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

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**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 10m wind speed and vertical gradient of specific humidity. 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, within 0.25°-2.5°, is up to 40 W m-2 / m s-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 simulations. This study can help to improve heat fluxes parameterization and correct the related model bias in the higher-resolution models.

**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 and aspects (Moore et al., 2014; Zolina and Gulev 2003; Small et al. 2008). 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. THFs 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, energizing storms and anchoring the major storm tracks along the western boundary current (Nakamura et al. 2004; Kwon et al., 2010). The THFs are also crucial drives of the variability of many ocean processes, such as deep convection in the subpolar waters (Holdsworth and Myers 2015). 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 climate studies. In reality, however, quantifying these THFs, is a big challenge in observation. The directly measured THFs tend to be infrequent, and highly localized, which are not sufficient to build global distributed THFs (Brunke et al. 2011). The global distributed THFs are normally estimated by bulk formulars. The bulk formulars parameterized THFs with easily achievable surface state variables and bulk transfer coefficients. 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 (Large and Pond 1981, 1982).

The original 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, 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 (details shown in section 2.2). Although widely used in the numerical models, previous studies repeatedly shown discrepancies of state-of-the art bulk algorithms lead to a large spread in the computed THFs (Blanc 1985; Zeng et al., 1998; Eymard et al., 1999; Brunke et al., 2003). Studies also noticed that even though the surface state variables vary with models and sources, the constants and parameter approximations in bulk formulas are rare changed (Josey et al., 2014; Berry and kent 2005). Brodeau et al., (2016) quantified the uncertainties and shown that the choice of algorithm is related to 15% of uncertainties in the evaluation of turbulent heat flux, and the approximations of skin temperature and saturation humidity can lead to an uncertainty of turbulent heat flux up to 10%.

The uncertainties of simulated THFs are not only associated with the chosen bulk algorithms and the simulated surface state variables, biases of THFs also arise with the increase of model resolutions. Normally, higher-resolution models are expected to give better simulations. When the horizontal resolution increased to finer than 100km, a significant improvement in large-scale circulation in atmosphere is identified (Roberts et al., 2018). When the resolution reaches kilometre scales, the clouds, deep convection and turbulent eddies can be resolved (Stevens et al., 2020). Concurrent with many significant improvements, 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. Wu et al., (2010, 2013) found that the evaporation and precipitation are more sensitive to net surface short-wave radiation in higher-resolution models. Ding et al., (2021) also found that with the increasing of model resolutions, the air-sea latent heat fluxes are enhanced along Kuroshio. These biases and changes in air-sea fluxes imply that the bulk algorithms suited for low resolution versions require further adjustment in higher resolution models.

The question is how to adjust the THFs’ parameterizations 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 models, data and methods. The results are shown in section 3 and discussed further in section 4. The conclusions are summarized in section 5.

**2. Models, data and methods**

**2.1.1 NorESM**

This work uses the Norwegian Earth System Model (NorESM) developed by the Norwegian Climate Center (Horowitz et al., 2003, Bentsen et al., 2012). NorESM is a globally fully coupled model for climate simulations. It is developed based on the Community Earth System model (CESM) (Vertenstein et al., 2012; Danabasoglu et al., 2019). The atmospheric component CAM6-Nor features advanced aerosol chemistry schemes (Kirkevåge et al., 2013, 2018). The ocean compontent Bergen Layered Ocean Model (BLOM; Bentsen, 2020) is an updated version of the isopycnal coordinate ocean model MICOM (Bleck et al. 1992). The ocean model has the isopycnic coordinate Hamburg Ocean Carbon Cycle (iHAMOCC) model for ocean biogeochemistry (Tjiputra et al., 2020). In current study, we use the NorESM with two resolutions. The low-resolution model (NorESM-LR) is 1.9° longitude and 2.5° latitude, and the high resolution (NorESM-HR) is 0.25° longitude and 0.25° latitude. In the vertical, the model has 32 hybrid-pressure layers and a “rigid” lid at 3.6 hPa (40 km).

**2.1.2 bulk formular in NorESM**

In 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. In the model, the above equations are solved by iteration.

**2.2 ERA5**

the fifth generation European Center for Medium-Range Weather Forecasts atmospheric reanalysis of the global climate (ERA5; Hersbach et al., 2018) with a spatial resolu-tion of 0.25° × 0.25°

**2.3 method**

2.3.1 Filter

2.3.2 Contributions of changes in latent heat flux

In which, and . The subscript H and L indicate variables in high-resolution and low-resolution model, respectively.

**3. The results**

**3.1 Annual mean precipitation**

In the tropical pacific, the annual 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 in the middle and eastern Pacific. There is also a weaker 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.

As shown in Fig. 1b, the rain belt in northern tropical Pacific is under evaluated in the low-resolution NorESM (NorESM\_LR), which is 1/3 weaker than in ERA5. In the southern hemisphere, the rain belt is over estimated in NorESM\_LR, with the amplitude and eastward extension resemble 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 (NorESM\_HR), the double-ITCZ problem is significantly modified (Fig. 1c). The southern branch of precipitation is much weaker than the northern one and has a shorter south-eastward extension to 200°E. The amplitude of the southern branch is very close to the ERA5. Despite the improved double ITCZ problem, excessive precipitation is produced in the northern branch of the ITCZ. The average precipitation amount is almost twice of that in the observation.

**3.2 Processes related to the tropical precipitation biases**

Considering the strong sea surface temperature (SST) – convection relationship in tropics (Samanta et al., 2019), we diagnosed the SST bias in the model. As shown in Fig.2b, in most of the tropical ocean, SST is warmer in NorESM\_LR than ERA5. Only in the cold tongue region, the SST has cold bias. The warmer bias in southern tropical pacific is consistent with the eastward extended precipitation and might contributed to the double ITCZ bias in NorESM. Nevertheless, the warmer bias in Indian ocean and in the northern branch of ITCZ is incoherent with the weaker precipitation in NorESM\_LR. This implies that the warmer SST is not the dominant factor of the precipitation bias in NorESM\_LR. In contrast, SST in NorESM\_HR is colder than ERA5 in most tropical Pacific, especially in the southeaster tropical Pacific. This may reduce precipitation in southern branch of TICZ and is related to the modified double ITCZ problem. In the northern Pacific, however, there is not clear warm SST bias to explain the over evaluated precipitation in region between 5°N-10°N. This means the excessive precipitation in NorESM\_HR is not directly forced by the SST beneath.

Since we cannot explain the excessive precipitation in NorESM-HR with the SST bias, we analysed the column integrated moisture transport convergence in the tropical ocean. Furthermore, the moisture transport convergence is separated into components due to anomalous circulation and the part due to anomalous specific humidity (Li and Ting 2017, Tian et al., 2018).

**3.2.1 Dynamic: Hadley circulation**

The tropical precipitation is strongly related to the Hadley circulation which transport water and fuels the convections. 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.

**3.2.2 Thermodynamic: Specific humidity and Latent 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.

**3.3 Variables relate to latent heat fluxes**

The increased specific humidity is directly related to stronger evaporation and then stronger latent heat flux

In the NorESM models, the latent (E) heat fluxes are calculated based on equation (6), in which is air density, is the specific heat at constant pressure, is volumetric heat capacity of air. is surface wind stress, is dimension-less bulk transfer coefficient for moisture and heat. and , they are the vertical gradient of humidity 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)

*3.3.1 Vertical gradient of specific humidity*

We express the vertical gradient of specific with the difference of surface and 2m air specific humidity (DQ=Qs-Q2m). As shown in Fig.6a, in ERA5, DQ has the large value along the ITCZ region and in the pacific warm pool region with the maximum up to 8 g/kg. The large DQ is related to strong latent heat flux and evaporation, moisture the air above.

In NorESM\_LR (Fig. 6b), DQ is under evaluated between 0-5°N in the eastern and western Pacific, and it is over evaluated in the ocean around the maritime continent and in the middle Pacific, with the maximum centred around 10°N and 5°S, respectively. The distribution of the bias in DQ is slightly patchy, but generally it is similar to the bias in latent heat flux in NorESM\_LR. This indicates that the bias vertical gradients of related humidity contribute partly to the latent heat flux bias in NorESM\_LR.

Unlike the Low-resolution model, in the ¼° NorESM (NorESM\_HR), the bias of DQ is along the equator in Indian ocean, Pacific and eastern Atlantic. And around 8°S in the southern Pacific. This indicated an enhanced and south shifted DQ in NorESM\_HR related ERA5. The region of positive bias in DQ is much smaller than that in latent heat flux, which is over estimated over almost the whole tropical ocean. This indicate that the bias of DQ is not the main restraint of the changes in latent heat flux in NorESM\_HR.

The different role of DQ in the two models indicate that the sensitivity of latent heat flux to the DQ are different in NorESM with different resolutions. Fig. 7 shows the regression coefficient of daily latent 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.

*3.3.2 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.

**3.4 Sensitivity of turbulent heat fluxes**

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.

**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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