NUMERICAL MODELING OF HEAT AND FRESHWATER FLUXES, TEMPERATURE, AND SALINITY AT THE JAPAN SEA SURFACE

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By using the MHI oceanic model, the upper layer currents, the surface heat and freshwater fluxes, the sea surface temperature (SST) and the sea surface salinity (SSS) are simulated for the Japan Sea. Given atmospheric variables are used for the surface heat and freshwater balance calculation; no restoration condition towards observed SST or SSS is applied. The simulated surface fluxes, SST, and SSS compare well with the long-term data. The simulated freshwater flux is directed from the air to the sea in spring and summer and from the sea to the air in autumn and winter. The simulation results suggest that SSS in the Japan Sea is mostly controlled by the surface freshwater flux and advection through the straits.

Key Words: numerical modeling, Japan Sea circulation, air – sea heat and freshwater fluxes, sea surface temperature and salinity

1. INTRODUCTION

The Japan Sea is situated in the monsoon region of temperate northeast Asia and is subjected to strong wintertime continental winds. At the same time, there is a throughflow of subtropical water from the Korean (Tsushima) Strait northeastward towards Tsugaru and La Perouse (Soya) Straits (Fig. 1). Observation evidence suggests that the excessive surface cooling is balanced by the inflow of warm water. This suggestion is confirmed by observation-based estimates of the heat budget in the Japan Sea. Improved long-term monthly heat flux components based on modern bulk formulas are presented by Hirose *et al.*¹⁾ According to their results, wintertime surface heat flux exceeds 300 W/m² (Watts/m²) in the southern area with the increased sea surface temperature (SST); it is about 100 W/m² in the northwest Japan Sea¹⁾. The summertime surface heat

gain is of 120 – 130 W/m². It is also estimated that the heat balance in the Japan Sea can be maintained by an inflow of 2.2 Sv (Sverdrup; 1 Sv = 10⁶ m³/s) of subtropical water from the Korean Strait¹⁾.

Evaporation rate can be estimated as 180 – 200 mm/month for December – January in the southeast Japan Sea and less than 100 mm/month in the northwest Japan Sea, based on the latent heat flux reported by Hirose *et al.*¹⁾ Precipitation over the Japan Sea is considerably inferior in winter. According to GPCP (Global Precipitation Climatology Project) data²⁾, it is of 30 – 50 mm/month for the northwest Japan Sea and of 80 – 120 mm/month for the southeast Japan Sea in January. In contrast, summertime precipitation exceeds evaporation, averaging 150 mm/month and 20 – 30 mm/month over the Sea, respectively^{1),2)}. Precipitation – evaporation residual (surface freshwater flux), advection from the Korean Strait, and continental run-off determine sea surface salinity (SSS).

Fig. 1 Model bottom topography (m) of the Japan Sea. Contours are shown every 250 m. Korean (Tsushima) Strait (1), Tsugaru Strait (2), La Perouse (Soya) Strait (3), Tatarsky (Mamiya) Strait (4), East Korea Bay (5), and Peter the Great Bay (6) are shown. The northwest and southeast areas of the Japan Sea are separated by the dashed line.

The intermediate and deep waters in the Japan Sea are isolated from the adjacent basins by shallow thresholds, because the depth of the straits is less than 150 m. These waters are rich in oxygen, because they are ventilated within the Japan Sea. The surface water loses buoyancy due to wintertime cooling and salinity increase caused by evaporation excess. The surface water subducts and moves with the currents in the intermediate or deep layers. The formation of intermediate waters was studied numerically by Kim and Seung³⁾ and by Yoon and Kawamura⁴⁾. They emphasize the significance of realistic simulation of surface heat and freshwater fluxes, temperature, and salinity in studies on the formation of intermediate and deep waters. Realistic surface salinity is especially important, as the salinity variation is low in the intermediate and, especially, deep waters of the Japan Sea, being within the range of 33.9–34.2 ppt (parts per thousand) at the depths below 200 m⁵⁾.

The restoration to prescribed (climatic) surface temperature and salinity is applied as the surface buoyancy forcing in many modeling studies of the Japan Sea, for example, by Kim and Seung³⁾. However, the direct atmospheric forcing facilitates simu-

lations under varied external conditions. Yoon and Kawamura⁴⁾ applied the linearized Haney type boundary condition for heat flux⁶⁾, while the restoration condition remained for SSS. However, it is known that the linear condition underestimates the heat flux⁶⁾. The MHI (Marine Hydrophysical Institute) model was previously applied for modeling an impact of seasonal and decadal variation in the surface thermal forcing upon circulation patterns⁷⁾ and surface heat flux and SST⁸⁾ in the Japan Sea, with the application of the complete (non-linear) formulation for the surface heat flux, while the surface freshwater flux was neglected. In the present study, an attempt is made to simulate salinity variations in the Japan Sea entirely on the base of the external forcing, namely the surface freshwater flux and inflow from the Korean Strait in the south and from the Amur River Estuary in the north. The discharge of small rivers and continental run-off are still neglected and their impact will be analyzed separately in the following studies.

Realistic simulation of thermodynamic conditions at the Japan Sea surface is important for studies of the response of the Japan Sea circulation to climate change induced, in particular, by human activities. Better understanding of the vertical water exchange affected by surface density is important for analyses of spreading of pollutants discharged from the sea-coasts or shipwrecks.

2. SIMULATION SETUP

(1) Numerical model and simulation parameters

The MHI model developed by N.B. Shapiro⁹⁾ is the 3D primitive equation free surface model under the hydrostatic and Boussinesq approaches. It belongs to a class of layered and isopycnal models, in which the sea consists of a number of layers spreading one on another in the vertical. Interfacial surfaces between layers can freely move up and down and layers can deform, physically vanish (outcrop) and restore. Equations of the MHI model, vertically integrated within layers, are formulated at the beta-plane, with the x-axis directed from the west to east and the y-axis directed from the south to north. A specific feature of the MHI model is that the density (buoyancy) of any layer is allowed to vary with time. This approach allows an independent evolution of temperature and salinity for any layer. The present study is focused on temperature and salinity variations in the upper layer. Restoration condition is not applied for temperature or salinity, with the exception of the outflow ports where SST and SSS are relaxed towards those in the adjacent area.

Surface buoyancy forcing is provided by estimation of heat and freshwater fluxes from the bulk formulation⁶⁾ with the use of atmospheric variables and temperature in the upper (first) model layer as SST. The surface heat flux Q_0 is defined as follows:

$$Q_0 = Q_s + Q_b + Q_h + Q_e \quad (1)$$

where Q_s is the short-wave solar radiation taken from long-term data, Q_b is the long-wave radiation, and Q_h and Q_e are turbulent, sensible and latent, heat fluxes, respectively. They are calculated from the following bulk formulas⁹⁾:

$$Q_b = -\sigma I(T_a + 273.15)^4, \quad (2)$$

$$Q_h = \rho_a C_p c_H V_a (T_a - T_1), \quad (3)$$

$$Q_e = \rho_a c_E V_a L [r_h e(T_a) - e(T_1)] 0.622/p_a, \quad (4)$$

$$I = 0.985 \{0.39 - 0.05[r_h e(T_a)]^{1/2}\} [1 - 0.6(n_0)^2], \quad (5)$$

$$e(T) = 10^{964051 - 2353/(T+273.15)}, \quad (6)$$

where T_a is the air surface temperature, T_1 is temperature in the first model layer (the simulated SST), n_0 is the cloudiness, V_a is the wind speed, $e(T)$ is the saturated vapor pressure, r_h is the relative humidity, $\sigma = 1.36 \times 10^{-12}$ cal/s/cm²/°C⁴ is the Stephan-Boltzman constant, $\rho_a = 1.25 \times 10^{-3}$ g/cm³ is the air density, $C_p = 0.24$ cal/g/°C is the air specific heat, $L = 595$ cal/g is the latent heat of vaporization, $p_a = 1013.25$ mb is the surface pressure, $c_H = c_E = 1.4 \times 10^{-3}$ are the turbulent exchange coefficients.

The surface freshwater flux R_0 is defined as the difference between precipitation P and evaporation E ($P - E$ residual), with the former taken from observation data and the latter being proportional to the latent heat flux and, thus, providing coupling of heat and freshwater fluxes:

$$R_0 = P - E, \quad (7)$$

Therefore, the surface buoyancy flux M_0 is defined as follows⁹⁾:

$$M_0 = \alpha Q_0 - \beta R_0, \quad (8)$$

where α and β are the coefficients of thermal expansion and saline contraction, respectively. To estimate the upper layer depth, the Kraus – Turner type (slab type) model¹⁰⁾ is employed, based on the balance of turbulent kinetic energy (TKE) in the upper mixed layer. The TKE equation is vertically integrated over the mixed layer as follows⁹⁾:

$$h_1 [M - W_1 (b_1 - b_k)] = 2m(u^*)^3, \quad (9)$$

where h_1 is the mixed layer depth, $M = a(M_0 + M_1 + \epsilon)$, $M_1 = \mu(b_1 - b_k)$ is the buoyancy flux at the mixed layer base, μ is the diapycnal exchange coefficient, ϵ is the background dissipation, $W_1^- (< 0)$ is the up-

ward entrainment rate at the mixed layer base, k is an inner layer contacting with the upper layer, b_1 and b_k is buoyancy in the upper and k th layer, respectively, u^* is the dynamic friction proportional to wind speed, and a and m are adjustable constants.

The TKE equation supplemented by the continuity equation provides the closure for estimation of the mixed layer depth and entrainment/subduction rate at its base. In particular, the seasonal deepening or shallowing of the mixed layer is simulated through entrainment or subduction of water from or to an inner layer underneath. The comprehensive formulation of the MHI model is done by Shapiro⁹⁾ and its application to the Japan Sea is discussed by Trusenkova *et al.*⁷⁾.

The model domain covers the Japan Sea from 34°15' to 52°N in the south-north direction and from 127°15' to 142°30'E in the west-east direction. The inflow ports are the Korean (Tsushima) Strait in the south and a border between Tatarsky (Mamiya) Strait and Amur River Estuary in the north; the outflow ports are the Tsugaru and La Perouse (Soya) Straits in the east (**Fig. 1**). The Japan Sea is divided into 12 layers in the vertical, with the horizontal mesh size of 1/8°. The bottom topography is adopted from navigation maps (**Fig. 1**). Islands cannot be resolved in coastline for the chosen mesh but they are represented in topography⁷⁾.

Starting from flat interfaces, the MHI model was integrated with the time step of 7.5 min for 12 years. The initial characteristics of the layers are shown in **Table 1**. Note, that temperature, salinity, and other variables in the upper layer get adjusted to the surface forcing after a year and half of the initial spin-up, as shown below.

The isopycnal bi-harmonic viscosity used in the momentum equations is equal to 2.5×10^8 m⁴/s. This is equivalent to the harmonic viscosity of 7 m²/s. The isopycnal harmonic diffusion coefficient for diapycnal diffusion coefficient at interfacial surfaces

Table 1 Simulation parameters

Initial depth of interfacial surfaces (m)	10, 25, 50, 75, 100, 150, 250, 350, 500, 700, 900
Initial buoyancy in layers (cm/s ²)	3.8, 2.6, 1.8, 1.4, 1.1, 0.95, 0.85, 0.785, 0.755, 0.735, 0.725, 0.717
Initial temperature in layers (°C)	18, 13.5, 9.5, 6.75, 4.3, 2.8, 1.75, 1, 0.6, 0.335, 0.195, 0.08
Initial salinity in layers (ppt)	33.5, 33.8, 34.01, 34.03, 34.04, 34.05, 34.06, 34.07, 34.07, 34.07, 34.07, 34.07
Monthly volume transport from the Korean Strait (Sv; 1 Sv = 10 ⁶ m ³ /s)	1.05, 0.8, 0.8, 0.8, 0.8, 1.3, 2, 2.6, 2.85, 2.7, 2.4, 1.85

between the layers is equal to 5×10^{-4} m/s. The latter is equivalent to the harmonic diapycnal diffusion coefficient of 5×10^{-6} m²/s if applied in the conventional level model with the level spacing of 100 m in the vertical (see Trusenkova *et al.*⁷⁾ for detail).

Monthly volume transport in the Korean Strait is taken after geostrophic estimates of Leonov based on observations performed in 1950s¹¹⁾. The outgoing transport is distributed between the Tsugaru and La Perouse Straits, with the ratio of 2 to 1. Temperature and salinity in the Korean Strait are taken from the long-term data.

The accepted transport (**Table 1**) is lower than that employed in the previous simulations with the MHI model (2 – 3 Sv)^{7,8)}. Its annual mean is equal to 1.66 Sv, lower than that estimated from heat balance considerations (2.2 Sv; see **Section 1. Introduction**). In late summer to autumn, the inflow transport of 2.4 – 2.85 Sv is close to that accepted in other modeling studies of the Japan Sea. It is set up low in late winter through early summer (0.8 Sv from February through May), providing decreased supply of warm water in the Korean Strait. On the other hand, the accepted wintertime air temperature corresponds to the mild climatic conditions of the end of 20th century (see **Section 2.(2)**). The thermal forcing applied seems to be unbalanced but this allows us to compare an impact of two factors upon the thermal regime. Another reason is our intention to compare various factors (inflow from the Korean Strait, surface buoyancy forcing, baroclinicity, etc.) affecting the surface circulation in the Japan Sea. This is the subject of another paper¹²⁾.

Volume transport entering the Sea in the north from Amur River Estuary is very small and has no considerable effect upon the circulation outside the adjacent area. However, it is often thought to determine salinity along the whole northwest coast⁴⁾. Run-off and small river discharge from the adjacent land should be also important, especially in the shallow northern Tatarsky Strait, but it is hard to obtain quantitative estimates because observations are very scarce in this poorly populated area with the severe climate. Estimation of wintertime volume transport from Amur River Estuary is based on the unique survey in 1973¹³⁾. For this reason, we accept the inflow from Amur River Estuary and neglect the continental run-off. River discharge to Peter the Great Bay (**Fig. 1**) is known but also neglected here. This will allow us to discriminate effects of the Amur River water and adjacent land run-off. There is a northward flow out of the Japan Sea along the eastern coast and southward flow into the Sea along the western coast¹³⁾. Net transport direction is determined by wind and local sea surface slope. In winter and spring, a resulting flow of 1700 m³/s directed

from Amur River Estuary to the Japan Sea¹³⁾ is accepted in the simulation, with temperature and salinity also taken after Yakunin¹³⁾.

(2) Atmospheric data applied in the simulation

In this study, monthly atmospheric variables are used for parameterization of surface buoyancy flux and momentum flux (wind stress). Some variables are taken from the historical data and some from the data for 2000. The atmospheric data are available in frame of NCEP (National Centers for Environmental Prediction, USA) Reanalysis Project. They were retrieved from the Web site of the NOAA-CIRES Climate Diagnostics Center, Boulder, Colorado, USA (<http://www.cdc.noaa.gov/>).

Monthly wind stress is calculated from the 6-hour NCEP surface wind. Distinct patterns of winter and summer monsoon are revealed, with northern and northwest winds from September through March and southern, southwest, and southeast winds from April through August, respectively. Wind forcing is much stronger in winter than in summer. Typical wind direction is shown in **Fig. 2**; applied wind stress is discussed in detail by Trusenkova & Ishida¹²⁾. At the same time, variations of atmospheric pressure can have a direct impact on sea surface height (SSH). However, this effect is neglected in the simulation and surface pressure is considered constant in the momentum equations. This can be justified by low spatial gradients of the large-scale monthly wind (and atmospheric pressure) applied.

Short-wave solar radiation is taken from long-term data. It is close to that used by Hirose *et al.*¹⁾. The following atmospheric variables are applied in the bulk formulas for calculation of the other kinds of fluxes. Cloudiness is taken from long-term data. Surface air temperature is averaged for the period from the mid 1980s through mid 1990s, providing

Fig. 2 Monthly wind speed (m/s) for February (a) and July (b) over the Japan Sea. Contours are shown every 1 m/s. Arrows show prevailing wind directions.

mild surface thermal conditions in winter. Monthly wind speed and relative humidity are calculated from the 6-hour NCEP reanalysis data.

Monthly wind speed is higher over the Japan Sea than over the adjacent land, with a gradient zone along the coast (Fig. 2), as the wind strengthens over the water surface due to decreased friction. In April – November, the wind speed is the highest in the area between 42°N and 48°N (Fig. 2b). Strong wind with the speed of more than 8 m/s persists over the Sea from November through March. In December – March, the wind speed is the highest in the area between 38°N and 44°N. The wind speed is low over northern Tatarsky Strait even in winter. Monthly air relative humidity over the Japan Sea is moderate in winter (55 – 70%) and high in summer (75 – 85%), with only slight spatial variation.

Monthly precipitation applied for the estimation of the surface freshwater flux is based on the GPCP (Global Precipitation Climatology Project) data²⁾ averaged for 23 years from 1979 through 2001, providing conditions of early 1990s. As discussed in the Section 1, precipitation over the Japan Sea is high in summer and low in winter, with its spatial and temporal variation conditioned by summer and winter monsoon winds. From September through April, precipitation increases from the northwest coast towards the southeast coast and is the highest along the Honshu coast (Fig. 3a). In summer, precipitation is high along the southwest and southeast coasts, especially over the East Korean Bay (up to 260 mm/month in July), and it is decreased over the central Sea and Tatarsky Strait (Fig. 3b).

Ice model is not used but the surface fluxes will be put equal to zero in the grid locations where and when the surface water reaches the freezing point.

Fig. 3 Monthly precipitation (mm/month) for February (a) and July (b) over the Japan Sea and adjacent land. Contours are shown every 10 mm/month.

3. RESULTS

(1) Simulated surface currents in the Japan Sea

Surface currents simulated after 12 years of integration are consistent with those obtained in the simulations performed without direct wind stress forcing under buoyancy forcing only⁷⁾. They are also consistent with simulations performed by other kinds of numerical models, such as in studies of Kim and Seung³⁾ and Yoon and Kawamura⁴⁾, and compare well with the known circulation schemes derived from observation data, such as that of Uda¹⁴⁾. The simulated surface and deep circulation is discussed in detail in Trusenkova & Ishida¹²⁾, with only a brief summary provided here.

Fig. 4 represents the surface circulation as sea surface height (SSH) patterns for every month on the 12th year of integration. The surface currents are also represented by velocity vectors for January and July in Fig. 5. The anticyclonic circulation characterized by the sea surface rise develops in the southern subtropical region (Figs. 4 and 5). The Tsushima Warm Current (TWC) flows along the Honshu coast and the East Korea Warm Current (EKWC) flows along the Korea Peninsula coast until it separates from the coast at 37°30'–39°N (Fig. 5). SSH is decreased in the northern subarctic region where several cyclonic gyres develop (Fig. 5), revealed by local SSH depressions throughout the year (Fig. 4). The cold currents (Shrenk Current, Primorye (Liman) Current, and North Korean Cold Current) flow southwestward along the western coast from northern Tatarsky Strait to Korea Peninsula (Fig. 5). The subarctic and subtropical regions are separated by Subarctic Front stretched out across the Japan Sea (Figs. 4a – 4l and 5). SSH rises in the northern tip of Tatarsky Strait in April – May (Figs. 4d and 4e) due to the inflow of buoyant fresh water from Amur River Estuary.

Surface currents in the entire Japan Sea strengthen from spring through late summer and weaken from autumn through late winter (compare Fig. 5a and 5b). Accordingly, SSH spatial gradients are the highest in August (Fig. 4h) when the surface circulation is the most intense. They are the lowest in February when the surface circulation is the weakest (Fig. 4b). In particular, the surface cyclonic gyres are stronger in the warm season than in the cold season (Fig. 4). Possible mechanisms are discussed by Trusenkova & Ishida¹²⁾.

Due to the external forcing applied, the simulated EKWC is strong in winter (Fig. 5a). The anticyclonic EKWC meander persists throughout the year in the southwest Japan Sea and another developments in the northwest Sea from October through De-

Fig. 4 Sea surface height (cm) for every month on the 12th year of integration. Contours are shown every 2 cm.

Fig. 5 Currents in the upper model layer on January 15 (a), and July 15 (b) on the 12th year of integration.

cember (Figs. 4j – 4l), extending towards the North Korea coast under the forcing of local anticyclonic wind stress curl¹²⁾. It will be shown that spatial patterns of the simulated surface heat and freshwater fluxes, SST, and SSS reflect the circulation patterns, due to heat and salt advection by the currents.

(2) Simulated and long-term heat fluxes at the Japan Sea surface

Figs. 6 and 7 represent the heat fluxes averaged over the Japan Sea surface in such a way that their positive values indicate that the Sea gains heat from the air and negative values indicate that the Sea loses heat to the air. Fig. 6 shows monthly timeseries of the six kinds of fluxes for the period from 1st through 12th year of integration. The short-wave solar radiation (open circles in Fig. 6) is taken from the long-term data (see Section 2) and reveals an annual cycle but no interannual variation. The long-wave radiation emitted by the sea surface (thick gray line in Fig. 6) is calculated by the MHI model. The net radiative flux (closed circles in Fig. 6) is a sum of the climatic short-wave and simulated long-wave radiation. Two turbulent heat fluxes, sensible (squares in Fig. 6) and latent (diamonds in Fig. 6) fluxes are simulated by the MHI model. The net heat flux (see Section 2) is shown by the thick black line in Fig. 6.

The simulated surface fluxes manifest regular annual cycles, without any considerable change of regime after a year and half of the initial spin-up (Fig. 6). It is seen in Fig. 6 that at the Japan Sea surface, the summertime heat gain is less than the wintertime heat loss, in line with the observation-based estimates¹⁾. In December, the simulated long-wave ra-

diation exceeds short-wave solar radiation, resulting in the negative net radiative flux at the Japan Sea surface (Fig. 6). The observation-based net radiative flux is negative in December – February¹⁾.

For better comparison with observation data, Fig. 7 represents both simulated and long-term, observation-based, average surface heat fluxes for every month of a year. The former are additionally averaged from the 3rd through 12th year of integration (for every month of a year) and shown with solid lines. The latter are reproduced from Hirose *et al.*¹⁾ and shown with dashed lines. The long-wave fluxes are shown in thick gray lines, the sensible fluxes are shown with open squares, the latent fluxes are shown with closed circles, and the net heat fluxes are shown in thick black lines. As seen from Fig. 7, the simulated and observation-based fluxes are consistent with each other, although some quantitative disagreement occurs. Next, we compare all kinds of the simulated and observation-based fluxes in detail.

The simulated long-wave radiation (gray solid line) is somewhat higher in magnitude than the observation-based flux (gray dashed line), especially from February through July (Fig. 7). Still, they both show the same annual cycle: the simulated (observation-based) long-wave radiation has the lowest magnitude in August (July) and the highest magnitude in October – February (December – January).

Energy emitted by a body is proportional to the 4th power of its absolute (Kelvin) temperature. However, it is known that the long-wave radiation from the sea surface is the most sensitive to the cloud cover⁶⁾. Clouds reflect and also absorb and re-emit radiation, so a portion of the emitted radiation is re-absorbed by the sea surface, reducing the long-

Fig. 6 Average short-wave radiation, long-wave radiation, net radiative flux, sensible heat flux, latent heat flux, and net heat flux (W/m^2) during 12 years of integration. At this and subsequent charts, fluxes are negative if the sea loses heat to the air and positive if the sea gains heat from the air.

Fig. 7 Simulated (solid lines) and observation-based (dashed lines; reproduced from Hirose *et al.*¹⁾) longwave radiation (thick gray lines) and sensible (open squares), latent (closed circles), and net (thick black lines) heat fluxes (W/m^2) for every month of a year.

wave flux. Summertime cloud cover is 65 – 80%, with only slight variation over the Japan Sea, on monthly time-scale. In contrast, average cloudiness during the winter monsoon season is 60%, increasing from 50% near the northwest coast to 70% near the southeast coast. Seasonal variation of the long-wave flux (Fig. 7) follows that of cloudiness. The quantitative gap between two estimates of the long-wave radiation may be explained by the use of different cloudiness datasets. The simulated long-wave radiation does not reveal considerable spatial variation over the Japan Sea, in line with Hirose *et al.*¹⁾.

Two curves for the average sensible flux (open squares in Fig. 7) closely follow each other, although the simulated flux is lower in magnitude in January and February. The average simulated (observation-based) sensible flux is the highest (in magnitude) in December (January). In the northwest Japan Sea, the observation-based sensible flux is the highest in December¹⁾. In April - July, the average sensible flux is close to zero (Fig. 7).

Both simulated and observation-based latent heat fluxes (closed circles in Fig. 7) are directed from the sea to the air throughout the year and exceed the sensible flux in magnitude. They manifest the same pattern of seasonal variation and both are the highest in November (Fig. 7). In the beginning of a cold season, the air is cool (but not too cold and capable

to contain enough moisture), wind is rather strong, and SST is not yet low. These conditions result in the highest latent heat flux in November.

The quantitative agreement between two estimates of the average latent flux is the best from March through early May and from mid-August through October. The simulated latent flux is lower (in magnitude) than the observation-based flux during the period from mid-November through February and higher from mid-May through July (Fig. 7). The observation-based flux does not considerably change from May to July when it is the lowest (in magnitude) throughout a year. In contrast, the simulated latent flux is the lowest in late April - early May and its magnitude increases in June and July. The gap between two estimates is equal to 30 W/m^2 in July (Fig. 7). This gap could be associated with the difference in SST, air temperature, air relative humidity, or wind speed applied in the simulation and by Hirose *et al.*¹⁾. Two estimates of the sensible heat flux are close in summer (Fig. 7), so the disagreement in the latent flux may be attributed to the use in the simulation of relative humidity for 2000, while Hirose *et al.*¹⁾ used the long-term data.

Both long-term and simulated net heat fluxes (thick black lines in Fig. 7) are positive from April through August when the Japan Sea gains heat from the air. They both are negative from October through February when the Sea loses heat to the air. The reversal of the surface net heat flux occurs in March and September in both cases (Fig. 7). In both cases, the surface heat gain during the warm season is less than the surface heat loss during the cold season and the heat balance in the Japan Sea must be maintained by the inflow of warm water in the Korean Strait¹⁾.

Two estimates of the net heat flux are very close during the transition months from early March through April and from early September through mid-December. The net heat flux is the highest (in magnitude) in December in both cases but the simulated flux sharply decreases (in magnitude) from mid to late December, while the observation-based flux is still high in January (Fig. 7). The surface heat gain is the highest in May in both cases. This can be attributed to the highest insolation in May and explained by the fact that cloudiness over the Japan Sea is lower in May than in June and July¹⁾. From mid-May through late July, the observation-based net heat flux remains high, while the simulated flux drops in June and July (Fig. 7). This can be explained by the gap in the latent and long-wave fluxes, which, in turn, can be caused by the use of different long-term datasets for some atmospheric variables and by application in the simulation of some data for 1980 – 2000. Moreover, there is a

quantitative disagreement between different observation-based estimates based on different datasets¹⁾.

Fig. 8 represents spatial patterns of the simulated turbulent (sensible and latent) heat fluxes for one month of every season of the year. February, May, and November are chosen because they can be attributed to the mid-months of the maritime winter, spring, and autumn, respectively. However, the summertime spatial patterns are presented for July rather than for August (mid-month of the maritime summer) because the latent flux starts increasing in August, while the sensible flux does not considerably change from July to August (**Fig. 7**). The sensible and latent heat fluxes are proportional to the air – sea difference in temperature and vapor pressure, respectively. Vapor pressure, in turn, exponentially depends upon temperature (see **Section 2.1**), which makes the latent heat flux more sensitive to the air – sea temperature difference. Monthly air temperature applied in the simulation is characterized by the smooth spatial pattern over the Japan Sea, while the simulated SST reflects the circulation patterns due to heat advection by the currents (see below).

In spring and summer when the air – sea temperature difference is not high, the simulated sensible flux does not show considerable spatial variation over the Japan Sea (**Figs. 8c and 8e**), as is the case for the observation-based sensible flux¹⁾. Spatial

variation of the long-term latent flux is not high in spring and summer, either: it is less than 20 W/m^2 over the entire Japan Sea in July¹⁾. Due to the SST increase within the warm currents, the simulated latent flux is increased within the EKWC (along the southwest coast) in spring and summer and within the TWC (along the southeast coast) in summer (**Figs. 8d and 8f**). In particular, the simulated latent flux exceeds 100 W/m^2 within the EKWC meander in July. The simulated latent flux is decreased inside the cyclonic gyre between 43°N and 48°N in July and August (**Fig. 8f** for July).

In autumn and winter, the simulated sensible and latent heat fluxes are increased in the warm subtropical region (**Figs. 4, 8a, 8b, 8g and 8h**), as is the case for the observation-based fluxes¹⁾. The spatial distribution of both simulated fluxes reflects the circulation patterns. In particular, the high flux cores are simulated within the EKWC meanders in the southwest and northwest Japan Sea (**Figs. 4, 8a, 8b, 8g and 8h**). Both simulated and observation-based turbulent fluxes are decreased in the cold northwest Japan Sea. However, the wintertime sensible flux is increased in Tatarsky Strait (**Fig. 8a**), where air temperature is very low. The latent flux remains low in Tatarsky Strait, as the cold air cannot contain much moisture (**Fig. 8b**). Bowen ratio (sensible to latent flux ratio) is lower than unity over most of the Japan Sea both in the simulation and observation-

Fig. 8 Simulated sensible (a, c, e, g) and latent (b, d, f, h) heat fluxes (W/m^2) for February (a, b), May (c, d), July (e, f), and November (g, h) on the 12th year of integration. Contours are shown every 20 W/m^2 .

based estimates (Fig. 7). However, Bowen ratio exceeds unity in Tatarsky Strait in winter (Figs. 8a and 8b; Hirose *et al.*¹⁾).

The spatial patterns of the sensible and latent heat fluxes obtained with the use of the MHI model in the previous simulations⁸⁾, under the forcing of weak constant wind with the speed of 5 m/s, and presently (Fig. 8), under the monthly wind forcing, are consistent with each other. This implies the dominant role of the air – sea temperature difference in the spatial variation of the turbulent heat fluxes. Stronger monthly wind applied in the present simulation results in substantially higher flux magnitude.

(3) Simulated and long-term temperature at the Japan Sea surface

Temperature in the upper model layer (simulated SST) reveals regular seasonal variation, without any drift after 1.5 years of the initial spin-up (not shown). Seasonal variation and spatial patterns of the simulated SST are consistent with those of the long-term SST from the Oceanographic Atlas of Rostov *et al.*⁵⁾. The latter is based on the US Navy's Generalized Digital Environmental Model (GDEM) dataset. This dataset is different from that used by Hirose *et al.*¹⁾ for estimation of the surface heat flux. For comparison, both simulated and long-term SST is presented in Figs. 9 and 10.

Fig. 9 represents the simulated and long-term SST (the curves with circles) averaged over the Japan Sea. The simulated SST is shown with the solid line and the long-term SST is shown with the dashed line. (The explanation on the SSS also shown in Fig. 9 is presented below). Both simulated and long-term SST is the highest in August (22.3°C and 22.1°C, respectively) and the lowest in February (3.8°C and 5.6°C, respectively), lagging 1 to 3 months behind the extreme surface heat flux (Figs. 7, 9 and 10). The quantitative agreement between the two SST estimates is the best in August, at the top of the warm season. The simulated SST exceeds the long-term SST during the warming season from April through July, while the long-term SST exceeds the simulated SST during the cooling season from September through December (Figs. 9 and 10). The gap is from 2°C to 3°C in spring and from 3°C to 5°C in autumn. The excess of the simulated over long-term SST is consistent with the excess of the simulated over the observation-based latent and long-wave fluxes in June and July (Fig. 7).

The simulated and long-term net heat fluxes are close from March through May and from August through September, while the former is lower than the latter in June and July (Fig. 7). So, the gap in SST cannot be related to an excess of the simulated surface heat flux. Heat advection counteracts the

SST gap, as the inflow transport in the Korean Strait applied in the simulation is low from April through June and high from August through December (Table 1). However, temperature in the upper mixed layer is defined by its thickness no less than by the surface heat flux and heat advection. The seasonal deepening or shallowing of the upper layer is simulated with the TKE equation (see Section 2.(1)).

In August, the simulated and long-term SST is in good agreement (Fig. 9) and the simulated depth of the upper layer is about 10 m from late July through early August (not shown), consistent with observations. Too fast seasonal warming or cooling of the upper layer can indicate that it shallows too fast or deepens too slow during the transition months. Imperfect estimation of the mixed layer depth can be, in part, explained by application of monthly wind speed. Averaging on monthly time-scale attenuates atmospheric forcing of the sea, resulting in the shallower mixed layer during the warm season⁶⁾. The TKE model is a semi-empirical model with some adjustable constants (see Section 2.(1)), so further model tuning may be needed. This is a simple one-dimensional model with the assumption of linear buoyancy flux in the vertical, so other models with more complete physics would perform better¹⁰⁾.

Fig. 10 represents spatial patterns of the simulated and long-term SST for every other month of a year. The simulated SST is shown for the 12th year of integration. In every month of a year, both simulated and long-term SST decreases over the Japan Sea from the southeast towards northwest coast (Fig. 10). The simulated SST reflects circulation patterns due to heat advection by the currents. As estimated in the previous simulations with the MHI model⁸⁾,

Fig. 9 Average simulated (solid line) and long-term (dashed line; reproduced from Rostov *et al.*⁵⁾) SST (°C; circles) and SSS (ppt; squares) for every month of a year.

Fig. 10 Simulated (a-f) and long-term (g-l) SST (°C) for January (a, g), March (b, h), May (c, i), July (d, j), September (e, k), and November (f, l). Contours are shown every 1°C. Long-term SST is reproduced from Rostov *et al.*³⁾.

heat advection provides considerable contribution to the heat transport in the Japan Sea. The simulated SST is increased within the warm currents and decreased inside the cyclonic gyre (Figs. 4, 5 and 10). In particular, SST is increased in the EKW meanders in the southwest and northwest Japan Sea during the cold season (Figs. 4j – 4l, 5a, 10e and 10f). The spatial patterns of the simulated sensible and latent heat fluxes, in turn, reflect the circulation patterns within the warm currents (Figs. 8g and 8h).

The simulated SST is lower than the long-term SST⁽⁵⁾ in late winter – early spring over most of the Japan Sea. Temperature of water entering the Korean Strait is accepted as 13°C in February. In the southern Japan Sea, the long-term and simulated SST varies from 13° to 10°C and from 11° to 8°C, respectively, in February (Figs. 10a and 10g). Simulated convection depth varies from 70 to 150 m in the southern Sea, which is consistent with observations. The observation-based and simulated net heat fluxes are the same from September through December (Fig. 7). So, the wintertime gap between the simulated and long-term SST in the southern Japan Sea can be attributed to the decreased supply of warm water in the Korean Strait, accepted in this simulation (Table 1). The lower simulated SST in January and February results in lower turbulent and net heat fluxes, compared to observations (Fig. 7).

In February, the simulated SST is under 1°C in the northwest Japan Sea (outlined by dashed line in Fig. 1) and under 0°C and -1°C in southern and northern Tatarsky Strait, respectively (Fig. 10a). In contrast, the long-term GDEM SST⁽⁵⁾ is higher than 0°C in Tatarsky Strait and higher than 1°C in the northwest Sea (Fig. 10g). However, very cold surface water was reported in wide areas of the northwest Japan Sea in winter from 1996 through 2000^(15–17). Water colder than 1°C was found northward of 41°N between 132°E and 134°E in late February 1996⁽¹⁵⁾ and also in the most of the northwest Japan Sea in March 1999⁽¹⁶⁾ and in late February to early March 2000⁽¹⁷⁾. In 2000, water colder than 0°C was stretched out along the northwest coast from 46°N to Vladivostok. Northern Tatarsky Strait is covered by ice in winter but the freezing point is not reached in this simulation. The area adjacent to North Korea (westward of 130°30'E) is characterized by higher SST (Fig. 10a). It is separated from the cold region by the thermal front documented from observations and called the northwestern front⁽¹⁵⁾. Similar front was simulated by Yoon and Kawamura⁽⁴⁾ as a part of the recirculation inside the cyclonic gyre. In the present simulation, this front is associated with an anticyclonic meander of the EKW Extension⁽¹³⁾ (Fig. 5a). Warm water is located southwestward of this front inside the meander.

(4) Surface freshwater flux in the Japan Sea

Surface freshwater flux is defined as the precipitation – evaporation ($P - E$) residual. As evaporation is proportional to the latent heat flux, its realistic simulation justifies the simulated surface freshwater flux, because precipitation is taken from observations⁽²⁾. Fig. 11a represents average timeseries of the $P - E$ residual from the 1st through 12th year of integration (Figs. 11b and 11c are related to the simulated salinity and will be discussed below). Fig. 12 represents the $P - E$ annual cycle averaged from the 3rd through 12th years of integration. The surface freshwater flux ($P - E$ residual) is positive when directed from the air to the sea (precipitation excess). It is negative when directed from the sea to the air (evaporation excess).

The regular annual cycles of the average $P - E$ residual for 12 years of integration manifest an excess of evaporation in autumn and winter and an excess of precipitation in spring and summer. The $P - E$ residual varies from -80 mm/month to 80 mm/month throughout the year, with the annual mean close to zero (Figs. 11a and 12). Spatial patterns of the surface $P - E$ residual are shown in Fig. 13 for one month in every season of a year. July and December are chosen as the months of the extreme freshwater flux and March and September are the transition months of zero average flux (Fig. 12).

The freshwater flux from the air to the sea is the highest in mid-July, as July and August are the months of the highest precipitation and evaporation is lower in July than in August (Fig. 7). In summer, the spatial patterns of the $P - E$ residual and precipitation resemble each other (compare Fig. 3b and Fig.

Fig. 11 Simulated freshwater flux (a; mm/month), salinity in the upper model layer (b; ppt), and average salinity in the entire Japan Sea (c; ppt). In this and subsequent charts, the freshwater flux is negative if directed from the sea to the air and positive otherwise.

13b). In July, the $P - E$ residual is the highest along the Korea coast where precipitation is the highest. The surface freshwater flux exceeds 120 mm/month in this area, despite increased evaporation (proportional to the latent heat flux) within the EKWC (Fig. 8f). The $P - E$ residual is especially high in the East Korean Bay, exceeding 200 mm/month in July (Fig. 13b). Due to decreased precipitation, the $P - E$ residual is decreased in the central Japan Sea and Tatarsky Strait, being under 60 mm/month in July (Figs. 3b and 13b). An area of the increased freshwater flux centered at $44^{\circ}30'N$, $138^{\circ}30'E$ in the northern Japan Sea can be explained by the decreased evaporation (latent heat flux) there (Figs. 8f and 13b).

The highest excess of evaporation over precipitation is simulated for December (Figs. 12 and 13d), lagging a month behind the highest evaporation (latent heat flux; Fig. 7), as the precipitation decrease overbalances the evaporation increase from November to December. From October through February, the spatial patterns of the $P - E$ residual follow those of evaporation (latent heat flux). This can be even seen in the comparison of spatial distribution of the latent flux for November (Fig. 8h) and the $P - E$ residual for December (Fig. 13d).

Fig. 12 Average annual cycle of the simulated surface freshwater flux (mm/month).

In the subtropical region, the freshwater flux from the sea to the air mostly exceeds 120 mm/month, with the maximum of 220 to 240 mm/month in December, while it is under 60 mm/month in the subarctic region (Fig. 13d). Precipitation is increased along the Honshu coast during the winter monsoon from September through April (Fig. 3a). In this area, an excess of evaporation over precipitation is low (under 80 mm/month) even in the period of the highest evaporation (from October through December). Along the Honshu coast, precipitation exceeds evaporation from January through August, while evaporation exceeds precipitation over most of the Japan Sea from January through March (Fig. 13).

The transition from the upward (downward) surface freshwater flux to the downward (upward) flux occurs in March (September), the very months of the surface heat flux reversal (Figs. 12 and 7). This can be explained by the coupling of the latent heat and freshwater fluxes. In both transition months, evaporation exceeds precipitation in the southwest and central southern areas and precipitation exceeds evaporation in the eastern and northern areas (Figs. 13a and 13c), while the average freshwater flux is close to zero (Fig. 12). Precipitation, evaporation, and their residual are considerably lower in March than in September.

(5) Simulated and long-term salinity at the Japan Sea surface

Fig. 11b represents timeseries of salinity in the upper model layer (simulated sea surface salinity; SSS) averaged over the Japan Sea. The average simulated SSS manifests stable annual cycles from the 3rd through 12th year of integration, consistent with the annual cycles of the surface freshwater flux (Figs. 11a and 11b). Namely, SSS is the lowest in August, lagging a month behind the highest down-

Fig. 13 Simulated surface freshwater flux (mm/month) for the late March (a), mid-July (b), early September (c), and mid-December (d) on the 12th year of integration. Contours are shown every 20 mm/month.

ward freshwater flux in July. The upward freshwater flux is the highest in December and SSS is the highest from January through March (Figs. 12 and 11b). For comparison with observation data, both simulated and long-term SSS are shown in Fig. 9 for every month of a year, with the curves marked by closed squares. The simulated SSS is additionally averaged from the 3rd through 12th year of integration and shown with the solid line. The long-term GDEM SSS⁵⁾ is shown with the broken line.

The average simulated and long-term SSS reveals the same pattern of seasonal variation. In both cases, SSS is the lowest in August (August and September for the long-term SSS) and the highest in February and March, lagging 1 to 2 months behind the extreme P – E residuals (Figs. 9 and 12). The highest average SSS is equal to 34.06 ppt in both cases and the minima are 33.25 ppt and 33.56 ppt for the simulated and long-term SSS, respectively (Fig. 9). The summertime simulated SSS is lower than the long-term SSS, with the gap of 0.1 to 0.2 ppt in June and of 0.2 to 0.3 ppt in July and August (Fig. 9).

The summertime SSS gap may be caused by the difference in the surface freshwater flux. The simulated summertime latent heat flux and evaporation are not underestimated, as discussed above. At the same time, a long-term increasing trend in springtime precipitation is documented over the adjacent subtropical land area for the second half of the 20th century¹⁸⁾. In particular, precipitation increase is found over Kushu and southern Honshu Islands for March – May and over the East China Sea for June¹⁸⁾. The GPCP precipitation²⁾ applied in this simulation is averaged from 1979 through 2001 and, therefore, can be related to 1990. The long-term GDEM SSS⁵⁾ can be attributed to the mid-1960s because oceanographic observations in the Japan Sea have been mostly performed after 1950. The springtime and summertime simulated SSS, lower than the long-term SSS, can be partly explained by the long-term precipitation increase. Besides, in the transition months in spring (May to June) and autumn (November to December), the simulated SSS decreases or increases too fast, compared to the long-term SSS, as is the case for SST (Fig. 9) and due to the same causes (see Section 3.(3)). Still, considerable gap in the summertime SSS points out to an unsolved problem in the simulation.

Spatial patterns of the simulated and long-term SSS represented in Fig. 14 for every other month of a year are consistent with each other. From January through March, both are the most spatially homogeneous throughout the year and mostly exceed 34 ppt (Figs. 14a and 14b for February). SSS is the highest in the southern Japan Sea where evaporation is the highest (Fig. 13d) and the saline water (34.6 to 4.65

ppt) is advected from the Korean Strait. In both cases, wintertime SSS is decreased along the Honshu coast due to increased precipitation. From January through April, the long-term SSS is less than 34 ppt along the northwest coast from Tatarsky Strait to North Korea (Figs. 14g and 14h). The simulated SSS is decreased along the northwest coast in December but not in January or February (Figs. 14a and 14f).

From June to October, both simulated and long-term SSSs are increased in the central Japan Sea and decreased along the coasts and in northern Tatarsky Strait. Generally, the summertime SSS pattern mirrors that of the P – E residual: both simulated and long-term SSS are decreased where the residual is increased (Figs. 13b, 14d, and 14j). The exception is northern Tatarsky Strait where SSS is very low despite the decreased P – E residual, due to the inflow of the Amur River water.

The simulated SSS is very low in the East Korean Bay where the P – E residual is the highest (Figs. 13b and 14d). In summer, fresh surface water is supplied to the southwest Japan Sea from the Yellow Sea through the western Korean Strait. From July through September, the long-term SSS is the lowest (32.4 to 32.6 ppt) in the western Korean Strait and along the adjacent coast of southern Korea Peninsula, while it is equal to 32.8 – 32.9 ppt in the East Korean Bay (Fig. 14j). Boundary condition accepted in this simulation for the upper layer salinity in the western Korean Strait is 32.8 ppt in August, intentionally higher than the long-term SSS. The simulated SSS varies from 32.8 to 32.9 ppt along the southwest Korea coast and from 32.4 to 32.6 ppt in the East Korean Bay (Figs. 13b and 14d). This confirms that SSS in the East Korean Bay is determined by the surface freshwater flux.

Strong negative salinity anomaly, with SSS of 33 ppt, is simulated in northern Tatarsky Strait in March, caused by the fresh water inflow from Amur River Estuary (not shown). It expands in April (with SSS of 27 to 29 ppt; Fig. 14b) and May (27 to 29 ppt; not shown). With the cyclonic gyre strengthening in May¹²⁾, negative salinity anomaly is advected southwestward along the northwest coast no farther west than 134°E. The anomaly gradually diffuses, with SSS reaching values of 30 to 31 ppt by June, 32 to 33 ppt by July, 33.4 to 33.6 ppt by October, and 33.8 to 33.9 ppt by December, and finally disappears in late January (Fig. 14).

In northern Tatarsky Strait, the long-term SSS is the lowest from March through July, which can be attributed to the inflow of the fresh Amur River water (Figs. 14h – 14j). The long-term SSS is decreased along the northwest coast from Tatarsky Strait to Korea Peninsula throughout the year (Figs.

Fig.14 Simulated (a-f) and long-term (g-l) SSS (ppt) for January (a,g), March (b,i), May (c,d), July (d,j), September (e,k), and November (f,l). Countours are shown every 1 ppt. Long-term SSS is reproduced from Rostov *et al.*³⁾

14g – 14l). The “fresh” zone off Vladivostok (the sea area adjacent to Peter the Great Bay; **Fig. 1**) is important for the formation of intermediate water of decreased salinity. This zone is often attributed to the southwestward advection of the Amur River water^{3),4)}. However, this simulation provides no evidence of direct advection of the SSS anomaly from Tatarsky Strait to the area off Vladivostok. The decreased SSS off Vladivostok should be associated with the fresh water supply from Peter the Great Bay although it is not reproduced in this simulation because this source of fresh water is neglected. SSS also decreases during ice melting in northern Tatarsky Strait and small bays along the northwest coast.

As assimilation of oceanographic data is not applied, the simulated salinity is entirely controlled by the external forcing (precipitation – evaporation residual and inflow from the Korean and Tatarsky Straits), although some freshwater sources (small river discharge and land run-off) are neglected. Evaporation is estimated by the MHI model. Precipitation and salinity in the inflow ports (Korean and Tatarsky Strait) are given. SSS in the outflow ports (Tsugaru and La Perouse Straits) is relaxed towards that in the adjacent sea area. It is interesting to consider how the salt budget in the Japan Sea is maintained in this modeling setup.

Fig. 11c represents timeseries of average simulated salinity in the entire Japan Sea for 12 years of integration. During this period, it varies in the narrow range from 34.05 to 34.07 ppt and reveals seasonal variation with the amplitude of 0.005 ppt, synchronized with the SSS seasonal signal (**Figs. 11b** and **11c**). The annual mean salinity slightly increases from the 1st through the 4th year of integration and decreases afterwards, dropping from 34.065 ppt to 34.055 ppt by the 12th year of integration, while the simulated SSS reveals no interannual trend (**Fig. 11b**). So, this trend is, probably, caused by the surface water subduction into the intermediate and deep layers. Minor drift of average salinity implies that the periodical quasi-steady regime, with respect to salt content, is reached in this simulation performed under the forcing of external sources of fresh (or saline) water, without assimilation of oceanographic data on salinity.

4. CONCLUSIONS

(1) The simulated annual cycle and spatial patterns of the radiative and turbulent (sensible and latent) heat fluxes at the Japan Sea surface are in good agreement with the long-term fluxes¹⁾. The model Japan Sea loses heat to the air in the cold season,

with the highest flux in December, and gains heat from the air in the warm season, with the highest flux in May, as is the case in reality¹⁾. The reversal of the surface heat flux occurs in March and September in both cases. The simulated heat loss at the sea surface exceeds the gain and the heat balance in the Japan Sea must be maintained by the warm water inflow from the Korean Strait, in line with the observation-based results¹⁾. The quantitative agreement between the simulated and long-term¹⁾ surface fluxes is the best in the transition months from winter to summer and from summer to winter. Smaller magnitude of the simulated wintertime heat flux, compared to the long-term flux, can be explained by insufficient heat supply (volume transport) in the Korean Strait accepted in the simulation. Considerable summertime gap between two estimates can be partly explained by the use of different long-term datasets and by application in the simulation of some data for the late 20th century. The same is true for SST and SSS. However, the reproduction of the summertime heat flux and SSS needs improvement.

(2) Annual cycle and spatial patterns of the simulated and long-term SST are consistent with each other, even if the data applied in the simulation, data used by Hirose *et al.*¹⁾, and GDEM SST dataset⁵⁾ are different. The quantitative agreement is the best in August, at the top of the warm season. In the transition months from winter to summer and from summer to winter, the simulated upper layer warms or cools too fast. With the same simulated and long-term fluxes in these months, this gap can be explained by the difference in the upper layer depth. Steering the upper layer by application of the synoptic wind forcing and tuning the TKE model^{9),10)} can help in improving the results. Low wintertime heat supply from the Korean Strait accepted in this simulation results in the lower simulated SST and surface heat flux in the southern Japan Sea from January through March, compared to the long-term data^{1),5)}. The simulated SST in the northwest Japan Sea and Tatarsky Strait is in agreement with the data of wintertime oceanographic cruises in 1996 – 2000¹⁵⁾⁻¹⁷⁾.

(3) The simulated surface freshwater flux (precipitation – evaporation residual) is justified by the realistic latent heat flux (which evaporation is proportional to), as precipitation is prescribed. The simulated evaporation from the Japan Sea surface exceeds precipitation in autumn and winter, with the highest excess in December, lagging a month behind the highest evaporation in November. Precipitation exceeds evaporation in spring and summer, with the highest excess in July. The reversal of both freshwater and heat fluxes occurs in March and September. Spatial patterns of the surface freshwater flux mostly follow evaporation patterns in winter and

mirror precipitation patterns in summer. In the area adjacent to the Honshu coast, the freshwater flux is controlled by precipitation even in winter. The highest downward freshwater flux is simulated in July in the East Korean Bay.

(4) Seasonal variation and spatial patterns of the SSS simulated without assimilation of salinity data are consistent with those of the long-term GDEM SSS⁵⁾. The SSS spatial patterns mostly mirror those of the simulated freshwater flux, with the SSS decrease in the areas of the flux increase and vice versa. However, the highest wintertime SSS is associated with the inflow of the saline Kuroshio water from the Korean Strait. Both simulated and long-term SSS is decreased along the Honshu coast due to the increased precipitation. The simulated SSS is the lowest in the East Korean Bay in July due to the highest freshwater flux. Negative SSS anomaly persists there throughout the year. In northern Tatarsky Strait, SSS is decreased from March through July, despite the decreased surface freshwater flux. It can be explained by the fresh Amur River water, although the runoff from the adjacent land can be also important. However, the latter effect is hard to estimate. The decreased long-term SSS southward of Vladivostok, important for the formation of intermediate salinity minimum^{3),4)}, cannot be explained by the advection of the Amur River water and should be associated with the fresh water supply from Peter the Great Bay. The quantitative agreement between the simulated and long-term SSS is the best from January through March. The summertime SSS gap can be partly explained by the long-term precipitation increase¹⁸⁾ but still represents a problem to be solved in the future simulations.

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