Improvements in LICOM2. Part I: Vertical Mixing

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ABSTRACT

Better computational stability is achieved in an improved version of the National Key Laboratory of Numerical Modeling for Atmospheric Sciences and Geophysical Fluid Dynamics (LASG)/Institute of Atmospheric Physics (IAP) Climate Ocean Model, version 2 (LICOM2, the standard version), after improvements to the implementations of the vertical mixing, mesoscale eddy parameterization, and bottom drag schemes. The large warm biases of LICOM2 in the western Pacific Ocean and eastern Indian Ocean warm pool and on the east coast of the Pacific Ocean are significantly improved. The salinity bias in the tropical Pacific Ocean related to the warm bias of the warm pool is also alleviated. The simulation of the Atlantic meridional overturning circulation is improved because of enhanced vertical mixing in the high latitudes of the North Atlantic Ocean. The new version also presents a stronger Deacon cell, and thus a more powerful Antarctic Circumpolar Current that is closer to the observation, due to weaker southward mesoscale eddy transport in the Southern Ocean.

1. Introduction

An ocean general circulation model (OGCM) is composed of a dynamical core and several physical parameterization schemes. The dynamical core includes horizontal and vertical grids; horizontal and vertical discretization schemes; an advection scheme used for advecting tracers, such as potential temperature and salinity; a horizontal diffusion scheme; and a time stepping scheme. The necessary physical parameterizations in an OGCM are vertical mixing/turbulence schemes for the momentum and tracer equations, a bottom boundary layer scheme, a shortwave radiation penetration scheme, and a convective adjustment scheme (in a hydrostatic ocean model), among others. Thus, an OGCM is a complicated system, involving tens of thousands of lines of codes. Numerical stability for a long integration and good performance to reproduce observations are the two main targets that many OGCM developers pursue. The realization of the two main targets is determined by both scientific design and technical implementation of the scientific design in the model. The National Key Laboratory of Numerical Modeling for Atmospheric Sciences and Geophysical Fluid Dynamics (LASG)/Institute of Atmospheric Physics (IAP) Climate Ocean Model, version 2 (LICOM2; Liu et al. 2012), is the standard version of the model. LICOM2 has reasonable basic performance in terms of simulating sea surface temperature
(SST), sea surface salinity (SSS), and surface mixed layer depth (Liu et al. 2012). It serves as the ocean component of the Flexible Global Ocean–Atmosphere–Land System Model gridpoint, version 2.0 (FGOALS-g2; Li et al. 2013), and FGOALS, second spectral version (FGOALS-s2; Bao et al. 2013), part of phase 5 of the Coupled Model Intercomparison Project (CMIP5). The CMIP5 requires an OGCM to run for more than several thousand years, which is a great challenge to a model’s numerical stability. LICOM2 experienced some computational instability problems in some of its CMIP5 experiments, which led us to reexamine the technical implementation in the model.

A longitude/latitude grid is used by LICOM2. With the purpose to enlarge the time step and to increase the model’s numerical stability, a process-splitting-based finite difference dynamical core, an artificial island from 87.5° to 90°N and a 1–2–1 low-pass filter were adopted in the high latitudes of the model domains of LICOM2. Since an explicit algorithm was used for the barotropic equations, the fast-moving surface gravity wave and the convergent zonal grid together limited the time step of the barotropic equations. When encountering most of the computational instability problems in running CMIP5 experiments, the first two measures were reducing the time step and increasing the frequency of the low-pass filtering—that is, instabilities were attributed to the problems in the dynamical core. In fact, not only the dynamical core but also other factors, such as steep terrain, unreasonable parameters in the parameterization schemes of physical processes, and incorrect implementations of numerical algorithms and parameterization schemes, may have led to computational instabilities in an OGCM. This study identifies and then removes the sources of computational instability and to increase the time step, the region from 87.5° to 90°N in the Arctic Ocean (i.e., the North Pole and two neighboring zonal circles) is treated as an artificial island, and a 1–2–1 low-pass filter is applied to the latitude higher than 65°. The Arakawa B grid (Arakawa and Lamb 1977) is adopted, and the northemmost point is a scalar point for temperature/salinity. There are 30 layers in the vertical direction, with a 10-m resolution in the top 150 m. All variables, except for vertical velocity, which is defined on the integer levels, are defined in the middle of the integer levels (simply half levels hereinafter). The standard version adopts a splitting algorithm to separately solve the barotropic, baroclinic, and thermal/haline equations, and the leapfrog scheme with the Asselin time filter for time integration.

The physical parameterization schemes of LICOM2 are as follows. The vertical mixing scheme used in the standard version is the second-order turbulence scheme (Canuto et al. 2001, 2002, 2010), which includes shear-driven, wind-driven, double diffusion, and internal-wave-breaking-driven mixing. Three-dimensional (3D) diffusivities for heat, salt, and momentum at the interfaces of the 30 vertical layers are generated by this scheme. Note that the top and bottom interfaces of these 30 layers do not need vertical diffusivities, and the variables, such as temperature, salinity, and density, must be interpolated to the integer levels from the half levels. The mesoscale eddy parameterization from Gent and McWilliams (1990) is used. To keep the large vertical gradient of density in the surface mixed layer, the mesoscale eddy parameterization coefficients are tapered there (Large et al. 1997; Liu et al. 2012). For the short-wave radiation penetration, the horizontally constant double-exponent scheme from Paulson and Simpson (1977) is adopted. Details on other parameterization schemes used in LICOM2 can be found in Liu et al. (2012).

We use both LICOM2 (the standard version) and its improved version (named “LICOM2_imp”) in this paper. LICOM2_imp has corrected some physical parameterization schemes and coding issues in LICOM (Liu et al. 2012). A detailed description of the improvements

2. Model description, experiment design, and datasets used

a. Model description

The dynamical core of LICOM2 can be described as follows. LICOM2 utilizes the longitude/latitude grid, with an even 1° zonal resolution (i.e., 360 grid points in the zonal direction) and an approximate 1° meridional resolution that is refined to 0.5° in the equatorial region (i.e., 196 grid points in the meridional direction). The model domain is from 78.5°S to 87.5°N. To overcome the computational instability and to increase the time step, the region from 87.5° to 90°N in the Arctic Ocean (i.e., the North Pole and two neighboring zonal circles) is treated as an artificial island, and a 1–2–1 low-pass filter is applied to the latitude higher than 65°. The Arakawa B grid (Arakawa and Lamb 1977) is adopted, and the northemmost point is a scalar point for temperature/salinity. There are 30 layers in the vertical direction, with a 10-m resolution in the top 150 m. All variables, except for vertical velocity, which is defined on the integer levels, are defined in the middle of the integer levels (simply half levels hereinafter). The standard version adopts a splitting algorithm to separately solve the barotropic, baroclinic, and thermal/haline equations, and the leapfrog scheme with the Asselin time filter for time integration.

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We use both LICOM2 (the standard version) and its improved version (named “LICOM2_imp”) in this paper. LICOM2_imp has corrected some physical parameterization schemes and coding issues in LICOM (Liu et al. 2012). A detailed description of the improvements

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a. Model description
is included in the appendix for reference. Briefly, the improvements and corrections are related to the implementations of the vertical mixing scheme, the bottom drag, and the mesoscale eddy parameterization scheme. LICOM2_imp has much better computational stability than LICOM2. In a 400-yr stable integration, LICOM2_imp allows inclusion of all latitudes except the North Pole, but LICOM2 cannot. The main reasons behind the computational instability in LICOM2 are the incorrect interpolation and solving process in the second-order turbulence model (see the appendix). The simple mechanism is as follows: the incorrect linear interpolation and solving process led to unreasonable or incorrect vertical diffusivity, and resulted in computational instability.

b. Experiment design

To evaluate the performance of LICOM2_imp, two experiments are carried out: Exp CTRL using LICOM2 and Exp IMPV0 using LICOM2_imp, which has all the improvements (see the appendix). All other settings for the two experiments are the same. Both experiments are integrated for 400 years. The mesoscale eddy parameterization coefficients, which include the isopycnal mixing and thickness mixing coefficients (Liu et al. 2012), are both set to 500 m$^{-2}$ s$^{-1}$. The time steps for the barotropic, baroclinic, and thermal/haline modes are 60, 720, and 2880 s, respectively. The Asselin filtering coefficients for the center time step are set to 0.57 in all three modes. An artificial island with the region from 87.5° to 90°N exists in both experiments. Polar filtering is applied to the barotropic modes once every day and to the baroclinic modes once every 4 days. Each experiment starts from the same climate mean state of potential temperature and salinity observations used for the initialization are the World Ocean Atlas 2005 (WOA05; Locarnini et al. 2006; Antonov et al. 2006).

To evaluate the basic performance of LICOM2_imp, the following datasets are used. WOA05 SST and SSS datasets are used for validating model SST and SSS. To study the SST seasonal cycle in the equatorial Pacific Ocean, the SST dataset developed by Hurrell et al. (2008) from 1980 to 2005 is adopted. The observed zonal currents in the equatorial Pacific Ocean by Johnson et al. (2002) are also used. The observed Atlantic meridional overturning circulation (AMOC) at 26.5°N from April 2004 to October 2007 by the Rapid Climate Change–Meridional Overturning Circulation and Heatflux Array (RAPID–MOCHA) program (Cunningham et al. 2007; Johns et al. 2011) is taken as the reference value. The observed value of the Antarctic Circumpolar Current (ACC) across the Drake Passage by Cunningham et al. (2003) is used. The estimation of poleward heat transport by Ganachaud and Wunsch (2003) is taken. The mean dynamic ocean topography (MDOT) dataset from 1992 to 2002 (Maximenko et al. 2009) is used as the observation for sea surface height (SSH).

c. Datasets

The external forcing is the long-term-mean annual cycle (1979–93) based on the 24-h forecast of the 40-yr European Centre for Medium-Range Weather Forecasts Re-Analysis (ERA 40; Simmons and Gibson 2000). The potential temperature and salinity observations used for the initialization are the World Ocean Atlas 2005 (WOA05; Locarnini et al. 2006; Antonov et al. 2006). To evaluate the basic performance of LICOM2_imp, the following datasets are used. WOA05 SST and SSS datasets are used for validating model SST and SSS. To study the SST seasonal cycle in the equatorial Pacific Ocean, the SST dataset developed by Hurrell et al. (2008) from 1980 to 2005 is adopted. The observed zonal currents in the equatorial Pacific Ocean by Johnson et al. (2002) are also used. The observed Atlantic meridional overturning circulation (AMOC) at 26.5°N from April 2004 to October 2007 by the Rapid Climate Change–Meridional Overturning Circulation and Heatflux Array (RAPID–MOCHA) program (Cunningham et al. 2007; Johns et al. 2011) is taken as the reference value. The observed value of the Antarctic Circumpolar Current (ACC) across the Drake Passage by Cunningham et al. (2003) is used. The estimation of poleward heat transport by Ganachaud and Wunsch (2003) is taken. The mean dynamic ocean topography (MDOT) dataset from 1992 to 2002 (Maximenko et al. 2009) is used as the observation for sea surface height (SSH).

3. Preliminary evaluation of LICOM2_imp

In this section, we evaluate the performance of LICOM2_imp, in terms of SST, SSS, SSH, global meridional overturning circulation (GMOC), AMOC, heat transports in the global ocean and the Atlantic Ocean, the ACC, and zonal currents and the seasonal cycle of the SST anomaly in the equatorial Pacific Ocean. In addition, the contributions of the improvements of LICOM2_imp (see the appendix) are evaluated through comparing the results of Exp CTRL and Exp IMPV0. To prevent the analysis from being influenced by the decadal and interannual oscillations, a 50-yr average of the results from model years 351–400 is done if not explicitly noted.

a. SST and SSS

Figures 1a and 1b present the biases of annual-mean model SST relative to the WOA05. The global tropical SST exhibits a great degree of spatial variability. Large warm area-averaged biases existing in the western Pacific Ocean warm pool (20°S–20°N, 120°E–180°E) are reduced from 0.44°C in CTRL to 0.27°C in IMPV0. Both experiments have a cold bias in the central tropical Pacific Ocean and a warm bias in the eastern tropical Pacific Ocean, but IMPV0 shows a smaller area-averaged bias over these regions (20°S–20°N, 180°–60°E) (0.30°C) than CTRL (0.34°C). In the tropical Indian Ocean (20°S–20°N, 30°–120°E), the area-averaged biases in CTRL (0.25°C) are also improved in IMPV0 (0.20°C). The area-averaged biases in the tropical Atlantic Ocean (20°S–20°N, 60°W–30°E) are 0.48° and 0.40°C for both experiments. As shown in Fig. 1c, Exp IMPV0 has larger vertical thermal diffusivities (time averaged over 10-m depth) than Exp CTRL in nearly all the tropics, which reduce the warm biases in nearly all the tropics. Besides, the area-averaged bias over the region south of 60°S is 0.67°C in CTRL, which is larger than 0.64°C in Liu et al. (2012). This is because we do not use any restoring on SST in the sea ice regions, while Liu et al. (2012) did at the time scale...
of 30 days. With the improvements (see the appendix), LICOM2_imp slightly reduces these biases to about 0.62°C in IMPV0. In a word, large warm biases in CTRL are significantly reduced in IMPV0 in the western Pacific Ocean and eastern Indian Ocean warm pool region, and on the east coast of the Pacific Ocean. Moreover, the cold biases in the central tropical Pacific Ocean and tropical Atlantic Ocean are somewhat improved.

Figure 2 presents the biases of annual-mean SSS relative to the WOA05. Both experiments give reasonable
SSS. The biases in most regions are lower than 0.5 psu, which are comparable to those in Liu et al. (2012). However, a few large biases could be found in CTRL, such as the positive biases in the tropical Pacific Ocean and negative biases in the Tasman Sea. Note that the western part of positive biases in the tropical Pacific Ocean of CTRL is related to the warm biases in the western Pacific Ocean warm pool. LICOM2_imp alleviates these biases in IMPV0.

b. Meridional overturning circulation and heat transports

Figure 3 presents the GMOC and AMOC from the two experiments. The meridional overturning circulations in Fig. 3 are the sums of resolved transports and unresolved (parameterized) mesoscale eddy isopycnal transports. In both experiments, the North Atlantic Deep Water (NADW) is located between the depths of 500 and 3000 m and north of 34°8′S (Figs. 3b and 3d). The maximum NADW in Exp CTRL is 16.49 Sv (1 Sv = 10⁶ m³ s⁻¹) at the depth of 731.40 m and 37°N in Fig. 3b, while that in Exp IMPV0 is 16.97 Sv at the depth of 1021.67 m and 39°N (Fig. 3d). The strengthening of the NADW in IMPV0 is mainly attributed to vertical-mixing-scheme-related issues (see the appendix)—that is, the replacement of the nonconvergent iteration of the analytic method and the corrected vertical interpolation before the vertical mixing scheme being performed. Accompanying the improvements in the vertical mixing, the mixed layer depths of Exp IMPV0 shows a significant enhancement in the Labrador Sea (Fig. 4; compare to CTRL), which further leads to a stronger NADW. It should be noted that, although the difference of mixed layer depth between the two experiments in the Labrador Sea is smaller than that in the Irminger Sea, the maxima of the mixed layer depth of both experiments are in the Labrador Sea. All these implementations, especially the analytic solving of the nonlinear equation in the second-order turbulence model, lead to stronger and more reasonable vertical mixing in the high latitudes.

The RAPID–MOCHA program has provided the vertical profile of the AMOC at 26.5°N (Cunningham et al. 2007; Johns et al. 2011) (solid curve in Fig. 5). The observed maximum transport has the strength and
variability of about $18.5 \pm 4.9$ Sv (Johns et al. 2011). The maximum AMOC transports at 26.5°N are 14.15 and 15.32 Sv for Exps CTRL and IMPV0, respectively (dashed curves in Fig. 5). However, the depth of the simulated maximum in CTRL (731.40 m) is shallower than that in the observation (1000 m). The observed NADW extends to a depth of about 4500 m, while that in each experiment extends to a depth of only <3000 m (see the zero crossing of the AMOC), which requires further investigation and improvement.

In the deepest part of the ocean (between 3000 and 4500 m), there exists the Antarctic Bottom Water (AABW). The AABW originates from the cold descending water along the Antarctic continent in the boreal summer. Since the Antarctic continent is surrounded by the Atlantic Ocean, Indian Ocean, and Pacific Ocean, the AABW contributes to both the GMOC and AMOC (Fig. 3). For the Atlantic Ocean, the AABW in Exp CTRL has its maximum (negative) value of 1.79 Sv (Fig. 3b), while that in Exp IMPV0 has its maximum (negative) value of 3.14 Sv (Fig. 3d). Here a negative AABW refers to a northward-moving water mass. For the global ocean, Exp CTRL has the maximum (negative) value of 1.50 Sv (Fig. 3a), while that in Exp IMPV0 has the maximum (negative) value of 2.29 Sv (Fig. 3c). Both modeled values for the AABW in the global ocean are outside the range of available observations (Orsi et al. 2002)—that is, $8.1 \pm 2.6$ Sv (negative). According to the values for the AABW.

![Fig. 3. Annual-mean, zonally integrated overturning streamfunction (Sv) in the (a),(c) global ocean and the (b),(d) Atlantic Ocean from (a),(b) Exp CTRL and (c),(d) Exp IMPV0. The model output is taken from model years 351–400.](image-url)
in the Atlantic Ocean and the global ocean in the two experiments, the AABW in the Pacific Ocean and the Indian Ocean is moving in a southward direction, which is opposite to that of observation.

In the Southern Ocean (about 40°S–60°S), there exists the Deacon cell (Figs. 3a and 3c), which is a shallow, wind-driven overturning cell. The Deacon cell in each experiment has a maximum at the depth between the surface and 100 m (approximately the thickness of an Ekman layer). However, there is a marked difference in the strength of the Deacon cell of the two experiments—that is, 29.72 and 36.65 Sv for Exps CTRL and IMPV0, respectively. Theoretically, the Deacon cell is mainly modulated by the northward Ekman transport (induced by the westerly wind) and the parameterized eddy-induced southward fluxes (Farneti and Delworth 2010). Since the wind stress forcing for the two experiments is the same, it is inferred that the parameterized southward eddy transport in Exp CTRL is stronger than that in Exp IMPV0. Because the mesoscale parameterization coefficients for the two experiments are the same, the baroclinicity of temperature/salinity structure of Exp CTRL in the Southern Ocean is stronger than that of Exp IMPV0, which may be partially explained by the changes of the vertical mixing scheme.

As in Liu et al. (2012), the poleward heat transport in this paper also includes three parts—that is, the resolved meridional circulation, unresolved (parameterized) mesoscale transport, and diffusivity-driven transport. According to the estimations at limited latitudes by Ganachaud and Wunsch (2003, see their Table 1), both experiments (Fig. 6b) underestimate the heat transport at nearly all these latitudes except that Exp CTRL overestimates the transport at 47°N in the Atlantic Ocean (Fig. 6a). In the Southern Ocean (40°–80°S), the poleward heat transport in Exp CTRL (maximum: 0.70 PW at 51°S; Fig. 6a) is larger than that in Exp IMPV0 (maximum: 0.46 PW at 51°S; Fig. 6c), which is explained by the increased parameterized mesoscale eddy transport (southward) (see the discussion about the Deacon cell in section 3c). In the Northern Hemisphere of the global ocean, the heat transport in CTRL peaks at 22°N with 1.36 PW (Fig. 6a), while that in IMPV0 peaks at 27°N with 1.45 PW (Fig. 6c).
In the northern part of the Atlantic Ocean, the maximum heat transport in CTRL is 0.77 PW (Fig. 6b), while that in IMPV0 is 0.85 PW (Fig. 6d). The increased AMOC due to the enhanced vertical mixing in the North Atlantic Ocean is the leading cause for the strengthening of poleward heat transport in the Northern Hemisphere.

c. The ACC and SSH

The annual cycle of the ACC across the Drake Passage in each experiment is shown in Fig. 7. The ACC in each experiment has its maximum in September, which is winter in the Southern Hemisphere. Note that the annual-mean ACC simulated by Liu et al. (2012) is about 81 Sv. The annual-mean ACC transports in Exps CTRL and IMPV0 are 91.26 and 108.93 Sv, respectively. Since the observed value for the ACC is 134 Sv (Cunningham et al. 2003), the ACC simulated by LICOM2_imp outperforms the LICOM2 greatly. To understand the difference of the ACC between CTRL and IMPV0 in more detail, the distribution of SSH for the two experiments is presented in Fig. 8. According to the MDOT (Maximenko et al. 2009; Fig. 8a), both experiments (Figs. 8b and 8c) capture the basic pattern of SSH. From the difference of SSH between IMPV0 and CTRL (Fig. 8d), it is found
that CTRL has a higher SSH near 60°S and a lower SSH near 40°S. These differences lead to a weaker SSH gradient in the Southern Ocean in CTRL, which is consistent with a lower ACC across the Drake Passage in the model.

d. Zonal currents and SST seasonal cycle in equatorial Pacific Ocean

In the observation of Johnson et al. (2002), the equatorial zonal currents (Fig. 9a) present a westward flow in the upper 50 m with its maximum value of 0.38 m s⁻¹ (position: 5-m depth and 140°W), which is the South Equatorial Current (SEC), and an eastward flow (ascending toward east) between 50 and 200 m with its maximum value of 1.11 m s⁻¹ (position: 85-m depth and 125°W), which is the Equatorial Undercurrent (EUC). Both experiments (Figs. 8b and 8c) capture the basic pattern of the zonal currents. However, the two models overestimate the SEC (CTRL: 0.96 m s⁻¹; IMPV0: 0.91 m s⁻¹) and underestimate the EUC (CTRL: 0.44 m s⁻¹; IMPV0: 0.45 m s⁻¹). Besides, both CTRL and IMPV0 present the same positions for the velocity maxima—that is, 5 m and 136°W for the SEC, and 105 m and 130°W for the EUC. To be consistent with the following additional sensitivity experiments, the analysis period for the zonal currents is set to years 391–400.

The important role of eddy viscosity on the EUC was emphasized (Maes et al. 1997; Large et al. 2001; Richards and Edwards 2003). And it was suggested that an eddy viscosity of the order 10³ m² s⁻¹ is required to simulate the structure of the EUC. The eddy viscosity used in both models is 10⁴ m² s⁻¹. An additional experiment is carried out, with the same configuration as Exp IMPV0 but with a reduced eddy viscosity of 3 × 10³ m² s⁻¹. This low eddy viscosity experiment is initialized using the output of IMPV0 at the end of year 390. It is found that the low-eddy-viscosity experiment (Fig. 9d) captures the maximum strength of the SEC (maximum: 0.53 m s⁻¹) and EUC (maximum: 0.84 m s⁻¹) with smaller biases.

![FIG. 7. Annual cycle of the net volume flux across the Drake Passage (56°–69°S along 69.5°W; Sv) from Exp CTRL (blue) and Exp IMPV0 (red), where the positive sign is for eastward transport. The model output is taken from model years 351–400.](image_url)

![FIG. 8. Annual-mean SSH (m) from the (a) observation, (b) Exp CTRL, (c) Exp IMPV0, and (d) SSH difference between IMPV0 and CTRL. The MDOT dataset (Maximenko et al. 2009) from 1992 to 2002 is used as the observation of SSH, and the model output is taken from model years 351–400.](image_url)
Also, the positions of the two maxima are improved, which are 5 m and 142°W for the SEC and 85 m and 126°W for the EUC.

The seasonal cycle of the SST anomalies in the equatorial Pacific Ocean (2°S–2°N) is given in Fig. 10. As shown by the observation (Hurrell et al. 2008; Fig. 10a), the SST has a warm core (2.94°C) in March and a cold core (2.19°C) in September at 95°W (eastern Pacific Ocean). The observed seasonal cycle is well reproduced in both CTRL (Fig. 10b) and IMPV0 (Fig. 10c). However, the warm core in each experiment lags the observation by one month, while the cold core leads the observation by one month. Besides, the amplitude of the warm core is underestimated in the models (CTRL: 2.09°C; IMPV0: 2.05°C), while that of the cold core is overestimated (CTRL: 2.59°C; IMPV0: 2.63°C).

### 4. Summary and conclusions

Efforts have been made to improve the numerical stability and basic performance of LICOM2. Through studying scientific design and checking technical implementations of the model, the sources of numerical instability have been eliminated. An improved version, LICOM2_imp, with better long-term integration stability is obtained. The major improvements of LICOM2_imp over LICOM2 are the proper implementations of the vertical mixing scheme, the bottom drag at the equator, and the mesoscale eddy parameterization. Two 400-yr experiments have been carried out to study the contributions of these improvements. The annual mean fields of the last 50 years of the experiments are used to evaluate model performance. LICOM2_imp outperforms...
LICOM2 in simulating the annual-mean SST and SSS, the AMOC, and the ACC. Most of these improvements can be attributed to the changes in the vertical mixing scheme. Besides, both versions of LICOM2 have comparable performance in simulating the zonal currents and seasonal cycles of SST anomalies in the equatorial Pacific Ocean.

Among the four improvements, the numerical improvement made to the Canuto scheme (Canuto et al. 2001, 2002, 2010) is the most important one. First, the iteration method is usually applied to vertical mixing schemes for obtaining solutions, including the Canuto scheme and the K-profile parameterization (KPP) scheme (Large et al. 1994). This method, however, suffers from unexpected bad convergence in some special cases (especially on the computational platform of Tansuo100 of Tsinghua University), which is a problem hidden in the implicit vertical mixing computation and therefore hard to identify. It is due to this implicit problem that the original Canuto scheme gets an incorrect solution and produces weak vertical mixing (see Fig. 1c). This problem affects the basic performance of the ocean models that use the Canuto scheme. LICOM2 is one of these models suffering from the weak vertical mixing by the Canuto scheme. Therefore, the identification of the problem and the numerical improvement on it may not only benefit the basic performance of LICOM2 but also benefit that of other ocean models with the Canuto scheme. In particular, the improved Canuto scheme presented in this manuscript provides a new choice for other ocean models on vertical mixing schemes.

Second, similar to the Canuto scheme, the KPP scheme, a commonly used vertical mixing scheme, also employs the iteration methods to obtain the solution, and therefore may also be subject to the aforesaid problem. Efforts to solve this problem can refer to the numerical improvements made to the vertical mixing in LICOM2.

Finally, the numerical improvement made to the vertical mixing in LICOM2 gives us a clue to the overestimation of the AMOC by FGOALS-g2, a coupled climate system model with LICOM2 as its ocean component. As Huang et al. (2014b) pointed out, the obvious overestimation of AMOC by FGOALS-g2 was attributed

Fig. 10. Seasonal cycle of SST anomalies (°C) in the equatorial Pacific Ocean (2°S–2°N) for (a) the observation, (b) Exp CTRL, and (c) Exp IMPV0. The observation is from Hurrell et al. (2008) for the period of 1980–2005, and the model output is taken from model years 351–400.
to an artificial background vertical mixing added to the ocean component LICOM2. However, no reason about the use of the artificial background vertical mixing was given in previous works. This study reveals that the implicit bad convergence problem hidden in the Canuto scheme results in weak vertical mixing in LICOM2, which gives a reasonable explanation that the artificial background vertical mixing is used to reduce the underestimation of vertical mixing by the Canuto scheme. Meanwhile, the numerical improvement made to the Canuto scheme leads to an enhancement of vertical mixing in LICOM2, which indicates a possibility to abandon this artificial background vertical mixing in FGOALS-g2.

The incorporation of the four improvements achieved in this study, especially the numerical improvement made to the Canuto scheme, into FGOALS-g2 will hopefully reduce some of its existing biases, including the overestimation of the AMOC.

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APPENDIX

Improvements over LICOM2

a. Problems in LICOM2

Even with the artificial island and polar filtering, the standard version of the model, LICOM2, still experienced computational instability in some experiments of CMIP5, which led to our careful examination of the codes. As a result, four unreasonable or incorrect implementations of some physical parameterization schemes are identified.

1) In the second-order turbulence model from Canuto et al. (2001, 2002) in the standard version, the vertical mixing coefficients are functions of the dynamical time scale of the turbulence. The nonlinear equation solved by the recommended Wegstein iteration method in Canuto et al. (2001, 2002) has been replaced by a quasi-analytic method. The iteration method is found to be nonconvergent (on the Tansuo100 computational platform of Tsinghua University) when the Richardson number is very small, which leads to incorrect weaker vertical mixing. If we update the scheme to Canuto et al. (2010), a cubic equation with one unknown needs to be solved to obtain the time scale. We replace the iteration method by an analytic method (Dunham 1990; Fan 1989). The details about the analytic method for the cubic equation (Canuto et al. 2010) are as follows.

Let the cubic equation have the following expression:

$$ax^3 + bx^2 + cx + d = 0 \quad (a \neq 0), \quad (A1)$$

where $x$ is the unknown and all the others are known real coefficients. Mark

$$B = \frac{bc}{6a^2} - \frac{b^3}{27a^3} - \frac{d}{2a} \quad \text{and} \quad C = \frac{c}{3a} - \frac{b^2}{9a^2}; \quad (A2)$$

three roots of the equation are obtained—that is,

\begin{align*}
    x_1 &= -\frac{b}{3a} + \sqrt[3]{B + \sqrt{B^2 + C^2}} \\
    &\quad + \sqrt[3]{B - \sqrt{B^2 + C^2}}, \quad (A3) \\
    x_2 &= -\frac{b}{3a} + \frac{-1 + \sqrt[3]{i}}{2} \sqrt[3]{B + \sqrt{B^2 + C^2}} \\
    &\quad + \frac{-1 - \sqrt[3]{i}}{2} \sqrt[3]{B - \sqrt{B^2 + C^2}}, \quad \text{and} \quad (A4) \\
    x_3 &= -\frac{b}{3a} + \frac{-1 - \sqrt[3]{i}}{2} \sqrt[3]{B + \sqrt{B^2 + C^2}} \\
    &\quad + \frac{-1 + \sqrt[3]{i}}{2} \sqrt[3]{B - \sqrt{B^2 + C^2}}. \quad (A5)
\end{align*}

It is the real root presented in (A3) that we are trying to obtain.

2) In the mesoscale eddy parameterization scheme used to partly compensate for the low horizontal resolution of the model, an unmatched meridional spacing of the $\nu$ grid (i.e., the grid for the meridional velocity) on the Arakawa B grid was utilized, which was mistakenly set to be the same as that of the $T$ grid (i.e., the grid for temperature). Using the meridional spacing of the $\nu$ grid does not cause a problem by itself, if the model employs a uniform grid. However, the meridional grid in LICOM2 varies to a finer one in the equatorial region. This may lead to errors in the equatorial region.

3) The effect of the Coriolis force on the bottom drag along the equator is not correctly represented in the standard version. The deflection by the Coriolis effect should be zero at the equator. However, the bottom drag along the equator was incorrectly considered in
the Northern Hemisphere and, as a result, the Coriolis effect deflected the ocean current to the right on the equator.

4) An incorrect interpolation formula was applied to the vertical mixing scheme at the baroclinic step. As mentioned in the introduction, when the vertical mixing scheme is called at the baroclinic step, the temperature, salinity, and density must be interpolated to the integer levels from the half levels in the vertical direction. Assume that the scalar variable $T$ at two neighboring half levels are $T_{k-1/2}$ and $T_{k+1/2}$, and $D_{k-1/2}$ and $D_{k+1/2}$ are the thicknesses from the $(k-1)$th level to the $k$th level and from the $k$th level to $(k+1)$th level, respectively, where $k$ is the index for the integer levels and its direction is downward as $k$ increases, while $[k+(1/2)]$ is the index for the half levels between two neighboring integer levels with indices of $k$ and $(k+1)$. The term $T_k$, the approximate value of $T$ at the $4$th level, can be obtained using the linear interpolation

$$T_k = T_{k-1/2} D_{k-1/2} + D_{k+1/2}$$

$$+ T_{k+1/2} D_{k-1/2} + D_{k+1/2}$$

$$= T_{k-1/2} + (T_{k+1/2} - T_{k-1/2})$$

$$\times \frac{D_{k-1/2}}{D_{k-1/2} + D_{k+1/2}}.$$  \hspace{1cm} (A6)

However, in the standard version $T_k$ was incorrectly calculated as

$$T_k = T_{k-1/2} - (T_{k+1/2} - T_{k-1/2})$$

$$\times \frac{D_{k-1/2}}{D_{k-1/2} + D_{k+1/2}}.$$  \hspace{1cm} (A7)

b. Solutions

After the above-mentioned problems have been identified, improvements are obtained:

1) Instead of the original iteration method, an analytic method is employed to solve the nonlinear equation in the second-order turbulence module, which correctly produces stronger vertical mixing (Canuto et al. 2001, 2002, 2010).

2) The unmatched meridional spacing of the $v$ grid in the mesoscale eddy scheme is corrected.

3) The Coriolis effect on the bottom drag along the equator is removed.

4) The error in the linear interpolation formula for the vertical mixing scheme at the baroclinic step is removed, and the correct interpolation formula (A7) is employed. The influence of this correction on the vertical mixing cannot be neglected.

REFERENCES


