Resolution sensitivity of tropical turbulent air-sea heat fluxes and precipitation in NorESM models

**Abstract**

We evaluated the ¼° model NorESM1.3 in which the well-known “double-ITCZ problem” in the Pacific is mitigated. However, excessive precipitation is produced in the northern branch of the ITCZ. The excessive precipitation is consistent with overevaluated latent heat flux in the tropical ocean. Further analysis shows that in NorESM1.3, the latent heat flux is too sensitive to the surface wind. The increased sensitivity in the ¼° model is partly contributed by small-scale air-sea interaction. The sensitivity of latent heat flux to surface wind, with the scale finer than 2.5°, is up to 40 (Wm-2 / ms-1), which is almost twice of that with scale coarser than 2.5°. This study identified the extra air-sea interaction resolved by higher-resolution models and the impact on the climate state of the model simulation. This can help to improve heat fluxes parameterization and correct the related model bias.

**1. Introduction**

Turbulent air-sea heat fluxes (THFs), i.e., the latent and sensible heat flux, influence the variability and climatology of the atmospheric and oceanic processes at all scales (). Surface latent heat flux is the heat extracted from the ocean when seawater evaporates. This heat is released to the atmosphere, when the vapor condenses, forming clouds. Likewise, sensible heat flux is the heat extracted from the ocean associated with an air-sea temperature difference. The THFs are crucial drives of the variability of many ocean processes, such as deep convection in the subpolar waters (Holdsworth and Myers 2015). THFs also have strong impact on the atmospheric circulation by heating or cooling the lower atmosphere.

As seawater evaporates, the ocean surface cools; and when the moisture later condenses into cloud droplets, the heat is released, warms the atmosphere. This moistening and warming make the air buoyant, driving low-level baroclinicity and atmospheric convection, causing wind convergence at the surface and divergence aloft. Patterns of surface heat fluxes also affect large-scale atmospheric circulation patterns, with deep convection over the thermal equator forming the upward branch of the “Hadley Cell” that drives trade winds. The rising branch is aligned with the surface heat of the atmosphere associated with stronger surface heat fluxes. The surface wind patterns, e.g., the easterly trades in the tropics, in turn drive the ocean general circulation. Therefore, the THFs represent the direct communication between the ocean and atmosphere.

Considering the importance of THFs, its accurate estimation is critical for a wide range of weather- and climate related studies. In reality, however, quantifying these THFs, is a big challenge. In observation, the directly measured THFs tend to be idealized, infrequent, and highly localized. These observed data are not sufficient to build global distributed THFs (Brunke et al. 2003, 2011). The global distributed THFs are normally estimated by bulk formulars. The bulk formulars parameterized THFs with surface state variables, which are easily measurable and widely available, and a bulk transfer coefficient. The bulk transfer coefficients typically depend on the wind speed, the stability of the atmospheric surface layer, and the adjustment of atmospheric scalars of the standard height through the flux-profile relationships. Originally, the bulk parameterizations are developed for voluntary observing ship and buoy data, but it is also widely used to compute THFs by surface state variables derived from reanalysis datasets (Josey et al. 2013) and numerical models (Large and Yeager 2009).

In the numerical models, e.g., NorESM, the THFs are generally estimated using air-sea differences in the mean “bulk” state variables simulated at the surface and at some height within the surface layer. The bulk aerodynamic method links the turbulent fluxes to mean air-sea velocity, temperature and humidity difference using transfer coefficients:

Where is specitic heat at constant pressure; and are the fluctuating along-wind and cross-wind velocity components, respectively; and is the fluctuating potential temperature; , , and are the bulk transfer coefficients (known as drag coefficient) for momentum, latent heat and sensible heat, respectively; S is the scalar wind speed relative to the ocean surface current. , , , and are the air-sea differences in the along wind, crosswind, specific humidity and potential temperature, respectively.

The bulk transfer coefficients, , are computed at a height :

Where is von kármán’s constant. indicates the roughness length for momentum, evaporation, or heat, respectively. The momentum roughness length () varies with the surface fluxes over oceans, and the roughness of evaporation and heat are taken as constants. The transfer coefficients also depend on the integrated flux profiles (), which themselves depends on the stable conditions of atmospheric surface layer. The stable conditions vary with the surface fluxes over oceans.

The parameterizations of bulk formular are widely used in numerical models with the parameters tuned to fit each individual model. However, for a given model, the differences in THFs arose and increased when the model resolutions increased to a certain extent. This implies that the parameterizations suit for low resolution models are no longer properly to be used in the higher resolution versions. Mathematically, numerical models use essentially approximated equations, and truncation errors decrease with grid sizes. Thus, higher-resolution models are normally expected to give better simulations. However, the multi-model evaluation by Vannière et al. (2018) shows a strengthening of the global hydrological cycle with increased model resolution by an increase in surface latent-heat flux driven by more outgoing long-wave radiation and less out-going short-wave radiation at the top of the atmosphere. This is because evaporation and precipitation are more sensitive to net surface short-wave radiation (Wu et al., 2010, 2013). This is also found in Ding et al., (2021), which found that the air-sea latent heat fluxes are enhanced along Kuroshio. This is due to a stronger near-surface wind and drier surface air in high-resolution models.

Reducing biases in THfs is important for improving long-term weather and climate predictions.

The question is how to adjust the THFs’ parameterization, which works well in low resolution models, to fit the higher resolution models, and would such a change bring the model any closer to observation? As a first step to solve this problem, here, we addressed the sensitivity of computed THFs to model resolution and the impact on hydrological cycle in NorESMs.

This paper is organized as the follows: In section2, we describe the data and methods. The results are shown in section 3 and discussed further in section 4. The conclusions are summarized in section 5.

**2. Data and model experiments**

**2.1 models**

**2.2 method**

**2.3 ERA5**

**3. The results**

**3.1 Biases in the tropical precipitation**

In the tropical pacific, the mean precipitation is concentrated in the Intertropical Convergence Zone (ITCZ). In ERA5 (Fig. 1a), the rain belt of the northern tropical Pacific distributes along 5°N-10°N with the maximum up to 15 mm/day around 200°-250°E. There is also a rain belt in the southern tropical pacific, which extend from western Pacific warm pool south-eastwards towards 20°S, 200°E with the maximum up to 12mm/day.

In the low-resolution NorESM(NorLR), the precipitation is around 10 mm/day in northern tropical Pacific, which is 1/3 less than in ERA5. In the southern hemisphere, the NorLR has a strong and long branch, which resembles its Northern Hemisphere ITCZ. This too zonally elongated southern Pacific rainband in model simulation is known as the double-Intertropical Convergence Zone problem. The double ITCZ problem is a significant and persistent bias existing in the last several generations of climate models (Hwang et al. 2013).

In the super-high-resolution NorESM (NorSR), the double-ITCZ problem is likely to be solved. As shown in Fig.1c, the southern branch of precipitation is much weaker than the northern one and has a shorter south-eastward extension to 200°E. It is very close to the ERA5. Despite the improved double ITCZ problem, excessive precipitation is produced in the northern branch of the ITCZ. The amplitude is almost twice of the observation. This huge bias might be related to the feedback between energy budget and circulation patterns. Therefore, we devoted to understand it in chapter 3.2 the energy budget and in chapter 3.3 the circulation related physical processes.

**3.2 The shift of Bowen Ratio**

Bowen ratio is defined as the ratio of sensible heating flux and latent heating flux (eq.1). The evaporative fraction (EF) in eq.2 is related to the reciprocal of Bowen ratio. EF indicates the relative roles of evaporation in the total turbulent heat flux. A larger Bowen ratio is related to a smaller EF, indicating less water vapor generated by given turbulent heat flux, indicating drier atmosphere and less precipitation.

(1)

(2)

In observation, Bowen ratio in tropical ocean has a large value above 1/11 along ITCZ region. The maximum value, up to 1/10, locates in the warm pool and cold tongue region. Out of the ITCZ region, the tropical Bowen ratio is relatively lower, below 1/15 in ERA5.

In the low-resolution model (NorLR), the Bowen ratio is over estimated. The simulated Bowen ratio is larger than 0.14 over the whole ITCZ region and only a small place has the value close to observation, less than 1/10. This indicates a drier bias of the simulated atmosphere, which is consistent with less precipitation in Fig.1b.

The bias of Bowen ratio in NorLR is largely reduced in NorSR in the view point of amplitude. The values are below 0.1 in most area of the tropical ocean. However, the simulated Bowen ratio in western Pacific and Indian Ocean is much smaller in NorSR related to ERA5. The smaller Bowen ratio indicates an overestimation of evaporation for a given turbulent heat flux. This might be related to the excessive precipitation in Fig.1c.

**3.2.1 equilibrium Bowen ratio**

In the tropical air-sea surface, when the surface and the air at the reference level are saturated, the Bowen ratio approaches the equilibrium Bowen ratio (). In this condition, it is assumed that the flux of moisture from the boundary layer to the free atmosphere is sufficient to just balance the upward flux of moisture from the surface so that the humidity at the reference height is in equilibrium at the saturation value. is inversely proportional to the rate of change of the saturation mixing of water vapor with temperature (eq.3).

(3)

(4)

(5)

The saturation vapor pressure is exponential dependence of the temperature (eq.4 and 5). This indicates that the saturation specific humidity is increased rapidly with temperature, so that equilibrium Bowen ratio decreases exponentially with temperature.

Consider the importance of temperature, we compared the SST in ERA5 (Fig.3a) with the models. As shown in Fig.3b, SST in NorLR is warmer than that in ERA5 in a large region of tropical ocean, however, the Bowen ratio in NorLR is overestimated. This is opposite with the case of equilibrium Bowen ratio above. On the opposite side, the SST in NorSR is slightly colder than ERA5, the Bowen ratio is smaller than in ERA5. This is also inconsistent with the above analysis based on surface temperature. Therefore, it worth to believe that equilibrium Bowen ratio is not a proper hypothesis in both NorLR and NorSR. Thus, we analysed the latent and sensible heat flux separately.

**3.2.2 Latent and sensible heat flux**

As shown in eq.1, the changes in Bowen ratio can be contributed by both sensible and latent heat flux. As shown in Fig.4c, NorLR simulates a larger sensible heat flux, especially in the eastern Pacific. The bias of latent heat flux (Fig. 4d) is relatively small. In the eastern Pacific, the latent heat flux is under evaluated. As a result of the larger sensible heat flux and smaller latent heat flux, NorLR has a large Bowen ratio. On the contrary, In the NorSR the sensible heat flux is reduced in the Indian ocean and western Pacific. The latent heat flux is increased in a large region of the tropical ocean. Both the reduced sensible heat flux and increased latent heat flux contribute to a smaller Bowen ratio in NorSR in Fig 2c. Consider that the sensible heat flux in increased in the eastern Pacific in NorSR, the decreased Bowen ratio there is dominated by the increased latent heat flux.

Comparing NorSR and NorLR, the bias of sensible heat flux is reduced in NorSR, However, the overestimation of latent heat flux is a new bias in NorSR. This might be related to the increased model resolution and might lead to high precipitation in Fig. 1c.

**3.2.3 Variables relate to latent and sensible heat fluxes**

Previous analysis shows that the excessive Bowen ratio in NorLR is dominated by the over estimation of sensible heat flux. The reduced Bowen ratio in NorSR is contributed by larger latent heat flux over the whole tropical ocean. To understand the bias of latent and sensible heat fluxes, all the factors related to the latent and sensible heat fluxes are analysed.

In the NorESM models, the latent (E) and sensible (H) heat fluxes are calculated based on equation (5) and (6), in which is air density, is the specific heat at constant pressure, is volumetric heat capacity of air. is surface wind stress, and are dimension-less bulk transfer coefficient for moisture and heat. and , they are the vertical gradient of humidity and temperature between sea surface and lowest air level. Subscripts *s* an *A* indicate values for the sea surface and the air at lowest model level.

(6)

(7)

*a) Vertical gradient of temperature*

Considering that the models have no output for the lowest model level temperature, we express the vertical gradient of temperature with the difference of surface and 2m air temperature (DT=Ts-T2m). As shown in Fig.5a, in ERA5, the large value of DT follows the ITCZ, with the maximum up to 2.5°C in the warm pool region. The large DT indicates that in the tropical ocean, especially in the warm pool region, the ocean warms the air above via both sensible and latent heat flux.

In NorLR (Fig. 5b), DT is under evaluated in northern Indian ocean and western Pacific, and it is over evaluated in Indian ocean, maritime continent, eastern Pacific and Atlantic. This pattern resembles the bias in sensible heat flux in Fig. 4c an latent heat flux in Fig. 4d. The similar distribution of the bias in DT and the sensible and latent heat flux in NorLR indicate that the vertical gradients of temperature and related humidity are the dominant factors of turbulent heat flux bias in low-resolution NorESM model.

Unlike the Low-resolution model, in the super-high-resolution model NorSR, the bias of the turbulent heat flux is incoherent with the bias in DT. In NorSR, DT is under evaluated up to 50% (Fig. 5c), however, NorSR simulates stronger sensible heat flux in eastern Pacific and Atlantic. The latent heat flux is also over estimated over almost the whole tropical ocean. This indicate that the bias of DT is not the main restraint of the changes in turbulent heat flux in NorSR.

The different role of DT in the two models indicate that the sensitivity of turbulence heat flux to the DT are different in the models. As shown in Fig.6a, the regression coefficient of daily sensible heat flux is close to or below zero in ITCZ region, warm pool region and tropical Indian Ocean. This implies negative feedback between DT and turbulent heat flux, i.e. strong turbulent heat flux enhances the SST cooling, and in turn reduce DT. As shown in Fig. 6bd and Fig. 6ce, the negative feedback is over evaluated in NorLR. The bias of the sensitivity is small in NorSR. As shown in Fig.6f-j, the sensitivity of latent heat flux to DT is also over evaluated in NorLR and the bias in NorSR is smaller. Due to stronger feedback between DT and the turbulent heat flux, they have stronger correlation in NorLR.

*b) Surface wind*

The surface wind in ERA5 in Fig.7a shows that the wind speed is relative weak in the equatorial region (5°N-5°S). Especially, in the western Pacific warm pool, the wind speed is under 4m/s. The wind speed is larger in the eastern Pacific out of the equatorial region, with the maximum around 8-9m/s. The direction of the wind shows a clear convection zone along 5°N, which is consistent with the precipitation belt in Fig 1a.

In NorLR (Fig.7a), the wind speed in the eastern Pacific is weaker than in ERA5. The cross-equator flow is reduced in the cold tongue region. The weaker wind speed is opposite with the over estimation of sensible heat flux in NorLR. This further confirms the previous analysis that the amplified turbulent heat flux in NorLR is dominated by the larger DT.

In NorSR, the wind speed is stronger than ERA5 over the almost the whole tropical ocean. The wind direction shows an enhanced cross equator flow in eastern Pacifci and stronger convection along 5°N. This might be related to a large-scale circulation anomaly, which will be addressed later. Since the air-sea temperature difference in NorSR is smaller than ERA5, the excessive sensible and latent heat flux is dominated with the stronger surface wind.

Considering that the amplified sensible heat flux in eastern Pacific and larger latent heat flux in the whole tropical ocean is opposite with the smaller Ts-Ta, the increased sensible and latent heat flux is dominated by the stronger surface wind.

*c) Sensitivity of turbulent heat fluxes to surface wind*

The sensitivity of turbulent heat flux to surface wind is calculated by regressing the daily sensible and latent heat flux to the surface wind. As shown in Fig. 8, the sensitivity of sensible heat flux is weaker in NorLR than in ERA5. This indicate that the excessive sensible heat flux in NorLR is dominated by the larger air-sea temperature difference. In NorSR, the sensitivity of sensible heat flux is slightly stronger, especially along 5°N in the eastern Pacific. Consider the weaker air-sea temperature difference, the sensible heat flux in NorSR is dominated by the surface wind speed. Consistent with stronger wind speed in pacific along 5°N, the sensible heat flux has a large value belt there.

The sensitivity of latent heat flux is similar as that of sensible heat flux. The sensitivity is smaller in NorLR than in ERA5, but it is larger in NorSR around of 5°N over the whole Pacific and most of Indian ocean. The over estimation is specific in NorSR and is not seen in NELR. Due to stronger wind speed in north of tropical pacific, the larger sensitivity of latent heat flux is related to the large latent heat flux in NorSR.

Therefore, due to larger sensitivity of latent heat flux to surface wind and the stronger surface wind, the latent heat flux is over evaluated in NorSR. The changes in sensible heat flux have a smaller amplitude than the latent heat flux, the Bowen ratio is small in NorSR. A smaller Bowen ratio indicate more turbulent heat flux are utilized to generate a wetter atmosphere, which is a favourable condition for precipitation.

**3.3 The physical processes related to the precipitation biases**

**3.3.1 Moisture transport**

**3.3.2 Walker and Hadley circulation**

Based on eq. 6 and 7 latent and sensible heat flux are related to transfer coefficient and . The transfer coefficientare the function of surface wind speed (eq. 7)

(7)

and in eq. 6 and 7 are affected by advection, which is also related to the surface wind. To understand the variables that related to the surface wind, we analysed the simulation of Walker circulation and Hadley circulation.

As shown in Fig.7a, the walker circulation in ERA5 has a strong ascending branch in the western Pacific (75°E-175°E), and a mild descending branch between (225°E-275°E). Consistent with the circulation, the specific humidity is converging at the lower levels of the ascending branch and transported upward.

The Walker circulation in NorLR is much weaker than in ERA5. The ascending branch is suppressed west of 100°E and is enhanced between 100°E-150°E. The descending branch is also offsetted by an upward bias around 200°-250°E. Related to the circulation bias, the convergent and rising specific humidity is eastward shifted with the maximum around 950hPa-750hPa. The bias of moisture air in NorLR is consistent with less precipitation in Indian ocean and more precipitation in southern tropical Pacific east of 200°E.

In NorSR, however, the upward branch is enhanced in 50°-75°E and supressed in 125°-175°E. This bias implicates a westward shift of upward branch. The westward shift of rising branch is consistent with more precipitation in Indian Ocean and a shorter eastward extension of the southern branch of ITCZ. In eastern Pacific, there is no descending flow in the meridional averaged circulation (Not shown). This is because the over evaluated rising flow in 200°-250°E overwhelmed the descending flow in the averaging processes. To know the details of the circulation structure, we analysed the local Hadley cell in eastern Pacific (120°E-180°E,.50°S-50°N) and western Pacific (180°E-270°E, 50°S-50°N), respectively.

In ERA5, there is a strong ascending branch in western Pacific (west of 180°E) between 10°S-10°N. The descending flow is not symmetric with respect to the equator. In the southern Hemisphere, there is a mild descending branch at 30°S-40°S. In the northern hemisphere, the descending flow is not clear. In the eastern Pacific, there is a narrow strong ascending flow in the north of the equator, around 3°-13°N. On the southern and northern of the rising flow is a widely distributed descending movement. Since the eastern Pacific is dominated by the descending flow, the rising movement is overwhelmed during the meridional average, as shown in walker circulation (Fig. 7a).

In NorLR (Fig. 7e), Local Hadley cell of western Pacific has a clockwise bias in 30°S-5°N. The direction of bias is opposite to the mean state, indicating a weaker local Hadley circulation in the western Pacific. Correlated to the reduced subsidence branch in the southern Hemisphere, the specific humidity is over evaluated, with the maximum around 20°S at 850 hPa. In the eastern Pacific (Fig. 7f), the bias also shows a clockwise cell, which is much stronger than in the western Pacific. The clockwise bias leads to a weakened ascending branch around 10°N and a fake upward flow around 10°S. Related to the ascending flow in NorLR, more moisture concentrated and more precipitation generated around 5°S-20°S east of 180°E. This is known as the double-ITCZ bias in NorLR.

In contrast with NorLR, the bias of the local Hadley circulation in NorSR is anti-clockwise in both the western Pacific and eastern Pacific. In the western Pacific, the ascending branch is supressed in 10°S-0° and enhanced around 5°N, indicating an enhanced and northward shifted upward branch in western Pacific. In the eastern Pacific, there is also an anti-clockwise circulation bias in 20°S-10°N, indicating an overestimated Hadley cell. With the enhanced ascending branch around 10°N, specific humidity concentrated and convected upward, this is consisted with the excessive precipitation in the northern branch of ITCZ. The enhanced subsidence flow pushes cold and dry air downward. This is consistent with the dry equator region in NorLR.

Since the overestimated rising air and the wider distributed subsident flow can feedback to each other, it’s worth to figure out the possible trigger of the anomalous vertical circulation. Normally, the excessive convection in the tropical ocean is related to a warmer SST, but the SST in NorSR is slightly colder than that in NorLR (Fig. 3c,e), indicating that the excessive convection in NorSR is not SST forced. We thus analysed the radiation on the top of the atmosphere instead.

**4. Conclusion and discussion**

Less medium-level cloud is simulated in SHRM-> strong top colling (OLR)-> Strong Hadley circulation-> Stronger subsidence branch of Hadley cell -> Dry surface humidity-> large air-sea humidity difference -> large latent heat flux -> The shift of ocean Bowen ration towards to lower values

Strong Hadley circulation -> strong precipitation

The understand evaluation of the medium-layer clouds might be related to the convection or other physical frame in the model, it may also directly be caused by the increase of horizontal resolution. To understand this, further work is required with several model experiments.

The decreased DT is due to the air-sea heat and momentum exchange is enhanced when the model resolution is increased.

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Hypothesis to be test:

Higher resolution increased the air-sea interaction

🡪 and , reduced

🡪 latent and sensible heat flux shift from heat forcing to wind forcing

🡪 coupling among surface wind speed, local evaporation, convection, local Hadley circulation is enhanced

* Regress precipitation to latent heat flux (evaporation)
* Regress local Hadley circulation to latent heat flux

🡪precipitation and Hadley circulation bias at 5N°

AGCM:

Atm:

CAM6-Nor with low resolution, L, M, H: settled

Lingling’s version, change the cloud brightness. See if it can change the circulation.

Oct:

Coupled with same SST

云参数，parameters\_tunable.90

NorESM2/banben/components/cam/src/physics/cam

LinglingSuo

NESM2 Low 1\*1

NorCPM: NorESM2 + data assimualation 2019.12-2022.3