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Evaluation of a global ocean general circulation model; The Lat-Lon-Cap (LLC90) configuration of the MITgcm

Bernawis Lamona¹, Judith Hauck² and Christoph Völker²

¹Department of Oceanography, Institut Teknologi Bandung, Indonesia
²Alfred Wegener Institute, Germany

E-mail: lamona@fitb.itb.ac.id

Abstract. An evaluation of a general circulation model from Massachusetts Institute of Technology (MITgcm) with Lat – Lon – Cap (LLC90) configuration was done. Simulation of 100 years yields the annual means of potential temperature, salinity, meridional overturning stream function and transport of several throughflows. A reference run with widened Davis strait from the model was used to compare with observational WOA09 data, and the MITgcm has reproduced reasonably reliable data. The patterns of potential temperature can resemble the WOA09, however not so for the salinity fields. An experiment to simulate the model in 5 different tuning set were then proceeded. For the potential temperature and salinity fields, the North Pacific and the Southern Ocean still has significant difference to the WOA09. The difference was in the range of -6 to 5 °C for the potential temperature and -0.5 to 3 psu for the salinity field. The meridional overturning circulation stream function was still lower than the observational (± 17 Sv); the MITgcm simulates it 10 ± 1 Sv in the experiment. The root mean square (RMS) of the difference were calculated, the surface layer has the largest magnitude of difference due to the forcing dynamics. The best simulation, which has the least difference to WOA09, was the one with the original bathymetry and smallest vertical diffusivity coefficient, 1 x 10⁻⁵ m² s⁻¹.

1. Introduction

MITgcm is widely used by oceanographers in wide spectrum of expertise; physics, biology, chemistry, and ocean-atmosphere interactions among others. The uniqueness of this general circulation model is that it employs the non-hydrostatic equation and fluid isomorphism to simulate fluid phenomena of both ocean and atmosphere, from the small scale of convection with 100m range to the general circulation of the world ocean on a scale of thousands of kilometers. In our simulations we used a resolution of 1°.

Marshall [1] elaborated ocean models based on hydrostatic primitive equations (HPE), quasi hydrostatic (QH) and non-hydrostatic (NH) equations. In HPE, the vertical momentum equation is reduced to hydrostatic balance and an approximation is made with approximated Coriolis force and a shallow atmosphere approximation is made. With this, in large scale, the terms omitted in HPE are negligible, but on a small scale they become large enough to become a problem. For the application to
global circulation, an HPE model is very accurate but it will break down somewhere between 10 to 1 km as the horizontal scale of the motion becomes comparable with its vertical scale.

In QH, the precise balance between gravity and pressure gradient forces is relaxed. The Coriolis force is treated exactly by including the cosine of latitude and becomes significant when approaching the equator; this provides the model a complete angular momentum. And in NH, the incompressible Navier Stokes equations were used.

Marshall [1] simulated small, meso and large scale with the three models and drew conclusions that there are equations that are more accurate than HPE which can be formulated and implemented e.g. QH and NH. These two are more versatile because they can simulate both small and large scale.

MIT scientists then developed milestones on the development of this general circulation model: [2],[3],[4] in aspects of efficiency of ocean modelling using the non-hydrostatic equation, new treatment of Coriolis terms in C-grid models, the using of parallel processor to run models, and construction of the adjoint MIT ocean general circulation model , respectively.

Currently, as in MITgcm manual by Adcroft, [5], the NH equation that is employed by the model is as below. Density,

\[ \rho = \rho_0 + \rho' \]  

Variations with depth \( \rho_0 \) are negligible and compressible in \( \rho' \) are:

\[ \rho_0 = \rho_c \]

\[ \rho' = \rho(\theta, S, p_0(z)) - \rho_0 \]  

This leads to “semi-compressible” Boussinesq equations:

\[ \frac{D\tilde{v}_h}{Dt} + f\tilde{k} \times \tilde{v}_h + \frac{1}{\rho_c} \nabla z p' = \tilde{F} \]

\[ \epsilon_{nh}\frac{dw}{dt} + \frac{g \rho' \rho_c}{\rho_c} + \frac{1}{\rho_c} \frac{1}{\rho_c} \frac{\partial p'}{\partial z} = \epsilon_{nh} \tilde{F}_w \]

\[ \nabla_z \cdot \tilde{v}_h + \frac{\partial w}{\partial z} = 0 \]

\[ \rho' = \rho(\theta, S, p_0(z)) - \rho_c \]  

\[ \frac{d\theta}{dt} = Q_\theta \]

\[ \frac{dS}{dt} = Q_s \]

Where \( \tilde{v}_h \) (u,v,0) is horizontal component of velocity (on pressure surface), f is Coriolis force; \( 2\Omega \sin \phi \), \( \rho_c \) is constant reference density of water, F is force, \( w = \frac{dp}{dt} \) is vertical velocity in p-coordinate, \( \epsilon_{nh} \) is coefficient of non-hydrostatic to make eq.(4) not negligible, \( \theta \) is potential temperature, S is salinity, and Q is heat rate per unit mass.

Griffies, [6] have done comparative study by using seven models i.e. NCAR-POP, FSU-HYCOM, GFDL-MOM, GFDL-HIM, Kiel-ORCA, KNMI-MICOM, and MPI (see section 5 for acronyms). The study was within the frame of Coordinated Ocean-ice Reference Experiments (COREs). They aimed the COREs to become a tool to learn the behaviour of global ocean-ice models forced by common atmospheric dataset. In their work they evaluated the simulation results from the seven models, which were:
- the annual mean temperature, salinity
- anomaly of SST and SSS
- monthly values of heat content
- annual mean sea-ice area
- sea-ice concentration
- upper ocean temperature on the equator in the Pacific
- upper ocean zonal velocity component on the equator in the Pacific
- maximum mixed layer depth
- anomalous zonal-mean decadal mean of potential temperature and salinity
- Atlantic Basin meridional overturning stream function
- global meridional overturning stream function
- annual mean of volume transports.

The second phase of COREs (COREs-II) was conducted by Danabasoglu, [7] focusing in the North Atlantic, more specifically in the AMOC (Atlantic Meridional Overturning Circulation). AMOC has large heat and salt transport that influence the North Atlantic climate and finally affected the global climate through the ocean-atmosphere interaction. AMOC plays an important role in decadal and longer time scale of the climate variability, hence it will be important aspect to develop model to predict the earth’s future climate in that time scale. The COREs-II experiment was a hindcast simulation to provide reconstruction of the AMOC behavior the past decades/centuries. They evaluated more or less the same parameters as Griffies, [6] have done, but with 13 different models, namely : MOM4p1, MICOM, NEMO 3.1.1, 3.2, 3.3, 3.4, HYCOM 2.2, GOLD, GISS Model E2-R, INMOM, MITgcm, MOVE/MRI.COM 3, POP2 simulated by 18 research institutions. Some institutions used the same model with different configurations.

Hauck, [8] used MITgcm to investigate seasonally different carbon flux changes in the Southern Ocean in response to the Southern Annular Mode (SAM). They did not include the Arctic in that work. In 2015, they modified a global configuration, namely LLC90 (lat lon cap) provided by Forget, [9] to include the Arctic into the model. This paper will discuss the evaluation process of the simulation result and the tuning of some parameters to reproduce the closest possible salinity and potential temperature to the observational data.

2. Model
The Massachusetts Institute of Technology general circulation model (MITgcm) was set up globally with a 1° resolution. The ocean water column was set to 50 layers, with 10m increment from 5 m to 115 m, and larger increment down to 5906m depth. A thermodynamic and dynamic sea-ice model [10] was used. One bathymetry product from LLC90 configuration provided by Forget, [9] was applied. The Davis Strait was widened on purpose to avoid the model break down because of ice flood.

The reference run was set up as follow. The diffusive rate coefficient was $2 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$. The Davis Strait was widened on purpose to avoid numerical problems with too strong ice formation. Some other changes at the Arctic and Anarctic coast were also applied to prevent too strong build-up of sea ice. The model then simulated for 100 years, the first 90 years is considered as the model spin-up. The last ten years were then calculated to get the annual mean for the analysis. The analysis was mostly done by using the gcmfaces, a Matlab package to analyze MITgcm provided by [11]. This reference run was then used to evaluate by comparing the data it reproduced to the observational World Ocean Atlas 2009 (WOA09).

3. Evaluation of the reference run
In this project we only evaluate the potential temperature and salinity field in global scale and in Atlantic basin, the zonal mean of those fields, the meridional overturning stream function and the transports volume in several throughflows.
We observe the difference between the global salinity and potential temperature fields of WOA09 and the simulated field by the model from the reference run as in Figure 2 and 3.

![Figure 1](a) Surface salinity and potential temperature fields from WOA09 (a,c), and from MITgcm reference run result (b,d).
Figure 2. Salinity and potential temperature fields of 1000m depth from WOA09 (a,c) and MITgcm reference run result (b,d).

Figure 3. Annual zonal mean potential temperature and salinity (a,c) from WOA09 and (b,d) from MIT gcm reference run.
Figure 3 shows the comparison between zonal mean of potential temperature and salinity from WOA09 and from the reference run. We needed to interpolate the model results because WOA09 has 33 depth layers while MITgcm has 50 depth layers. Therefore, we interpolated the MITgcm depth layers into WOA09's to make those comparable. We can say from these Figures that MITgcm is able to reproduce the potential temperature and salinity reasonably well. Further on we would like to calculate the difference between model and data in a plot to get a more detailed look.

Figure 4(a) is showing some transports calculated from the reference run. We can see that the value for Davis Strait is out of range of $-3 \sim 1$ Sv; probably due to the widened bathymetry of it in the model configuration to prevent the piling-up of ice, and 4(b) the volume of meridional overturning. More to discuss in section 5.

![Figure 4](image)

**Figure 4.** (a) the transport in several throughflows of (from left to right); Bering Strait, Davis Strait, Denmark Strait, Iceland Faroe, Faroe Scotland, and Indonesia Throughflow, respectively. (b) the meridional stream function.

4. Experiment

We continued to the next step from model evaluation i.e. to conduct some tuning experiment. We did five more simulations with different sets of parameters. One change was to change the bathymetry to the original bathymetry that was provided in the LLC-90 configuration as in [9]. Other parameter that was altered was the vertical diffusivity and delta Z of sea-ice growth calculation. We named the runs as run5,6,7,8, and 9, respectively.

The following table is the summary of the tuning parameter set in the runs of our experiment. We refer to as “original bathymetry” for that of ecco configuration (bathy_ecco_llc_90x50_min2pts.bin in the table). For the newly set up bathymetry with widen Davis Strait and some changes in Arctic and Antarctic coast, we refer to as “new bathymetry” (bathymetry.llc in the table). BL79 in run 8 is a vertical diffusivity scheme suggested by Bryan, [12]. In run9, we enabled a flag in SEAICE_OPTIONS.h and altered the code in seaice growth.f to limit the effect of sea-ice on the non-linear free surface to a maximum of one full grid box. In all other runs this flag was not set, hence the effect of sea-ice on the non-linear free surface was not limited.

**Table 1.** Summary of the tuning parameter set in simulation experiment

| Run | Bathymetry                  | Initial T/S       | Vertical diffusivity ($m^2 s^{-1}$) | Delta Z in seaice_growth.F |
|-----|-----------------------------|-------------------|-------------------------------------|----------------------------|
| reference | bathymetry_llc.bin       | llc90_12months.bin | $2 \times 10^{-5}$               | not set                    |
| 5   | bathy_ecco_llc_90x50_min2pts.bin | llc90_12months.bin | $1 \times 10^{-5}$               | not set                    |
4.1 Global

Figure 5. The difference (°C) between observational SST WOA09 and MITgcm for run 5, 6, 7, 8.

We can see in Figure 5, the value difference of observational SST data from WOA09 and MITgcm from run 5 to 8. We can see some high difference mostly in the North Atlantic. The difference is 2 to 4°C, where the MITgcm is colder than WOA09. We also get a notable difference at the Southern Ocean, where the MITgcm is warmer than WOA09. However it is not so clear here, with some areas are too warm and some areas are too cold. The largest difference is in the North Atlantic and also in the Arctic where WOA09 is consistently warmer.
Figure 6. The difference between observational potential temperature at 1000m depth WOA09 and MITgcm for run 5, 6, 7, 8.

At 1000m depth as in Figure 6, in general, we have the MITgcm data about 2°C warmer than WOA09. We can see in Figure 6 that run 5 is the best run with the least difference to WOA09 among other runs. At this depth the model is less affected from the surface forcing, as expected for a 100 year run.
Figure 7. The difference (psu) between observational surface salinity WOA09 and MITgcm for run 5, 6, 7, 8.

As for the surface salinity, we have the most contrast difference (still) in the North Atlantic; the MITgcm is less saline than WOA09 up to 5 psu. In the Arctic Ocean in contrary, the MITgcm is more saline than WOA09 up to 5 psu.
Figure 8. The difference between observational 1000m depth salinity WOA09 and MITgcm for run 5, 6, 7, 8.

Opposite to the surface where the MITgcm is more saline in the Arctic Ocean, at 1000m depth the MITgcm is more saline at Southern Ocean. Run 5 still hold the least difference to WOA09. From these results we would like to give more focus to North Atlantic Ocean because there is constant significant difference.

4.2 Atlantic basin

The zonal mean of potential temperature from all runs almost look identical (Figure 9). The model can resemble the WOA09 quite well, however the surface to 1000 m depth layer isotherms in the South Atlantic are deeper than WOA09. In contrary, the isotherms in the North Atlantic seem shallower. At the surface of low latitude, the MITgcm is warmer than WOA09 –and of high latitude it is colder. The isotherms of 2-3°C pattern from the models are also different. Those stay at depth of 2000-1000m through the equator, while in the WOA09 those stay at much lower, below 3000m depth.

![Figure 9](image-url)

Figure 9. The zonal mean of potential temperature from (a). The WOA09 observational data. (b) MITgcm run 5.

In Figure 10 we show the difference of zonal mean potential temperature between the WOA09 and MITgcm. This Figure sustains that MITgcm is warmer WOA09 at the low latitude surface layer and
colder the high latitude surface layer. It takes longer time for the deep layer to experience the dynamic forcing from the surface layer.

We continue to look at the zonal mean salinity vertical distribution of WOA09 as in Figure 11 (a). Here MITgcm does not seem to reproduce the data very well. From surface to 1000 m depth we have two high salinity cores of 36-37 psu from 40°S to equator, less saline in equator due to high precipitation, then continues from near equator toward 40°N. The Northern core is almost 300% larger than the Southern one. MITgcm reproduced these cores at least 2 psu less than WOA09, with the Southern core is much larger than the Northern one. The WOA09 also has tongue of 34-34.5 psu at the Northern part, which MITgcm failed to reproduce. Rather it produced the tongue at the Northern part of Atlantic basin. The rest of salinity gradient of MITgcm is not as smooth as in WOA09 with more significant difference.

![Image](https://example.com/image1.png)

Figure 10. The difference of zonal mean potential temperature between WOA09 and MITgcm from (a), (b), (c), (d) run 5, 6, 7, and 8, respectively.
Coming from these result, we checked on the difference between WOA09 and MITgcm (Figure 12). We found two contrast cores of highest difference from surface to 1500m depth. In the southern part the higher difference takes place at the 34-34.5 psu salinity tongue. MITgcm is more saline than WOA09. In the northern part it takes place at the high salinity core of WOA09. This extended up to 60⁰N. The difference value is at least 1.5 to 2.5 psu, the MITgcm is less saline than WOA09.

**Figure 11.** (a). The zonal mean salinity in Atlantic basin of WOA09. (b). The same for MITgcm run 5.

**Figure 12.** The zonal mean salinity difference between WOA09 and MITgcm from (a), run 5 (b) run 6, (c), run 7 and (d) run 8.
4.3 Meridional overturning circulation

Our reference run did not reproduce well the North Atlantic meridional overturning (see section 5. Discussion). Run 5 gave almost 10 Sv of the overturning and all the rest runs; 6,7,8,9, resulted in smaller volume than observational value of 17 Sv [1] of the overturning. Run 8 and 9 are not shown due to insignificant difference with run 6 and 7.

![Figure 13](image)

Figure 13. Meridional overturning from MITgcm (a). run 5, (b). run 6, (c). run 7. Other runs are not shown due to similarity to run 6 and 7. In unit of Sverdup (10$^6$ m$^3$s$^{-1}$).

All the simulations of experiment return the value for the Davis Strait back in the scale range, because we used the original bathymetry again with more realistic Davis Strait width. The transport volumes among runs are not very much different; hence we only show result of run 5 in Figure 14. However these transports are still not showing seasonality, due to the values come from the annual mean.
4.4 Root Mean Square

To further analyze the magnitude of the difference, we calculated the root mean square (RMS) of it. Figure 15 are the RMS of the potential temperature and salinity difference for each depth layers of Atlantic basin. The RMS peak of potential temperature is at 10m depth, and from 0 to 5m depth for salinity. All runs are almost identical for the RMS as in Figure 15(a) and (b). To check the magnitude along the latitude, we chose surface, 1000 m and 3000 m depth layers. For run 5 (Figure 15 (c)) the low latitudes have the smallest RMS difference in all layers and the difference are larger in the north and south of the Atlantic basin. Layer 3000m depth is the most stable RMS and only peaks in the north Atlantic. Run 6,7,8,9 have similar RMS for the difference, hence we only show run 8 in Figure 15(d). The magnitude of difference is larger than in run 5. At surface, the RMS is peaked higher in the north than in the south of Atlantic. In contrary at 1000m depth, it peaks higher in south than north. For 3000m depth it is the same as run 5, less difference with peak only in north of Atlantic. Surprisingly, the RMS of salinity difference is only large enough at the surface layer as in Figure 15(e).
Figure 15. Root Mean Square of (a) Atlantic basin potential temperature difference and (b) salinity. (c), (d). RMS of surface, 1000m, 3000m depth layers of potential temperature difference along the latitude from run 5 and 8. Other runs are quite similar to run 8, hence not shown. (e), the same for salinity difference, run 5.

For the global scale, Figure (16) shows us that the potential temperature difference magnitude is 0 to 1.5 °C, peaked at near 0° and 300° longitude for the surface layer in run 5. In run 6, the 1000 m is mostly with larger difference than the surface. The deep layer of 5000m gives the least difference. Other runs are relatively similar to run 6, hence not shown here.

As for the salinity, the surface layer is always the highest difference through all longitude. At 1000 m depth the model is less different than WOA09 except at longitude of 0° to 40°E. All runs are mostly having similar pattern of RMS value, and will be discussed in section 5.
5. Discussion and concluding remarks
The model in general, from all the 7 experiments is having more difficulties in the North Atlantic and in the fronts of Southern Ocean as seen in Figure 5,6,7. [13] mentioned at least the separation of the Gulf stream, the strength and structure of meridional transport, the mode water formation and subduction, the nonlinear dynamics of eddies and interbasin exchange with the Mediterranean and Nordic seas are difficult to model. They suggested that direct simulation of wind-driven and the MOC are not very successful yet due to high uncertainty being the sensitivity of the models. As for the Southern Ocean, is location of several fronts i.e the Subtropical Front (STF), South Subtropical Fronts (SSTF), Subantarctic Fronts (SAF), and Polar Fronts (PF) according to Belkin and Gordon [14]. These fronts’ dynamics are also difficult to simulate.

From the reference run, the volume transport of Davis Strait was out of range, with mean value -4.66 Sv. Davis Strait is the main conduit from the North Atlantic to Arctic Sea, exchanging not only sea water but fresh water as well from the Canadian glacial and West Greenland. Cuny et al. [15] found the net volume of -2.6 ± 1Sv, and fresh water of -92 ± 34 mSv from three consecutive years of moorings. Some moorings and gliders have been deployed to measure the transport volume within 2004 - 2010 and resulted in -1.6 ± 0.5 Sv, with -93 ± 6 mSv of liquid freshwater and sea ice export -10 ± 1mSv [16]. Our range was -3 to 1 Sv, hence with the new bathymetry in which Davis Strait was widened with some change in North Atlantic and Arctic coast we get the volume exceeding this range. In our experiment we used the original bathymetry and the transport is back in range, -0.2 to -0.5 out of the reproduced data. Apparently this number is still too low compared to the observational data.

Roemmich and Wunsch [17] calculated the net northward meridional transport of North Atlantic from transatlantic hydrographic sections in 1981 along latitudes 24.5° and 36.25°N. The data collected was obtained from 215 stations of CTD probe and water samples for other tracers. They came up with the number of 17 Sv of surface and intermediate water above $\sigma_t = 36.82$. Ganachaud [18] conducted inverse study on observational (WOCE, World Ocean Circulation data) estimates and obtained number of 16 ± 2 Sv transport at 48°N. Other inverse studies are from Ganachaud and Wunsch [19]; 15 ± 2 Sv at 42°N and Lumpkin, Speer [20]; 16 ± 2 Sv at 48°N. Lumpkin, Speer [21] and Lumpkin, Speer [20] had 13 ± 2 Sv at 42°N. Most of our experiment simulations resulted with less than 10 Sv for this transport. If we compare to other models as in Griffies et al. [6], in which they run the simulation for

Figure 16. (a), (b). RMS of global potential temperature difference (WOA09-MITgcm) of surface 1000m, 5000m in run 5 and 7, respectively. Run 6 and 8 are not shown due to quiet similarity to run 8. (c),(d). The same for salinity difference.
500 years; we would like to show out some numbers of their models’ annual mean Atlantic meridional overturning stream function values at year 100 as follows:

- NCAR-POP (National Center Atmospheric Research – Parallel Ocean Program) : ±10 SV
- FSU-HYCOM (Florida State University—Hybrid Coordinate Ocean Model) : ±8.5 SV
- GFDL-MOM (Geophysical Fluid Dynamic Laboratory - Modular Ocean Model) : ±12.5 Sv
- GFDL-HIM (GFDL - Hallberg Isopycnal Model) : ±9.5 Sv
- Kiel-ORCA (Kiel University – Ocean model configuration of the NEMO ocean code) : ±10.5 SV
- MPI (Max Planck Institute) : ±12.5 SV
- KNMI-MICOM (Royal Neterlands Meteorological Institute – Miami Isopycnal Coordinate Ocean Model) : ±10 SV

The pattern of Root Mean Square tends to be larger at the surface layer due the dynamics of the forcing at this layer. RMS is also larger at 0-40⁰E longitude comes from the Mediterranean Sea, even in depth layers of 1000m and 5000m. It is more difficult for the model to reproduce data from this sea because of its semi enclored position gives it unique range of temperature and salinity values and different forcing.

Out of the 5 simulation experiment, run 5 holds the best run as it has the least difference to WOA09.

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