Contribution of Oceanic Circulation to the Poleward Heat Flux

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Abstract Oceanic contribution to the poleward heat flux in the climate system includes two components: the sensible heat flux and the latent heat flux. Although the latent heat flux has been classified as atmospheric heat flux exclusively, it is argued that oceanic control over this component of poleward heat flux should play a critically important role. The so-called swamp ocean model practice is analyzed in detail, and the critical role of oceanic circulation in the establishment of the meridional moisture transport is emphasized.

Key Words poleward heat flux; climate; air-sea coupled model; freshwater flux; hydrological cycle; swamp model

Poleward heat flux in the climate system has been one of the main focuses of climate study. As the technology improves, heat flux estimates have been improved over the past decades. A recent study by Trenberth and Caron (2001) showed that the atmospheric component of the poleward heat flux dominates the total poleward heat flux at high latitudes. As stated by the authors in the abstract of their article, 'At 35° latitude, at which the peak total poleward transport in each hemisphere occurs, the atmosphere transport accounts for 78% of the total in the Northern Hemisphere and 92% in the Southern Hemisphere. In general, a much greater portion of the required poleward transport is contributed by the atmosphere than the ocean, as compared with previous estimates.' This quoted argument implies that the oceans would be unimportant for high-latitude climate in terms of carrying the heat flux much needed for warming up high latitudes. Is this true? In particular, to the south of 35°S the oceans cover 90% of the total surface area, but, according to the statement cited above, they would contribute only 8% of the total poleward heat flux. It seems incomprehensible how the oceans could play such a minor role in transporting heat flux poleward.

1 Definition of Poleward Heat Flux

Oceanic heat flux has been discussed in many papers and textbooks. As both instruments and numerical models are improved, poleward heat fluxes are better

diagnosed. For the most updated information, the reader is referred to the review by Bryden and Imawaki (2001). The theme of this note is to explore the physical meaning of the poleward heat flux as presented in many recent studies, such as Trenberth and Caron (2001); and we will assume the heat flux data presented in their study are accurate enough for our purpose. As the technology improves in the near future, the estimates of different components of the heat flux will be improved; however, our discussion here should be independent of the specific data used.

The main point is that, although heat flux data may be further improved, there seems to be a fundamental problem in the definition of poleward heat fluxes that gives rise to misconception and similar mistakes have been made by many people. From thermodynamics, it is well known that heat flux cannot be well defined for a system that has a net mass gain (or loss) through its lateral boundaries. For example, Trenberth and Caron (2001) made a point that the poleward heat fluxes in the South Pacific and Indian Oceans are not well defined because of the existence of the Indonesian Throughflow.

The same caution should be applied to both the oceans and atmosphere because they are not closed systems in terms of mass — there is water exchange between them. In fact, the latent heat flux associated with the hydrological cycle in the form of evaporation and precipitation is one of the essential ingredients of the heat flux in the atmosphere. The mass flux associated with evaporation and precipitation has been traditionally ignored in heat flux calculations by oceanographers for the following reasons: First, such a small mass flux is rather hard to identify from the classical

dynamic calculation based on either a reference level or by other means. Second, most oceanic general circulation models have been based on the Boussinesq Approximations; thus, the dynamic effects associated with evaporation and precipitation are simulated in terms of the virtual salt flux condition at the upper surface, while the mass flux associated with evaporation and precipitation is totally ignored. As oceanic general circulation models are improved, the natural boundary conditions (Huang, 1993) are now commonly used as the boundary condition for the salinity balance. As a result, the mass flux associated with e-vaporation and precipitation is exactly accounted for. A more accurate definition of heat flux is, thus, necessary, and to present such a definition is the main goal of this note. A simple definition of heat flux that can be used for an oceanic section with a net mass flux is given in the Appendix.

Strictly speaking, the poleward heat flux in the climate system should be defined in terms of three components, as shown in Fig.1:

1) Sensible heat flux in the oceans

2) Sensible heat flux in the atmosphere and

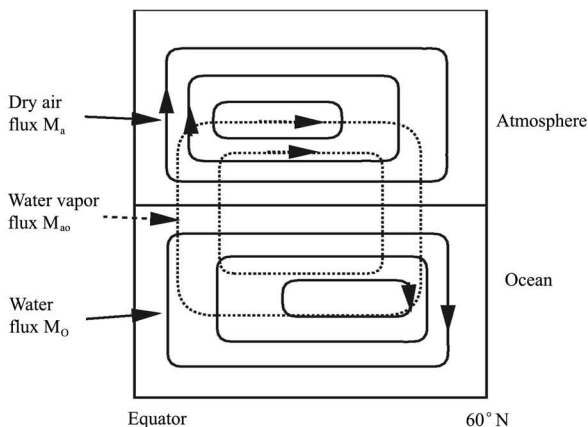3) Latent heat flux in the atmosphere-ocean-land coupled system.

Fig.1 Sketch of the atmosphere-ocean coupled climate system.

Note that it is probably more accurate to use the term of energy transport for the ocean, as suggested by Warren (1999). As discussed by Warren, the difference between the more accurate definition of energy flux and the heat flux based on the mean specific heat is no more than a few percent. Similarly, the heat flux in the atmosphere should be called the energy flux, including the contribution due to the gravitational potential energy transport, because in a compressible atmosphere the gravitational potential energy can be converted into internal energy. Thus, the sensible heat flux in the atmosphere in our discussion is actually the total 'heat flux' minus the latent heat flux.

The role of land in latent heat flux transport is relatively minor, compared with the other components. The oceans cover more than 70% of the total earth surface area. In addition, the evaporation rate from

the land is much smaller than from the oceans. Thus, the latent heat flux loop is basically an atmosphere-ocean coupled mode, with the land plays a minor additional role. In this note we will focus on the role of the dynamic oceans.

Note that over the latitudinal range from the equator to the pole there is a continuous exchange of heat between these three loops; however, discussion of the details of such exchange is beyond the scope of this note.

As an example, we calculate the freshwater flux through the air-sea interface. We use the evaporation and precipitation rates over the world oceans given by Da Silva *et al.* (1994) as reproduced in Fig.2. This dataset includes river run-offs, so the globally-integrated evaporation and precipitation rates are balanced. Near the equator precipitation dominates, especially between 0° and 10°N. However, evaporation dominates over the subtropics and thus produces a net water vapor flux that is transported by the atmosphere to high latitudes. At high latitudes (roughly 40° off the equator) precipitation dominates.
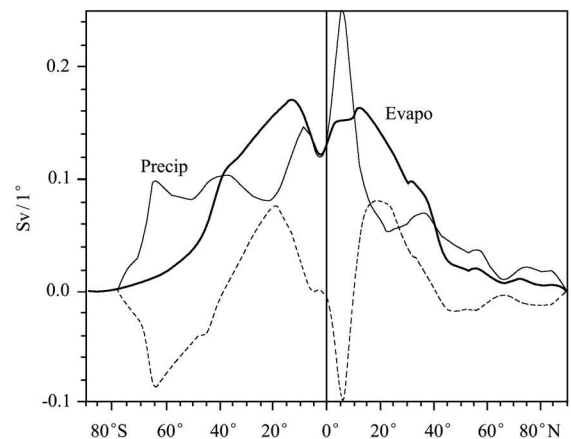

Fig.2 Water vapor sources and sinks due to evaporation and precipitation, integrated zonally, in units Sv/1° (1Sv = $10^9 \, \mathrm{kg\,s^{-1}}$) from Da Silva *et al.* (1994).

However, this dataset seems to give rise to a poleward freshwater flux that is too large in the Southern Hemisphere. As an alternative we use the water vapor flux calculated from the atmospheric circulation models (Gaffen *et al.*, 1997). The freshwater flux reported in this note is based on an average obtained from 25 atmospheric general circulation models. This poleward water vapor flux, M_w, is the major mechanism of poleward heat flux in the climate system (Fig.3).

The poleward heat flux associated with the water vapor cycle is closely related to the latent heat content of water vapor:

$$H_f = (L_e - h_w) M_w$$

where $L_e = (2500.8 - 2.3\,T) \mathrm{J}/g$ is the latent heat content of water vapor that depends on temperature only very slightly, h_w is the enthalpy of the returning water in the ocean, which is much smaller than L_e and

can be neglected in the calculation. The heat flux associated with the water vapor flux should be attributed to the ocean-atmosphere-land coupled system.
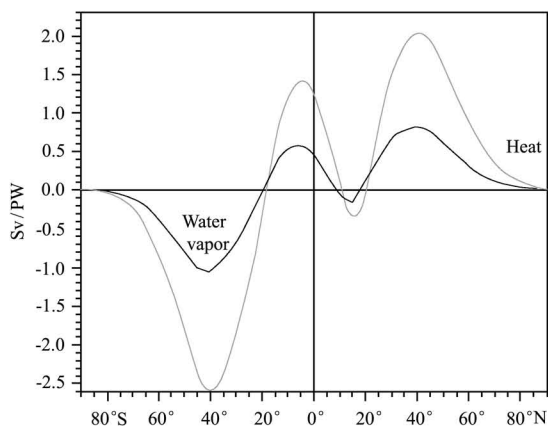

Fig.3 Northward water vapor flux (Sv) and associated latent heat flux (PW) due to evaporation and precipitation, as calculated from atmospheric general circulation models.

However, the mass flux associated with the water vapor flux is two orders of magnitude smaller than other mass fluxes in the climate system (Table 1). This is one of the reasons why this mass flux has been over-looked in many papers discussing the climate. On the other hand, this seemingly tiny mass flux is associated with a large heat flux. One kilogram of water vapor can deliver 2.5×10^6 J of heat. When one kilogram of water is cooled down by $10°C$, the heat released is 4.18×10^4 J. Thus, water vapor is 60 times more efficient than water in transporting heat.

In this note we use Sverdrup as a unit for mass flux. Sverdrup has been widely used as a volume flux unit in oceanography, and $1 Sv = 10^6 \, m^3 s^{-1}$. Since sea water density is very close to $10^3 \, kg m^{-3}$, as a mass flux unit we have $1 Sv = 10^9 \, kg \, s^{-1}$. As shown in Table 1, Sverdrup is a very convenient unit for the climate system because mass fluxes in both the oceans and atmosphere have the same order of magnitude, although the density of water is one thousand times larger than air (Held, 2001)[1]. The mass flux in the Jet Stream is inferred from the angular momentum. According to Peixoto and Oort (1992), the maximum angular momentum in the Northern Hemisphere is about $9.6 \times 10^{25} \, kg m^2 \, s^{-1}$. Assuming the mean latitude is at 30°N, this gives a rough estimate of 500 Sv for the mass flux associated with the westerlies in the Northern Hemisphere.

Table 1   Mass transport maxima in the climate system[†]

Current systems	Gulf Stream	Kuroshio	ACC	Meri. Cell in atmosphere	Jet Stream	Poleward water vapor
Mass flux/Sv	150	130	134	200	500	1

Note: [†] The sources of the values: Gulf Stream, Hogg and Johns (1995); Kuroshio, Wijffels *et al.* (1998); ACC (Antarctic Circumpolar Current), Rintoul *et al.* (1998); meridional overturning rate and the Jet Stream in the atmosphere, Peixoto and Oort (1992); poleward water vapor flux, Gaffen *et al.* (1997).

However, some people may say that the ocean plays a rather passive role in this loop, so the latent heat transport is basically an atmospheric process. They argue that the ocean transports water back to lower latitudes, without much heat content change along this part of the loop, while the atmospheric component of this loop is associated with the heat transport and release.

A close examination reveals that both the ocean and the atmosphere are equal partners in this heat-transport loop. If we have an imaginary system with a mass transport loop going through the atmosphere-ocean, where the atmosphere picks up heat from lower latitudes (within the atmosphere) and delivers it to higher latitudes, as shown in Fig.4 a, then the ocean would be a passive component indeed.

However, in reality the latent heat transport loop actually starts below the sea surface because evaporation takes place at the air-sea interface. At the molecular level, evaporation is a process in which water molecules with kinetic energies much higher than the average cross the air-sea interface. Since the water molecules left behind have less mean kinetic energies, at the macroscopic level evaporation draws heat from the ocean as shown in Fig.4 b.

This heat transport loop consists of four parts as shown in Fig.4 b: First, evaporation that absorbs heat from water; Second, the atmospheric transport leg during which there might be a very small decline in heat content; Third, precipitation during which latent heat is released to the atmosphere; and Fourth, the last leg during which water molecules return to lower latitudes in the ocean with a very small and gradual increase in heat content. Accordingly, both the atmospheric and oceanic legs can be viewed as passive, while the major action takes place below the sea surface and at the site of water vapor condensation. Thus, the heat transport process starts from the ocean, and it ends in the atmosphere, so the latent heat flux loop is really a coupled mode.

In the Northern Hemisphere the water vapor flux peaks at 40°N, with a mass flux of 0.84 Sv and a corresponding poleward heat flux of 2.1 PW, Fig.3. In the Southern Hemisphere, the poleward water vapor flux reaches its maximum at 40°S, with a mass flux of 1.05 Sv and a corresponding poleward heat of 2.64 PW. It is clear that the latent heat flux associated with evaporation and precipitation over the world oceans consists of a substantial portion of poleward heat flux

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① Held, 2001. Personal communication.

in the current climate system. From the hydrological cycle alone, we can estimate the poleward heat flux associated with latent heat as a major component of the poleward heat flux (Schmitt, 1995). Such a large poleward heat flux is, of course, intimately connected with the general circulation in the oceans.

a) An idealized climate system with a passive ocean

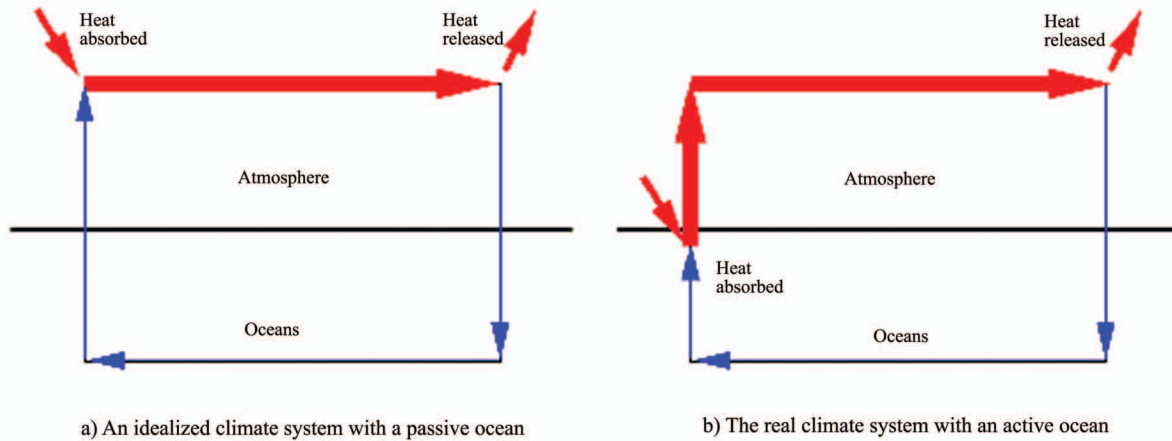b) The real climate system with an active ocean

Fig.4 Sketch of the latent heat flux loop, with the heavy lines indicating transport legs with high heat contents.

After introducing the poleward heat flux associated with water vapor cycle, the poleward heat fluxes in the current climate system are shown in Fig.5. The sum of these three fluxes is the total poleward heat flux diagnosed from satellite measurements. The oceanic sensible heat flux is the oceanic heat flux from the NCEP data (Trenberth and Caron, 2001). Although the atmospheric sensible heat flux is still the largest component in both hemispheres, it is not much larger than the other components.
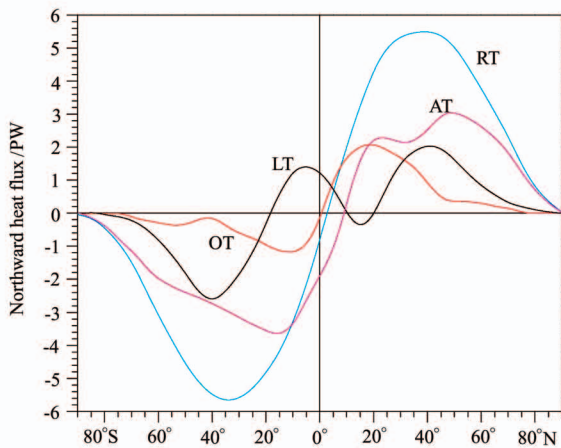
Fig.5 Northward heat flux in PW: RT indicates the heat flux required for radiation balance as diagnosed from satellite data; AT indicates the atmospheric sensible heat flux; OT indicates the oceanic sensible heat flux; and LT indicates the latent heat flux associated with the hydrological cycle in the atmosphere-ocean coupled system.

2 Heat Flux Divergence

Caution must be taken in the interpretation of the poleward heat flux. Heat flux itself may not be very important – what really matters is the divergence of the heat flux. For example, at 55°S the atmospheric sensible heat flux is larger than the latent heat flux. However, a close examination reveals that the latent heat flux divergence is larger than that of the atmospheric sensible heat flux (Fig.6).
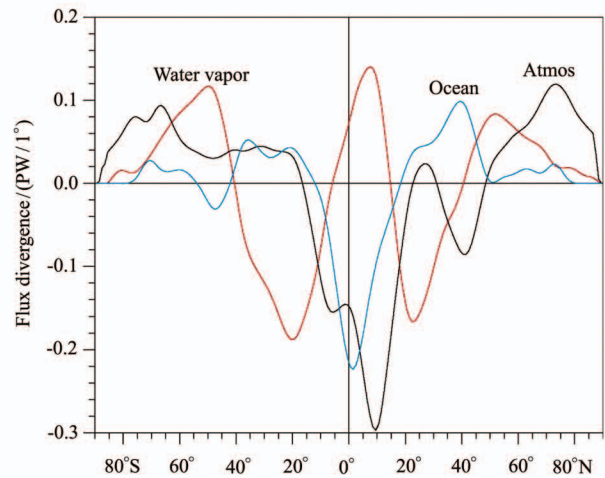
Fig.6 Heat flux divergences. Atmos indicates the atmospheric sensible heat flux divergence; ocean indicates the oceanic sensible heat flux divergence; and water vapor indicates the latent heat flux divergence associated with the hydrological cycle in the atmosphere-ocean coupled system.

At 35°S the atmospheric sensible heat flux is 3 times larger than the oceanic sensible heat flux; however, the divergence of both fluxes is about the same. In other words, their contribution to warming up the local climate is roughly equal. Therefore, heat flux divergence is a better quantity in diagnosing the contribution to local heat balance.

Similarly, in the Northern Hemisphere, the oceanic sensible heat flux divergence is the dominant source of heat south of 40°N, north of this latitude the divergence of latent heat flux becomes more important, and north of 50°N the atmospheric sensible heat flux domi-

nates.

In both hemispheres, poleward heat flux is carried by three components that work like a relay team. In the subtropics the oceanic sensible heat flux is the dominating contributor to the poleward heat flux divergence, and in mid latitudes the latent heat flux divergence is the dominating contributor. Finally, in high latitudes the atmospheric sensible heat flux divergence dominates (Fig.6). The situation in the Southern Hemisphere is an excellent example demonstrating the roles of these three components. The most noticeable term is the latent heat flux divergence that dominates over mid latitudes. The atmospheric sensible heat flux divergence plays a major role south of 70°S, where the Southern Ocean meets the Antarctica. The heat flux balance for a given water column is:

Short wave radiation + divergence of oceanic sensible heat flux = Air-sea heat fluxes (latent heat + long wave radiation + sensible heat).

It is important to note that the divergence of the oceanic sensible heat flux is relatively small in the oceanic heat flux balance at any given location (Fig.7). In comparison, latent heat released to the atmosphere is much larger than the divergence of the oceanic sensible heat flux at any given location. The latent heat flux released to the atmosphere is, thus, a continuous source of heat for the atmospheric circulation. Although the atmosphere may play an important role in sending heat to higher latitudes, without the continuous heat support from the oceans, the warm climate we enjoy there could not have been realizable at all.
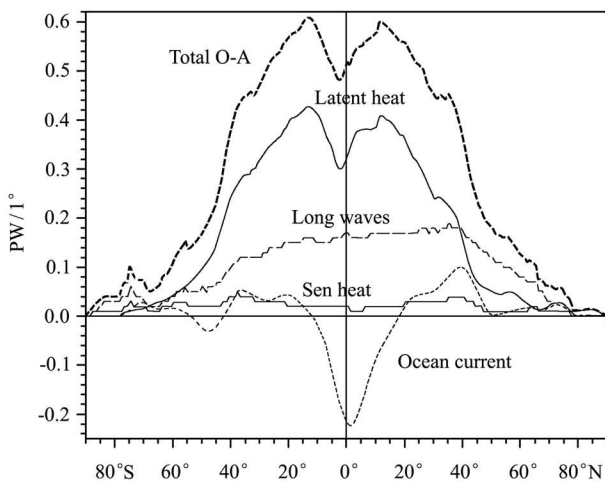

Fig.7 Heat fluxes from the ocean to the atmosphere, integrated zonally, and the divergence of the oceanic sensible heat flux, labeled as ocean current.

It is well-known that latent heat release from the ocean is very sensitive to the sea surface temperature. According to the commonly used bulk formula

$$Q_e = \rho_a L_e C_E (Q_s - Q_a) U_r,$$

where L_e is the latent heat content of the water vapor, C_E is the Dalton number, U_r is the wind speed, and

$$Q_s = Q_0 \exp\left(\frac{7.63 T}{273 + T} \right)$$

is the moisture vapor pressure at the sea surface, Q_0 is the saturated water vapor pressure at 0℃, T is the sea surface temperature in ℃. From this formula, latent heat flux is primarily controlled by the wind speed and the sea surface temperature, thus, changes in sea surface temperature can alternate the latent heat flux substantially. In the events of climate change, surface temperature and salinity changes will certainly affect the latent heat flux, and thus the entire climate system.

3 The Hydrological Cycle

In the climate system, solar radiation is primarily absorbed by the ocean. The atmosphere is primarily heated through evaporation and latent heat released to the atmosphere, plus the so-called greenhouse gases' absorption of heat in the atmosphere itself. As shown in Fig.7, the amount of latent heat flux from the ocean to the atmosphere is huge. The total amount of evaporation is 12.45 Sv, so the latent heat associated with evaporation is about 31.1 PW. As the oceanic circulation changes, the latent heat flux will be substantially modified; thus, we should study how this might happen.

The important role of the atmosphere-ocean hydrological cycle associated with evaporation and precipitation has been overlooked in the oceanographic literature. In oceanic heat flux calculations, the poleward heat flux associated with the hydrological cycle has been neglected, and it was incorrectly attributed to the atmosphere alone. Thus, what most oceanographers calculated in the past should have been called the oceanic sensible heat flux.

On the other hand, meteorologists have claimed that the poleward heat flux associated with the hydrological cycle is entirely due to atmospheric processes, and has very little connection with the ocean circulation. Such an entire separation of heat fluxes is inconsistent with physics. Without the oceanic currents to bring heat northward, there would be no high surface temperature in the Gulf Stream (or Kuroshio) region. As a result, there would be no strong latent heat release in these areas.

It is well known that the latent heat flux is a major component of poleward heat flux in the atmospheric circulation; however, this component has long been thought of as primarily an atmospheric process. The reason for such a misconception is, at least partially, due to the seemingly small amount of mass associated with this water vapor flux. The globally integrated meridional water vapor flux is on the order of 1 Sv, which is much smaller than the mass flux of other components of the general oceanic atmospheric circula-

tions (Table 1). Thus, the return branch of the global hydrological cycle, which takes place in the oceans, has been overlooked. This misconception is wide spread in many documents, papers, and books on climate.

First, although the main branch of the global hydrological cycle is through the oceans, many existing scientific programs related to the hydrological cycle completely ignore the oceanic component of the cycle.

Second, the thermohaline circulation in many existing climate models is not realistically simulated. For example, the oceanic component in many climate models is represented in terms of the so-called swamp ocean, *i.e.*, a very shallow layer of water (50 m or slightly deeper) with no current.

In such models the atmosphere takes up water vapor at low latitudes and transports it to high latitudes. After releasing the latent heat, the water has been discarded from the model, and little thought has been given to the dynamic consequences of this seemingly small amount of water flux. Obviously, an ocean without currents would be unable to transport fresh water from high latitudes, where precipitation prevails with the fresh water flowing back to lower latitudes, where evaporation prevails. Thus, without the oceanic currents, the subtropical ocean would eventually dry up, with all the water piling up at high latitudes – an equilibrium impossible to reach. It is very unlikely that when a quasi-steady state is reached such models can reproduce the strong thermohaline circulation in the oceans. Thus, the sea surface conditions, such as temperature, salinity, and air-sea heat fluxes, obtained from such models may be quite different from the present climate.

Many of the existing climate models have an oceanic component based on the so-called Boussinesq approximations. In such models, the ocean is treated as an incompressible fluid environment, and the salinity balance in the model is controlled by the so-called virtual salt flux condition at the sea surface, or the traditional technique of requiring the surface salinity relaxed to the climatological mean surface salinity. Changes in the water vapor cycle, thus, in such models have no dynamic consequence at all.

As the oceanic general circulation models are improved, the dynamic role of freshwater flux will be simulated more realistically. Among many other features, evaporation and precipitation can drive the so-called Goldsbrough-Stommel circulation (Huang and Schmitt, 1993) that is barotropic and with a horizontal mass flux on the order of a Sverdrup. This circulation is much smaller than the wind-driven or thermohaline circulation, therefore it has been neglected in most discussion about ocean circulation and climate. However, the seemingly small freshwater flux associated with evaporation and precipitation is responsible for establishing the haline circulation in the oceans. In fact, evaporation and precipitation alone (without

wind stress and heating) can induce a three-dimensional baroclinic circulation, two orders of magnitude stronger than the barotropic Goldsbrough-Stommel circulation, *i.e.*, the same order of strength as the wind-driven or the thermal circulation (Huang, 1993).

Thus, evaporation and precipitation comprise one of the major forces in setting up the oceanic general circulation. As the climate system shifts to a new state, the freshwater flux through the ocean-atmosphere system will change in response. As a result, the thermohaline circulation will be in a different state, and the sea surface temperature distribution will be different from the present-day distribution. Consequently, the surface air-sea heat fluxes shown in Fig.7 can be dramatically different. A comprehensive understanding will require much more study of the ocean-atmosphere coupled system.

4 The Feedback Between the Hydrological Cycle and the Meridional Thermal Circulation

One of the possible reasons for neglecting the freshwater flux in heat flux analysis could be the misconception that freshwater flux in the ocean is so small that it plays a minuscule role in controlling the heat flux. As will be shown shortly, however, the freshwater flux through the oceans plays an important role in controlling the strength of the meridional overturning and poleward heat flux.

Due to the evaporation at low latitudes and precipitation at high latitudes, there is a meridional salinity gradient. The meridional density gradient due to the salinity difference is opposite to the meridional density gradient due to the thermal forcing in the upper oceans. Thus, in the present climate setting, the hydrological cycle is a brake for the poleward heat transportation carried by the meridional thermally-forced circulation in the oceans. In other words, the poleward latent heat flux associated with the water vapor cycle has a negative feedback on the oceanic sensible heat flux.

If there were no hydrological cycle, the meridional overturning cell and the associated poleward heat flux in all five ocean basins would be intensified. As an example, in the North Atlantic the surface density difference between the equator and high latitudes is about $\Delta\sigma = 3.39$, based on the Levitus and Boyer (1994) data, Fig.8a. If there were no evaporation and precipitation, there would be very little salinity difference in the oceans, so we set salinity to a constant value of 35. Assuming the surface temperature distribution remains unchanged – a very bold assumption that is unlikely to be true in the real world – the corresponding surface density difference would be increased to $\Delta\sigma = 5.39$, Fig.8b. Due to the increase of the north-south density difference both the meridional overturning rate and poleward heat flux should increase.

Equator   70°N   Equator   70°N

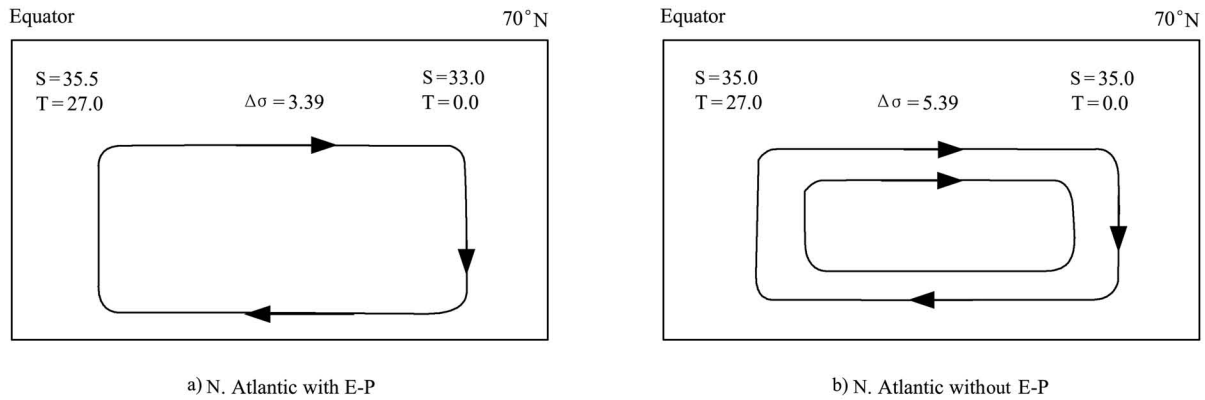a) N. Atlantic with E-P   b) N. Atlantic without E-P

Fig.8 Two models for the North Atlantic. (a) the standard model, forced by observed sea surface temperature and salinity; (b) a conceptual model with uniform salinity.

We carried out four numerical experiments for the Atlantic in order to explore the dynamic role of freshwater flux in the meridional circulation. The MOM2 (Pacanofsky, 1995) was used with a horizontal resolution of $2.5° \times 2°$ and 15 levels vertically, with a constant diapycnal mixing rate of $0.4 \times 10^{-4} \mathrm{m}^{-2} \mathrm{s}^{-1}$. The model's temperature and salinity were initialized with the Levitus and Boyer (1994) data, and in each experiment the model was run for 1 000 yr to reach a quasi equilibrium.

In experiment A both the surface temperature and salinity were relaxed toward the monthly mean climatology, along with the Hellermann and Rosenstein (1983) monthly mean wind stress data. In experiment B salinity was set to a constant value of 35, but all the other parameters remained the same as in experiment A.

In experiments C and D the density effect due to temperature was ignored. Thus, the salinity difference was the only force regulating the circulation. In order to calculate the poleward heat flux, temperature was treated as a passive tracer, with an upper bound-

ary condition such that the sea surface was relaxed toward the monthly mean climatology. In experiment C the surface salinity was subject to the surface relaxation condition, assuming that the climatological monthly mean sea surface salinity remained unchanged. In experiment D, the model's salinity was subject to the virtual salt flux diagnosed from the steady circulation obtained from experiment A.

The meridional overturning circulation (MOC) in the second experiment was 68% higher than in experiment A (Table 2), quite close to our scaling analysis. The poleward heat flux (PHF) in the second experiment was 21% higher than in experiment A, somewhat lower than the prediction from the simple scaling.

In experiments C and D, the meridional overturning cell reversed, and the associated meridional heat flux was equatorward, as shown in Table 2. The amount of equatorward heat flux was about 0.2 PW for the case with virtual salt flux. Thus, our numerical experiments demonstrated that the haline circulation is an important factor controlling the poleward heat flux.

Table 2   Four numerical experiments for the model Atlantic

	A	B	C	D
Temperature	SST relaxation	SST relaxation	Treated as a tracer	Treated as a tracer
Salinity	SSS relaxation	Set to S=35	SSS relaxation	Virtual salt flux
MOC (Sv)	12.58	21.10(+68%)	−13.27	−18.70
Poleward heat flux (PW)	0.72	0.87(+21%)	−0.07	−0.23

It is much more important to examine the poleward heat flux profile and the associated heat flux divergence. It is clear that in experiment B, the ocean transported much more heat to high latitudes. In particular, the poleward heat flux divergence was much larger for latitudes higher than 30°N (Fig.9). Thus, in the case without a hydrological cycle, the oceanic current will transport more heat poleward and release it to higher latitudes, partially compensating for the decline of poleward heat flux caused by the lack of latent heat flux associated with the hydrological cycle.

We also ran other experiments for the global oceans. The model had a low horizontal resolution of $4° \times 4°$ and 15 layers. The Indonesian Throughflow was included in these low-resolution simulations. The model's temperature and salinity were initialized with the Levitus and Boyer (1994) data and in each experiment. The model was run for 1 000 yr to reach a quasi equilibrium.

In the standard experiment both the surface temperature and salinity were relaxed toward the monthly mean climatology while the Hellermann and Rosen-

stein (1983) monthly mean wind stress data were used. In the second experiment salinity was set to a constant value of 35, but all other conditions remained the same as in the standard experiment. The poleward heat flux was calculated, using the definition discussed in the Appendix, so the heat flux in both the South Pacific and Indian is uniquely defined.

It is clear that by turning off the salinity effect, the total northward heat flux in the Northern Hemisphere increases, while it declines in the Southern Hemisphere (Fig.10). Most interestingly, the northward heat flux in the North Pacific increases substantially. This increase of heat flux at mid latitudes is probably associated with the newly-created thermal mode of the meridional circulation in the North Pacific, caused by the lack of salinity effect that opposes the thermal mode. The northward heat flux in the Atlantic declines due to the diminishing effect of the global conveyor belt that is driven, at least partially, by the salinity difference between the Pacific and Atlantic. As a result of the decline of the global conveyor belt, the poleward heat flux in the South Indian Ocean declines. Note that the heat flux in the South Atlantic is now southward, as required by the thermal mode in this basin.
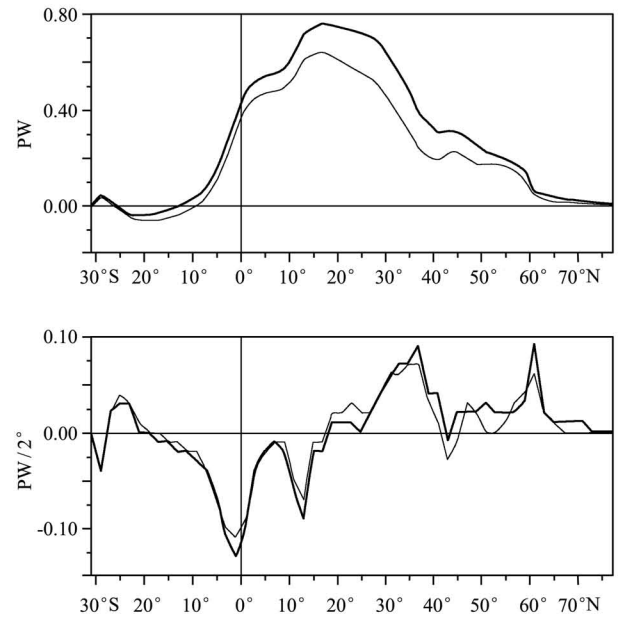

Fig.9 Poleward heat flux (upper panel) and its divergence (lower panel) (due to advection) diagnosed from the standard model with both temperature and salinity relaxed (thin line) and the model with uniform salinity (S = 35, heavy line).

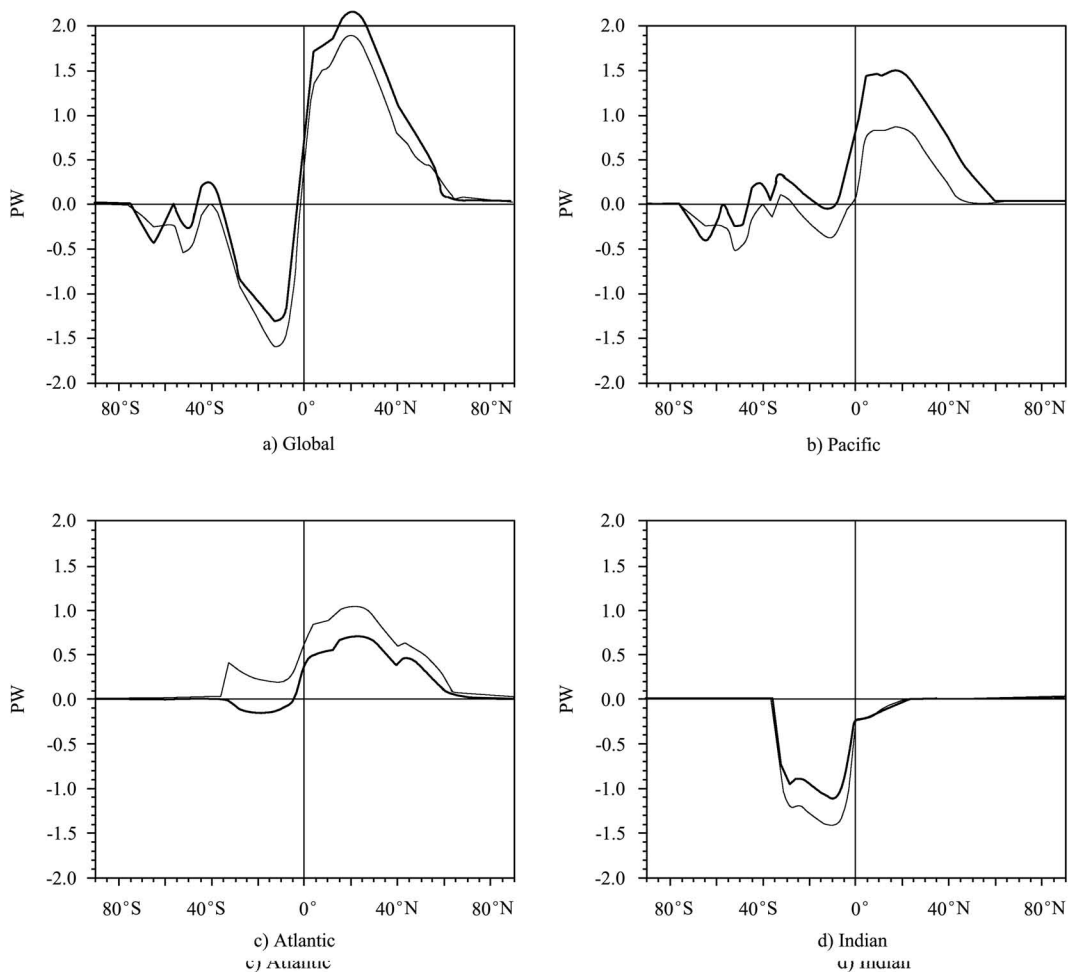
a) Global

b) Pacific

c) Atlantic

d) Indian

Fig.10 Poleward fluxes in the world oceans, diagnosed from a model based on observed SST and SSS (thin lines) and a model based on relaxation to observed SST and with constant salinity (heavy lines).

5 The Implication of a Swamp Ocean of No Current

The role of the oceans in the ocean-atmosphere coupled system is illustrated in Fig.11a. In the current climate, there are three components (the atmospheric sensible heat, the oceanic sensible heat, and the latent heat for the coupled system) in this climate heat conveyor, working together as a relay team. Although the atmosphere is responsible for sending heat to high latitudes, it might be misleading to state that the atmosphere heat transport alone is responsible for the high latitudes' poleward heat flux at high latitudes.
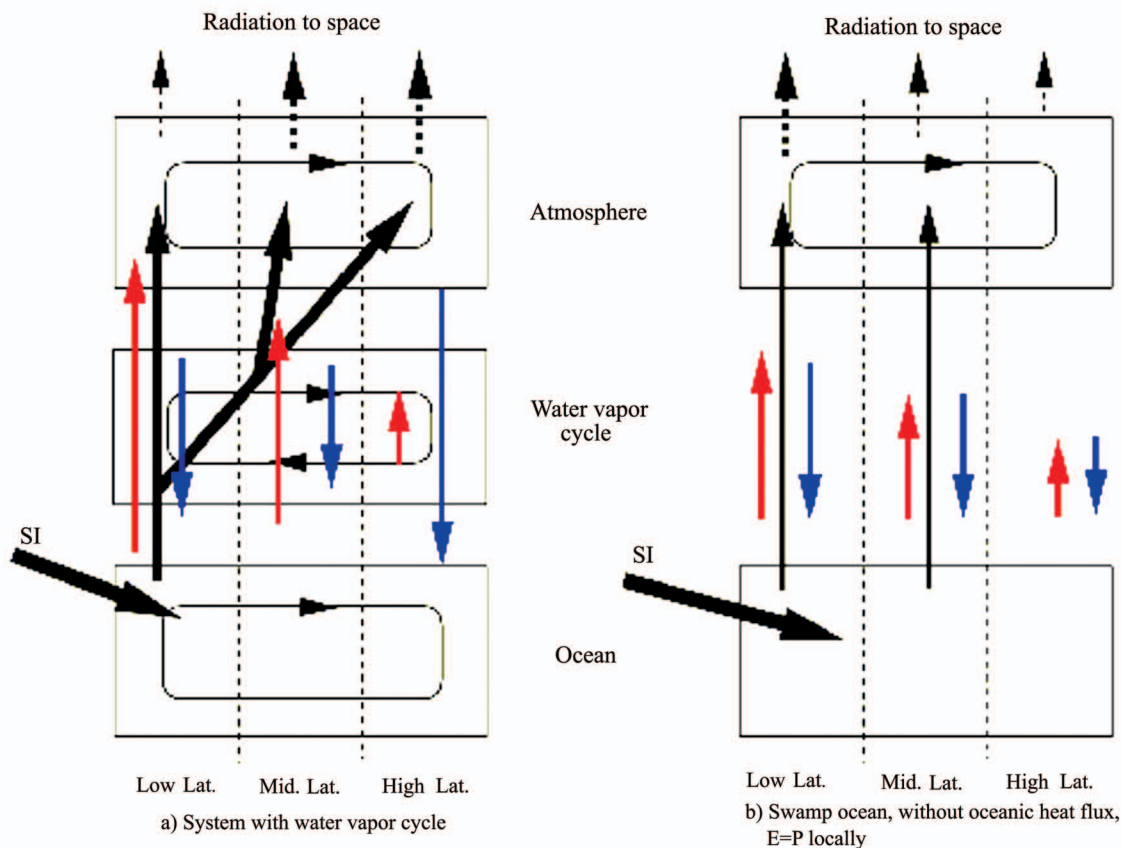

Fig.11 Sketches for (a) climate model with oceanic current and (b) climate model with a swamp ocean. SI is the solar isolation.

If there were no evaporation and precipitation closely related to the thermohaline circulation in the oceans, there would be no strong latent heat flux in the climate system. Thus, the whole earth would be cooled down substantially. Without evaporation and precipitation, there would be no strong salinity gradient in the oceans. With less heat flux transport from low latitudes, temperatures at high latitudes should be lower than they are under the present climate conditions. Due to the greater-than-present meridional temperature difference and almost zero salinity difference, the oceanic sensible heat flux would be intensified, which can partially compensate for the loss of poleward heat flux from the lack of the hydrological cycle, as shown in experiment B.

In some climate models, the oceanic component has been treated as a swamp, *i.e.*, with no currents. Although it is clear that such a model would not allow heat flux transport associated with currents, a far stronger constraint on the climate system has not been clearly defined. A close examination reveals the following constraints on such an oceanic model component:

1) No currents or meso-scale eddies, so there would be no horizontal heat flux because molecular heat diffusion is negligible.

2) No currents to transport fresh water horizontally. As a result,

ⅰ) Evaporation and precipitation would be locally balanced.

ⅱ) No latent heat flux in the atmosphere.

So although a swamp ocean can still play the role of absorbing shortwave radiation from the sun and heat up the atmosphere through the latent heat release associated with evaporation, long-wave radiation and sensible heat flux, there would be no sensible heat flux in the oceans or latent heat flux in the atmosphere. The only mechanism of heat transport would be the dry atmospheric circulation. It is speculated that the climate in such a fictitious planet would be dramatically colder than the climate on our planet.

6 Summary

In this note we discuss the important role of oceanic circulation in the climate system. The major points can be summarized as follows:

1) It is more appropriate to separate the poleward heat flux into three components: the atmospheric sensible heat flux, the oceanic sensible heat flux, and the atmosphere-ocean-land coupled latent heat flux.

2) It is important to examine the divergence of these fluxes and their interaction.

3) It is very important to explore how changes in the oceanic circulation affect the surface conditions, such as surface temperature and the air-sea heat fluxes, and the global climate conditions on Earth.

4) The hydrological cycle plays an important role in setting up the sea surface temperature and salinity; thus, the so-called swamp ocean model cannot correctly simulate the oceanic role in climate.

Appendix: Definition of the oceanic sensible heat flux

Traditionally, the poleward oceanic heat flux is defined as

$$H_o = \iint \rho C_p v \theta \, \mathrm{d}x \mathrm{d}z \ ,$$

where ρ is the *in situ* density, C_p is the specific heat under constant pressure, v is the meridional velocity, and θ is the potential temperature. It is well known that if the system has a net mass flux through a section, the heat flux calculated from this formula may depend on the choice of the temperature scale. In general, heat flux itself has no direct physical meaning, and the most important quantity associated with the heat flux is its divergence. There are two criteria in defining heat flux: First, the divergence of heat flux should satisfy the local heat balance equation. Second, the heat flux should depend on the local properties only.

The best-known example of the non-uniqueness of the oceanic heat flux is the circulation in the South Pacific and Indian Oceans. Due to the Indonesian Throughflow, the poleward heat flux in the South Pacific and Indian Oceans is not uniquely defined. To overcome the complication due to the net mass flux through the sections, many different approaches have been used. For example, Zhang and Marotzke (1999) have proposed a procedure in which the heat flux associated with the westward flow in the Strait is used to evaluate the poleward heat flux at sections south of the Strait. Although the heat flux definition they proposed satisfies the first criterion, it does not satisfy the second. In fact, their definition includes a term that depends on the thermal condition in the cross section of the Throughflow. If the condition within the Strait changes, the heat flux evaluated at sections south of the Strait will also change, even if the circulation and water properties at the given section south of the Strait do not change.

To overcome such a problem, thus, a much simpler definition is recommended:

$$H_o = \iint \rho C_p v (\theta - \theta_0) \, \mathrm{d}x \mathrm{d}z \ ,$$

where θ_0 is a reference potential temperature which can be chosen arbitrarily, and the most obvious choice is the global-mean potential temperature, which is approximately $2\,^\circ\mathrm{C}$. This definition satisfies the two criteria listed above. We will analyze the meaning of the heat flux defined in this way as follows:

In the oceans, there is indeed a net mass flux through any given section due to either evaporation/precipitation or Throughflow.

$$H_o = H_{o,0} + H_{o,1},$$
$$H_{o,0} = \iint \rho C_p v (\theta - \theta_0) \mathrm{d}x \mathrm{d}z - H_{o,1},$$
$$H_{o,1} = \iint \overline{\rho v} C_p (\bar{\theta} - \theta_0) \mathrm{d}x \mathrm{d}z,$$

where $\overline{\rho v} = \iint \rho v \mathrm{d}x \mathrm{d}z / \iint \mathrm{d}x \mathrm{d}z$ is the mass flux averaged over the section. The first component, $H_{o,0}$, is the oceanic sensible heat flux through this section, carried by the circulation without a net mass flux. By its definition, it is independent of the choice of the reference temperature. On the other hand, the second component, $H_{o,1}$, clearly depends on the choice of the reference temperature, so it should be isolated from the first component. In the oceans, the net mass flux through a given section is due to either evaporation/precipitation or an inter-basin loop current, such as the Indonesian Throughflow. We will separate the net mass flux through each section into two parts:

$$\overline{\rho v} = \overline{\rho v}_{\mathrm{loop}} + \overline{\rho v}_{\mathrm{emp}},$$

where $\overline{\rho v}_{\mathrm{loop}}$ is the zonal-mean meridional mass flux associated with the loop current going through the whole basin and the total amount of mass flux associated with this component should be constant at any zonal section, such as the Indonesian Throughflow; $\overline{\rho v}_{\mathrm{emp}}$ is the zonal-mean meridional mass flux associated with evaporation and precipitation. This component of the net mass flux through the section is associated with the water vapor cycle in the atmosphere, so it should not be counted as part of the oceanic sensible heat flux.

Accordingly, the poleward sensible heat flux in the Atlantic can be defined as:

$$H^A_{o,\,\mathrm{sen}} = \iint_A \rho C_p v (\theta - \theta_0) \mathrm{d}x \mathrm{d}z - H^A_{o,\,\mathrm{emp}},$$
$$H^A_{o,\,\mathrm{emp}} = C_p \iint_A \overline{\rho v}_{\mathrm{emp},\,A}(y)[\bar{\theta}_A(y) - \theta_0] \mathrm{d}x \mathrm{d}z$$
$$= C_p M_{\mathrm{emp},\,A}(y)[\bar{\theta}_A(y) - \theta_0],$$

where $M_{\mathrm{emp},\,A}$ is the zonally-integrated meridional mass flux due to evaporation/precipitation. Similarly, the poleward heat flux in the Pacific and Indian Oceans are defined as:

$$H^P_{o,\,\mathrm{sen}} = \iint_P \rho C_p v (\theta - \theta_0) \mathrm{d}x \mathrm{d}z - H^P_{o,\,\mathrm{emp}},$$
$$H^P_{o,\,\mathrm{emp}} = C_p M_{\mathrm{emp},\,P}(y)[\bar{\theta}_P(y) - \theta_0],$$
$$H^I_{o,\,\mathrm{sen}} = \iint_I \rho C_p v (\theta - \theta_0) \mathrm{d}x \mathrm{d}z - H^I_{o,\,\mathrm{emp}},$$
$$H^I_{o,\,\mathrm{emp}} = C_p M_{\mathrm{emp},\,I}(y)[\bar{\theta}_I(y) - \theta_0].$$

The total oceanic sensible heat flux is the sum of these three fluxes:

$$H_{o,\,\mathrm{sen}} = H^A_{o,\,\mathrm{sen}} + H^P_{o,\,\mathrm{sen}} + H^I_{o,\,\mathrm{sen}}.$$

The sum of water fluxes due to evaporation and precipita-

tion, including river run-offs into all the oceans, should be equal to the water vapor flux to the atmosphere. Thus, the sum of the evaporation and precipitation in three basins is

$$M_{emp} = M_{emp,A} + M_{emp,P} + M_{emp,I} .$$

The heat flux associated with this mass flux is:

$$H_{o,emp} = C_p [M_{emp,A} \bar{\theta}_A(y) + M_{emp,P} \bar{\theta}_P(y) +$$
$$M_{emp,I} \bar{\theta}_I(y)] - C_p M_{emp} \theta_0 .$$

This heat flux is independent of the choice of the temperature scale.

This example demonstrates that a working definition of heat flux should be independent of the specific choice of the temperature scale. In order to achieve this goal, one has to separate the net mass flux through a given section, and link this net mass flux with the corresponding return flow in the climate system, *i.e.*, the water vapor flux in the atmosphere:

$$H_{emp}^{air\text{-}sea} = - M_{emp}[L_e + C_p (\bar{\theta}_{atmos}(y) - \theta_0)] + H_{o,emp},$$

where the first term on the right-hand side indicates the heat content flux in the atmospheric branch of the water vapor loop, and the second term is the heat content flux in the oceanic branch of the water cycle. This definition is, of course, independent of the choice of the temperature scale.

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