The Ocean-Atmosphere System II: Oceanic Heat Budget

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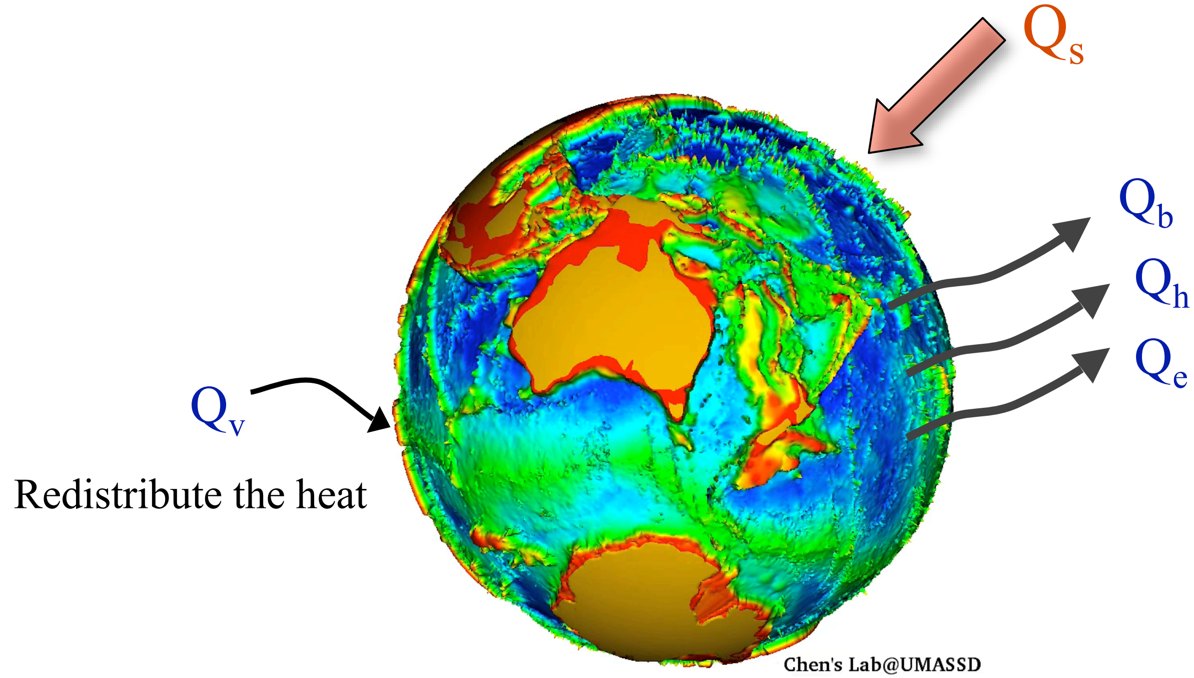
MAR 555 Lecture 2: The Oceanic Heat Budget



Q_s: The short wave energy radiated from the sun (the shortwave radiation)
Q_b: The net long-wave energy radiated back from the ocean (the longwave radiation);
Q_e: The heat loss by evaporation (latent heat flux);
Q_h: The sensible heat loss by conduction;
Q_v: The heat transfer by currents (advection and convection)

Assuming that the ocean is a closed system, then the net heat flux at the sea surface ΔQ equals to

$$\boxed{\Delta Q = Q_s - Q_b - Q_e - Q_h}$$

Considering no global warming tendency, for a long-term averaging,

$$\Delta Q = 0 \longrightarrow Q_S = Q_b + Q_e + Q_h$$

For a short-term, the net oceanic heat flux varies daily with diurnal variation of solar radiation, etc. For the real ocean, due to the global warming,

$$\Delta Q \neq 0$$

Question:

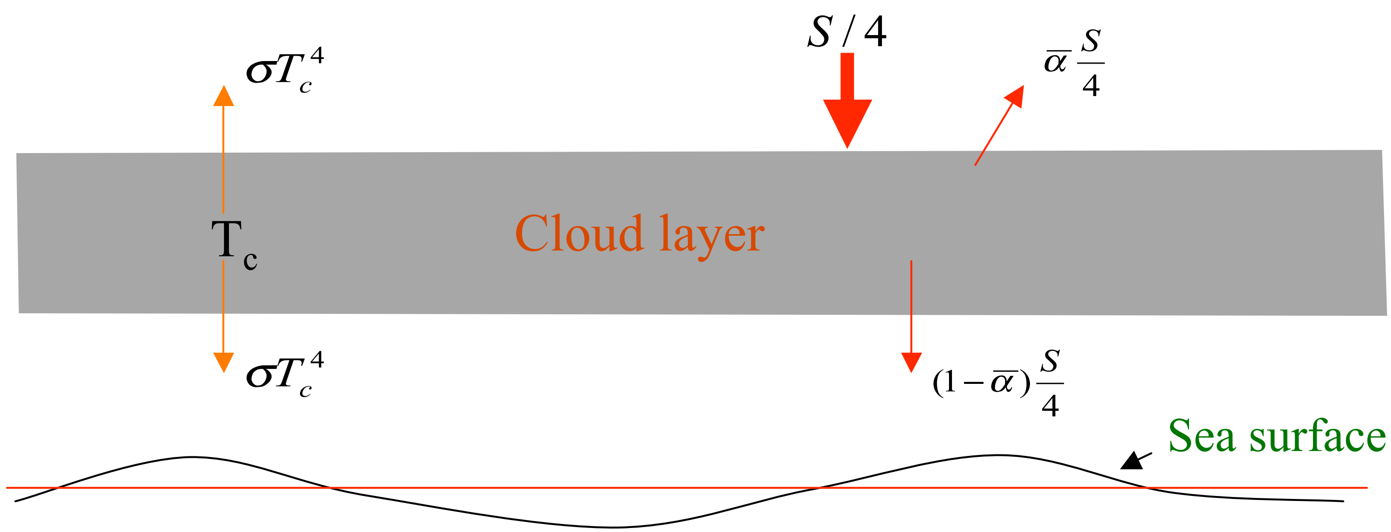
What are the physical processes controlling the heat flux?

Q_s: The shortwave radiation (unit: W/m^2)

Observational results:

Only 51% of the sun's shortwave radiation energy can get onto the sea surface:

$$51\% \Rightarrow \begin{cases} 28\%: \text{ directly from the sun after the reduction due to the cloud} \\ 23\%: \text{ from sky radiation and longwave radiation of the atmosphere} \end{cases}$$



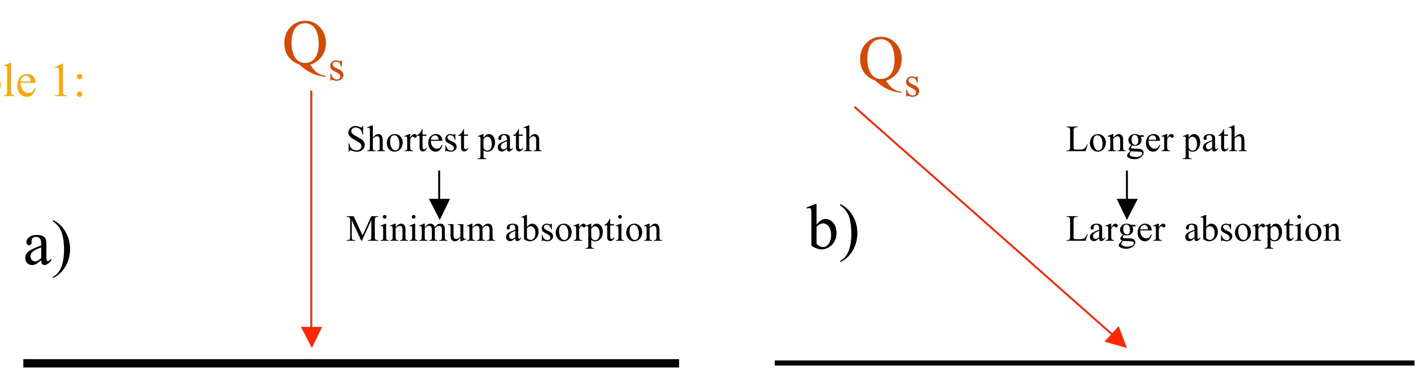
Total energy received at the sea surface is

$$\boxed{Q_S = Q_{S_1}(\text{direct}) + Q_{s_2}(\text{indirect}) = (1-\overline{\alpha})\frac{S}{4} + \sigma T_c^{4}}$$

The factors affecting Q_s:

- The length of the day: varies with seasons and geographic latitude;
- The absorption of atmosphere (gas molecules, dust, water vapor);
- Effects of cloud: reduce the average amount of energy reaching the sea surface through the absorption and scattering by the cloud)

Example 1:



Example 2:

Empirical equation:

$$Q_{S_C} = \frac{S}{4}(1-0.09c)$$

c: The proportion of cloud cover in the sky, which is estimated by an unit of eight divisions.

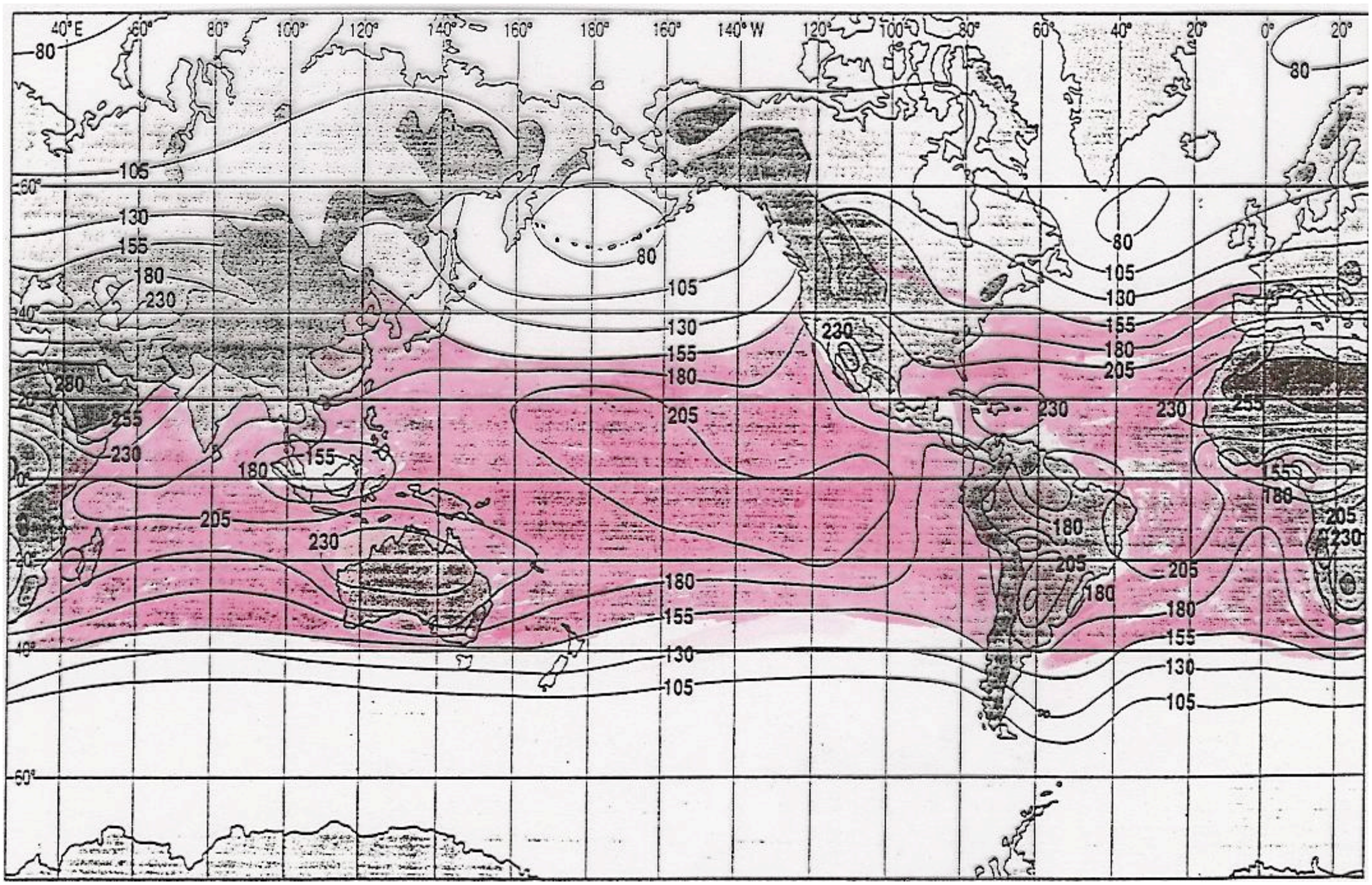
If the sky is completed covered by the cloud, c=8,

$$Q_{S_C} = \frac{S}{4}(1-0.09c) = \frac{S}{4}(1-0.09\times 8) = 0.28\frac{S}{4}$$

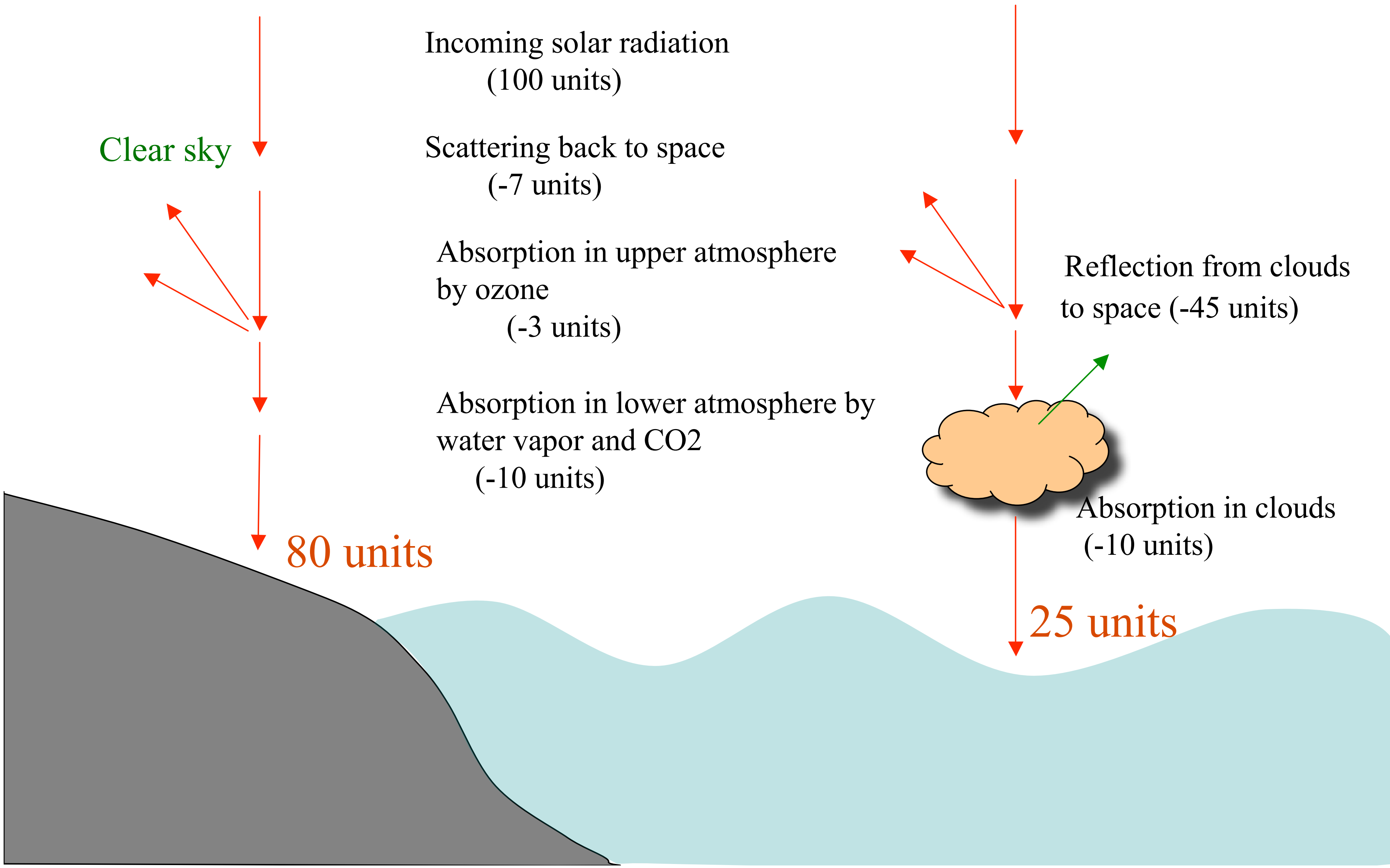
In the real ocean, the distribution of Q_s varies with latitude. The annual averaged values:

$$Q_S = \begin{cases} < 150-200\ \mathrm{W/m^2} & \theta > 20^0\,N \\ 200-255\ \mathrm{W/m^2} & \text{-}20^{\circ}\,S \le \theta \le 20^{o}\,N \\ < 200\ \mathrm{W/m^2} & \theta < \text{-}20^{\circ}\,S \end{cases}$$

Annually averaged Q_s(W/m^2)



Question: Why is Q_s larger on the continent than over the ocean?



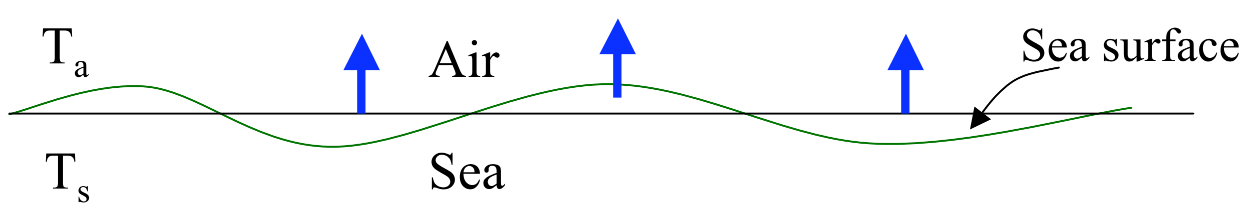
Q_b: The net long-wave radiation

All bodies with a temperature above absolute zero can emit radiation. The strength of such an emission depends on body's temperature: it is greater when the temperature of the body is higher. Also, as the temperature of the body becomes higher, the radiation spectrum is shifted toward shorter wavelengths.

In the real ocean, the energy emitted by the oceanic surface is estimated by a black body with a sea-surface temperature T_s. The net upward longwave flux Q_b equals to the difference between sea and air longwave radiations, i.e.:

$$\boxed{Q_b = \sigma(T_s^4 - T_a^4)}$$

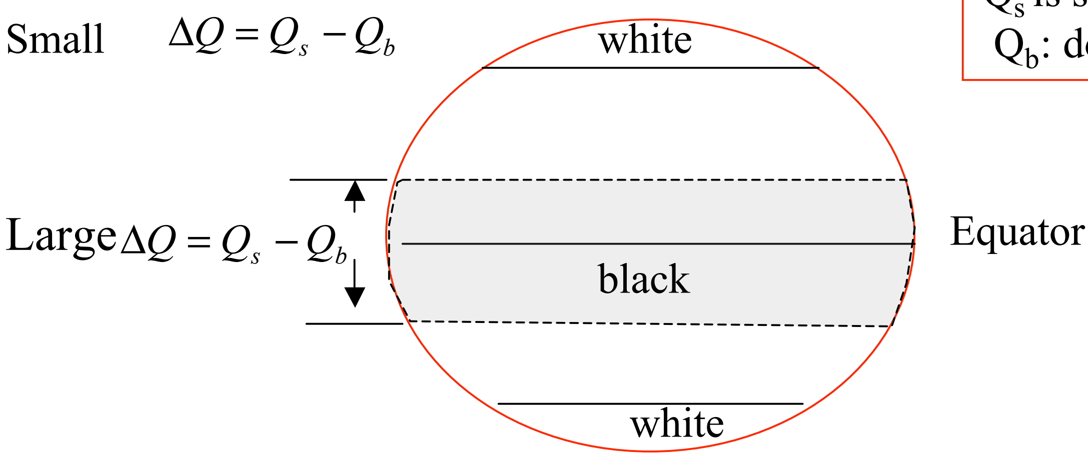
Ts: the water temperature at the sea surface
Ta: the air temperature at the sea surface
σ: Stenfan's constant



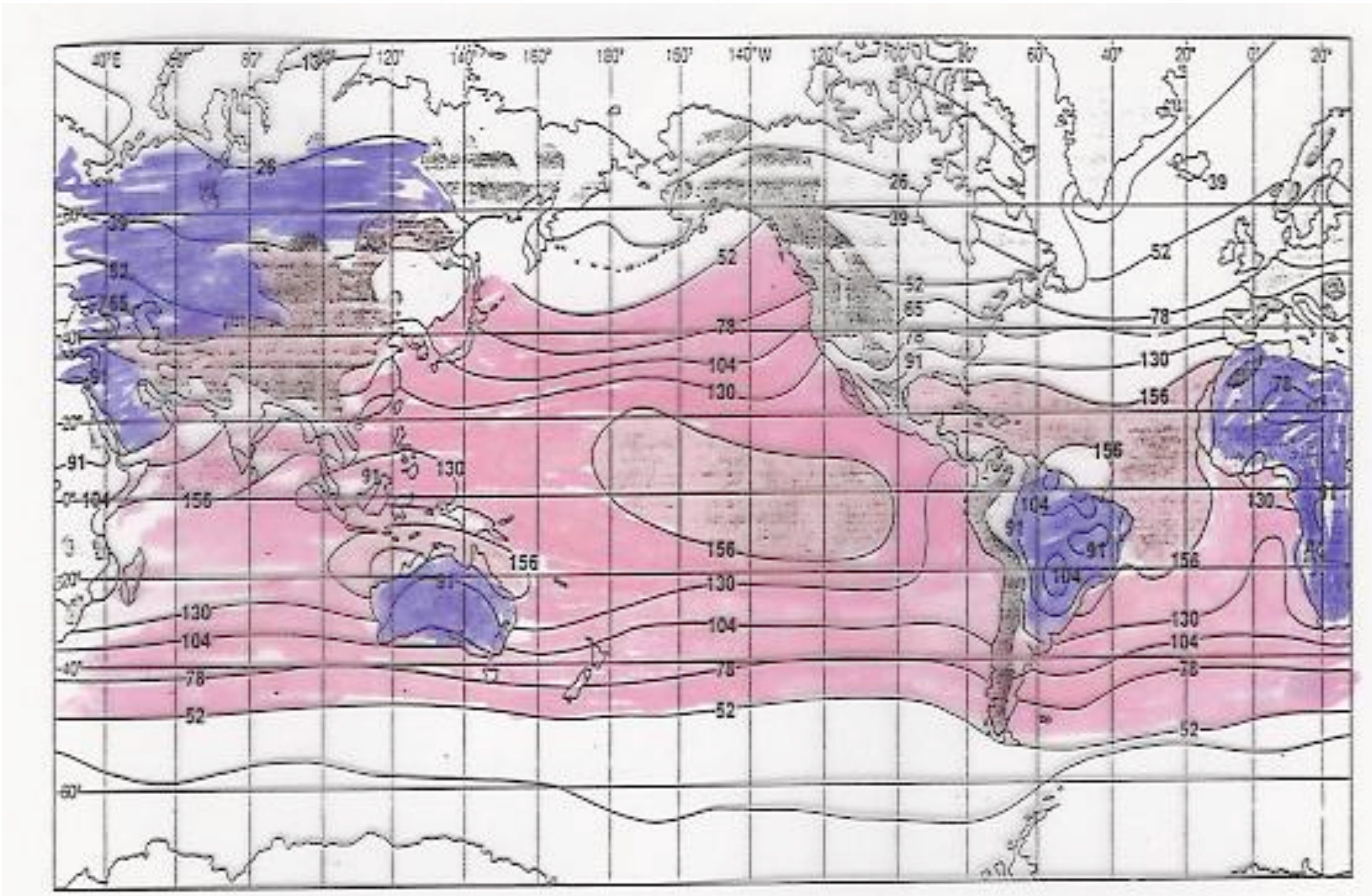
General evidences:

- $Q_b \sim 50\ W/m^2$;
- As the cloud coverage becomes larger, Q_b decreases. However, Q_b doesn't change much for ice and water;
- Q_b doesn't change much either daily or seasonally because neither sea temperature or relative humidity over the sea changes much in these time scales.
- $(Q_S - Q_b)_{\text{water}} > (Q_s - Q_b)_{\text{ice}}$, $(Q_s - Q_b)_{\text{low latitude}} > (Q_s - Q_b)_{\text{high latitude}}$

Q_s is small at high latitude and large at low latitude
Q_b: does not change much with latitude



$$Q_S - Q_b$$



40°E
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160°
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140°W
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20°
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20°
60°
40°
20°
0°
20°
40°
60°
26
39
52
65
78
91
104
130
156

Q_h: Sensible heat loss

1) The water is much denser than the air

$$\rho_{air} \sim 1.2 - 1.3 \quad \text{kg/m}^3, \qquad \rho_{\text{sea water}} \sim 1.025 \times 10^3 \ \text{kg/m}^3$$

so,

$$\frac{\rho_{\text{sea water}}}{\rho_{\text{air}}} \sim 10^3 \longrightarrow$$

The air-sea interface is very stable
if only the density difference is considered

However, the large temperature difference can cause a heat transfer from the warm region to the cold region. Such a heat flux from the ocean is called sensible heat loss or the heat loss by conduction.

Two conduction processes:

1) Molecular heat conduction; 2) Eddy (turbulent) diffusion

For the molecular case:

$$Q_h = -c_p k \frac{\Delta T}{\Delta z}$$

k: the molecular conductivity of heat;
c_p: the specific heat of the air at constant pressure
$\Delta T/\Delta z$: the air temperature gradient at the sea surface

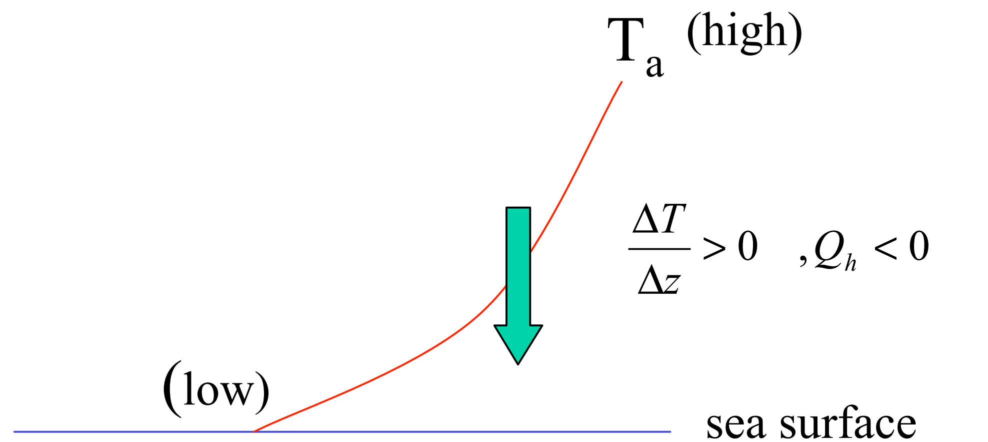
For the eddy case:

$$Q_h = \rho_a c_{pa} \overline{w'T'} \quad (= -c_{pa} A_z \frac{\Delta T}{\Delta z})$$

w' and T' are turbulent fluctuation of vertical air velocity and temperaute
ρ_a is the air density, and c_{pa} is the specific heat of the dry air
A_z is the vertical eddy viscosity.

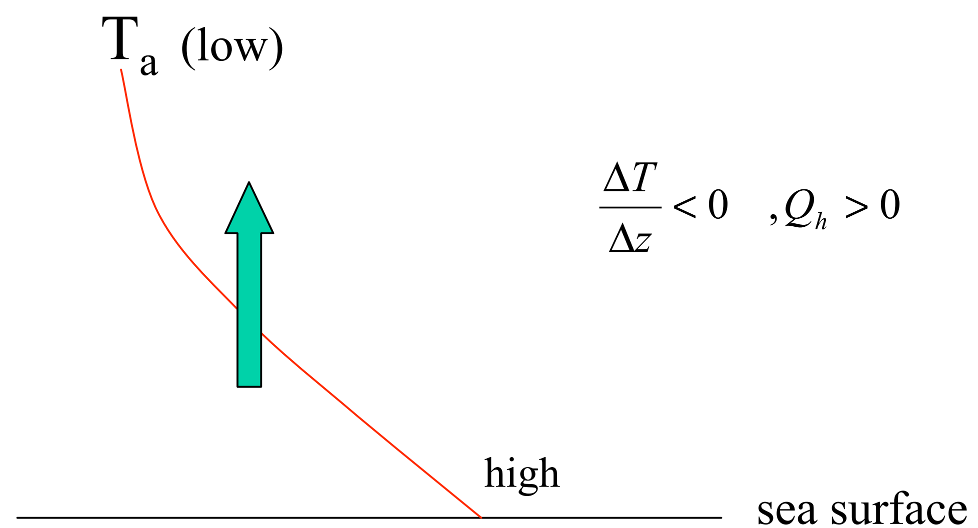
Example:

a)



The ocean gains heat

b)



The ocean losses heat

Q_e: The heat loss by evaporation (latent heat flux)

The latent heat flux can be estimated by a formula as

$$Q_e = \rho_a L_e \overline{w'q'}$$

w' and q' are turbulent fluctuation of vertical air velocity and water vapor mixing ratio.
L_e is the latent heat of evaporation of the sea water, which can be estimated as

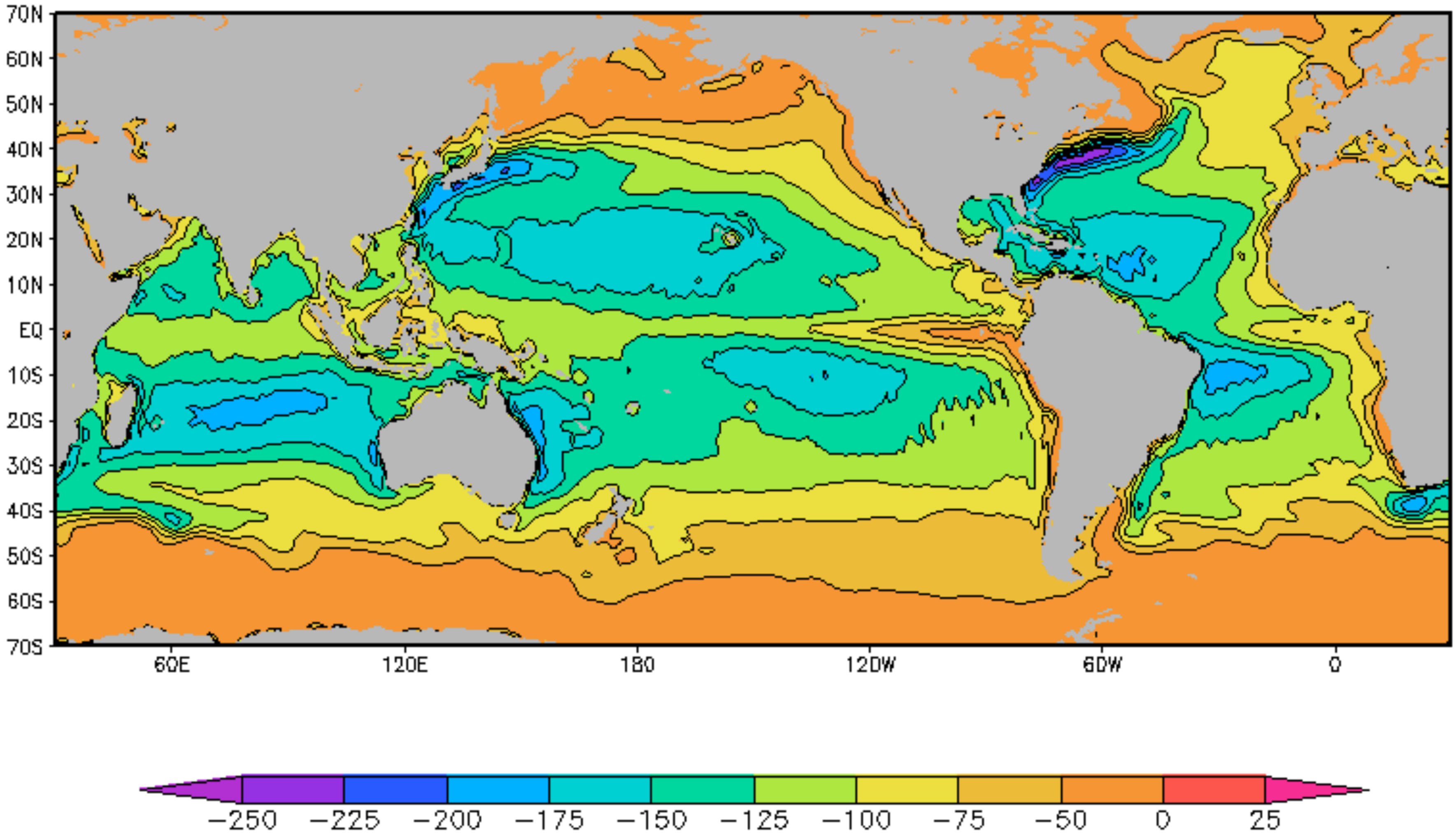
$$L_e = (2.501 - 0.00237 T_s) \times 10^6 \text{ J/kg}$$

In the real estimation,

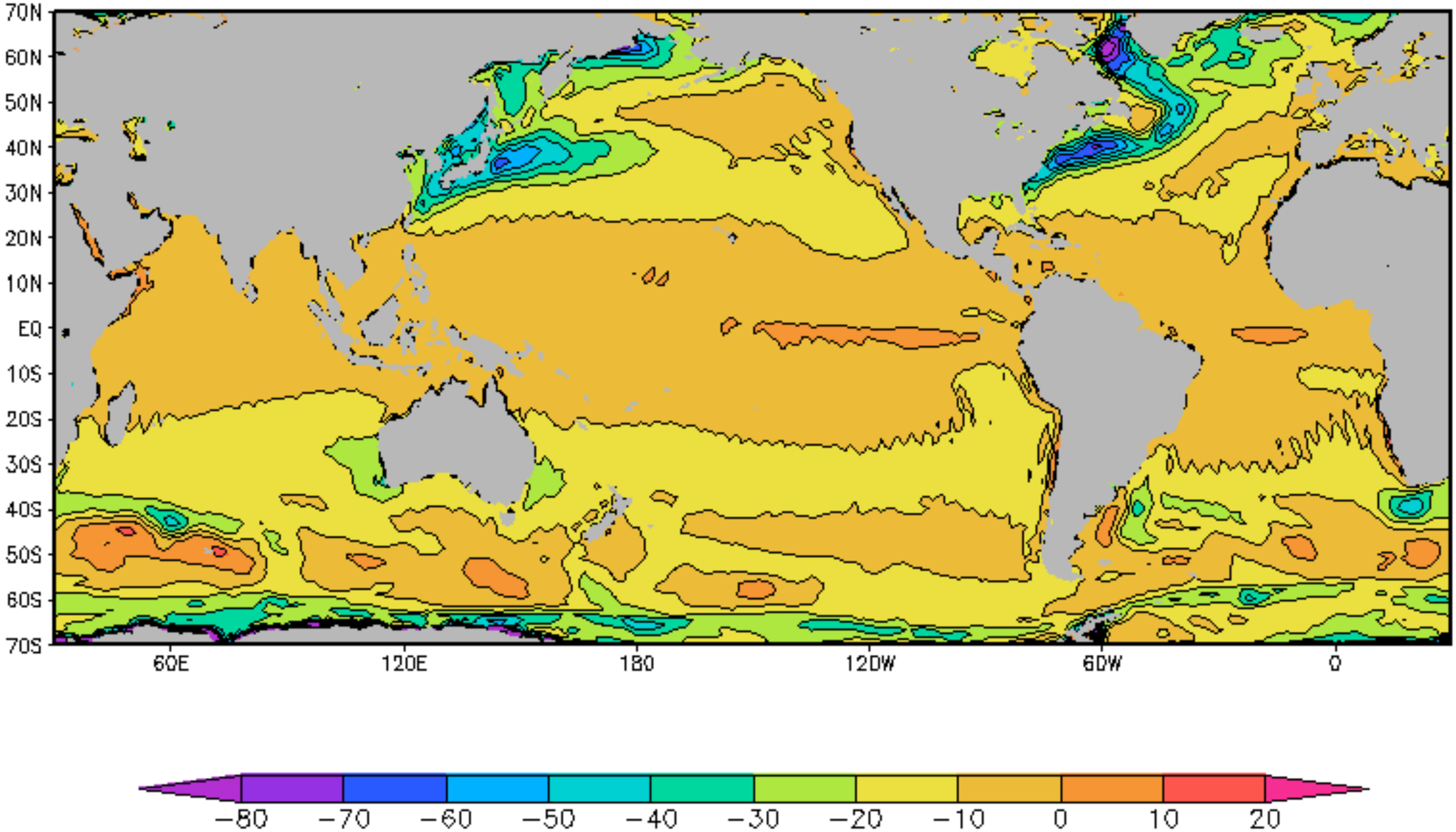
$$Q_e = \rho_a L_e u_* q_*$$

u_* and T_* are the Monin - Obukhov similarity scaling parameters of the frictional velocity, air temperature fluctuation.

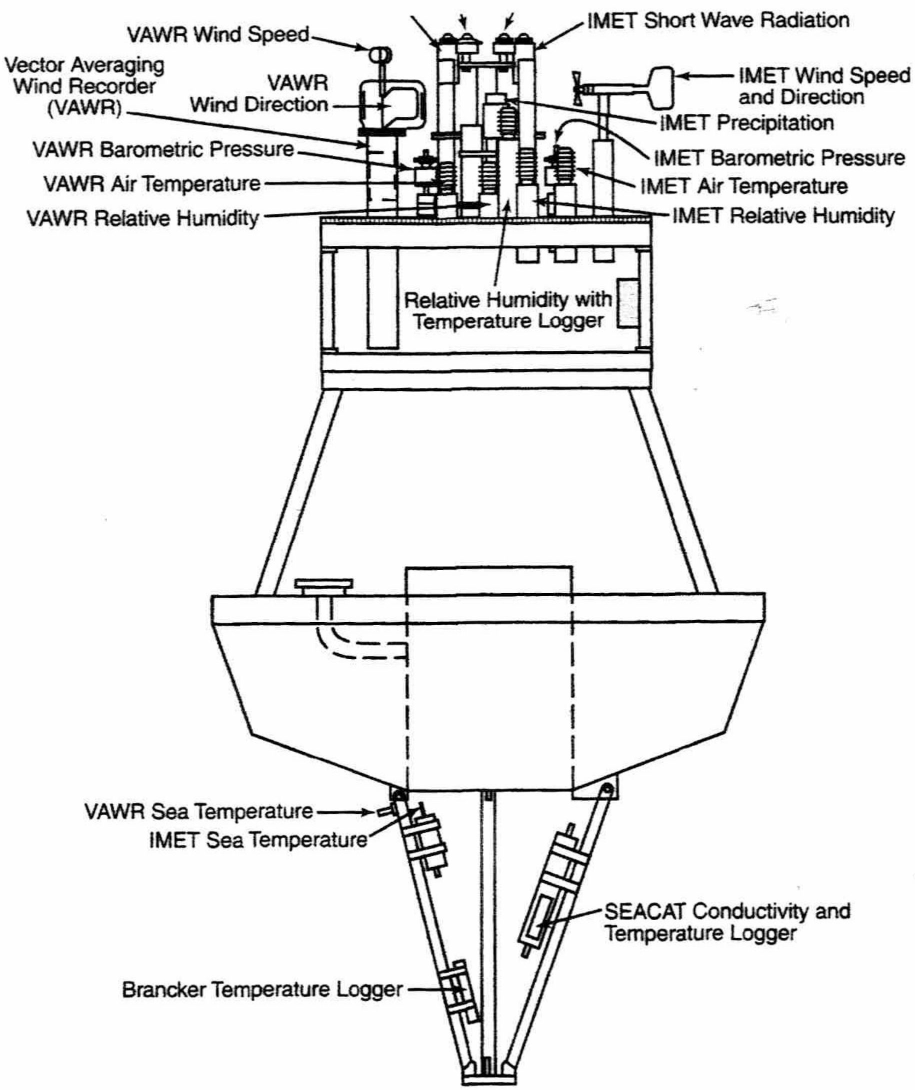
Annual Latent Heat (W/m^2)



Annual Sensible Heat (W/m^2)



Heat flux measurements at the sea surface



Critical issues needed to be addressed in order to provide an accurate estimation of Q_e and Q_h:

1) Determine accurately u_* under various stable or unstable conditions of the thermal structure in the near-sea atmosphere boundary;

2) Calculate accurately the air-sea temperature difference at the sea surface;

3) Estimate the contribution of precipitation to the surface cooling

Issue 1): u_* depends on the sea surface roughness and wind speed. How to calculate the roughness and take convective velocity into account?

Issue 2): The air temperature is measured at a certain height above the sea surface from the metrological buoys and ships, how to determine the true interfacial air-sea temperature difference?

Issue 3): The true interfacial air-sea temperature difference is sensitively influenced by the precipitation. Since the heavy rainfall usually companies with an atmospheric frontal passages, how to determine its contribution to the surface cooling?

Observational evidences:

a) The maximum Q_e occurs on the west side of the ocean and winter because of the existence of the west boundary warm current and large temperature gradient in winter;

b) At mid-latitude in the western North Pacific Ocean, Q_e is about 145 W/m^2. In the western North Atlantic, Q_e is about 195 W/m^2;

c) Q_e and Q_h are usually large during the storm passages.

Relation between Q_e and Q_h

In the ocean, Q_h and Q_e depends on the air-sea temperature difference and wind speed at the sea surface. These two fluxes are correlated each other and their relationship can be given as

$$\boxed{Q_h = RQ_e}$$

Where *R* is the so-called Bowen's ratio, which is equal to

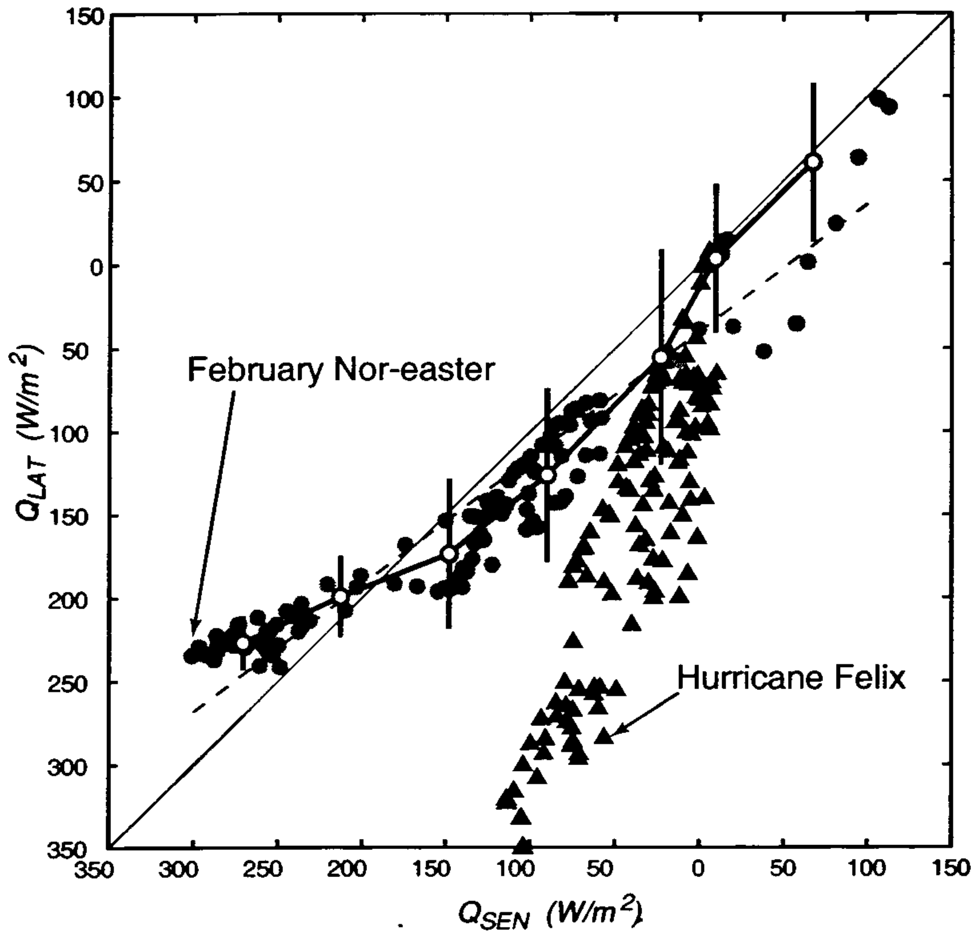
$$R = 0.062(T_s - T_a)/(e_s - e_a)$$

e_s and e_a are the sea water saturated vapor pressure and actual air vapor pressure.

Question:

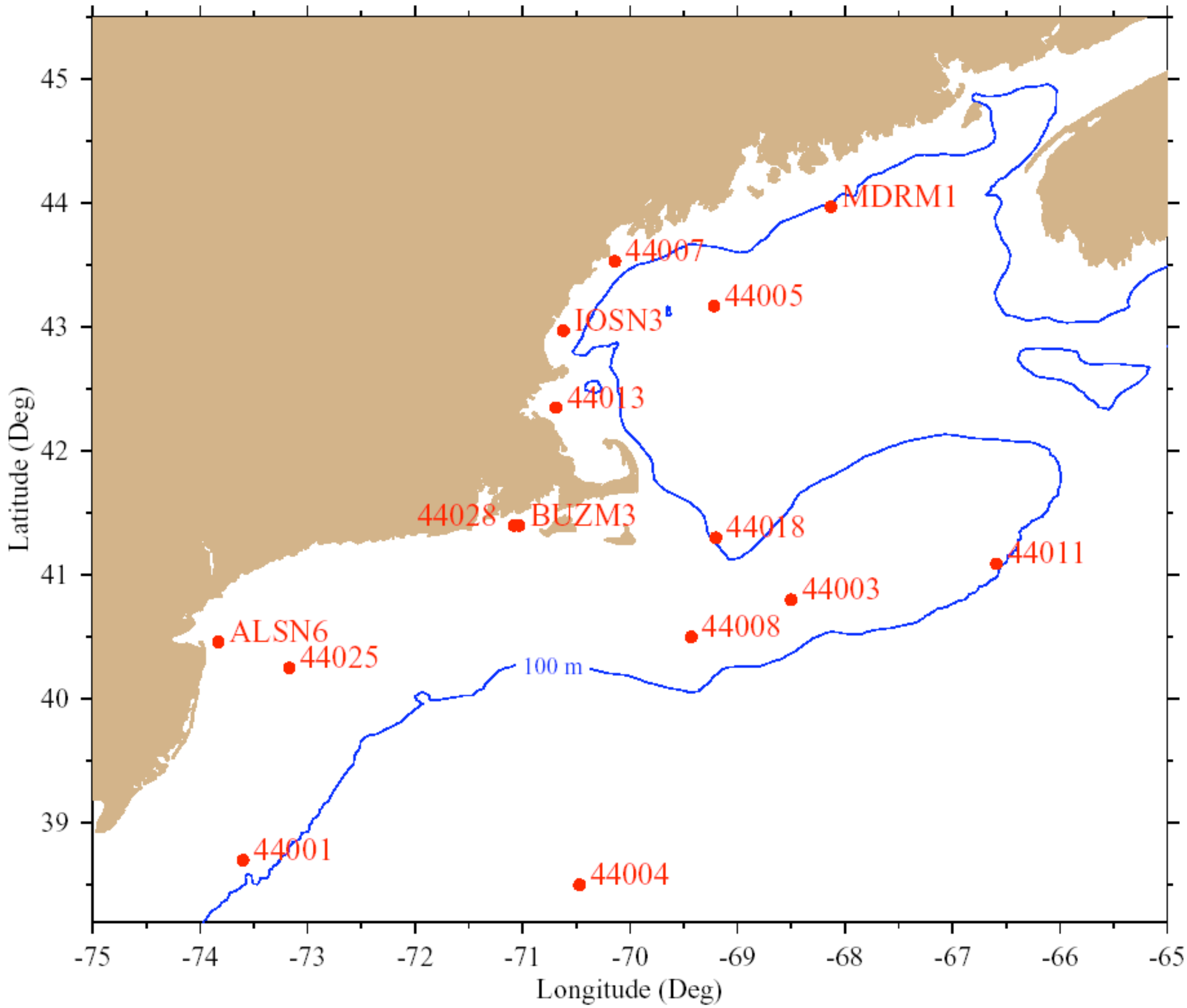
Is this relationship always valid?

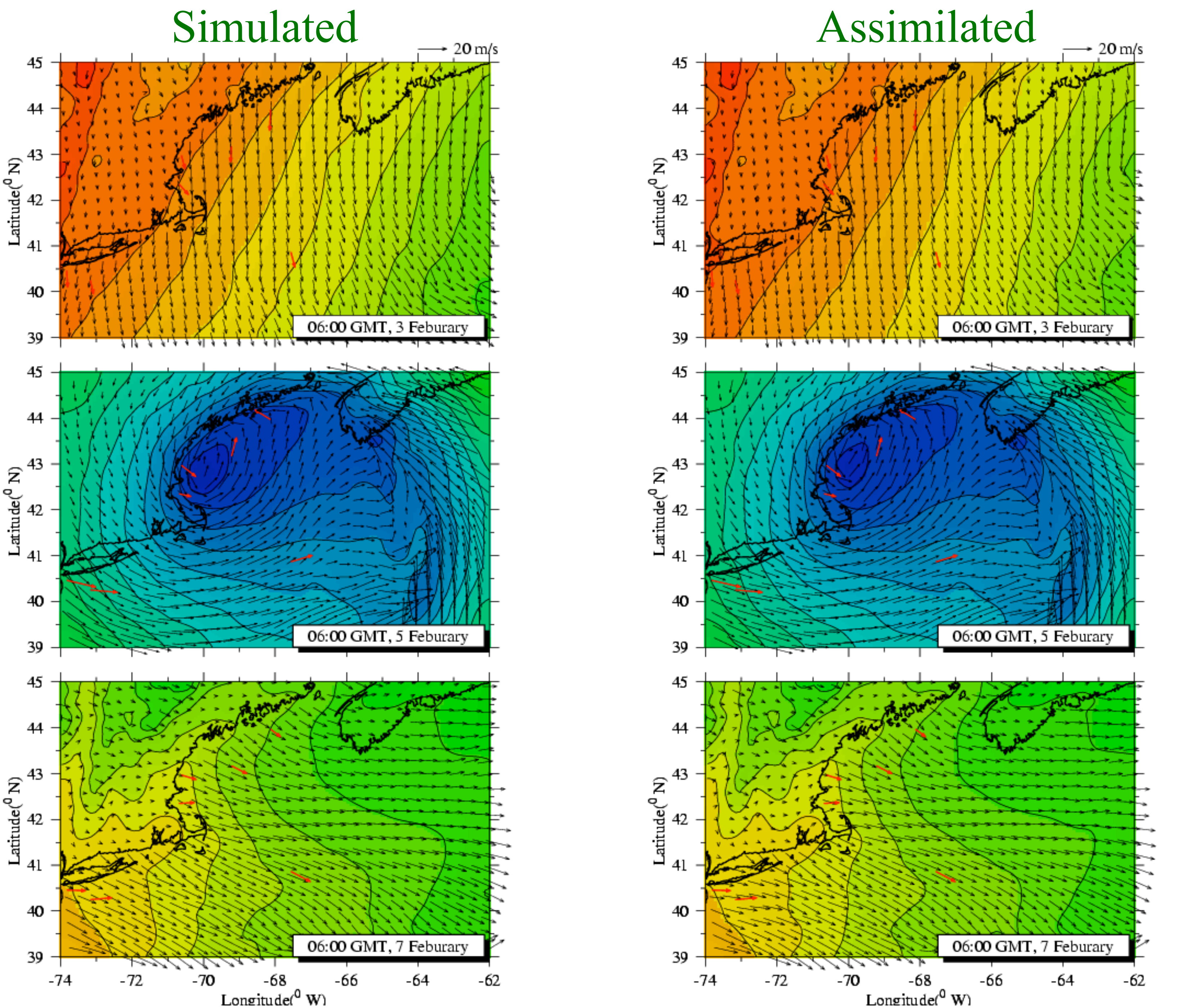
Example:

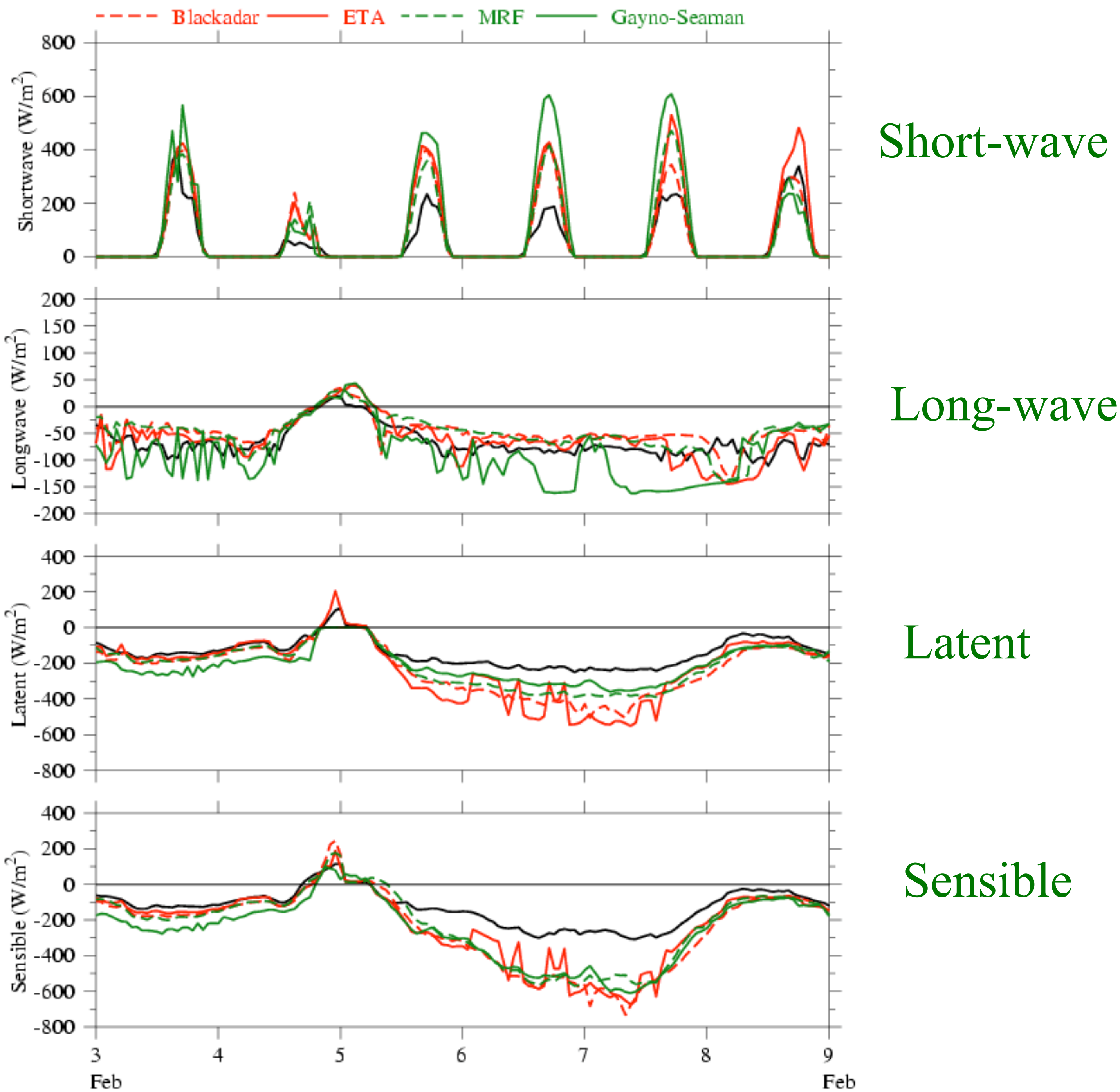


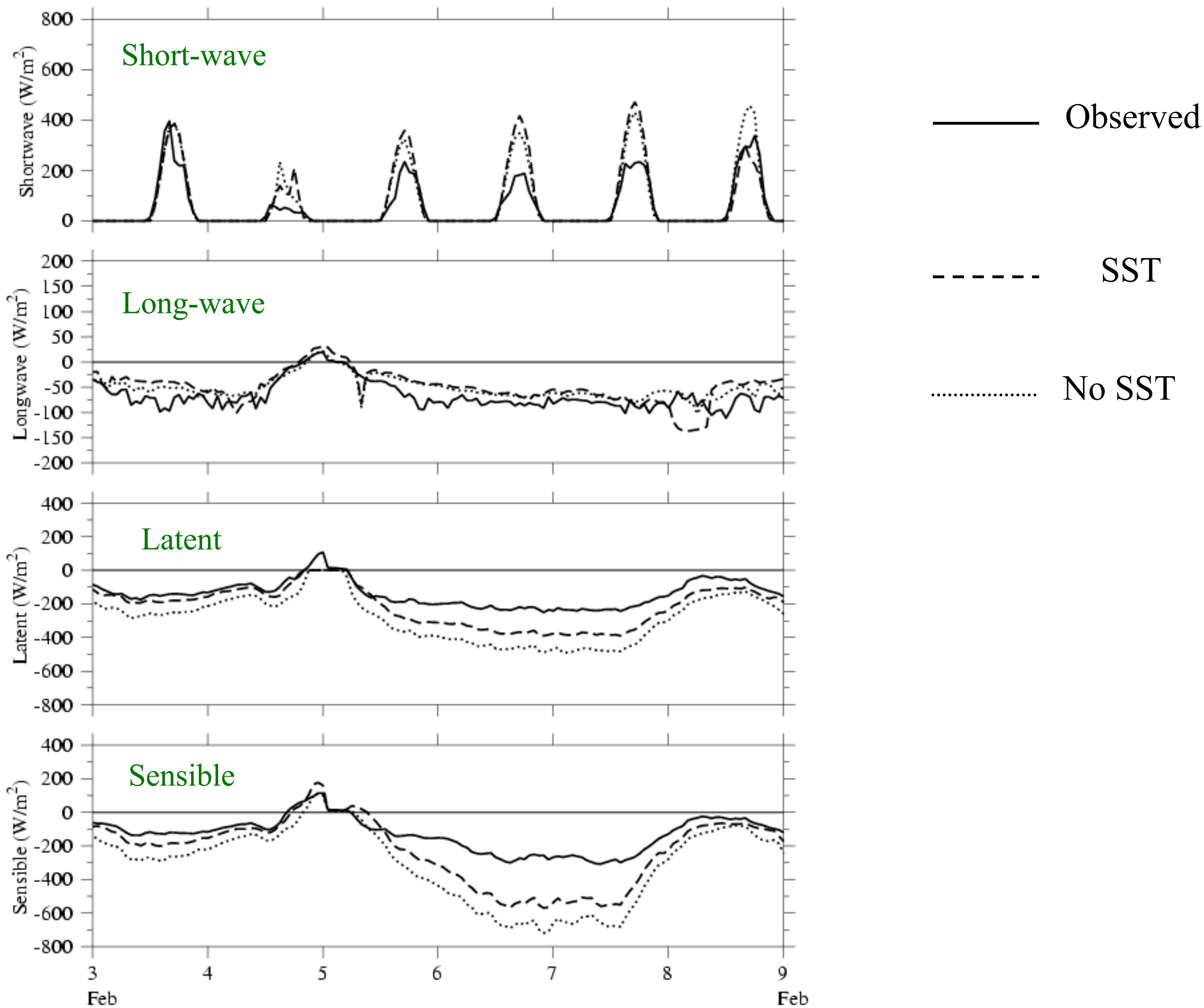
Scatter plot of Q_{sen} versus Q_{lat} for February 3-7 nor'easter (dot) and August 18-23 tropic storm Felix (diamonds), 1995 (Beardsley et al. 2003).

The Gulf of Maine Buoy Sites




Simulated
Assimilated
20 m/s
06:00 GMT, 3 Feburary
06:00 GMT, 5 Feburary
06:00 GMT, 7 Feburary
Latitude(0 N)
Longitude(0 W)
45
44
43
42
41
40
39
-74
-72
-70
-68
-66
-64
-62

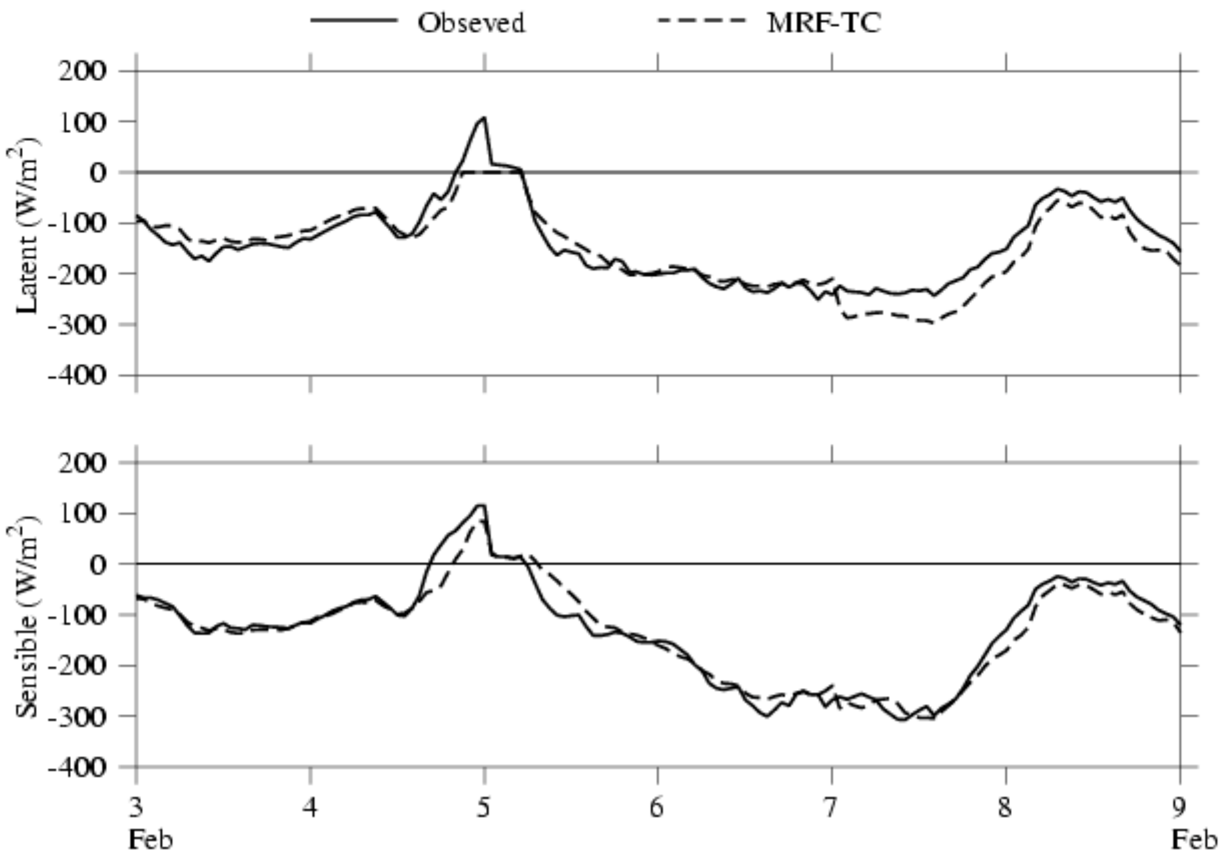

Blackadar
ETA
MRF
Gayno-Seaman
Shortwave (W/m^2)
Longwave (W/m^2)
Latent (W/m^2)
Sensible (W/m^2)
Short-wave
Long-wave
Latent
Sensible
3 Feb
4
5
6
7
8
9 Feb



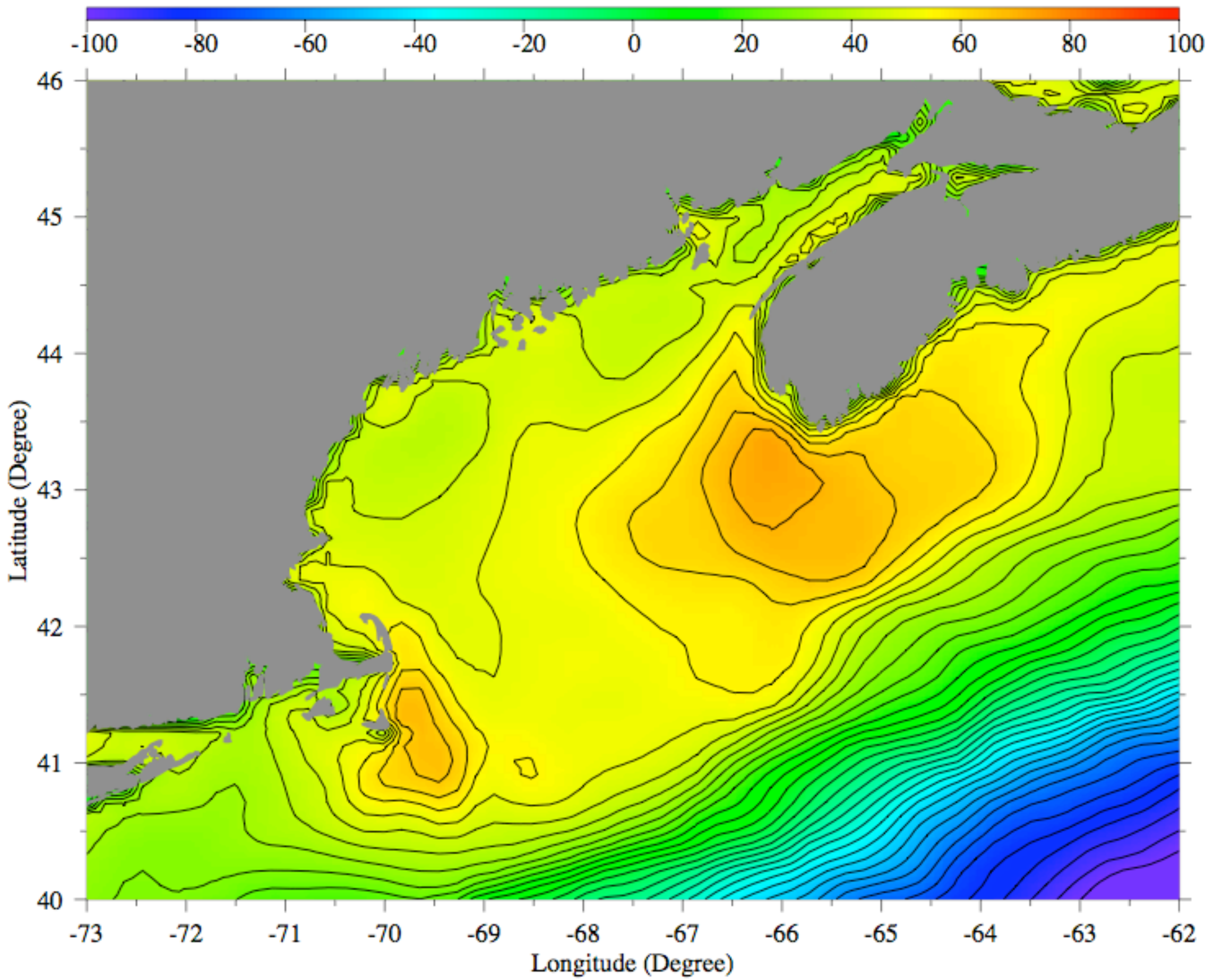
Short-wave
Long-wave
Latent
Sensible
Shortwave (W/m2)
Longwave (W/m2)
Latent (W/m2)
Sensible (W/m2)
Observed
SST
No SST
3 Feb
4
5
6
7
8
9 Feb

The TOGA/COARE (TC) heat flux algorithm developed by Fairall et al. (1996):

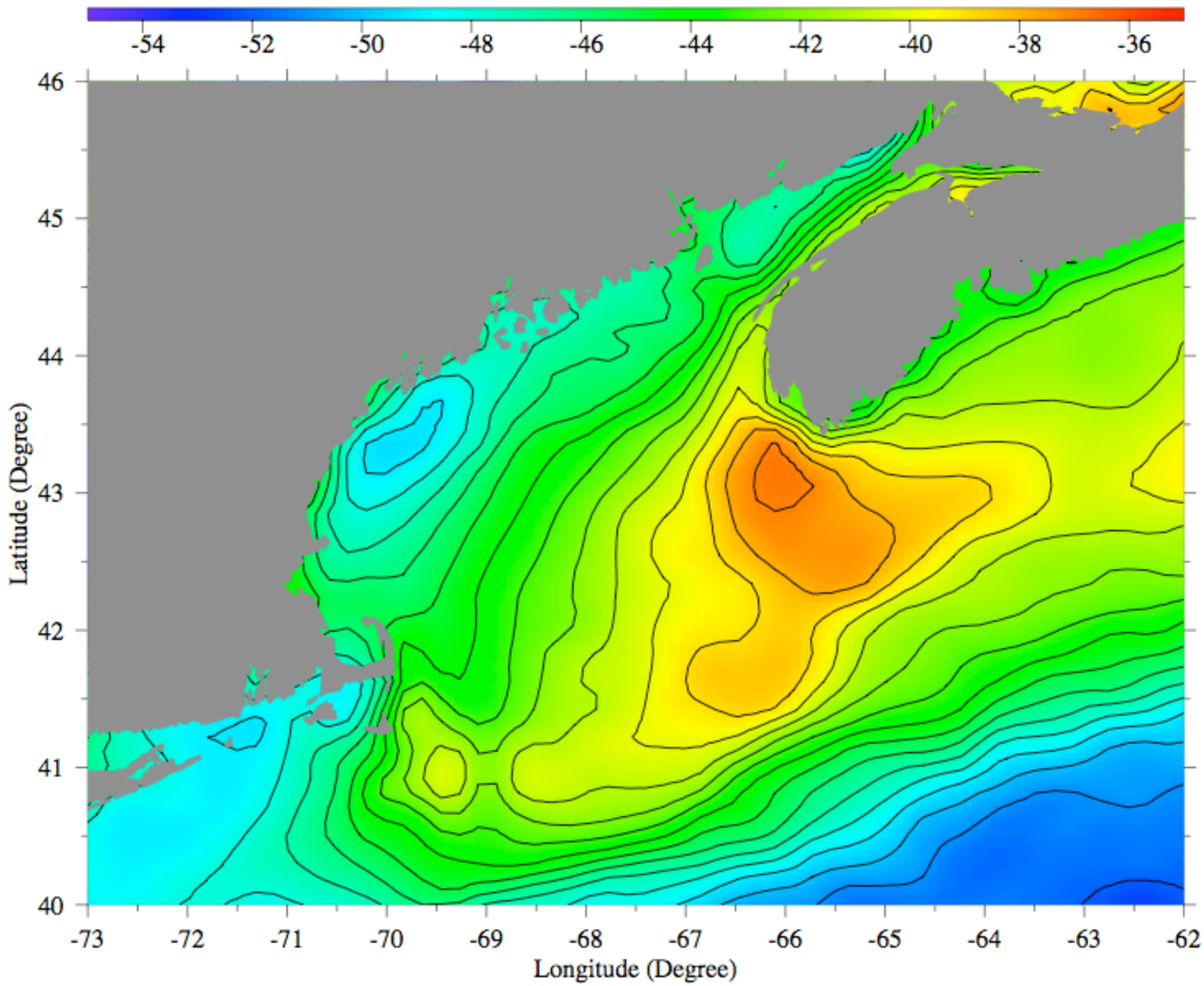
- More accurate skin water temperature;
- Better paramterization of surface roughness;
- More realistic vertical profiles for stable and unstable weather conditions
- Accountable wind gustiness



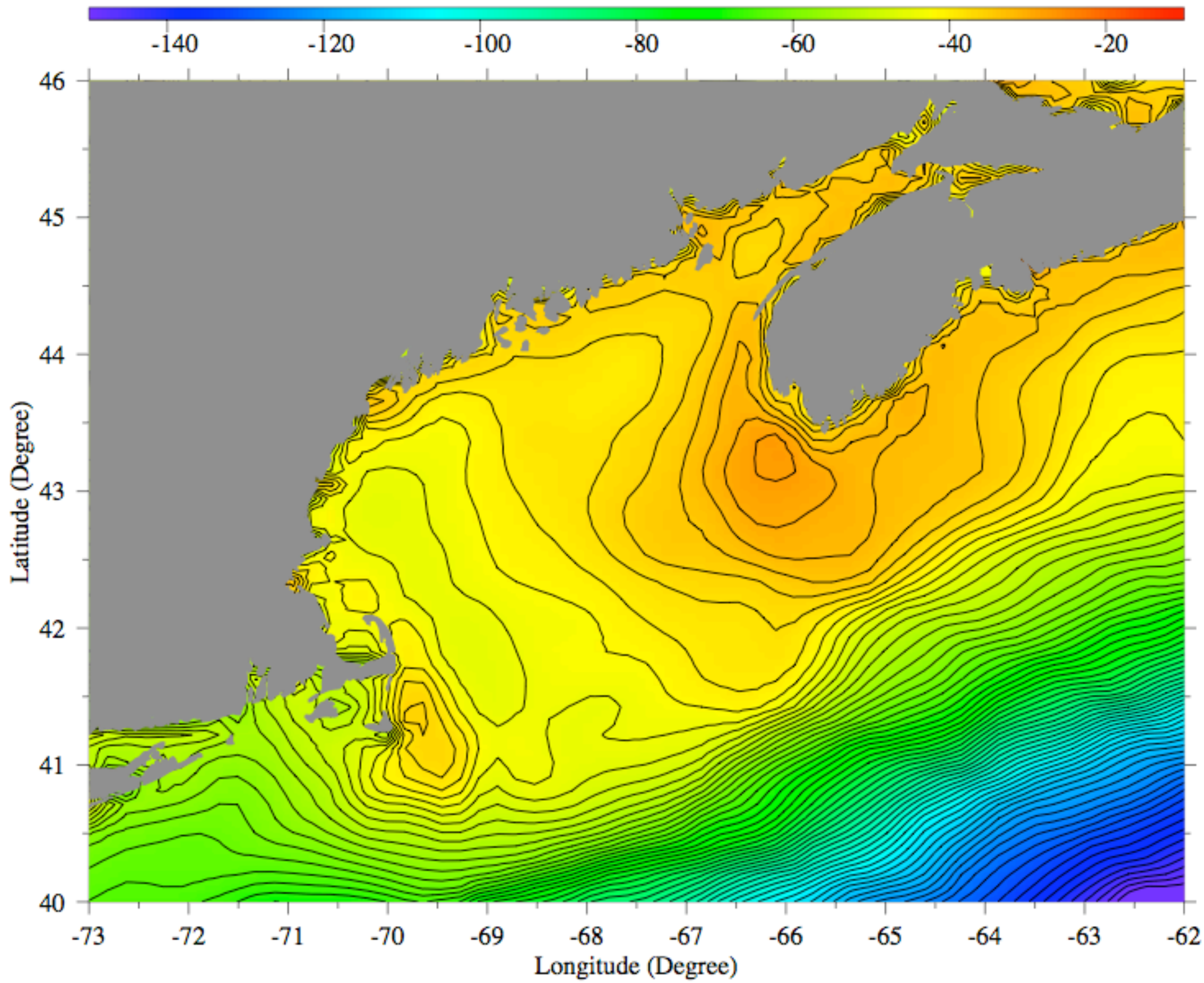
Annual net surface heat flux (W/m^2) averaged over 1978-2004



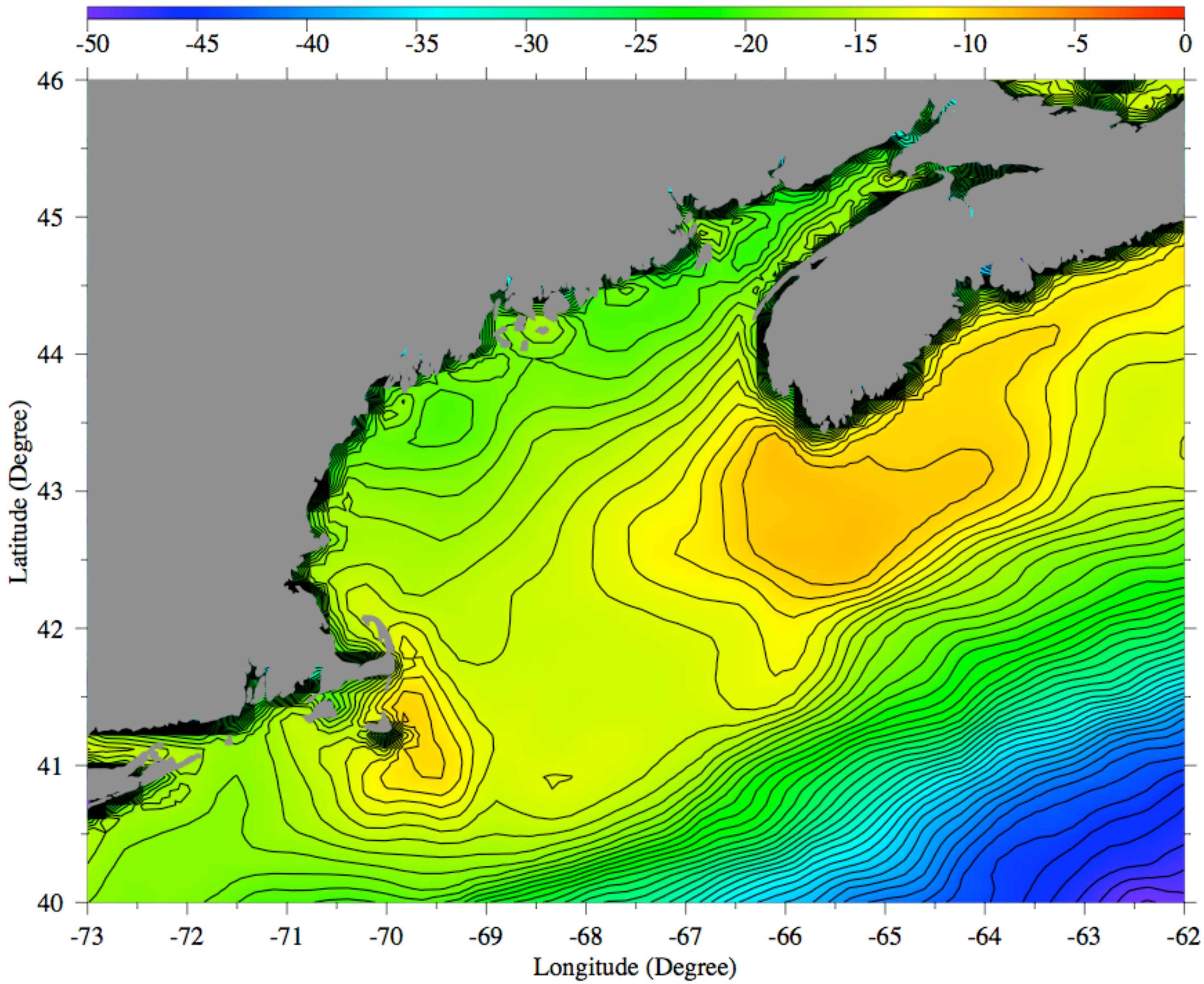
Annual longwave heat flux (W/m^2) averaged over 1978-2004



Annual latent flux (W/m^2) averaged over 1978-2004



Annual sensible flux (W/m^2) averaged over 1978-2004



Suggested papers:

1) Fairall, C. W., E. F. Bradley, D. P. Rogers, J. B. Edson, and G. S. Young, 1996. Bulk parameterization of air-sea fluxes for tropic ocean global atmosphere coupled-ocean atmospheric response experiment. Journal of Geophysical Research, 101(C2), 3747-3764.

2) Beardsley, R., S. Lentz, R. A. Weller, R. Limeburner, J. D. Irish, and J. B. Edson, 2003. Surface forcing on the southern flank of Georges Bank, February-August, 1995. Journal of Geophysical Research, 108, C11, 8007. Dot: 10.1029/2002JC001359.