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"Tracing & Modelling the Ocean Variability"
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Atmosphere-Ocean Interactions

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ATMOSPHERE-OCEAN INTERACTIONS

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1. FLUXES AT THE SURFACE

2. TROPICAL ATMOSPHERE-OCEAN INTERACTIONS

   ENSO

3. MID-LATITUDE ATMOSPHERE-OCEAN INTERACTIONS

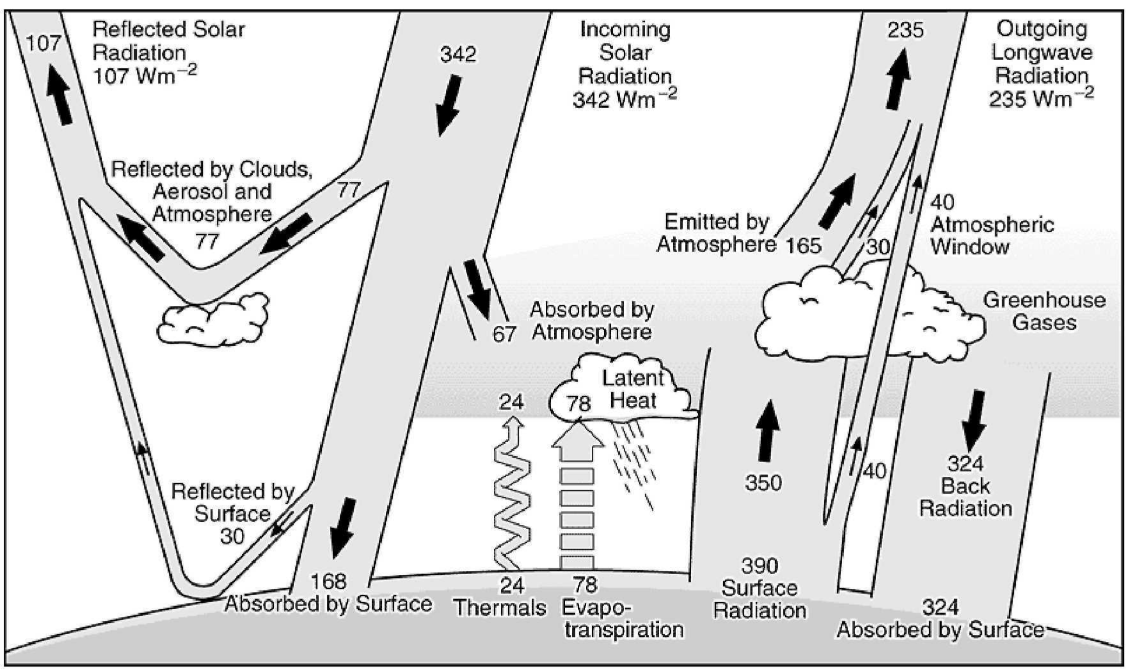
1. FLUXES AT THE SURFACE

► **INTERACTIONS BETWEEN THE ATMOSPHERE AND THE OCEAN TAKE PLACE THROUGH THE INTERCHANGE OF FLUXES AT THE SURFACE**

► **ULTIMATE SOURCE OF ALL FLUXES IS THE SOLAR CONSTANT**

$$F_{SOLAR} = 1367\ W/m^2$$

[Amount of available solar flux at top of atmosphere = ¼ F_{SOLAR} =342W/m^2]



► SURFACE BALANCE (ALL FLUXES POSITIVE UPWARD):

$$-R_{NET} = LE + S + Q \qquad (1)$$

WHERE

$$R_{NET} = F_{SOLAR}(0) + F_{IR\ NET}(0) \qquad (2)$$

Where $F_{IR\ NET}(0) = \varepsilon\ \sigma\ T_S^4 + F_{IR}^{\downarrow}(0)$

LE = LATENT HEAT OF EVAPORATION INTO ATMOSPHERE

S = SENSIBLE HEAT INTO ATMOSPHERE

Q = SENSIBLE HEAT INTO OCEAN

► AN INTERPRETATION OF (1) IS THAT THE NET RADIATION AT THE SURFACE THAT DOESN'T GO INTO EVAPORATING WATER OR SENSIBLE HEATING MUST GO INTO THE OCEAN

► ALL THE TERMS IN (1) AND (2) CAN BE PARAMETERIZED BY QUANTITIES MEASURED ON A SHIP

E.G.

$$F_{SOLAR}(0) \sim F_{BOOK}(0)\,(1 - c_1 n)$$

where n is cloud fraction,

$$F_{IR\ NET}(0) \sim \varepsilon\, \sigma\, T_S^4\, (a - bq_a^{\,r})(1 - c_2 n)$$

$$LE = \rho C_L L |v| (q_{sat}(0) - q_a)$$

where C_L is a drag coefficient, $q_{sat}(0)$ is the saturation vapor mixing ratio at the surface at temperature T_S, L is the latent heat of condensation, and q_a is the vapor mixing ration measured on ship,

$$S = \rho C_S |v| (T_S - T_a),$$

where C_S is a drag coefficient, and T_a is the temperature measured on ship ("anemometer level").

► **29 W/m^2 EVAPORATES 1mm/day**

► **50 W/m^2 INTO A 50 m MIXED LAYER HEATS 1K IN 50 DAYS (OR 100 W/m^2 INTO 100m MIXED LAYER HEATS .6K/MONTH). 100 W/m^2 INTO THE ATMOSPHERE HEATs 1k/day.**

► **BELOW, WE WILL LOOK AT THE INDIVIDUAL COMPONENTS OF THE SURFACE HEAT BUDGET FROM THE SOUTHHAMPTON ATLAS: JOSEY ET AL, 1999.**

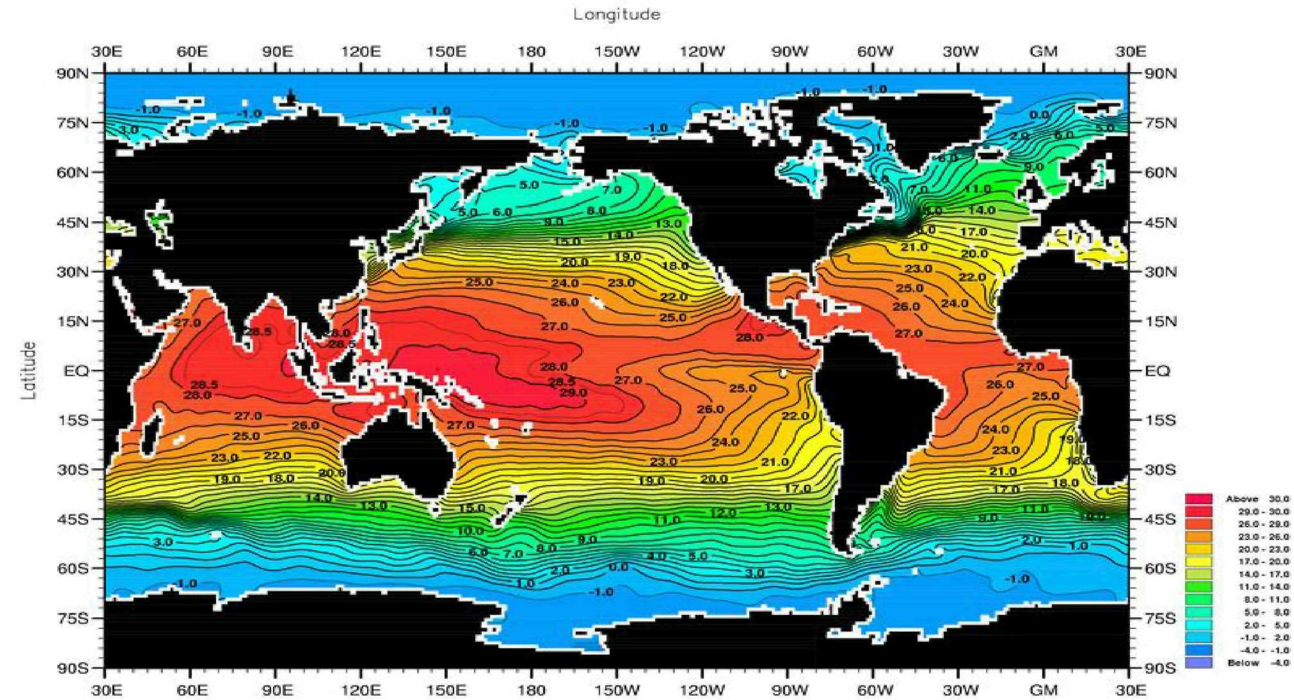
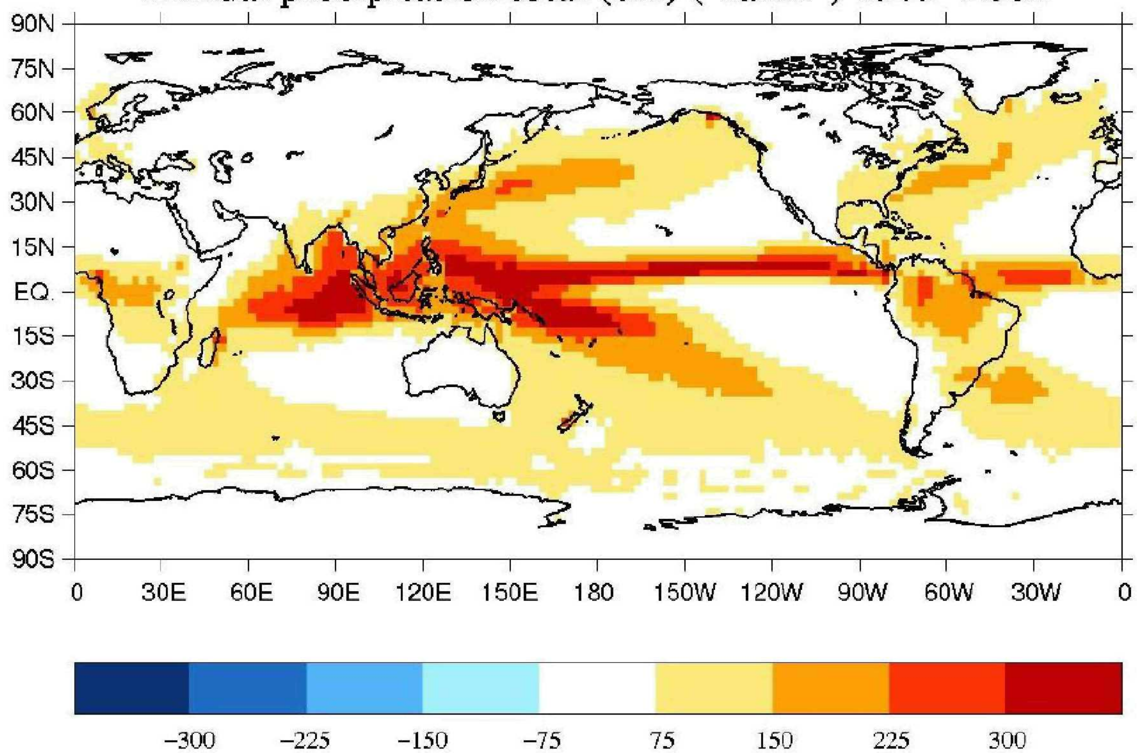
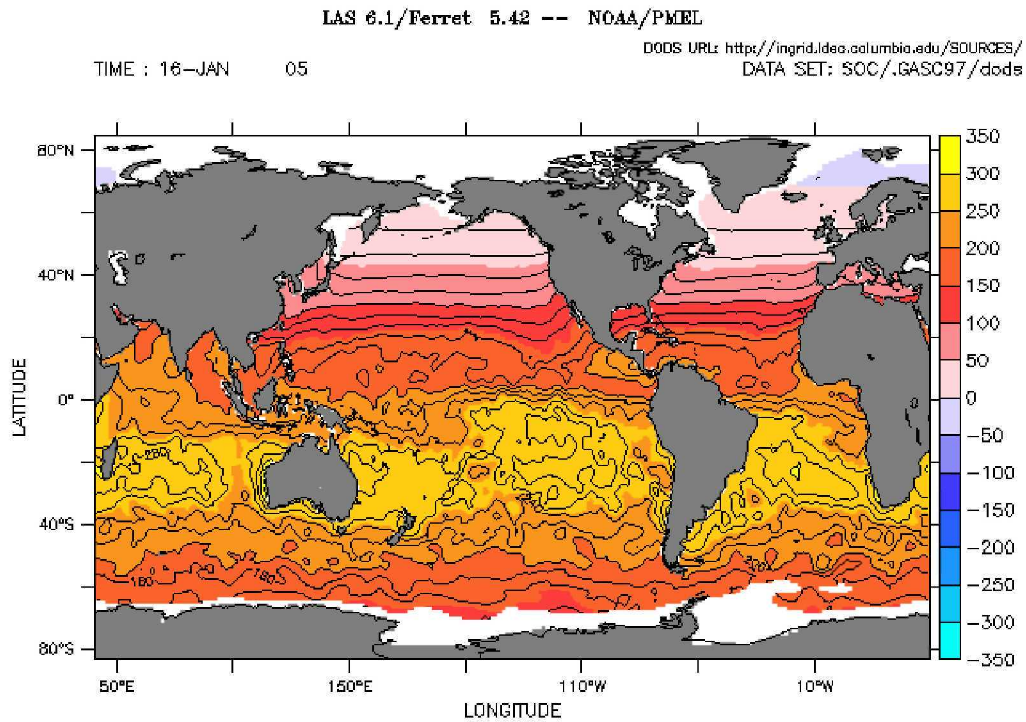


Fig. A2-1. Annual mean temperature (°C) at the surface.

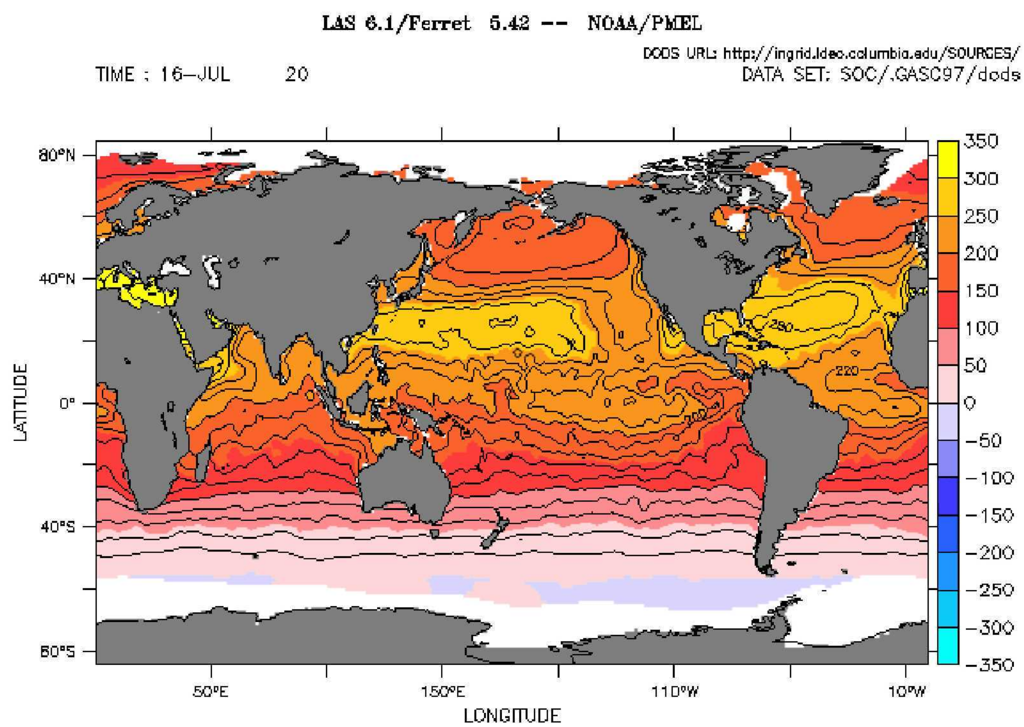
Minimum Value= -1.93 Maximum Value= 29.93 Contour Interval: 1.00

Annual precipitation total (cm) ("rain1") 1979–2001

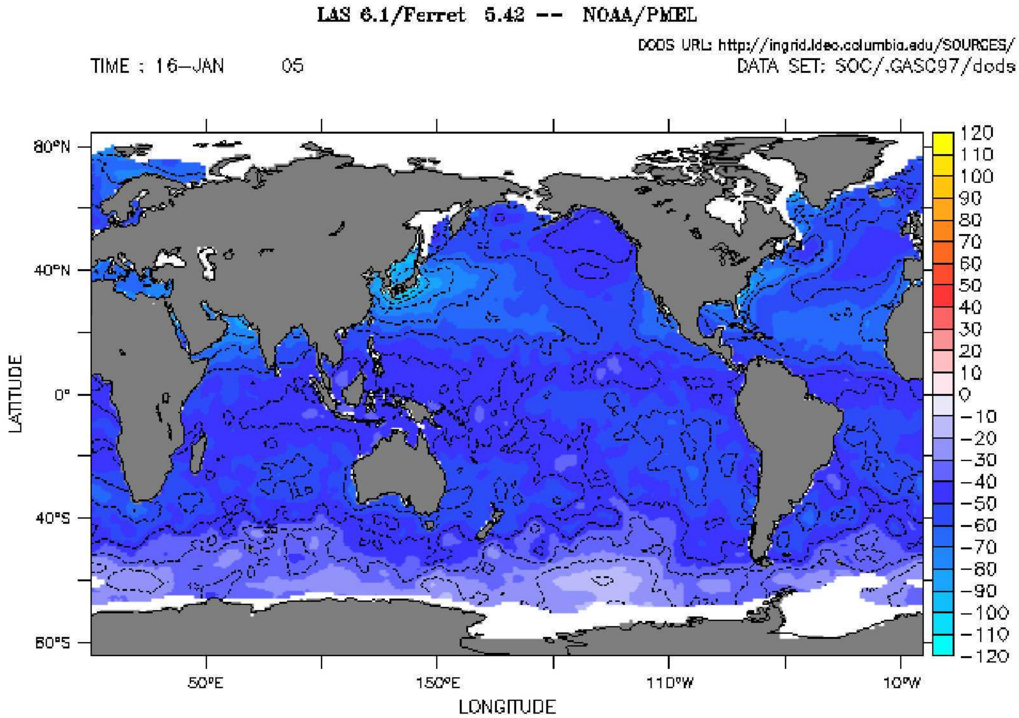




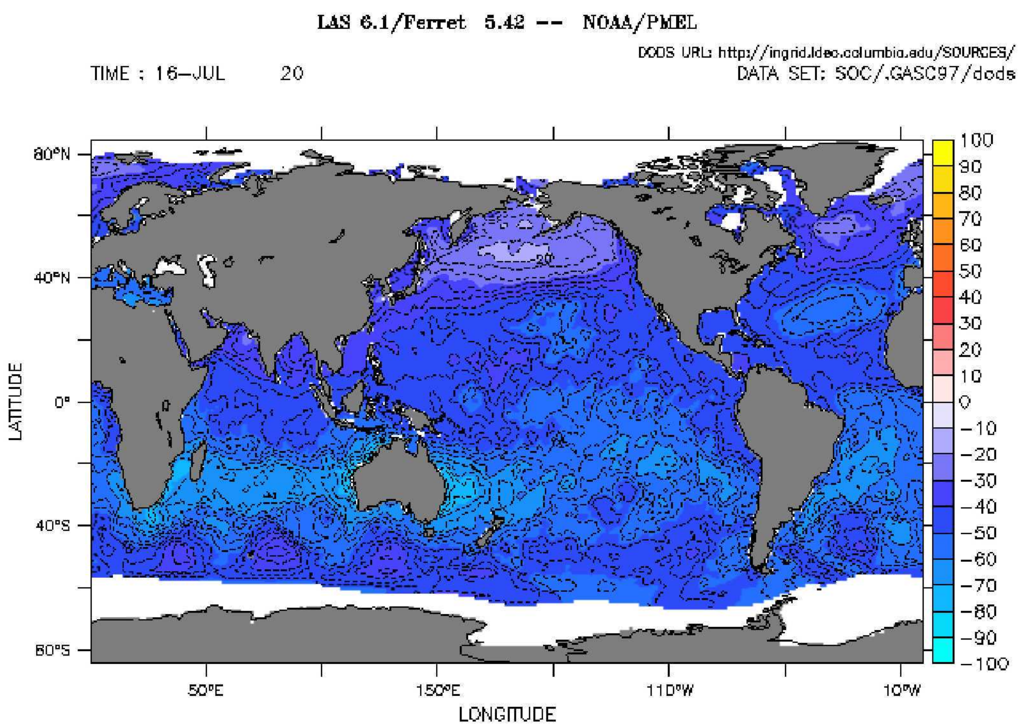
SW radiation net surface (Positive for heat gain by the ocean) (W/m2)



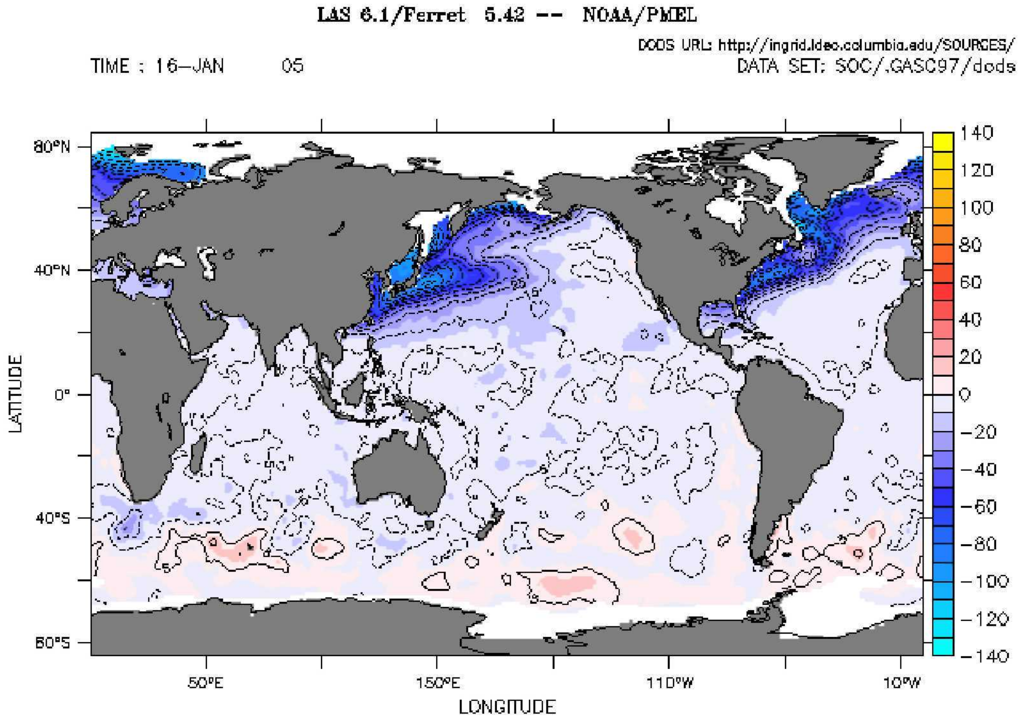
SW radiation net surface (Positive for heat gain by the ocean) (W/m2)



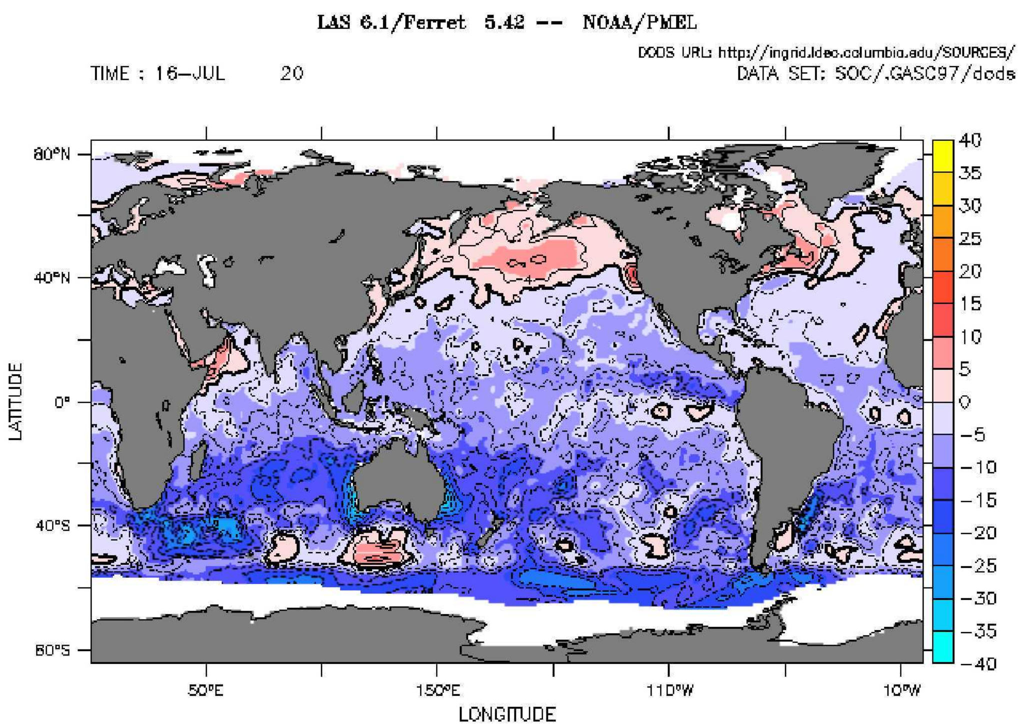
LW radiation net surface (Positive for heat gain by the ocean) (W/m2)



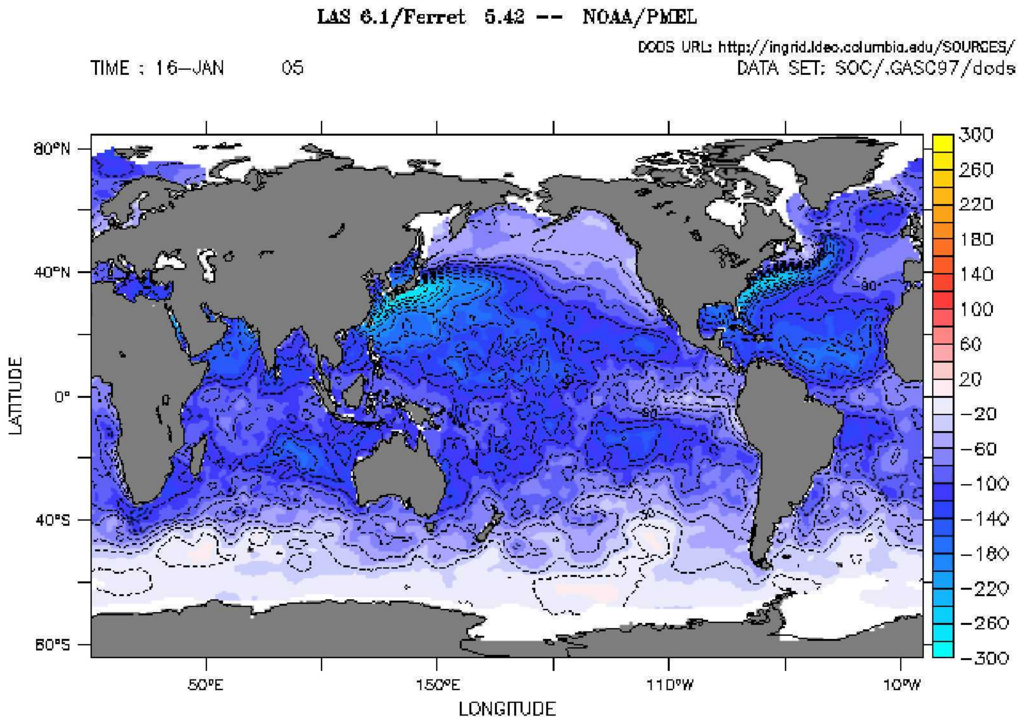
LW radiation net surface (Positive for heat gain by the ocean) (W/m2)



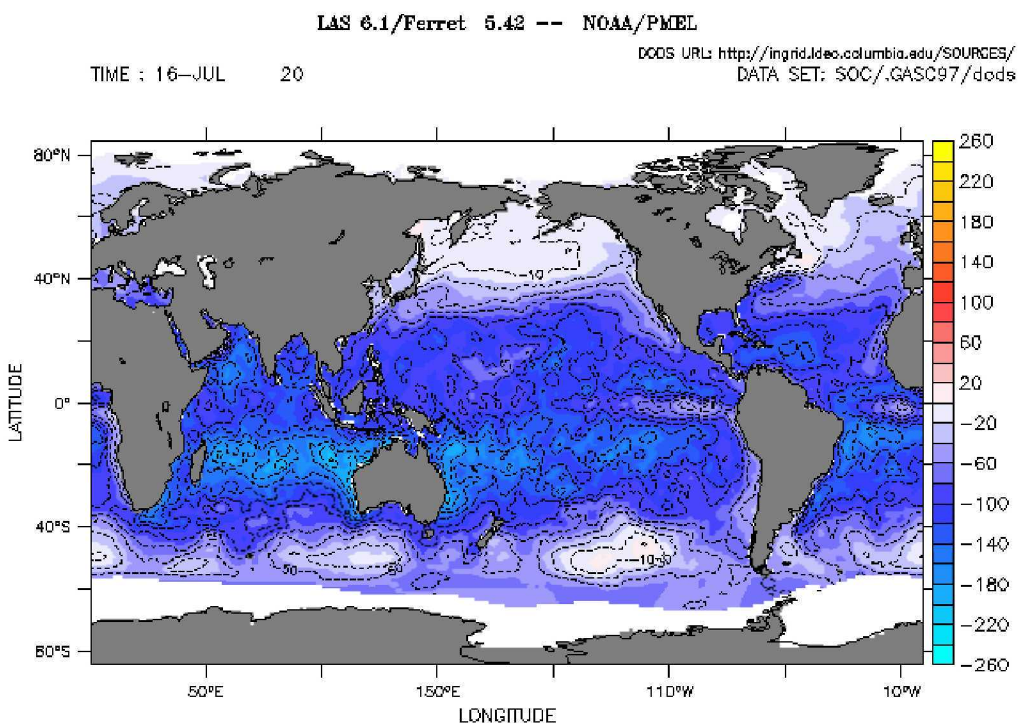
Heat flux sensible surface (Positive for heat gain by the ocean) (W/m2)



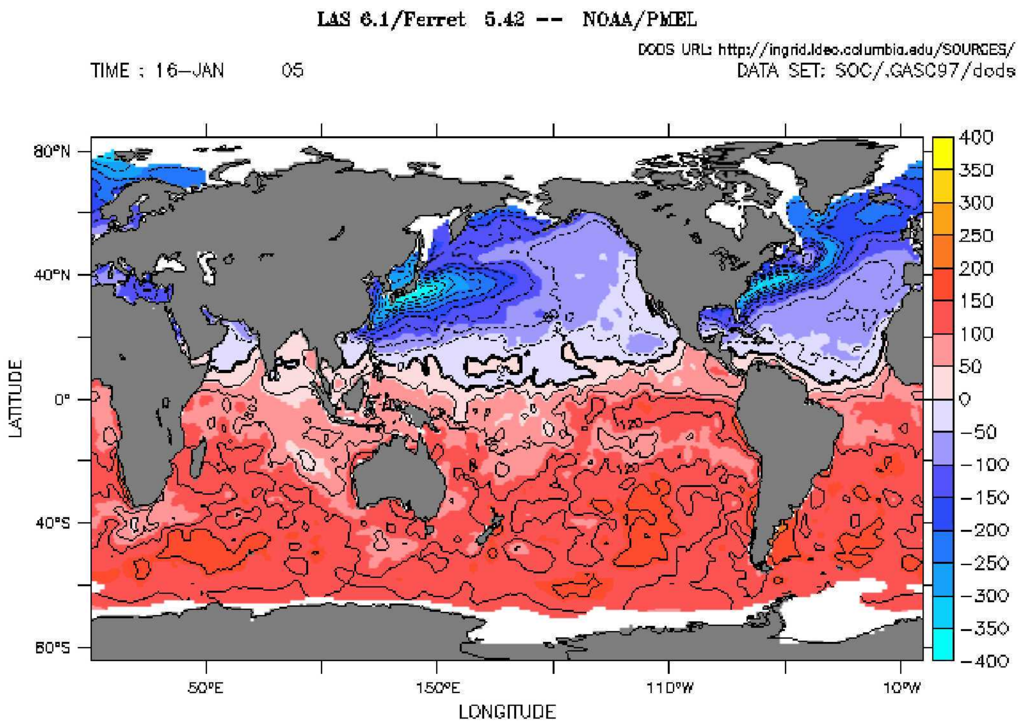
Heat flux sensible surface (Positive for heat gain by the ocean) (W/m2)



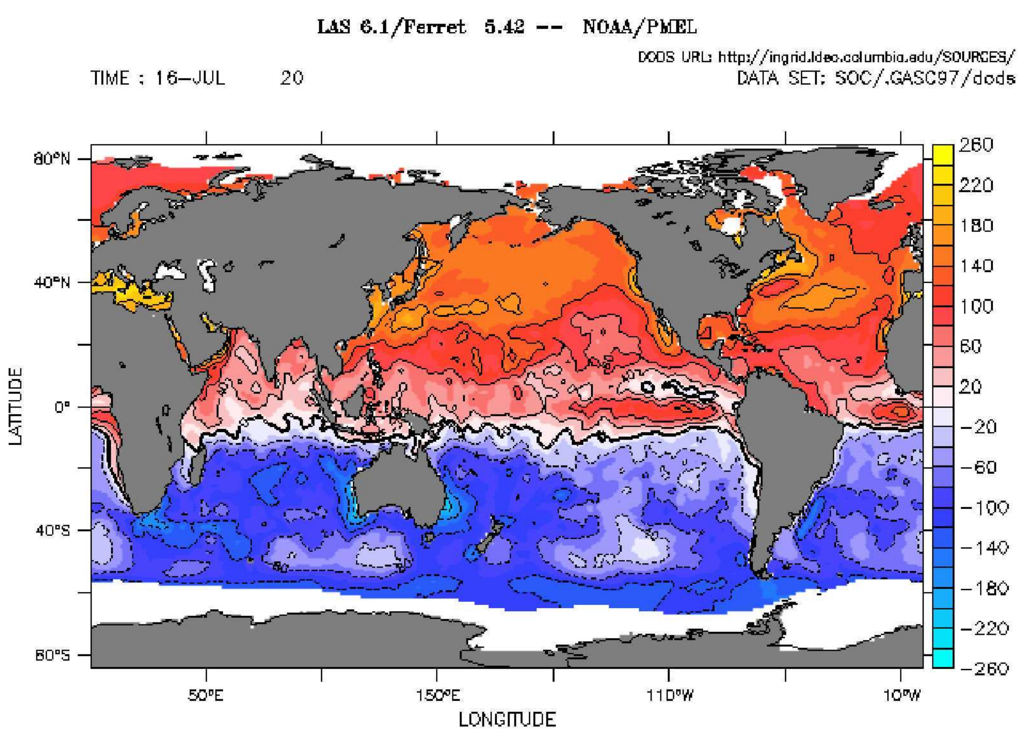
Heat flux latent surface (Positive for heat gain by the ocean) (W/m2)



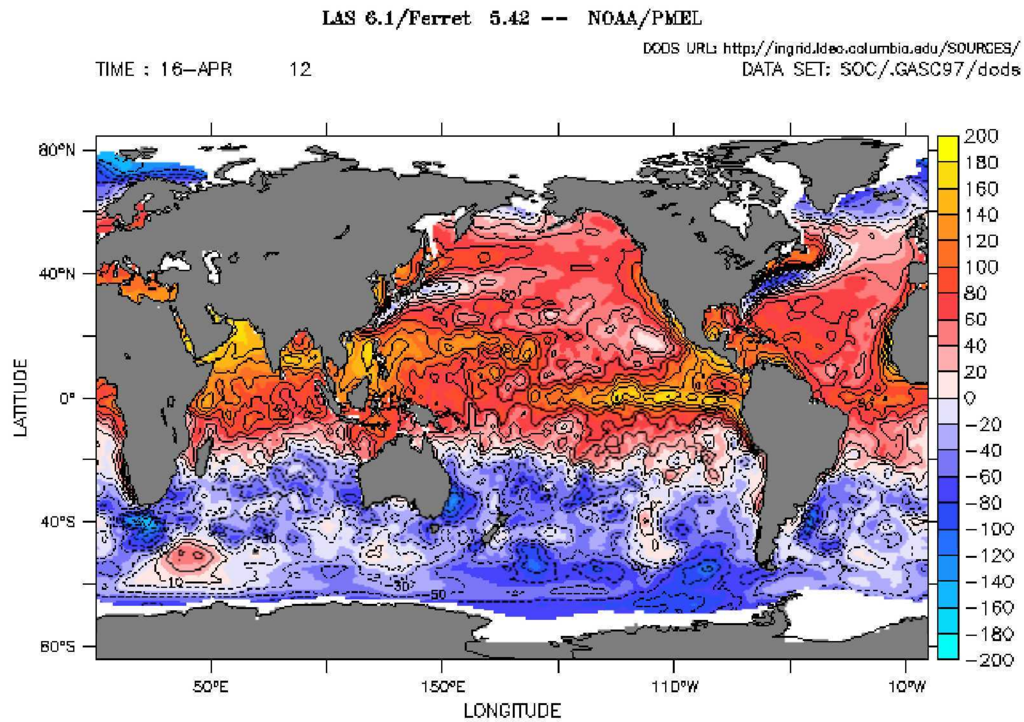
Heat flux latent surface (Positive for heat gain by the ocean) (W/m2)



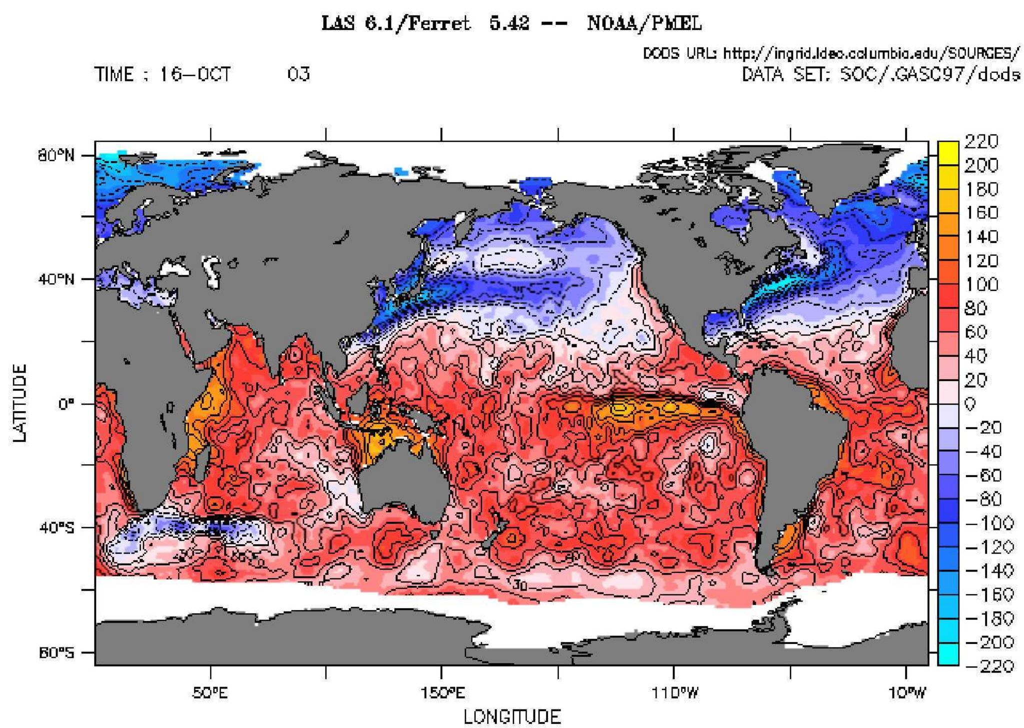
Net heat flux (positive for heat gain by the ocean) (W/m2)



Net heat flux (positive for heat gain by the ocean) (W/m2)



Net heat flux (positive for heat gain by the ocean) (W/m2)



Net heat flux (positive for heat gain by the ocean) (W/m2)

► NOTE THAT SUCH CHARTS ARE EXTREMELY INACCURATE—THE GLOBALLY INTEGRATED ANNUALLY AVERAGED HEAT FLUX INTO THE OCEAN IN THE SOC ATLAS IS 30 W/m^2

► ONCE THE HEAT GOES THROUGH THE SURFACE INTO THE OCEAN, IT CAN ONLY BE STORED OR DIVERGED AWAY:

$$-Q = \text{STORAGE} + \nabla \cdot [\text{TRANSPORT}]$$

► IF WE KNOW THE ANNUALLY AVERAGED FLUXES OVER THE ENTIRE OCEAN [STORAGE=0] AND HAVE A BOUNDARY CONDITION FOR TRANSPORT, WE KNOW THE TRANSPORT.

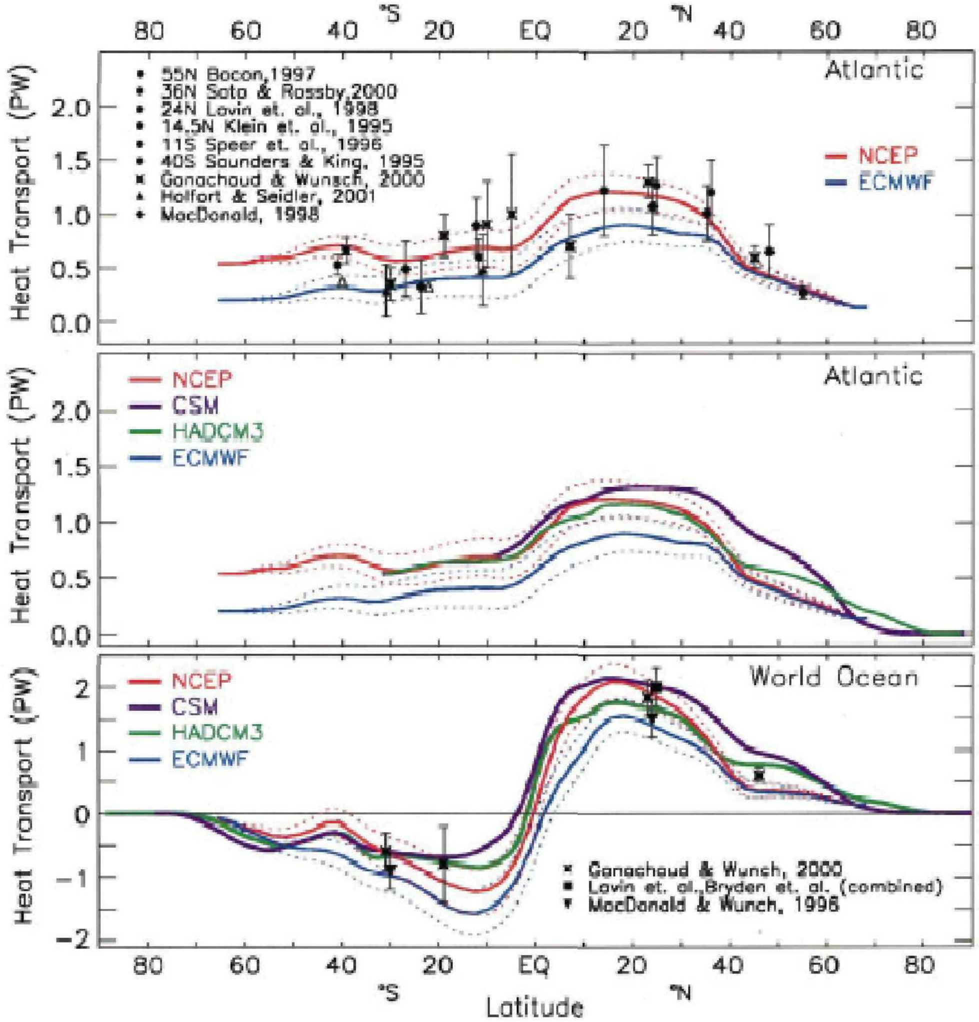


FIG. 6. The northward ocean heat transports from the NCEP-derived and ECMWF-derived products are compared (top) for the Atlantic Ocean with direct ocean estimates from sections, as identified in the key. The dashed curves show the ±1 std err for the derived transports. Where given in the original source, error bars are also plotted and the symbol is solid. Slight offsets in latitude are introduced where overlap would otherwise occur. Several sections are not exactly along a latitude circle, notably those for Bacon (1997) at ~55°N and the Saunders and King (1995) section along 45°S (South America to 10°E) to 35°S (Africa), plotted at 40°S. (middle) Comparison of the derived results with transports from the HADCM3 (years 81–120) and CSM (years 250–299) coupled models for the Atlantic. (bottom) Results for the global ocean along with those from Macdonald and Wunsch (1996) at 24°N and 30°S, and at 24°N the combined Lavin et al. (1998) and Bryden et al. (1991) and for Ganachaud and Wunsch (2000).

Trenberth and Caron, 2001.

2. TROPICAL ATMOSPHERE-OCEAN INTERACTIONS

► **OVER THE TROPICAL PACIFIC, HAVE DEEP CONVECTION WHENEVER SST>28° C**

► **THEREFORE THE PROBLEM OF ATMOSPHERE-OCEAN COUPLING IS TO FIND WHERE SST>28ºC.**

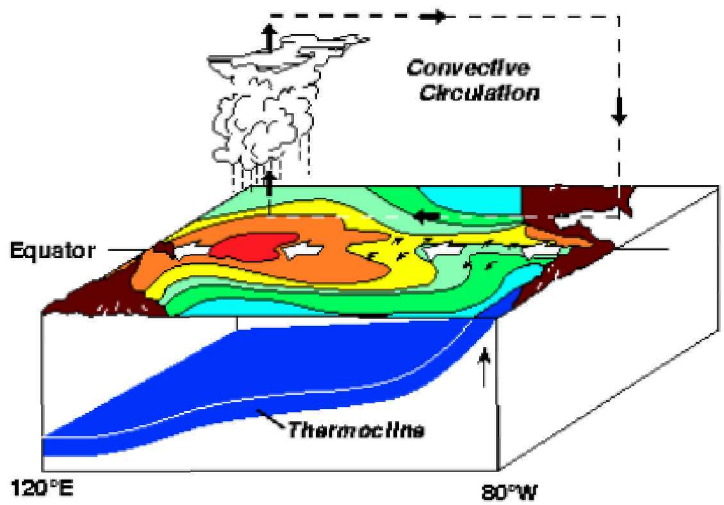
► **SST IS CALCULATED AS THE TEMPERATURE OF THE MIXED LAYER:**

$$\rho CH[\frac{\partial T}{\partial t} + v \bullet \nabla T + w_e \Delta T] = -Q$$

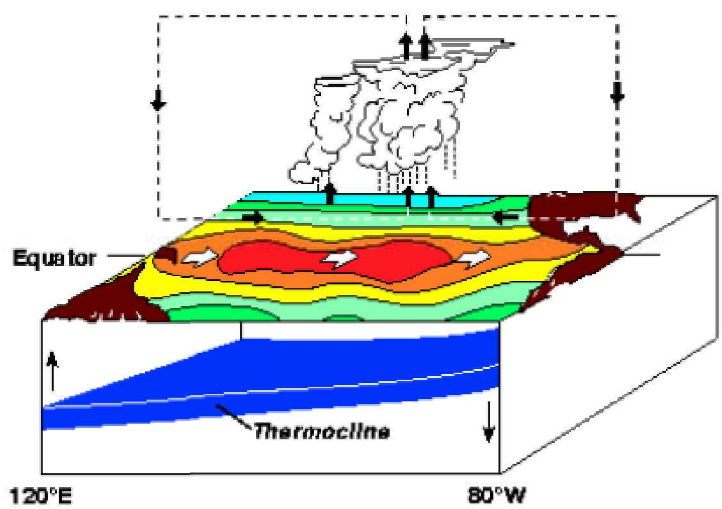
WHERE w_e is the entrainment velocity through the bottom of the mixed layer.

► **THE PROBLEM OF ENSO IS THEREFORE TO CALCULATE THE MUTUAL EVOLUTION OF THE SST AND THE ATMOSPHERIC FIELDS OF PRECIPITATION AND WINDS.**

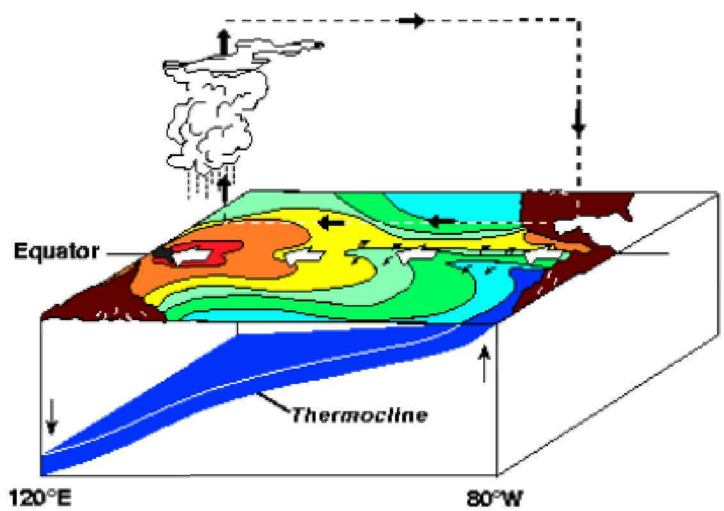
Normal Conditions



El Niño Conditions



La Niña Conditions



3. MID-LATITUDE ATMOSPHERE-OCEAN INTERACTIONS

► MID-LATITUDE A-O INTERACTIONS CHARACTERIZED BY WEAK EFFECT OF SST ANOMALIES ON ATMOSPHERE

► MODEL BY BARSUGLI AND BATTISTI :

$$dT_a/dt = -aT_a + bT_o + N(t)$$

$$\beta dT_o/dt = cT_a - dT_o$$

where β=40, a=1.12, b=.5, c=1, d=1.08.

The free (linear) system can be written as

$$D[T]/dt = A\,[T]$$

Where [T] is a state vector and A is the evolution matrix,

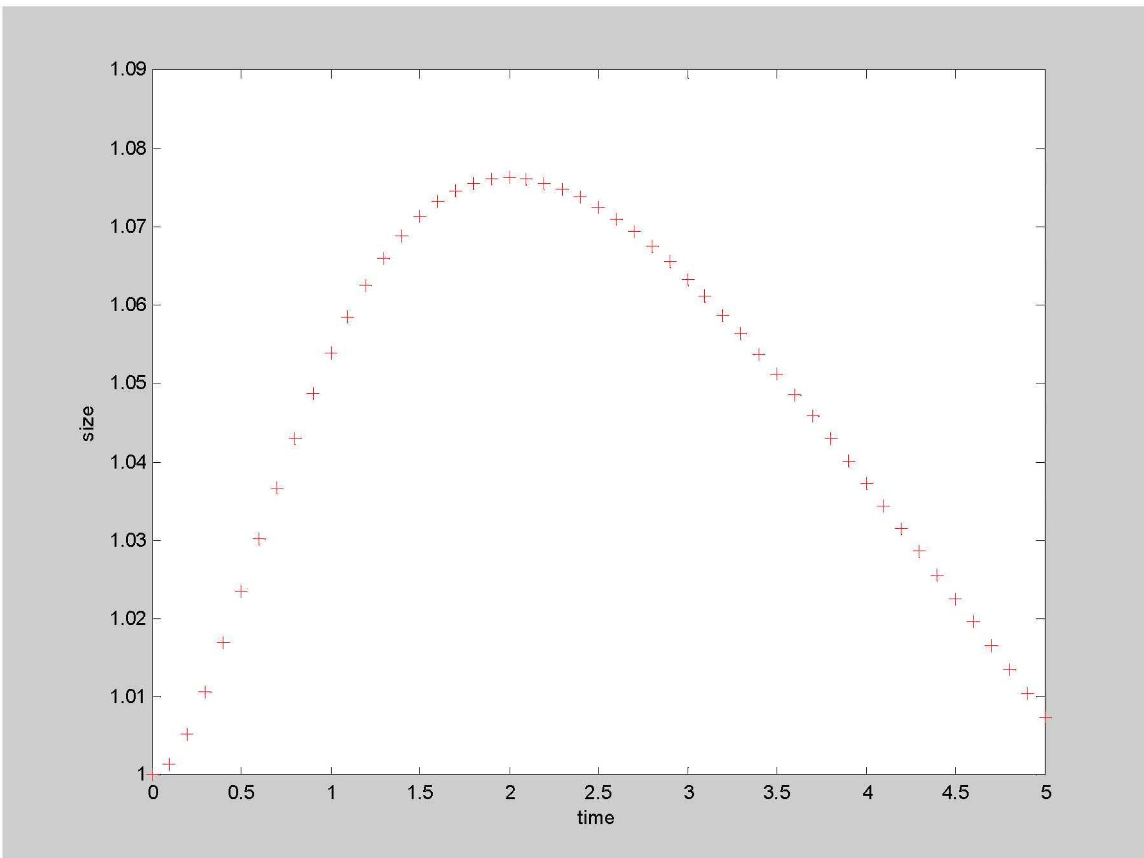
With solution:

$$[T]\,(t) = R\,(t,t_o)\,[T]\,(0)$$

where $R\,(t,t_o) = \exp[A(t-t_o)]$ is the "propagator."

► **FOR THE ABOVE PROBLEM, THE EIGENVALUES OF A ARE NEGATIVE SO THAT ALL MODAL SOLUTIONS DECAY.**

BUT, THE PROPAGATOR IS NON-NORMAL ($RR^+ \neq R^+R$) SO THAT CERTAIN INITIAL CONDITIONS GROW BEFORE DECAYING.



[The initial condition is the first singular vector, the maximum, growth is the square root of the first singular value]

► **NON NORMAL SYSTEMS (E.G. ATMOSPHERE-OCEAN COUPLED) ALLOW A LARGE AMOUNT OF VARIANCE FOR RELATIVELY SMALL STOCHASTIC FORCING.**

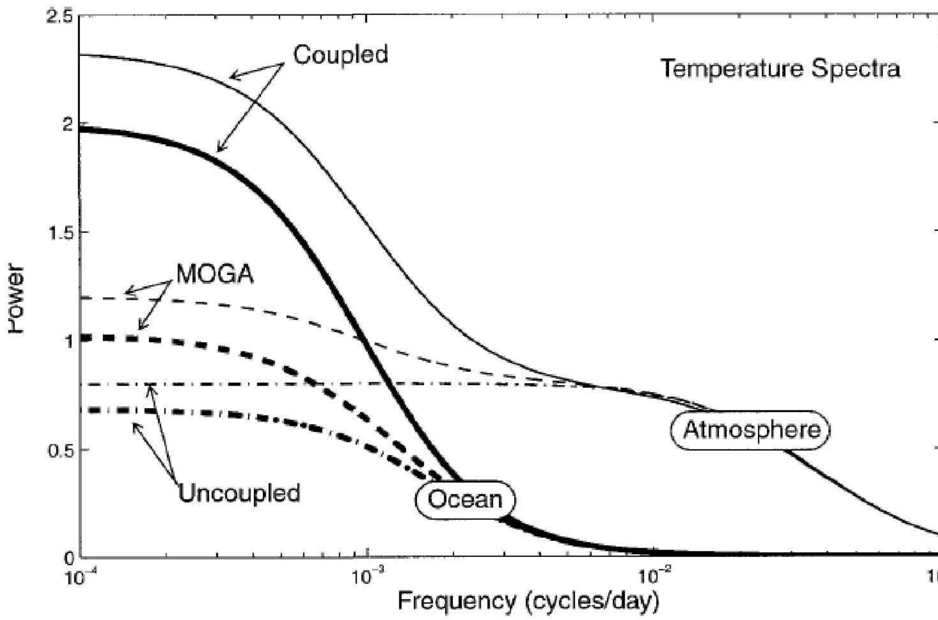


FIG. 4. Power spectra of atmosphere and ocean temperature for the coupled, MOGA, and uncoupled cases. The standard parameters (see Table 1) are used.

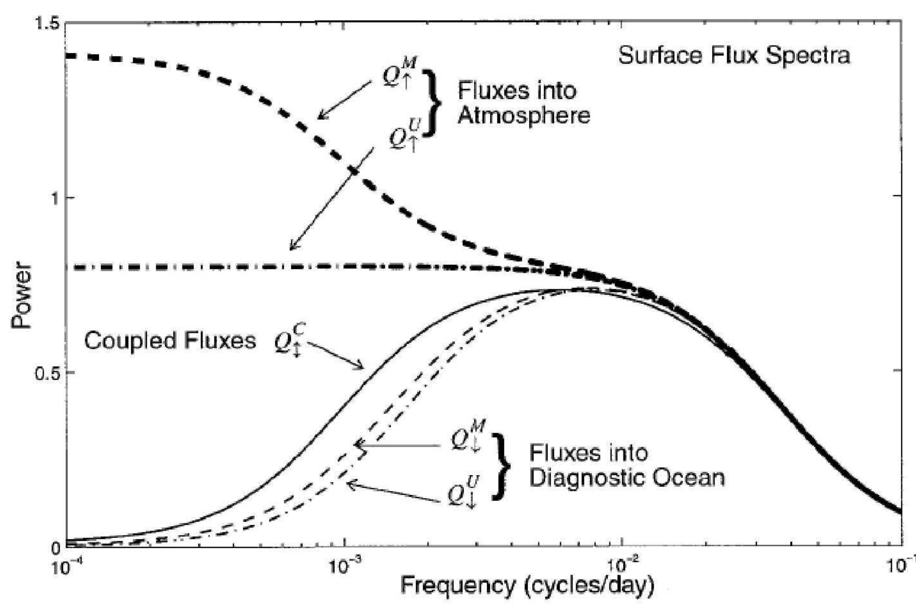


FIG. 7. Power spectra of surface flux, defined as $Q = T_s - T_o$, for coupled, MOGA, and uncoupled cases using the standard parameters. The fluxes into the atmosphere (↑), the fluxes into the diagnostic ocean model (↓), and the coupled fluxes (↕) are shown. The symbols are defined in section 3c.

► **THE GENERAL INTERPRETATION IS THAT COUPLING IN MID LATITUDES INCREASES THE VARIANCE AT LOW FREQUENCIES BY REDUCING THE THERMAL FLUXES WHICH DAMP THE SYSTEM**

REFERENCES:

Barsugli, J.J., and D.S. Battisti, 1998: The basic effects of atmosphere-ocean thermal coupling on midlatitude variability. *J. Atmos. Sci.,* **55**, 477-493.

Josey, S.A., E.C. Kent, and P.K. Taylor, 1999: New insights into the ocean hear budget closure problem from analysis of the SOC air-sea flux climatology. *J. Climate*, **12**, 2857-2880.

Trenberth, K. E., and J. M. Caron, 2001: Estimates of meridional atmosphere and ocean heat transports. *J. Climate*., **14**, 3433-3443.