Relative role of atmosphere and ocean in the global heat budget: tropical Atlantic and eastern Pacific

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(Received 31 August 1976; revised 15 November 1976)

SUMMARY

Calculations of the oceanic heat budget for the tropical Atlantic and eastern Pacific, in conjunction with satellite measurements of net radiation at the top of the atmosphere, permit a partitioning of heat export performed by the oceanic water body v, the atmospheric column. For individual Marsden squares, the divergence of heat transport within the oceanic water body ranges from +82% (export) to −120% (import) of the net radiative input to the system at the top of the atmosphere. Oceanic heat export is particularly conspicuous in the realm of the cold currents off the west coasts and in the cold water tongue immediately to the south of the equator.

1. INTRODUCTION

The tropics at large are the region of energy input to the global circulation. The net radiative heat gain through the upper boundary of the planet earth must be redistributed to higher latitudes: over the continents transports can be effected only within the atmosphere, whereas in the vast tropical sea areas, atmosphere and hydrosphere cooperate in the heat export to other parts of the globe. Satellite measurements of net radiation at the top of the atmosphere and computations of the oceanic heat budget for the Atlantic and eastern Pacific that have recently become available, warrant a reappraisal of the relative importance of tropical atmosphere and ocean in global energetics.

2. DATA

More than seven million ship observations taken over the tropical Atlantic and eastern Pacific during 1911–70 were obtained from the National Climatic Center at Asheville, North Carolina. Each observation includes several parameters: sea level pressure, wind direction and speed, sea surface and air temperature, dew point, total cloudiness, and others. Data have been compiled into a climatic atlas with a 1°-square spatial resolution (Hastenrath and Lamb 1977). An atlas of the oceanic heat budget is in preparation. Satellite-derived net radiation data for the top of the atmosphere over a few years have been published recently (Vonder Haar and Ellis 1974), but these refer to the middle of 10°-squares.

3. BASIC THEORY

A heat-budget scheme for the atmosphere–ocean system is presented in Fig. 1. The budget equation for the system as a whole can be written

\[ R_{\text{top}} = Q_{vo} + Q_{ta} + Q_{vo} + Q_{ao} \]  

The l.h. term signifies the net radiation at the top of the atmosphere. In the r.h. terms the subscripts v and a denote divergence of heat transport and storage, respectively, the subscripts o referring to atmosphere and ocean.
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Figure 1. Heat-budget scheme for atmosphere-ocean system. $R_{N_{\text{top}}}$ and $R_{N_{\text{Sfc}}}$ denote net radiation at the top of the atmosphere and at the ocean surface, respectively; $\text{Rad} = R_{N_{\text{top}}} - R_{N_{\text{Sfc}}}$; $Q_t$, divergence of horizontal heat transport; $Q_1$, heat storage, subscripts $a$ and $o$ referring to atmosphere and ocean; $Q_s$, $Q_e$, sensible- and latent-heat flux at air-sea interface.

The heat-budget equation for the atmospheric column reads

$$Q_{ea} + Q_{ta} = (R_{N_{\text{top}}} - R_{N_{\text{Sfc}}}) + Q_s + Q_e$$

where the r.h. terms denote the net radiative cooling of the atmospheric column, and latent- and sensible-heat transfer at the ocean surface, respectively.

For the heat budget of the ocean

$$R_{N_{\text{Sfc}}} = Q_s + Q_e + Q_{vo} + Q_{lo}$$

the l.h. term being the net radiation at the ocean surface. This in turn is equal to the sum of net shortwave radiation, $SW_{N_{\text{Sfc}}}$, and net longwave radiation, $LW_{N_{\text{Sfc}}}$, at the surface.

Addition of Eqs. (2) and (3) yields Eq. (1), consistent with Fig. 1. Storage terms are vanishingly small for the year as a whole.

4. Net radiation at the top of the atmosphere

Only annual-mean values taken from Vonder Haar and Ellis have been used in Fig. 3 and Table 1. While no error limits are given explicitly, tolerance is here intuitively estimated as about 10%. Annual values denote a heat gain for the entire tropical belt, with largest amounts in the equatorial region and some decrease towards the subtropics.

5. The oceanic heat budget

Reference is made to Eq. (3). For the computation of net shortwave radiation, $SW_{N_{\text{Sfc}}}$, various empirical formulae have been suggested (e.g. Budyko 1958), using total cloudiness ($C_T$) and latitude as input. In the present study use was made of the theoretical framework presented by Bernhardt and Philipps (1958).

The net shortwave radiation at the ocean surface is the resultant of five flux components:

$$SW_{N_{\text{Sfc}}} = SW_{\text{down-sfc}}(\text{dir,clear}) + SW_{\text{down-sfc}}(\text{diff,clear}) + SW_{\text{down-sfc}}(\text{diff,cloudy}) + SW_{\text{up-sfc}}(\text{dir}) + SW_{\text{up-sfc}}(\text{diff})$$

These are the downward-directed direct solar radiation for the clear portion of the sky, the downward-directed diffuse solar radiation for the clear and for the cloudy portions of the sky, and the upward-directed direct and diffuse shortwave radiation, respectively.
The downward-directed direct solar radiation for the clear portion of the sky is given by Bernhardt and Philipps (1958) as

\[ SW_{\text{down-sfc}}(\text{dir}, \text{clear}) = (1 - C_T) \frac{I_0'}{p' \sin \delta} \cos \theta \left(0.907/(\cos \theta)^{0.018}\right) T^{1/4} \text{w/m}^2 \]  \hspace{1cm} (5)

The zenith angle of the sun, \(\theta\), is determined by the declination \(\delta\), the latitude \(\phi\), and the hour angle \(\omega\):

\[ \cos \theta = \sin \delta \sin \phi + \cos \delta \cos \phi \cos \omega \]  \hspace{1cm} (6)

The value for the solar constant \(I_0 = 1352 \text{ W/m}^2\), the ratio of true to mean distance of the sun \(p\), and the solar declination, \(\delta\), for the middle of each calendar month, were taken from List (1968).

\(T\) is the so-called 'old' Linke turbidity factor. A spatially varying \(T\) by calendar month was calculated from surface specific humidity \(q\) (g kg\(^{-1}\)) according to

\[ T = 1.40 + 0.136q \]  \hspace{1cm} (7)

Figure 2. Annual heat export within the ocean. Isopleths from 1°-square analysis of 1911-70 ship data, in \text{W/m}^2.

In specifying the dependence on \(q\) in Eq. (7) use was made of conventional empirical relationships between absorption and precipitable water and between precipitable water and surface humidity. For the contribution by dust a constant value \(0.40\) was adopted, due to lack of information on the spatial pattern and seasonal variation of dust loads over the Atlantic and eastern Pacific (Bernhardt and Philipps).

To obtain monthly means of \(SW_{\text{down-sfc}}(\text{dir}, \text{clear})\), Eqs. (5) and (6) were applied by hourly intervals and values were then integrated over all sun angles. This was done for the middle of each calendar month and for 1°-latitude strips, with a turbidity factor spatially varying according to Eq. (7).

Values of \(\eta = \beta = 0.36\) were used in the calculation of \(SW_{\text{down-sfc}}(\text{diff}, \text{clear})\) and \(SW_{\text{down-sfc}}(\text{diff}, \text{cloudy})\). Here \(\eta\) is the ratio of diffuse radiation under completely cloudy sky to global radiation under cloudless sky; and \(\beta\) is the ratio of diffuse radiation to the difference between extraterrestrial radiation and direct radiation under cloudless sky. With these ratios, \(SW_{\text{down-sfc}}(\text{diff}, \text{clear})\) and \(SW_{\text{down-sfc}}(\text{diff}, \text{cloudy})\) were calculated using Eq. (5). The upward-directed shortwave flux components, \(SW_{\text{up-sfc}}(\text{dir})\) and \(SW_{\text{up-sfc}}(\text{diff})\), were then computed using an albedo of the ocean surface of 6% (List).
$SW_{\text{Ndc}}$ was computed using the above procedure, rather than that due to Budyko (1958), because it accounts for spatial variation of water vapour turbidity and would allow for varying dust turbidity as pertinent information may become available. For comparative purposes both procedures were applied to actual data, with the following results. Global radiation under cloudless sky, $SW_{\text{down-sf}}(\text{dir,clear}) + SW_{\text{down-sf}}(\text{diff,clear})$, computed using the above method, is smaller than Budyko’s latitude-mean values by about 15 W m$^{-2}$. However, variations along a latitude circle resulting from the humidity pattern are of the same magnitude, but are ignored in the Budyko procedure. For global radiation with actual cloud cover, the above method yields values only about 5-10 W m$^{-2}$, a few per cent, smaller than Budyko’s procedure: Budyko uses a somewhat stronger reduction of radiation due to cloudiness. The two methods assume the same value for the albedo of the ocean surface. Consequently maps of net shortwave radiation with actual cloud cover differ by a few percent. This is well within the uncertainty of either method.

Net longwave radiation was computed from Brunt’s formula (Budyko):

$$LW_{\text{Ndc}} = 4\varepsilon\sigma T_w^4(0.39 - 0.056\sqrt{q}(-0.53C_p^2) + 4\varepsilon\sigma T_a^4(T_w - T_a)$$  \hspace{1cm} (8)

sea surface temperature, $T_w$, and air temperature, $T_a$, being in $K$; surface specific humidity, $q$, in g kg$^{-1}$; emissivity $\varepsilon = 1$; and Stefan-Boltzmann’s constant, $\sigma = 567.10^{-10}$ W m$^{-2}$K$^{-4}$.

Sensible- and latent-heat fluxes were calculated from the bulk aerodynamic equations:

$$Q_s = \rho C_D c_p(T_w - T_a)V$$  \hspace{1cm} (9)

$$Q_e = \rho C_D L(q_w - q_d)V$$  \hspace{1cm} (10)

Values of $\rho = 1.175 \text{ kg m}^{-3}$ and $C_D = 1.410^{-3}$ were used for air density and drag coefficient, respectively; $c_p$ is specific heat at constant pressure; and $L$ latent heat of evaporation. The saturation specific humidity, $q_w$, corresponding to the sea surface temperature $T_w$, was computed with reference to a salinity of 35 per mille. Scalar mean wind speed was used for $V$.

Of the observational data mentioned in section 2, the following elements thus served as input to the heat-budget calculations: total cloudiness, dew point, and pressure for $SW_{\text{Ndc}}$; sea surface and air temperature, dew point, pressure, and cloudiness for $LW_{\text{Ndc}}$; sea surface and air temperature, dew point, pressure, and scalar wind speed for $Q_s$ and $Q_e$.

Calculations were performed by calendar month, using 60-year-mean data. Covariance between elements can make the product of averages differ from the average of products, and $C_D$ depends on stability (Bunker and Worthington 1976). However, these effects may be of subordinate importance for the low-latitude oceans mapped here. In fact, Bunker’s (1976) annual maps of $Q_s$ and $Q_e$ are rather similar to the present charts: only in limited areas are his figures of $Q_s$ by about 15 W m$^{-2}$ larger than the present ones, a difference of
about ten per cent. However, Bunker's values of net radiation $SWLW'$, calculated from somewhat different empirical formulae, are systematically larger than in the present study. Consequently, his residual $(Q_v + Q_i)_o$ differs from Fig. 2 towards larger positive values. Within the error limits of calculation methods this could be accounted for by a smaller emissivity and somewhat different coefficients in Eq. (8). Annual maps of heat-budget components were constructed from the sets of twelve-monthly computations. All maps were machine-isoplethed and then redrawn by hand.

After calculation of the l.h. and the first two r.h. terms of Eq. (3), the sum of the third and fourth r.h. terms was obtained as a residual. This residual is mapped in Fig. 2 for the year as a whole, thus representing essentially the heat export within the oceanic water body. Superimposed on the broad decrease from the equatorial belt towards the subtropics, the cold water tongue immediately to the south of the equator stands out as an effective exporter of heat, with much smaller values in the band of maximum cloudiness and rainfall around $5-10^\circ N$.

![Figure 4](image_url)  
**Figure 4.** Atlantic, latitudinal and seasonal variation of residual heat storage and export within the ocean, $(Q_v + Q_i)_o$, in W m$^{-2}$. Dot pattern denotes positive areas.

Seasonal variations of the residual heat export/import plus storage/depletion, $(Q_v + Q_i)_o$, can be appreciated from Fig. 4. Values are generally largest in the respective spring and summer seasons, and small in winter towards the subtropics, where they may become negative. Large storage/export results not only in southern hemisphere summer, but remarkably also in the cold water tongue immediately to the south of the equator during northern hemisphere summer and autumn. By contrast, residual $(Q_v + Q_i)_o$ is then very small in the band of abundant cloudiness and rainfall around $5-10^\circ N$.

Errors in the estimation of the l.h. and the first two r.h. terms of Eq. (3) for the annual mean may have typical magnitudes of 10–20, 10–15, and 1–2 W m$^{-2}$. The error in the residual $(Q_v + Q_i)_o$ is accordingly estimated to be of the order of 20–30 W m$^{-2}$. It may be recalled that Bunker's (1976) figures for this term differ towards larger positive values.
6. Conclusions

The relative importance of the ocean in the heat export to other parts of the globe is illustrated in Fig. 2 and Table 1 in terms of the spatial pattern, and in Fig. 3 by means of meridional profiles. Because of the comparatively modest spatial variation of net radiation at the top of the atmosphere, the ratio of oceanic to total heat export follows a pattern similar to \((Q_\text{a} + Q_\text{b})_0\). For individual Marsden squares (Table 1) of the tropical Atlantic and eastern Pacific the divergence of heat transport within the oceanic water body ranges between +82% (export) and −120% (import) of the net radiative heat input to the system at the top of the atmosphere.

The present results cannot be readily compared with Oort and Vonder Haar (1976) since their figures refer to complete latitude bands.

Positive and negative areas of residual \((Q_\text{a} + Q_\text{b})_0\) can be partly related to major surface currents. However, Bunker and Worthington have given examples of large regional heat export/import which take place by mechanisms other than organized surface currents. Oceanic heat import is most conspicuous in the Gulf Stream domain, and export is particularly large in the cold Canary, Benguela, and Humboldt current regions, as well as in the tongue of cold water immediately to the south of the Atlantic and Pacific equator. This is mainly a consequence of the reduced latent-heat flux characteristic of these environments. By contrast, the ocean contributes relatively little to the total heat export in the band of abundant cloudiness and precipitation near 5–10°N, where sea surface temperature is high. In this region, a substantial portion of heat disposal takes place within the atmosphere.

The relative partitioning between oceanic and atmospheric heat export in Table 1 and Figs. 2 and 3 is contingent upon the representativeness of satellite-measured net radiation at the top of the atmosphere and the calculations of the oceanic heat budget. However, the present error-limits are rather smaller than for data available in earlier studies. With this reservation and with the exception of the extended northern hemisphere band of maximum cloudiness and rainfall, the hydrosphere of the tropical Atlantic and eastern Pacific is found to play a substantial role in the redistribution of heat to other parts of the globe.

Acknowledgments

This study was supported by the Office for Climate Dynamics of the National Science Foundation. P. Lamb and P. Guetter did the computer programming.

References


