e.g., rigid lid models, or models with a virtual tracer flux) have zero contributions to these transport induced heat fluxes.

  For precipitation and evaporation, the heat flux associated with water transport across the ocean boundaries generally provides a global net cooling of the ocean, since evaporation transfers water away from the ocean at a temperature typically higher than precipitation adds water. In a steady state, this net heat loss is compensated by ocean heat transport, and the ocean heat transport is in turn balanced by atmospheric heat transport. However, most atmospheric models do not carry the temperature of its moisture field, thus precluding the atmospheric component in a climate model from representing the heat transport. There is a resulting non-conservative heat budget in the simulated climate system.

  The size of the atmospheric heat transport associated with the temperature of its moisture field is small relative to the atmospheric heat fluxes associated with phase changes of water in the atmosphere. It is the dominance of heating associated with phase change that has motivated atmospheric modelers to ignore the temperature of its moisture field. In contrast, ocean water must be tagged with its temperature field in order to simulate the ocean fluid, providing a generally nonzero heat associated with water that passes across ocean boundaries. The reader is referred to Section 3 of the coupled model paper from Delworth et al. (2006), where an estimate for the heat flux is given. Section D.2 of the ocean model comparison paper by Griffies et al. (2009) also provides a discussion of this heat flux.

4.5.1 Boundary fluxes of water mass

- rainfall_flux
- snowfall_flux
- water_evaporation_flux
- water_flux_into_sea_water_from_rivers
- water_flux_into_sea_water_from_icebergs
- water_flux_into_sea_water_due_to_sea_ice_thermodynamics
- water_flux_into_sea_water
- water_flux_into_sea_water_without_flux_correction
- water_flux_correction

These fluxes (Table 2.5) aim to present the analyst with sufficient information to perform a water mass budget on the liquid ocean. Models that employ a virtual salt flux, and so do not allow for the transfer of water mass across the liquid ocean boundary, will report zero for each of these fields. The following presents some general comments.

- Liquid runoff is defined as liquid water that enters the ocean from land, such as through rainwater in rivers, or snow and ice meltwater in rivers. It may also incorporate melt water from sea ice and icebergs.
- An iceberg model exports a certain amount of calved land ice away from the coasts. It is thus important to record where the icebergs melt (horizontal position and depth), hence the suggestion to include iceberg melt (Table 2.5).

We now present detailed comments about each of the fields. As discussed in the first bulleted item at the beginning of Section 4.5, the fluxes, which may be defined only over a portion of each ocean grid cell, are normalized by the full area of each ocean grid cell. As a result, the product of the ocean horizontal area and the flux will render the full mass entering or leaving the liquid ocean.

- rainfall_flux: The mass flux of liquid precipitation from the atmosphere entering the ice-free portion of an ocean grid cell.
- snowfall_flux: The mass flux of frozen precipitation from the atmosphere entering the ice-free portion of an ocean grid cell.
- water_evaporation_flux: This is a flux that is positive for water leaving the liquid ocean. It measures the rate at which water vapor leaves the liquid ocean and enters the atmosphere, through the ice-free portion of an ocean grid cell.
- water_flux_into_sea_water_from_rivers: This field measures the mass of liquid water runoff entering the ocean from land-surface boundaries.
- water_flux_into_sea_water_from_icebergs: The solid mass that enters the ocean from land-ocean boundaries will eventually melt in the ocean. This melt may occur just at the ocean-land boundary, be distributed seawards by a spreading scheme, or participate in the transport via icebergs. It is this melt that is requested.
- water_flux_into_sea_water_due_to_sea_ice_thermodynamics: This is the contribution to liquid ocean mass due to the melt (positive flux) or freezing (negative flux) of sea ice.
- water_flux_into_sea_water: This is the net flux of liquid water entering the liquid ocean.
- water_flux_correction: This field stores the mass flux due to flux corrections.
- water_flux_into_sea_water_without_flux_correction: This field stores the mass flux due to physical processes absent the flux corrections.

4.5.2 Boundary fluxes of salt

- virtual_salt_flux_into_sea_water_due_to_rainfall
- virtual_salt_flux_into_sea_water_due_to_evaporation
- virtual_salt_flux_into_sea_water_from_rivers
- virtual_salt_flux_into_sea_water_due_to_sea_ice_thermodynamics
- virtual_salt_flux_correction

- virtual_salt_flux_into_sea_water
- downward_sea_ice_basal_salt_flux
- salt_flux_into_sea_water_from_rivers

These fluxes (Table 2.6) aim to present the analyst with sufficient information to perform a salt budget on the liquid ocean. The following presents some details about the fields. Note that for models using real water fluxes, the virtual salt flux fields will all be zero.

- virtual_salt_flux_into_sea_water_due_to_rainfall: This field measures the virtual salt flux associated with liquid precipitation.
- virtual_salt_flux_into_sea_water_due_to_evaporation: This field measures the virtual salt flux associated with the evaporation of liquid water.
- virtual_salt_flux_into_sea_water_from_rivers: This field measures the virtual salt flux associated with the liquid runoff from land processes.
- virtual_salt_flux_into_sea_water_due_to_sea_ice_thermodynamics: This field measures the virtual salt flux associated with the melting or freezing of sea ice.
- virtual_salt_flux_correction: This field measures the virtual salt flux arising from a salt flux correction.
- virtual_salt_flux_into_sea_water: This field measures the total virtual salt flux entering the ocean.
- Salt transport from sea-ice to the ocean is measured in the field downward_sea_ice_basal_salt_flux.
- Rivers may contain a nonzero salinity, in which case salt_flux_into_sea_water_from_rivers will be zero.

4.5.3 Boundary fluxes of heat

- upward_geothermal_heat_flux_at_sea_floor
- temperature_flux_due_to_rainfall_expressed_as_heat_flux_into_sea_water
- temperature_flux_due_to_evaporation_expressed_as_heat_flux_out_of_sea_water
- temperature_flux_due_to_runoff_expressed_as_heat_flux_into_sea_water
- heat_flux_into_sea_water_due_to_snow_thermodynamics
- heat_flux_into_sea_water_due_to_sea_ice_thermodynamics
- heat_flux_into_sea_water_due_to_iceberg_thermodynamics
- surface_net_downward_longwave_flux
- surface_downward_latent_heat_flux
- surface_downward_sensible_heat_flux
- surface_net_downward_shortwave_flux
- downwelling_shortwave_flux_in_sea_water
- heat_flux_correction

These fluxes (Table 2.7) aim to present the analyst with sufficient information to perform a heat budget on the liquid ocean. The following presents some details about the fields.

- upward_geothermal_heat_flux_at_sea_floor: Some ocean model components planned for CMIP5 employ geothermal heating through the ocean model's bottom surface. This heating is typically unchanging over the course of climate simulations, but there may be some efforts to allow for time dependence of the geothermal heating, in which case the monthly heat flux should be archived. It is assumed that models considering a geothermal heating will inject this heating at the sea-floor.
- temperature_flux_due_to_rainfall_expressed_as_heat_flux_into_sea_water: This field measures the heat carried by the transfer of rainfall into the liquid ocean, with the heat computed with respect to $0°C$. This heat is computed as
$$\text{rainfall heat (W/m}^2) = Q_{\text{rain}}\, C_p\, T_{\text{rain}}, \tag{4.19}$$
where Q_{rain} is the rainfall mass flux in kg/(m^2 sec), C_p is the rainfall heat capacity, and T_{rain} is the temperature of rainfall in degrees Celsius. Most climate models choose the rainfall temperature to equal the ocean sea surface

temperature. The reason for this assumption is that atmsospheric models tend not to carry the temperature of their moisture field. But this assumption need not be applied for more complete atmospheric models.

For models employing a virtual tracer flux, in which there is no mass or volume transport of water across the ocean surface, the field temperature_flux_due_to_rainfall_expressed_as_heat_flux_into_sea_water is zero.

Note that since we measure this heat flux with respect to $0°C$, there is no need to record an analogous heat flux due to snowfall, if we assume the snow enters the ocean at $0°C$.

- temperature_flux_due_to_evaporation_expressed_as_heat_flux_out_of_sea_water: This field measures the heat carried by the transfer of water away from the liquid ocean through the process of evaporation. This heat is distinct from latent heat flux, and it is computed with respect to $0°C$ in the following manner:

$$\text{evaporation heat } (\text{W/m}^2) = Q_{\text{evap}}\, C_p\, T_{\text{evap}}, \tag{4.20}$$

  where Q_{evap} is the evaporative mass flux in $\text{kg/(m}^2\,\text{sec)}$, C_p is the water heat capacity, and T_{evap} is the temperature of evaporating water in degrees Celsius, with T_{evap} generally equal to the ocean sea surface temperature.

  For models employing a virtual salt flux, in which there is no mass transport of water across the ocean surface, the field temperature_flux_due_to_evaporation_expressed_as_heat_flux_out_of_sea_water is set to zero.

- temperature_flux_due_to_runoff_expressed_as_heat_flux_into_sea_water: This field measures the heat of runoff that enters the liquid ocean, with respect to $0°C$. This heat is computed as

$$\text{runoff heat } (\text{W/m}^2) = Q_{\text{runoff}}\, C_p\, T_{\text{runoff}}, \tag{4.21}$$

  where Q_{runoff} is the liquid runoff mass flux in $\text{kg/(m}^2\,\text{sec)}$, C_p is the ocean heat capacity, and T_{runoff} is the temperature of liquid runoff in degrees Celsius. Note that this "runoff" mass flux may include melt from sea ice and melt from icebergs, in which case the name "runoff" is inappropriate, but is retained as a placeholder.

  For models employing a virtual tracer flux, in which there is no mass transport of water across the ocean surface, the field temperature_flux_due_to_runoff_expressed_as_heat_flux_into_sea_water will be zero.

- heat_flux_into_sea_water_due_to_sea_ice_thermodynamics: This field accounts for heat gain by the liquid ocean when sea ice is formed, or heat loss from the liquid ocean when sea ice melts. It also includes the conductive heat flux from the ice-ocean interface into the sea ice. Note that the field upward_sea_ice_basal_heat_flux was asked for in CMIP3, and this field is closely related to the field asked for here. Whereas CMIP3 asked for the total heating of ice, in Watts, we ask for heating per horizontal area of ocean an ocean grid cell.

  The reason to bundle two heat fluxes together is that many ocean-ice models partition these terms in different manners. To expose some details, consider the discussion in Winton (2000), where we here consider heating of the liquid ocean to be positive. Equation (23) in Winton (2000) then reads

$$F_b = M_b - 4\,K\,(T_f - T_i)/h_i, \tag{4.22}$$

  where F_b is the net heat flux into the liquid ocean associated with sea ice melt and formation, as well as conduction. The term $M_b > 0$ is the heat flux entering the liquid ocean during the formation of sea ice, whereas $M_b < 0$ is the heat flux lost from the liquid ocean upon melting sea ice. The second term on the right hand side is the conductive term, with K the thermal conductivity of sea ice. A value of $K = 2.03\,\text{W/(m °C)}$ is typical. T_f is the freezing temperature of seawater, with the ice-ocean interface assumed to be constantly at this temperature. T_i is the temperature of the sea ice. Finally, h_i is the sea ice thickness. The conductive term contributes a negative heat flux to the liquid ocean when the freezing temperature T_f is greater than the ice temperature T_i, and a positive heat flux for an oppositely signed temperature difference.

- heat_flux_into_sea_water_due_to_iceberg_thermodynamics: Icebergs transport calved land ice from the land into the ocean. A rudimentary "iceberg" model may simply be the insertion of calving land ice/snow into the ocean. More realistic iceberg models are being considered for climate modeling (Jongma et al., 2009). Melting of the icebergs into the liquid ocean is associated with a transfer of the latent heat of fusion from liquid ocean, and so represents a cooling of the liquid ocean in regions where the icebergs melt. It is this heat flux that is to be archived in the field heat_flux_into_sea_water_due_to_iceberg_thermodynamics.

- heat_flux_into_sea_water_due_to_snow_thermodynamics: Snow entering the liquid ocean is assumed to melt upon transferring its latent heat of fusion from the ocean. This cooling of the liquid ocean is what is to be archived in heat_flux_into_sea_water_due_to_snow_thermodynamics.

- surface_net_downward_longwave_flux: This field measures the net downward flux of longwave radiation that enters the liquid ocean.

- surface_downward_latent_heat_flux: This field measures the net flux of latent heating associated with the phase change from liquid ocean to water vapor.
- surface_downward_sensible_heat_flux: This field measures the net downward flux of sensible heating acting on the liquid ocean.
- surface_net_downward_shortwave_flux: This field measures the net downward flux of shortwave heating that hits the liquid ocean surface.
- downwelling_shortwave_flux_in_sea_water: This field measures the downwelling flux of shortwave heating within the three-dimensional liquid ocean. Shortwave radiation penetrates into the ocean column, with this penetration of fundamental importance for many ocean processes.
- heat_flux_correction: This field records the heat flux correction acting at the liquid ocean surface.

4.5.4 Boundary fluxes of momentum

- surface_downward_x_stress
- surface_downward_y_stress
- surface_downward_x_stress_correction
- surface_downward_y_stress_correction

These fluxes (Table 2.8) aim to present the analyst with sufficient information to quantify the net momentum imparted to the liquid ocean surface from the overlying atmosphere, sea ice, ice shelf, etc. Components of the vector stresses are aligned according to the ocean model's native grid, as these should be the flux components applied to the ocean model's surface grid cells.

4.6 In support of vertical/dianeutral SGS parameterizations

Thus far, we have recommended some fields that provide insight into the workings of various ocean subgrid scale (SGS) parameterizations. Table 2.9 presents additional fields to further characterize the parameterizations and their impact on the simulation, with focus on the vertical/dianeutral SGS parameterizations. In Section 4.6, we present fields helping to characterize lateral SGS parameterizations in the ocean models (see Table 2.10). In both cases, these fields are new relative to the CMIP3 fields discussed in WGCM (2007).

There is one limitation of the fields requested here that is worth highlighting. Although we recommend saving diffusivities and work terms, what is of more fundamental importance for characterizing the effects the SGS has on a field is the flux that it produces. Many fluxes are formulated as a diffusivity times a gradient. However, some parameterizations of fluxes are not expressed as such. Indeed, downgradient diffusion is not a good model for many SGS processes. It is for this reason that we ask for "buoyancy work from sgs parameterization" rather than just that from diffusivity. More generally, the protocols for saving fluxes rather than diffusivities and work need to be developed as more nontrivial SGS parameterizations become common.

We propose that dominant scientific use of the fields discussed in this subsection are realized by archiving *just* the climatological monthly means computed from 01JAN1986 to 31DEC2005 for the CMIP5 "historical" experiment. We thus recommend that only these fields be archived for CMIP5.

4.6.1 Vertical/dianeutral tracer diffusivities (2)

- ocean_vertical_heat_diffusivity
- ocean_vertical_salt_diffusivity
- ocean_vertical_tracer_diffusivity_due_to_background
- ocean_vertical_tracer_diffusivity_due_to_tides

Vertical/dianeutral tracer diffusivities used in modern IPCC class models typically consist of a static background value and a dynamically determined value. For the background diffusivity, some modelers choose a globally constant value, whereas others impose spatial dependence. There is evidence that the background diffusivity influences such processes as ENSO variability and overturning strength in model simulations. Hence, it is very important to have this field archived.

There are an increasingly large number of physical processes used by IPCC-class models that affect the vertical tracer diffusivity. For example, vigorous mixing processes in the upper ocean are associated with large mixing coefficients; more quiescent processes in the ocean pycnocline region lead to much smaller coefficients; and enhanced mixing near the ocean bottom generally increases the mixing coefficients. Ideally, the mixing coefficients corresponding to each of the separate processes will be archived.

It is difficult to anticipate the full suite of physical processes affecting the vertical/dianeutral diffusivity. As a start, we identify the following processes that are commonly found in IPCC class models, whose corresponding diffusivities would be of use in the CMIP5 archive:

- Static background tracer diffusivity meant to parameterize the background internal wave field;
- Tidal induced tracer diffusivity, with all relevant tidal constituents contributing to the mixing (e.g., Simmons et al., 2004; Lee et al., 2006);
- Boundary layer diffusivity meant to parameterize mixing at or near the ocean boundaries.
- Total vertical/dianeutral diffusivity for temperature and salinity associated with all physical processes, including the background diffusivity.

The background, tidal, and boundary layer diffusivities are the same for temperature, salinity, and other tracers. The total diffusivities may differ, however, if including a parameterization of double diffusive processes. In many implementations of boundary layer processes, the effects from double diffusion are wrapped into the boundary layer diffusivities. Hence, we write the diffusivities κ for temperature and salinity in the following form

$$\kappa^{\theta} = \kappa_{\text{back}} + \kappa_{\text{tides}} + \kappa^{\theta}_{\text{boundary+dd}} \tag{4.23}$$

$$\kappa^{S} = \kappa^{S}_{\text{back}} + \kappa_{\text{tides}} + \kappa^{S}_{\text{boundary+dd}} \tag{4.24}$$

We request archival of the following diffusivities

- κ^{θ}= ocean_vertical_heat_diffusivity
- κ^{S}= ocean_vertical_salt_diffusivity
- κ_{back}= ocean_vertical_tracer_diffusivity_due_to_background
- κ_{tides}= ocean_vertical_tracer_diffusivity_due_to_tides,

from which the boundary layer plus double diffusion diffusivities can be deduced.

4.6.2 Vertical/dianeutral momentum diffusivities (2)

- ocean_vertical_momentum_diffusivity
- ocean_vertical_momentum_diffusivity_due_to_background
- ocean_vertical_momentum_diffusivity_due_to_tides
- ocean_vertical_momentum_diffusivity_due_to_form_drag

As for tracers, the vertical/dianeutral diffusivities (also called kinematic viscosities) used in modern IPCC class models typically consist of a static background value and a dynamically determined value. The dynamically determined value is often determined as a function of the tracer diffusivity, with an assumed Prandtl number. In general, it is important to archive the viscosities in order to help characterize the parameterizations.

We write the total vertical/dianeutral momentum viscosity as the sum

$$\nu = \nu_{\text{back}} + \nu_{\text{form drag}} + \nu_{\text{boundary}} + \nu_{\text{tides}}. \tag{4.25}$$

These viscosities arise from the following processes:

- Static background vertical/dianeutral viscosity, meant to parameterize the background internal wave field;
- Vertical/dianeutral viscosity aimed at parameterizing mesoscale eddy induced form-drag (Greatbatch and Lamb, 1990; Ferreira and Marshall, 2006);
- Planetary boundary layer processes (e.g., Large et al., 1994);
- Tidal induced vertical/dianeutral viscosity (e.g., Simmons et al., 2004; Lee et al., 2006).

We request archival of the following momentum viscosities

- ν= ocean_vertical_momentum_diffusivity
- ν_{back}= ocean_vertical_momentum_diffusivity_due_to_background
- ν_{tides}= ocean_vertical_momentum_diffusivity_due_to_tides
- $\nu_{\text{form drag}}$= ocean_vertical_momentum_diffusivity_due_to_form_drag

from which the boundary layer diffusivities can be deduced.

4.6.3 Rate of work done against stratification (2)

- tendency_of_ocean_potential_energy_content
- tendency_of_ocean_potential_energy_content_due_to_tides
- tendency_of_ocean_potential_energy_content_due_to_background

A vertical/dianeutral diffusivity impacts the solution only in regions where there are nontrivial vertical tracer gradients. A measure of the impact can be deduced by mapping the rate at which work is done against the stratification by the tracer diffusivity. This work against stratification also impacts the potential energy budget, hence the name for the variables. We recommend mapping this work rate per horizontal area as a three-dimensional field.[28]

4.6.4 Frictional dissipation of kinetic energy by vertical viscosity (2)

- ocean_kinetic_energy_dissipation_per_unit_area_due_to_vertical_friction

Friction dissipates kinetic energy, especially in those regions of nontrivial shear. In characterizing the manner in which viscous friction affects the energy budget in an ocean model (Wunsch and Ferrari, 2004), it is important to map the energy dissipation.[29] It is sufficient for most purposes to map the dissipation from just the total vertical viscosity, rather than to partition it into separate processes.

4.7 In support of lateral SGS parameterizations

In this section, we present fields helping to characterize lateral SGS parameterizations in the ocean, with Table 2.10 summarizing the fields. As for the vertical/dianeutral SGS parameterizations, we propose that dominant scientific use of the fields discussed in this subsection are realized by archiving *just* the climatological monthly means computed from 01JAN1986 to 31DEC2005 for the CMIP5 "historical" experiment. We thus recommend that only these fields be archived for CMIP5.

4.7.1 Lateral tracer diffusivities (2)

- ocean_tracer_bolus_laplacian_diffusivity
- ocean_tracer_bolus_biharmonic_diffusivity
- ocean_tracer_epineutral_laplacian_diffusivity
- ocean_tracer_epineutral_biharmonic_diffusivity
- ocean_tracer_xy_laplacian_diffusivity
- ocean_tracer_xy_biharmonic_diffusivity

It is important to archive diffusivities used for neutral diffusion (Solomon, 1971; Redi, 1982) (laplacian and/or biharmonic), laplacian eddy-induced transport (Gent and McWilliams, 1990; Gent et al., 1995), and biharmonic eddy induced transport (Roberts and Marshall, 1998). In addition, some models employ an along-coordinate surface diffusion, either Laplacian or biharmonic, to stabilize the numerical instabilities inherent in certain implementations of rotated neutral diffusion (Griffies et al., 1998). We thus ask for the following diffusivities to be archived for CMIP5:

- ocean_tracer_bolus_laplacian_diffusivity = eddy induced transport diffusivity for a Laplacian operator;
- ocean_tracer_bolus_biharmonic_diffusivity = eddy induced transport diffusivity for a biharmonic operator;
- ocean_tracer_epineutral_laplacian_diffusivity = epineutral or isopycnal diffusivity for a Laplacian operator;
- ocean_tracer_epineutral_biharmonic_diffusivity = epineutral or isopycnal diffusivity for a biharmonic operator;

- ocean_tracer_xy_laplacian_diffusivity = along-coordinate diffusivity for a Laplacian operator;
- ocean_tracer_xy_biharmonic_diffusivity = along-coordinate diffusivity for a biharmonic operator.

Note that for isopyncal models, the distinction between epineutral and along-coordinate diffusivities is often blurred, though there is in general a distinction. Also note that the term "bolus" should be used synonymously with "eddy-induced", with the term bolus used for historical reasons.

4.7.2 Eddy kinetic energy source from Gent et al. (1995) (2)

- tendency_of_ocean_eddy_kinetic_energy_content_due_to_bolus_transport

An energetic analysis of the extraction of potential energy from the Gent and McWilliams (1990); Gent et al. (1995) scheme indicates that it affects an increase in the eddy kinetic energy (Aiki and Richards, 2008). The rate of eddy kinetic energy increase, per unit horizontal area is

$$\text{rate of eddy kinetic energy increase, per unit horizontal area, from GM (W/m}^2\text{)} = (\rho\,\mathrm{d}z)\,\kappa\,(N\,S)^2. \tag{4.26}$$

In this expression, N is the buoyancy frequency, S is the magnitude of the neutral slope, κ is the diffusivity setting the overall strength of the parameterization, $\rho\,\mathrm{d}z$ is the grid cell mass per horizontal area, with $\mathrm{d}z$ the cell thickness. In a Boussinesq model, the *in situ* density factor should be set to the constant Boussinesq reference density ρ_o used by the model.

4.7.3 Lateral momentum viscosities (2)

- ocean_momentum_xy_laplacian_diffusivity
- ocean_momentum_xy_biharmonic_diffusivity

We do not make the distinction between various methods used to compute the lateral momentum viscosities. Hence, we only recommend the total fields be archived from the ocean models in CMIP5:

- ocean_momentum_xy_laplacian_diffusivity = total lateral momentum Laplacian diffusivity
- ocean_momentum_xy_biharmonic_diffusivity = total lateral momentum biharmonic diffusivity.

4.7.4 Frictional dissipation by lateral viscosity (2)

- ocean_kinetic_energy_dissipation_per_unit_area_due_to_xy_friction

As for the vertical/dianeutral viscosity, we recommend archiving the maps of energy dissipation induced by the total lateral viscous friction.[30]

Notes

[9]As noted on page 47 of Gill (1982), with the exception of only a small percentage of the ocean, density in the World Ocean varies by no more than 2% from 1035 kg m^{-3}. Hence, ρ_o = 1035 kg m^{-3} is a sensible choice for the reference density used in a Boussinesq ocean climate model. However, some models may use a different value. For example, early versions of the GFDL ocean model (Cox, 1984) set ρ_o = 1000 kg m^{-3}.

[10]In a mass conserving non-Boussinesq fluid, the total mass of liquid seawater, $\mathcal{M} = \sum_{i,j,k} \rho\,\mathrm{d}x\,\mathrm{d}y\,\mathrm{d}z$, evolves according to the budget

$$\partial_t \mathcal{M} = \sum_{i,j} \mathrm{d}x\,\mathrm{d}y\,Q^{\mathrm{w}}, \tag{4.27}$$

where ρ the *in situ* density, $\mathrm{d}x\,\mathrm{d}y$ the horizontal area of a grid cell, $\mathrm{d}z$ the vertical thickness (generally time dependent), and Q^{w} (kg m^{-2} s^{-1}) is the net mass flux of water that crosses the liquid ocean boundaries, per horizontal cross-sectional area. Conservation of mass is a fundamental feature of a non-Boussinesq ocean which, unfortunately, is not respected by all non-Boussinesq models. Hence, it is useful to archive the total mass and boundary fluxes in order to quantify the level to which the model conserves.

[11]In a volume conserving Boussinesq fluid, the total volume of liquid seawater, $\mathcal{V}^{\mathrm{Bouss}} = \sum_{i,j,k} \mathrm{d}x\,\mathrm{d}y\,\mathrm{d}z$, evolves according to the budget

$$\partial_t \mathcal{V}^{\mathrm{Bouss}} = \sum_{i,j} \mathrm{d}x\,\mathrm{d}y\,Q^{\mathrm{w}}/\rho_o. \tag{4.28}$$

Conservation of volume is a fundamental feature of a Boussinesq ocean which, unfortunately, is not respected by all Boussinesq ocean models. Hence, it is useful to archive the volume and boundary flux fields to quantify the level to which the model conserves.

[12]The purpose of this endnote is to identify the differences in sea level reported from a Boussinesq model from that in a non-Boussinesq model. For this purpose, we note that mass conservation for a non-Boussinesq fluid column is given by

$$\partial_t \left(\int_{-H}^{\eta} \rho \, \mathrm{d}z \right) = -\nabla \cdot \mathbf{U}^{\rho} + Q^{\mathrm{w}}, \tag{4.29}$$

where η is the deviation of the sea surface from the $z = 0$ geopotential surface, $z = -H(x, y)$ is the ocean bottom,

$$\mathbf{U}^{\rho} = \int_{-H}^{\eta} \rho \, \mathbf{u} \, \mathrm{d}z \tag{4.30}$$

is the vertically integrated mass transport, and Q^{w} the mass of seawater per time per horizontal area crossing the ocean boundaries. This equation says that the mass per area of a fluid column has a time tendency determined by the convergence of mass carried by ocean currents into the column (the $-\nabla \cdot \mathbf{U}^{\rho}$ term), plus boundary fluxes of mass (the Q^{w} term). We can extract from the mass budget (4.29) the following equation for the sea level

$$\partial_t \, \eta = \frac{-\nabla \cdot \mathbf{U}^{\rho} + Q^{\mathrm{w}} - D \, \partial_t \, \overline{\rho}^{z}}{\overline{\rho}^{z}}, \tag{4.31}$$

where

$$\overline{\rho}^{z} = D^{-1} \int_{-H}^{\eta} \rho \, \mathrm{d}z \tag{4.32}$$

is the vertically averaged density in a column of seawater, and $D = H + \eta$ the thickness of the seawater column from the ocean bottom to the ocean surface. There are three terms on the right hand side of equation (4.31) that affect the sea level evolution. The mass convergence arises from dynamical currents moving mass around in a fluid column; the boundary flux of mass is known as a *eustatic term*; and the term $-D \, \partial_t \, \overline{\rho}^{z}$ gives rise to an increase in sea surface height when the vertically averaged density reduces (e.g., as from warming or freshening), with this change is known as a *steric effect*. Note that these three terms are weighted by the inverse of the column averaged density, so that the surface height changes more in regions of low density.

For a Boussinesq fluid, the seawater density factors in equation (4.31) are set to the constant reference density ρ_o, in which case the Boussinesq surface height evolves according to

$$\partial_t \, \eta^{\mathrm{Bouss}} = -\nabla \cdot \mathbf{U} + Q^{\mathrm{w}} / \rho_o. \tag{4.33}$$

Notably, there are no direct changes to η^{Bouss} from changes in the column averaged density. Instead, changes in density only affect the Boussinesq sea level indirectly as manifest through changes in the ocean currents affecting $-\nabla \cdot \mathbf{U}$. Such limitations must be recognized when using the sea surface height in Boussinesq models as an approximation to the sea level. For example, as pointed out by Greatbatch (1994), a horizontally uniform change in column averaged density will induce no changes to the currents, and so it will not alter η^{Bouss}. However, η will change from the steric term.

[13]To understand the basics of how the global mean sea level changes, consider the following relation between the total mass of liquid seawater, total volume of seawater, and global mean seawater density,

$$\mathcal{M} = \mathcal{V} \, \overline{\rho}, \tag{4.34}$$

where $\mathcal{M}$ is the total liquid ocean mass (equation (4.1)), $\mathcal{V}$ is the total ocean volume (equation (4.3)), and $\overline{\rho}$ is the global mean *in situ* density

$$\overline{\rho} = \frac{\sum \mathrm{d}x \, \mathrm{d}y \, \mathrm{d}z \, \rho}{\sum \mathrm{d}x \, \mathrm{d}y \, \mathrm{d}z}. \tag{4.35}$$

Temporal changes in total ocean mass are affected by a nonzero net mass flux through the ocean boundaries (equation (4.27))

$$\partial_t \, \mathcal{M} = \mathcal{A} \, \overline{Q^{\mathrm{w}}} \tag{4.36}$$

where $\overline{Q^{\mathrm{w}}} = \mathcal{A}^{-1} \sum \mathrm{d}x \, \mathrm{d}y \, Q^{\mathrm{w}}$ is the global mean mass per horizontal area per time of water crossing the ocean boundaries, with $\mathcal{A} = \sum \mathrm{d}x \, \mathrm{d}y$ the area of the global ocean surface. Temporal changes in the ocean volume are associated with sea level changes via

$$\partial_t \, \mathcal{V} = \mathcal{A} \, \partial_t \, \overline{\eta}, \tag{4.37}$$

where

$$\overline{\eta} = \mathcal{A}^{-1} \sum \eta \, \mathrm{d}x \, \mathrm{d}y \tag{4.38}$$

is the global mean sea level. Bringing these results together leads to the evolution equation for the mean sea level

$$\partial_t \overline{\eta} = \frac{\overline{Q^{\mathrm{w}}}}{\overline{\rho}} - \left(\frac{\mathcal{V}}{\mathcal{A}} \right) \frac{\partial_t \, \overline{\rho}}{\overline{\rho}}. \tag{4.39}$$

The first term in equation (4.39) alters sea level by adding or subtracting mass from the ocean, with the name *eustatic* associated with these processes. The second term arises from temporal changes in the global mean density; i.e., from *steric* effects. The steric effect is missing in Boussinesq models, so that the global mean sea level in a Boussinesq model is altered only by net volume fluxes across the ocean surface.

We can approximate each of the terms in equation (4.39) over a finite time Δt via

$$\Delta\overline{\eta} \approx \frac{\overline{Q^{\mathrm{w}}}\,\Delta t}{\overline{\rho}} - \left(\frac{\mathcal{V}}{\mathcal{A}}\right)\frac{\Delta\overline{\rho}}{\overline{\rho}}, \tag{4.40}$$

where the Δ operator is a finite difference over the time step of interest. In a Boussinesq model, we must compute each term on the right hand side in order to develop an accurate time series for the global mean sea level that includes steric effects. For a non-Boussinesq model, $\overline{\eta}$ is computed directly from the model's surface height field, since the prognostic model includes steric effects.

In either the Boussinesq or non-Boussinesq case, it is of interest to diagnose the steric term

$$\mathcal{S} \equiv -\left(\frac{\mathcal{V}}{\mathcal{A}}\right)\frac{\Delta\overline{\rho}}{\overline{\rho}} \tag{4.41}$$

in order to deduce the effect on $\overline{\eta}$ associated just with changes in global mean density, as distinct from changes in ocean mass. The steric effect is straightforward to diagnose from a model simulation, given temporal changes in the global mean density.

For CMIP5, we are interested in the change in sea level in a global warming scenario experiment, with respect to a reference state defined by the initial conditions of the experiment. In this case, the steric term at a time n is given by

$$\begin{aligned}\mathcal{S} &= -\left(\frac{\mathcal{V}^0}{\mathcal{A}}\right)\frac{\overline{\rho^n} - \overline{\rho^0}}{\overline{\rho^0}}\\ &= \left(\frac{\mathcal{V}^0}{\mathcal{A}}\right)\left(1 - \frac{\overline{\rho}^n}{\overline{\rho^0}}\right),\end{aligned} \tag{4.42}$$

where $\rho^0 = \rho(\theta^0, S^0, p^0)$ is the *in situ* density for a grid cell as determined by the grid cell's reference temperature, salinity, and pressure, $\rho^n = \rho(\theta^n, S^n, p^n)$ is the *in situ* density at time step n, and $\mathcal{V}^0$ is the reference volume of seawater.

[14]The time tendency for global mean density can be written as

$$\partial_t\overline{\rho} = \overline{\partial_t\,\rho} + \mathcal{V}^{-1}\iint \mathrm{d}x\,\mathrm{d}y\,(\rho(\eta) - \overline{\rho})\,\partial_t\,\eta, \tag{4.43}$$

where $\rho(\eta)$ is the density at the ocean surface. It is the first term in this equation that interests us when considering thermal effects, in which case we write

$$\overline{\partial_t\,\rho} = -\overline{\alpha\,\rho\,\partial_t\,\theta} + \overline{\beta\,\rho\,\partial_t\,S} + \overline{c_s^{-2}\,\partial_t\,p}, \tag{4.44}$$

where $\alpha = -\rho^{-1}\,\partial\rho/\partial\theta$ is the thermal expansion coefficient, $\beta = \rho^{-1}\,\partial\rho/\partial S$ is the haline contraction coefficient, and $c_s^2 = \partial p/\partial\rho$ is the squared sound speed. Given this expansion of the density tendency, we identify the thermosteric contribution to sea level as the following finite increment

$$\mathcal{S}^{\mathrm{thermo}} \equiv \left(\frac{\mathcal{V}}{\mathcal{A}}\right)\frac{\overline{\alpha\rho\,\Delta(\theta)}}{\overline{\rho}}. \tag{4.45}$$

As for the steric term given by equation (4.42), we identify a thermosteric term that is defined with respect to a reference state

$$\mathcal{S}^{\mathrm{thermo}} = \left(\frac{\mathcal{V}^0}{\mathcal{A}}\right)\frac{\overline{\alpha^{\theta^n}\,\rho^{\theta^n}\,(\theta^n - \theta^0)}}{\overline{\rho^0}}, \tag{4.46}$$

where $\alpha^{\theta^n} = \alpha(\theta^n, S^0, p^0)$ is the thermal expansion coefficient as a function of the evolving temperature and the reference salinity and reference pressure; likewise, $\rho^{\theta^n} = \rho(\theta^n, S^0, p^0)$. The reason to evolve the thermal expansion coefficient is that it is a strong function of the water mass properties. As the temperature of these water masses change under a changing climate, it is important to allow α to feel these changes. An alternative expression for the thermosteric term is motivated by the expression (4.42) for the steric term, so that we compute

$$\mathcal{S}^{\mathrm{thermo}} = \left(\frac{\mathcal{V}^0}{\mathcal{A}}\right)\left(1 - \frac{\overline{\rho(\theta^n, S^0, p^0)}}{\overline{\rho^0}}\right). \tag{4.47}$$

That is, the density in the numerator is computed as a function of the local temperature, with salinity and pressure held constant at their reference value.

[15]McDougall (2003) argues for the use of an alternative prognostic field called *conservative temperature*, which is the potential enthalpy divided by a reference heat capacity. Conservative temperature is indeed more conservative than potential temperature, and so provides a more solid foundation for prognosing heat movement in the ocean. However, for comparison to other models and to observational data, we still recommend that ocean components in CMIP5 archive their potential temperature field, whether the models consider this field as prognostic (most common situation) or diagnostic (as when conservative temperature is prognostic).

[16]According to the results of Griffies et al. (2009), one should *not* assume that all ocean models in CMIP5 are written with numerical methods that ensure the conservation of scalar fields such as mass, heat, and salt. One means to check for heat conservation is to compute the change in total heat over a specified time (say over a year), and compare that change to the total boundary heat input to the ocean system. The change in heat should agree to the heat input through the boundaries, with agreement to within numerical roundoff expected from a conservative model.

If there is a difference greater than numerical roundoff, then how significant is the difference? To answer this question, consider an order of magnitude calculation to determine the temperature trend that one may expect, given a nonzero net heat flux through the ocean boundaries. For simplicity, assume a Boussinesq fluid with constant volume (i.e., no net volume fluxes), so that the global mean liquid ocean temperature evolves according to

$$(\mathcal{V}\,\rho_o\,C_p)\,\partial_t\,\overline{\theta} = \mathcal{A}\,\overline{Q^{\mathrm{H}}} \tag{4.48}$$

where $\mathcal{V}$ is the liquid ocean volume, $\mathcal{A} = \sum \mathrm{d}x\,\mathrm{d}y$ is the surface area of the ocean, and $\overline{Q^{\mathrm{H}}} = \mathcal{A}^{-1}\sum Q^{\mathrm{H}}\,\mathrm{d}x\,\mathrm{d}y$ is the global average boundary heat flux. Typical values for the World Ocean yield $\mathcal{V}\,\rho_o\,C_p \approx 5.4\times10^{24}\,\mathrm{J}/^{\circ}\mathrm{C}$ and $\mathcal{A} = 3.6\times10^{14}\,\mathrm{m}^2$, leading to the decadal scale temperature trend

$$\frac{\Delta\overline{\theta}}{\text{decade}} \approx 0.02\,\overline{Q^{\mathrm{H}}}. \tag{4.49}$$

For example, with a $1\,\mathrm{W\,m^{-2}}$ average heating of the ocean over the course of a decade (the net signal from global warming is on the order of $1\,\mathrm{W\,m^{-2}}$), we expect a global mean temperature trend of roughly $0.02^{\circ}C$ per decade, or $0.2^{\circ}C$ per century. If there is an error in the balance (4.48), we may define a global mean heat flux

$$Q^{\mathrm{error}} \equiv \overline{Q^{\mathrm{H}}} - \mathcal{A}^{-1}\,(\mathcal{V}\,\rho_o\,C_p)\,\partial_t\,\overline{\theta}. \tag{4.50}$$

To translate the error in the net heating into an error in the temperature trend, use relation (4.49) to define

$$\frac{\Delta\overline{\theta}^{\mathrm{error}}}{\text{decade}} \approx 0.02\,Q^{\mathrm{error}}. \tag{4.51}$$

[17]As examples, note that CCSM plans to use 60 vertical levels for their ocean model in CMIP5, and GFDL will continue to use 50 levels in one of its coupled models, and 63 layers in another.

[18]Contrary to the notes in WGCM (2007), practical salinity, measured on the practical salinity scale, is *not* identical to absolute salinity, measured in g kg^{-1}. Practical Salinity of a sample of seawater is derived from a measurment of its conductivity, while absolute salinity is the total mass of dissolved material per mass of solution. Section 5 of Jackett et al. (2006) provides a discussion of the differences, and Millero et al. (2008) present a new definition of Standard Seawater (seawater of a specified reference composition of dissolved material). The composition of seawater does vary around the World Ocean, so the relationship between practical salinity S and absolute salinity S_A is rather complicated. If we ignore these compositional variations, and take seawater to always be of reference composition, then absolute salinity S_A is identical to reference salinity S_R, which is related to practical salinity S according to

$$S_R(\mathrm{g/kg}) = (35.16504/35)(\mathrm{g/kg})\ S(\mathrm{psu}). \tag{4.52}$$

Hence, seawater of practical salinity $S = 35$ corresponds to an absolute salinity of $S_R = 35.16504\,\mathrm{g\ kg^{-1}}$. Due to the ongoing work of the SCOR/IAPSO Working Group 127, it is anticipated that by the year 2011, the oceanographic community will employ a standard measure of salinity based on absolute salinity rather than practical salinity. In the meantime, we recommend that ocean models continue to assume their salinity variable indeed measures practical salinity. The only (minor) issue that arises with this interpretation is that fresh water concentration of a seawater parcel should be computed as

$$FW = (1 - 10^{-3}\,S_A), \tag{4.53}$$

where again S_A is the absolute salinity. Assuming the absolute salinity is the same as the reference salinity, then

$$FW \approx (1 - 1.0047\times10^{-3}\,S), \tag{4.54}$$

where $35.16504/35 \approx 1.0047$.

[19]The density factor ρ in a non-Boussinesq fluid becomes the constant ρ_o for Boussinesq fluids.

[20]The kinematic balance (4.31) for a non-Boussinesq fluid leads to the nontrivial divergence

$$\nabla\cdot\mathbf{U}^{\rho} = -\partial_t\,(D\,\overline{\rho}) + Q^{\mathrm{w}}, \tag{4.55}$$

whereas for a Boussinesq fluid, equation (4.33) leads to the divergence

$$\nabla\cdot\mathbf{U} = -\eta^{\mathrm{Bouss}}_{,t} + Q^{\mathrm{w}}/\rho_o. \tag{4.56}$$

[21]We consider the function

$$\psi^{U}(x,y) = -\int_{y_o}^{y}\mathrm{d}y'\;U^{\rho}(x,y'), \tag{4.57}$$

where the southern limit y_o is at Antarctica. Note that all intermediate ranges of latitude bands are included, so there are no shadow regions that may otherwise be isolated due to land/sea arrangements. By definition, the y derivative $\psi^{U}(x,y)$ yields the $\hat{\mathbf{x}}$-transport $\partial_y\,\psi^{U} = U^{\rho}$, yet the x derivative does not yield the $\hat{\mathbf{y}}$-transport due to the divergent nature of the vertically integrated flow. A complement function

$$\psi^{V}(x,y) = \psi^{U}(x_o,y) + \int_{x_o}^{x}\mathrm{d}x'\;V^{\rho}(x',y), \tag{4.58}$$

yields $\partial_x\,\psi^{V} = V^{\rho}$. In the special case of a Boussinesq rigid lid model absent surface water fluxes, ψ^{U} and ψ^{V} reduce to the single rigid lid barotropic streamfunction. In the more general case, comparison of ψ^{U} and ψ^{V} in climate model simulations at GFDL reveal that after just a few years of spin-up, patterns for the monthly means of ψ^{U} and ψ^{V} are very similar. This result provides evidence that much of the large-scale vertically integrated circulation is nearly non-divergent. In this case, either function ψ^{U} and ψ^{V} renders a useful map of the vertically integrated mass transport. Due to its simplicity, we recommend that ψ^{U} be archived.

[22]The mixed layer has near-zero gradients of θ, *salinity*, and density, as well as tracers such as CFCs. So most techniques to estimate the MLD rely on either a threshold gradient or a threshold change in one of these quantities, normally in potential temperature θ or density (see, for example Lorbacher et al., 2006; De Boyer Montegut et al., 2004; Monterey and Levitus, 1997). Relying solely on θ has the advantage of good observational data coverage, but this approach neglects salinity stratification associated with barrier layers (see e.g., Sprintall and Tomczak, 1992). In contrast, relying solely on density overlooks density-compensating changes in $\theta - S$, exaggerating the mixed layer depth.

The method we recommend for purposes of ocean model comparisons is that from Levitus (1982). Here, the MLD is defined based on meeting a "sigma-t" criterion. This method has the advantage that it is readily employed in off-line mode, thus supporting the use of monthly mean model fields, analogous to Levitus (1982).

[23]Large-scale vertical velocity is largely immeasurable in the real ocean. Hence, there is no observational motivation to archive vertical velocity. In contrast, horizontal velocity is measurable, thus motivating it being part of the CMIP5 archives.

[24]WGCM (2007) recommend partitioning the poleward overturning streamfunction into the Atlantic, Pacific, and Indian Oceans. However, to separate the Indian and Pacific Oceans is not sensible, since there is no meridional boundary separating these basins. Instead, the Atlantic-Arctic, Indian-Pacific, and World Ocean are physically relevant, and thus are recommended here. We make the same recommendation for the partioning of $\hat{\mathbf{y}}$-ward tracer transport into basins.

[25]We choose not to recommend plotting overturning on the neutral density coordinate from McDougall and Jackett (2005) in order to facilitate direct comparison of the density overturning streamfunction between isopycnal models, which are based on σ_{2000}, and non-isopycnal models.

[26]For the Gent et al. (1995) volume transport streamfunction in a Boussinesq fluid, we have

$$\begin{aligned}
\Psi^{\rm gm}(y,s,t) &= -\int_{x_a}^{x_b} \mathrm{d}x \int_{-H}^{z(s)} v^{\rm gm}\,\mathrm{d}z \\
&= \int_{x_a}^{x_b} \mathrm{d}x \int_{-H}^{z(s)} \partial_z\,(\kappa_{\rm gm}\,S^y)\,\mathrm{d}z \\
&= \int_{x_a}^{x_b} \mathrm{d}x\,\kappa_{\rm gm}\,S^y(z(s)),
\end{aligned} \tag{4.59}$$

where $\kappa_{\rm gm} > 0$ is the diffusivity, S^y is the $\hat{\mathbf{y}}$ neutral slope, and $\kappa_{\rm gm}\,S^y$ vanishes at the ocean bottom. As for the streamfunction Ψ defined by equation (4.17), we recommend archiving $\Psi^{\rm gm}$ on both depth/pressure levels and isopycnal (σ_{2000}) levels.

[27]The mass transport leaving the $\hat{\mathbf{y}}$-ward face of a grid cell is written here by $\mathcal{V}\,\mathrm{d}x = v\,\rho\,\mathrm{d}z\,\mathrm{d}x$, and so $C\,\mathcal{V}\,\mathrm{d}x$ measures the mass per time of tracer leaving the $\hat{\mathbf{y}}$-ward face. We now consider a decomposition of this transport by defining the basin average transport and basin average tracer concentration as follows

$$[\mathcal{V}] = \frac{\sum_i \mathrm{d}x\,\mathcal{V}}{\sum_i \mathrm{d}x} \tag{4.60}$$

$$[C] = \frac{\sum_i \mathrm{d}x\,C}{\sum_i \mathrm{d}x}, \tag{4.61}$$

along with the deviations from basin average

$$\mathcal{V} = [\mathcal{V}] + \mathcal{V}^* \tag{4.62}$$

$$C = [C] + C^*. \tag{4.63}$$

The discrete i-sum extends over the basin or global domain of interest, so that $\sum_i \mathrm{d}x\,\mathcal{V}$ is the total $\hat{\mathbf{y}}$-ward transport of seawater at this band at a particular ocean model vertical level. The resulting $\hat{\mathbf{y}}$-ward tracer transport becomes

$$\mathcal{H}(y,t) = \sum_i \sum_k \mathrm{d}x\,\mathcal{V}\,C \tag{4.64}$$

$$= \sum_i \sum_k \mathrm{d}x\,\left([\mathcal{V}]\,[C] + \mathcal{V}^*\,C^*\right), \tag{4.65}$$

where the k sum extends over the vertical cells in a column.

It is common to identify three components:

$$\text{y\_flux\_advect} = \sum_i \sum_k \mathrm{d}x\,\mathcal{V}\,C \tag{4.66}$$

$$\text{y\_flux\_over} = \sum_i \sum_k \mathrm{d}x\,[\mathcal{V}]\,[C] \tag{4.67}$$

$$\text{y\_flux\_gyre} = \sum_i \sum_k \mathrm{d}x\,\mathcal{V}^*\,C^*, \tag{4.68}$$

with

$$\text{y\_flux\_gyre} = \text{y\_flux\_advect} - \text{y\_flux\_over}. \tag{4.69}$$

This identity follows very simply when the advective flux takes on the form of either first order upwind or second order centered differences. It becomes more involved when considering higher order, or flux limited, advection schemes. In these general cases, this result serves as a defintion of the gyre component, so that the advective flux is built from the advection scheme used in the ocean model.

[28]The non-negative rate of work done against stratification by vertical/dianeutral diffusion of density is given by

$$\mathcal{P} \equiv \int \mathrm{d}V\, \kappa_{\mathrm{d}}\, \rho\, N^2, \tag{4.70}$$

where N^2 is the squared buoyancy frequency and κ_{d} is the vertical/dianeutral diffusivity corresponding to a particular SGS process. It is useful to map the integrand per unit area for each grid cell, which yields the rate of work per horizontal area for a grid cell of thickness $\mathrm{d}z$

$$\text{rate of work against stratification per horizontal area performed by vertical diffusion}(\mathrm{W/m^2}) = \kappa_{\mathrm{d}}\, \rho\, N^2\, \mathrm{d}z. \tag{4.71}$$

In this way, vertical integrals can be computed offline merely by taking a vertical sum. If we instead mapped the field $\kappa_{\mathrm{d}}\, \rho\, N^2$, then integrated amounts would not be computable accurately for those cases where the grid cell thickness is time dependent. These units also provide a means for directly comparing the work done by diffusion against the heat fluxes crossing the ocean boundaries.

Equation (4.70) assumes the heat and salt diffusivities are the same, which is the case for tidal and background diffusivities. For the full heat, $\kappa_{\mathrm{d}}^{\theta}$, and salt, κ_{d}^{S}, diffusivities, including double diffusion, we compute

$$\begin{aligned} &\text{rate of work against stratification per horizontal area performed by vertical diffusion}(\mathrm{W/m^2}) \\ &= -g\, \mathrm{d}z \left(\kappa_{\mathrm{d}}^{\theta}\, \frac{\partial \rho}{\partial \theta}\, \frac{\partial \theta}{\partial z} + \kappa_{\mathrm{d}}^{S}\, \frac{\partial \rho}{\partial S}\, \frac{\partial S}{\partial z} \right). \end{aligned} \tag{4.72}$$

[29]The energy dissipated at a grid box in a hydrostatic model from vertical/dianeutral viscous friction is $(\rho\, \mathrm{d}V)\, \kappa\, (\partial_z \mathbf{u} \cdot \partial_z \mathbf{u})$, where κ is the viscosity, $(\rho\, \mathrm{d}V)$ is the mass of the cell, and $\mathbf{u} = (u, v)$ is the horizontal velocity. Just as for the work done against stratification, it is useful to map the non-positive dissipation per horizontal area in a grid cell of thickness $\mathrm{d}z$, as given by

$$\text{dissipation per horizontal area from vertical friction}(\mathrm{W/m^2}) = -\kappa\, (\partial_z \mathbf{u} \cdot \partial_z \mathbf{u})\, \rho\, \mathrm{d}z. \tag{4.73}$$

In this way, vertical integrals can be computed offline merely by taking a vertical sum.

[30]The local energy dissipated in a hydrostatic model by a lateral Laplacian friction with isotropic viscosity A and anisotropic viscosity D (see Section 17.8.2 of Griffies, 2004) is given by the non-positive quantity

$$\mathcal{D} = -(\rho\, \mathrm{d}V) \left[A\, (e_{\mathrm{T}}^2 + e_{\mathrm{S}}^2) + 2\, D\, \Delta^2 \right], \tag{4.74}$$

where $e_{\mathrm{T}} = (\mathrm{d}y)\, (u/\mathrm{d}y)_{,x} - (\mathrm{d}x)\, (v/\mathrm{d}x)_{,y}$ and $e_{\mathrm{S}} = (\mathrm{d}x)\, (u/\mathrm{d}x)_{,y} + (\mathrm{d}y)\, (v/\mathrm{d}y)_{,x}$ are the deformation rates, θ is an angle that sets the alignment of the generally anisotropic viscosity (Large et al., 2001; Smith and McWilliams, 2003), $2\, \Delta = e_{\mathrm{S}} \cos 2\theta - e_{\mathrm{T}} \sin 2\theta$, and $\mathrm{d}x$ and $\mathrm{d}y$ are the horizontal grid elements. We recommend archiving three-dimensional maps of the dissipation per horizontal area for a cell of thickness $\mathrm{d}z$

$$\text{dissipation per horizontal area from lateral laplacian friction}(\mathrm{W/m^2}) = -(\rho\, \mathrm{d}z) \left[A\, (e_{\mathrm{T}}^2 + e_{\mathrm{S}}^2) + 2\, D\, \Delta^2 \right]. \tag{4.75}$$

As defined, vertical integrals can be computed offline merely by taking a vertical sum.

The local energy dissipated in a hydrostatic model by a lateral biharmonic friction is given by the non-positive quantity (see Section 17.9.2 of Griffies, 2004)

$$\mathcal{D} = -(\rho\, \mathrm{d}V)\, \mathbf{F} \cdot \mathbf{F}, \tag{4.76}$$

where $\rho\, \mathrm{d}V\, \mathbf{F}$ is the lateral Laplacian friction vector used to build up the biharmonic operator. As for the dissipation from vertical viscosity, we recommend mapping the dissipation per horizontal area for each grid cell of thickness $\mathrm{d}z$, as given by

$$\text{dissipation per horizontal area from lateral biharmonic friction}(\mathrm{W/m^2}) = -(\rho\, \mathrm{d}z)\, \mathbf{F} \cdot \mathbf{F}. \tag{4.77}$$

Again, vertical integrals can be computed offline merely by taking a vertical sum.

Chapter 5

Concluding Comments

We hope that this document presents the climate science community with a useful reference to help finalize the ocean model protocols used in CMIP5. Some of the material here will be of use for modelers developing the diagnostic tools in anticipation of their CMIP5 simulations, and some material will assist analysts aiming to understand precisely what was saved.

We nonetheless recognize that this document is incomplete. For example, there are numerous further model fields that could be of use for ocean model comparison research, and thus for supporting the IPCC model assessment process. We could also present far more scientific discussion of certain fields, thus better motivating their archival. We are very happy to consider future recommendations of fields that have been omitted here, acknowledging that this report should be revised for future comparisons (e.g., CMIP6). In turn, there is some balance required between asking for too much and too little. Anticipating some criticism of asking for too much, we reemphasize those areas where output reduction has been employed by, for example, saving certain fields for only the historical experiment, and saving only the 20 year monthly climatologies of certain subgrid scale parameters (Section 1.3). Such output reduction methods allow for the newly requested fields to represent only a modest increase in overall archive requirements.

It is useful to highlight certain questions that remain unanswered by this document. Developing workable answers for these questions is somewhat time critical as model resolutions are refined, and the use of more complex horizontal and vertical grids becomes common.

- **Grid Standards and Remapping**: We presented a discussion of spatial sampling in Sections 1.1 and 3.1.2, at which point we highlighted the incomplete status of a generally agreed upon grid specification standard. The absence of a standard greatly handicaps the community-wide ability to address issues such as remapping. It is of fundamental importance that the community agree on a grid specification standard in the very near future.
- **Eddy statistics**: Ocean model resolutions of $1/4^{\circ}$ or finer are actively being considered by various groups for AR5, with future assessments likely to see even more models of this eddying class. In this document, we made no comment on the needs of sampling fields to develop robust eddy statistics, with much work required to prescribe general recommendations.[31]
- **3d Term Balances**: A mechanistic understanding of ocean processes typically requires the analysis of detailed budget terms, in addition to boundary forcing. For tracer budgets, one generally requires grid cell tendencies and flux components from advection and all SGS processes. For momentum budgets, one may require all the forces and transport processes acting to alter momentum. Beyond tracer and momentum, there are budgets for diagnostic fields that may be of interest, such as potential vorticity. In this document, we did not propose saving such detailed budget terms for CMIP5, both due to time limitations in preparing this document, and anticipated limits to the CMIP5 archive. However, more careful thought may arrive at a middle ground. Two approaches seem prudent: (A) a limited number of 3d budget terms be saved that allow for a nontrivial process analysis, even if incomplete; (B) a full suite of fluxes and forces are saved over a very limited *intense sampling* period.
- **SGS Parameterizations**: As mentioned in Section 4.6, many SGS parameterizations are not represented as downgradient diffusion. Protocols for saving fluxes rather than diffusivities need to be developed, especially as nontrivial SGS parameterizations become common.
- **Vertical Coordinates for Analysis**: There are many ocean science questions that are best addressed within the kinematic framework of a particular vertical coordinate. The commonly used geopotential/pressure option is just one choice. Hence, we must continue to ask questions concerning what vertical coordinate(s) provides the most

physical insight for a particular diagnostic field. Answers to these questions will likely lead to recommendations for how to sample certain fields from the models. Sampling on alternative coordinate surfaces is a likely outcome of the growing sophistication of the ocean model output used in CMIP.

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