ANALYZED VERTICAL HEAT FLUXES OVER THE NORTHERN HEMISPHERE

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ABSTRACT

Methods for determining the observed diabatic heating Q of the atmosphere are reviewed. The direct method (measuring the components of Q) is less accurate than the indirect or residual method (measuring the atmospheric fluxes which balance Q). Zonal mean patterns of residual Q for the seasons of the northern hemisphere are presented. The estimated error is some 10^{-6} K s^{-1}.

An important part of the Q-budget is the vertical eddy heat flux. We discuss it in the form $\overline{\omega'h'}$ where h is moist static energy. This flux is mostly upward and converges in vertical direction. Its divergence is in approximate balance with the radiation flux divergence (radiative-convective equilibrium). $\overline{\omega'h'}$ is strongly influenced by subsynoptic processes. We review a method to determine it from the Q-budget with overspecification. This yields error estimates for the flux components. An important ingredient of the method is an objective technique for eliminating spurious divergence errors from the observed mass flux field.

Vertical climatological profiles of $\overline{\omega'h'}$ are presented for 48 equal area regions covering the northern hemisphere (4 latitude strips, 12 latitudinal sectors, 4 layers in the vertical). The results are discussed along with limitations of the method. The rms-error of each of the fluxes of the Q-budget is below 26 W m^{-2}.

1. INTRODUCTION

The first law of thermodynamics reads:

$$c_p \frac{T}{\Theta} \frac{d\Theta}{dt} = Q. \qquad (1.1)$$

T is actual temperature, Θ is potential temperature, c_p is specific heat at constant pressure. Q denotes the total diabatic heating. Its components are:

$$Q = Q_{RAD} + Q_{LH} + Q_{COND} + Q_{FRIC} + Q_{DIFF} \qquad (1.2)$$

They describe, respectively, heating by radiation, condensation (latent heating), conduction, friction and diffusion. The first two are large; they are due to the fact that the atmosphere is an open system (exchange of radiation and water substance). The last three are small and shall be neglected in the following; they are due to internal molecular processes of the system. Without the molecular terms the global atmospheric average of Q vanishes (Lorenz, 1967).

Q can be neglected in many cases of considerable meteorological significance (for example: short-term numerical weather forecasting). Nevertheless it is just the deviation from the idealized adiabatic case that eventually drives the motion systems of the planetary atmosphere. Thus knowledge of Q and its components and their distribution in space and time is material for an understanding of the general circulation. The purpose of this paper is to briefly review contemporary estimates of Q and its components. Main emphasis shall be on the vertical eddy heat flux components and on the errors involved.

2. ZONAL MEAN DIABATIC HEATING

There are two ways of estimating Q. The first (*direct method*) is to separately estimate its components according to eq. (1.2). This has been done by Newell et al. (1969) and by Newell et al. (1974). These authors split Q into Q_{RAD} plus Q_{LH} plus boundary layer heating (equivalent to vertical eddy flux convergence generated by time-averaging) and presented seasonal patterns for the zonal mean global atmosphere. Their results show consistent net heating in the inner tropics and net cooling in the subtropics and over the polar caps. However, the two estimates differ considerably in magnitude. In the extratropics the differences are particularly severe. For the summer season the 1969 paper reported lower atmosphere cooling in the latitude belt 30-60°N while the 1974 paper reported heating (compare Fig. 8 in the first paper and Fig. 7.18 in the second). These discrepancies are largely due to the different radiation employed in the two estimates. The uncertainties in the Q-fields are further accentuated by strong compensation between Q_{RAD} and Q_{LH} and by the unknown vertical distribution of the latter. For example, in the 1974 paper the authors were forced to distribute the surface precipitation in the vertical on the basis of preliminary cloud statistics and on model profiles of vertical motion. In summary, this approach is of limited value due to the limitations of the input data which cannot be measured but must be parameterized.

The second way of estimating Q (*indirect* or *residual method*) is to determine the left-hand side of eq. (1.1). This has been done by Hantel and Baader (1978) on the basis of the general circulation statistics of the MIT-Library (Oort and Rasmusson, 1971) along with a parameterization of the synoptic vertical eddy heat flux convergence (Hantel and Baader, 1976).

Fig. 1 shows the seasonal estimates of Q with an estimated error of 10^{-6} K s^{-1}. We note the following large-scale details:

a) The tropical zone (10°S-10°N) is a region of considerable net *heating* throughout the year, up to 20 units (10^{-6} K s^{-1}).

b) The subtropics are regions of net *cooling*, most pronounced in winter and spring, up to 10 units.

c) The midlatitudes show a more irregular pattern. By and large, there is tendency for net *heating*, most pronounced in spring and summer.

d) The polar cap is a region of considerable net *cooling*, with highest intensity in winter and spring, up to 25 units.

This four-latitude belt pattern can be qualitatively explained with the combined action of radiation and latent heating. Q_{RAD} is negative throughout most of the atmosphere (Dopplick, 1979) and exhibits a fairly smooth latitudinal distribution with little structure.

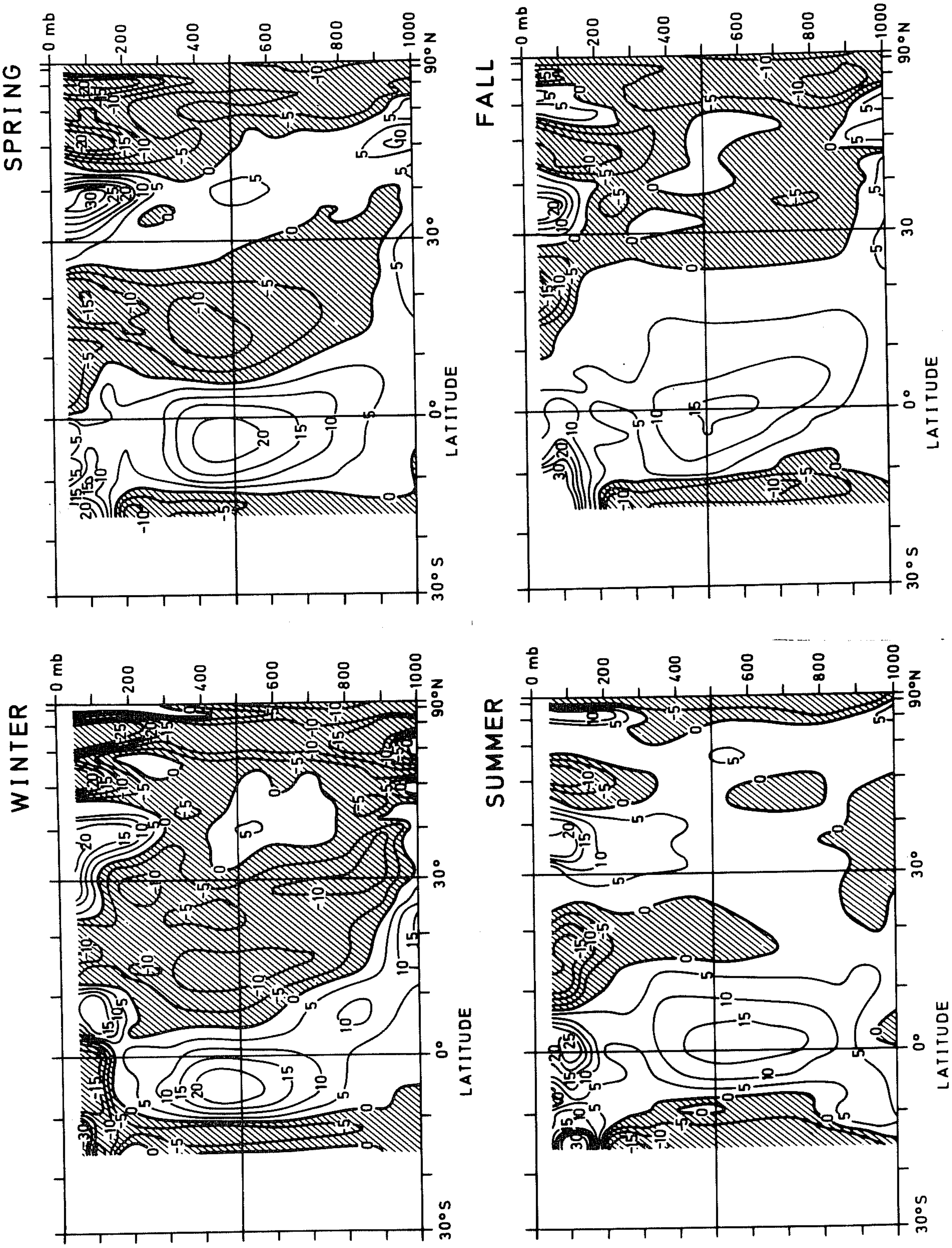

Fig. 1 Total diabatic heating (10^{-6}K s^{-1}) of zonal mean northern atmosphere determined from synoptic data by residual method.

The latitudinal structure in Fig. 1 is largely caused by Q_{LH} with maxima in the climatological rainbelts and minima in the dry zones (Hantel and Langholz, 1977). Consequently, Q_{LH} dominates in the tropics and midlatitudes, Q_{RAD} dominates in the subtropics and over the polar cap.

A conspicuous feature of Fig. 1 is the stratospheric heating between 30-40°N throughout the year. It cannot be explained by radiation or condensation processes. Since the maximum heating region coincides in latitude with the subtropical jet, Hantel and Baader (1978) speculated that there might be a downward heat flux of eddy potential temperature from the upper into the lower stratosphere and down across the tropopause. A vertical flux of about 20 W m^{-2}, if it converges within a 100 mb-layer, would account for heating rates of 20×10^{-6} K s^{-1} and could explain the feature.

Hantel and Baader compared their results with the estimates of the Newell group and concluded that the uncertainties of their indirect method are less severe than of the direct method employed by Newell et al. The principal limitation of the indirect method is that one cannot specify from observation or parameterization the subsynoptic contributions to the zonal and time average of $d\Theta/dt$. The estimates of Fig. 1 share this deficiency with the geographical Q-maps at certain levels compiled by Geller and Avery (1978) or Lau (1979).

3. HOW TO OBTAIN ERROR BARS FOR THE DIABATIC HEATING

The simplest method of estimating the error of a physical quantity is to measure the quantity twice. The difference is the error. When the same quantity is measured many times one considers the variance of the measurements as the expected statistical error of the quantity.

This principle of overspecification has not yet been applied in the existing estimates of Q, for various reasons. One is that there is a natural noise in climatological data, in addition to the observation error. It is not obvious a priori how to separate the two. In any case, the methods described above yield at most indirect error estimates. Neither the direct nor the indirect method has yielded objective error estimates.

In order to nevertheless come to an objective error estimate we shall employ the simple method of overspecification in a rudimentary way. We transform the first law (1.1) into the condensation-free form:

$$\frac{dh}{dt} \simeq Q_R \tag{3.1}$$

h is the *moist static energy*:

$$h \equiv c_p T + gz + Lq \tag{3.2}$$

Eq. (3.1) has been used with good success, for example by Yanai et al. (1973), to estimate the complete energy cycle including the subgrid-scale processes. Yanai's studies were limited to certain regions in the tropics.

In order to apply the h-equation for large-scale budgets we average (3.1) with respect to time ($\overline{}$ + '):

$$\underbrace{\frac{\partial \bar{h}}{\partial t} + \frac{\partial \overline{uh}}{\partial x} + \frac{\partial \overline{vh}}{\partial y} + \frac{\partial \bar{\omega}\bar{h}}{\partial p}}_{\text{synoptic}} + \frac{\partial \overline{\omega' h'}}{\partial p} = \underbrace{- g \frac{\partial \bar{R}}{\partial p}}_{\text{satellite + model}} \tag{3.3}$$

The first 4 terms on the left can be determined with synoptic data; note that $\overline{uh}$, $\overline{vh}$ comprise mean plus *synoptic eddy* contributions while $\bar{\omega}\,\bar{h}$ contains no eddies. The radiation flux R (in W m^{-2}, positive downward) on the right can be determined at the top of the atmosphere via satellite (e.g., Ellis and Vonder Haar, 1976) and at atmospheric levels through models (e.g., Dopplick, 1979).

It follows that (3.3) can be used for estimating the fifth term on the left which is due to synoptic plus *subsynoptic eddies*, as a residual. We shall do this through overspecification. (3.3) reads in vertically integrated form (2 vertical boxes of an atmospheric column):

$$\overline{\omega' h'}\Big|_{500\text{ mb}} = - \int_0^{500} \left(\frac{\partial \bar{h}}{\partial t} + \frac{\partial \overline{uh}}{\partial x} + \frac{\partial \overline{vh}}{\partial y} + \frac{\partial \bar{\omega}\bar{h}}{\partial p} + g\frac{\partial \bar{R}}{\partial p}\right) dp \tag{3.4}$$

$$-\overline{\omega' h'}\Big|_{500\text{ mb}} = - \int_{500}^{1000} \left(\frac{\partial \bar{h}}{\partial t} + \frac{\partial \overline{uh}}{\partial x} + \frac{\partial \overline{vh}}{\partial x} + \frac{\partial \bar{\omega}\bar{h}}{\partial p} + g\frac{\partial \bar{R}}{\partial p}\right) dp + LH + SH \tag{3.5}$$

LH, SH are latent and sensible heat flux at the earth's surface ($\simeq$ 1000 mb). By specifying them on the basis of semi-observed climatological data (Budyko), eqs. (3.4), (3.5) represent two independent budgets for the unknown $\overline{\omega'h'}|_{500\ mb}$. An estimate of the error can be gained in the following way (Fig. 2):

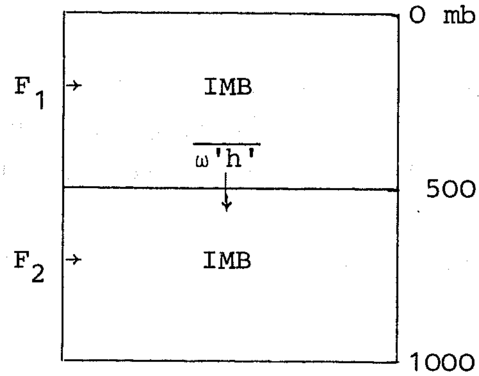

Fig. 2
Schematic sketch for over-specification of atmospheric energy flux data

a) Balanced budgets

Let F_1, F_2 denote the right-hand sides of eqs. (3.4), (3.5). In the steady state perfect data would yield $F_1+F_2=0$. $\overline{\omega'h'}$ describes the vertical exchange. Thus:

$$\begin{aligned} F_1 - \overline{\omega'h'} &= 0 \\ F_2 + \overline{\omega'h'} &= 0 \end{aligned} \qquad (3.6)$$

b) Non-balanced budgets

With actual data $F_1+F_2 \neq 0$. We assume that the imbalance is equally distributed in the vertical. Thus:

$$\begin{aligned} F_1 - \overline{\omega'h'} + IMB &= 0 \\ F_2 + \overline{\omega'h'} + IMB &= 0 \end{aligned} \qquad (3.7)$$

c) An example

Assume $F_1=9$, $F_2=-11$; the units are arbitrary. In this case (3.7) yields:

$$\begin{aligned} \overline{\omega'h'} &= (F_1-F_2)/2 = 10 \\ IMB &= -(F_1+F_2)/2 = \ \ 1 \end{aligned} \qquad (3.8)$$

We note the following points: First, the method of overspecification is applicable only if the emerging imbalance IMB is small compared to the residual quantity $\overline{\omega'h'}$ (high signal-to-noise ratio). Second, the example indicates that the results can be reasonable even if the a priori error is high. Third, the method can be generalized to an arbitrary number of boxes in the vertical and to a weighted vertical distribution of the anticipated imbalances. Fourth, the method works only in the present flux form but not in the non-integrated form (1.1) of the first law; the reason is that the budgets must be linked in some way.

4. ZONAL MEAN HEAT BUDGETS WITH ERROR ESTIMATE

The method just described has been applied by Hantel (1976) to the budgets of moist static energy in the form of eq. (3.3) after zonal averaging and vertical integration for the seasons of the northern hemisphere (Fig. 3). The main outcome of the study was that $\overline{\omega'h'}$ is

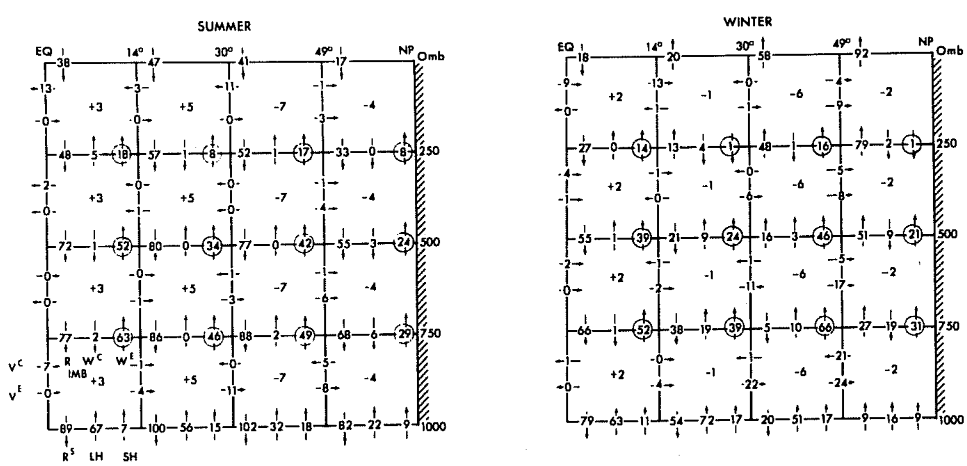

Fig. 3 Box budgets of h for solstice seasons, in units of 10^{14} W. To convert heat transports into fluxes divide by quarter of hemisphere's area ($.64 \times 10^{14}$ m^2). R: Radiation; V^C, W^C: meridional and vertical cell transports; V^E: meridional (synoptic) eddy transports; W^E (encircled): vertical (largely subsynoptic) eddy transports; center numbers: box imbalance.

upward in practically all cases. Further, this quantity is large and opposite to $\overline{R}$; this reproduces radiative-convective equilibrium of the atmosphere. The imbalance is of the order 5 units while $\overline{\omega' h'}$ (represented by W^E in Fig. 3; order of magnitude 50 units) is significantly above this noise level and thus deserves some confidence. Hantel obtained an rms-error of about 7 W m^{-2} for the box imbalances which probably is an underestimate.

5. BALANCING THE MASS FLUX FIELD

Any residual compilation of the heat flux divergence is hampered by the high level of zero energy. Even small errors in the mass flux field may cause large errors in the heat flux divergence. We shall demonstrate this with the following simple scale argument.

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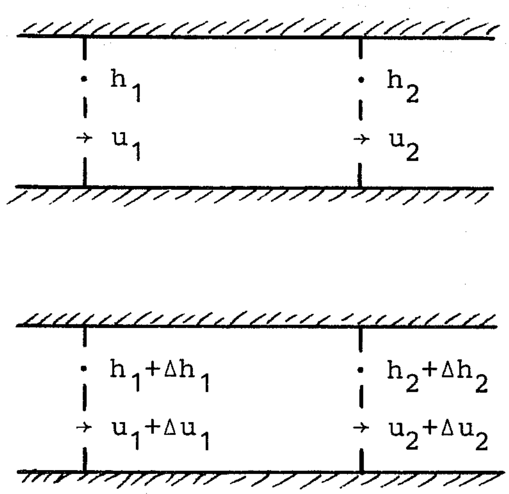

Fig.4
Schematic sketch for estimating error of heat flux divergence

Consider the steady horizontal transport across vertical box boundaries in the atmosphere (Fig. 4). h denotes moist static energy, u the horizontal wind.

a) Perfect data

For perfect data we would have nondivergent mass flux and thus:

$$u_2 - u_1 = 0 \qquad (5.1)$$

$$h_2 u_2 - h_1 u_1 = \Delta h\ u_2 = \Delta h\ u_1 \equiv \mathrm{DIV} \qquad (5.2)$$

where DIV denotes the perfect heat flux divergence.

b) Observed data

When the data are not perfect there is erroneous mass flux divergence:

$$(u_2+\Delta u_2) - (u_1+\Delta u_1) = \Delta u_2 - \Delta u_1 \equiv \Delta u \qquad (5.3)$$

The heat flux divergence, neglecting quadratic terms, is:

$$\left.\begin{aligned} &(h_2+\Delta h_2)(u_2+\Delta u_2) - (h_1+\Delta h_1)(u_1+\Delta u_1) \\ &\simeq \underbrace{\mathrm{DIV}}_{\simeq 300} + \underbrace{(h_2\Delta u_2-h_1\Delta u_1)}_{\substack{\simeq h\Delta u \\ \simeq 3000}} + \underbrace{(u_2\Delta h_2-u_1\Delta h_1)}_{\substack{\simeq u\Delta h \\ \simeq 300}} \quad \left(\frac{J}{g}\,\frac{m}{s}\right) \end{aligned}\right\} \qquad (5.4)$$

c) An example

Since the Δh, Δu are unknown we employ some speculation. Experience shows that $\Delta u \simeq u \simeq 10$ m/s while $\Delta h \simeq 0.1\ h \simeq 30$ J/g. The error caused by the erroneous mass flux divergence can thus be an order and more larger than DIV itself. Note that the present argument applies only to the cell flux components of the heat budget. The eddy heat flux components are not influenced by this effect.

We can eliminate the mass divergence impact upon the heat budget by correcting the mass flux for zero divergence. This corresponds to eliminating the term $h\Delta u$ which is the biggest in the budget (5.4). Various methods have been developed for this purpose. One simple but nontrivial and quite effective method has been recently designed and tested by Hacker (1980).

The method is sketched in Fig. 5. Consider a closed loop of 4 adjacent boxes with mass fluxes across the boundaries.

a) Perfect data

In this case we have:

$$\left.\begin{matrix} B-A=0 & D-C=0 \\ C-B=0 & A-D=0 \end{matrix}\right\} \qquad (5.5)$$

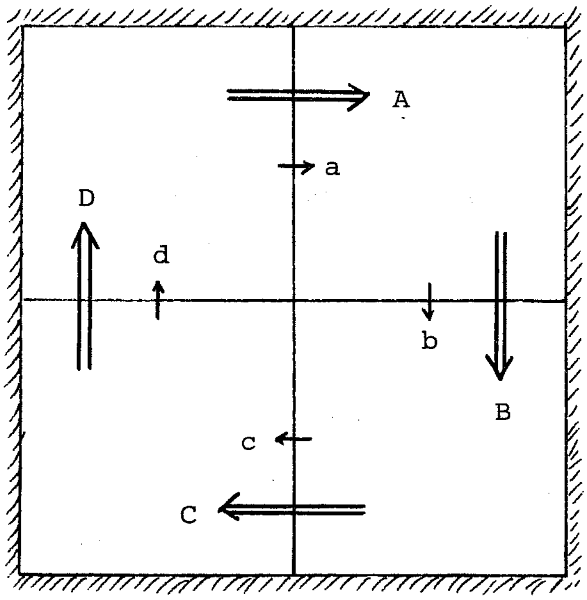

Fig. 5
Schematic sketch for objective mass flux correction.
A, B, C, D:
observed mass fluxes.
a, b, c, d:
small corrections.

b) Observed data

Observed data contain small unknown errors. This violates the budgets (5.5). Upon introducing corrections a, b, c, d we try to fulfill the budgets again. This transforms (5.5) into:

$$\left.\begin{aligned} (B+b) - (A+a) &= 0 \\ (C+c) - (B+b) &= 0 \\ (D+d) - (C+c) &= 0 \\ (A+a) - (D+d) &= 0 \end{aligned}\right\} \qquad (5.6)$$

However, the corrections cannot be calculated from (5.6) because the pertinent determinant vanishes.

c) Minimum correction method

Hacker now introduces the simple further condition that:

$$\frac{1}{2}\,(a^2+b^2+c^2+d^2) = \text{Minimum} \qquad (5.7)$$

This yields the specific equation:

$$a+b+c+d = 0 \qquad (5.8)$$

which, together with 3 equations chosen from the system (5.6), gives a unique solution for the four corrections.

We may note that (5.8) is equivalent to postulating that the corrections do not introduce additional circulation into the loop. We thus may call (5.8) a *zero vorticity condition*.

Hacker has shown that his method can be generalized to an arbitrary array of boxes in 3 dimensions. He has specifically applied the method to an array of 192 mass boxes (4 latitude strips, 12 longitudinal sectors, 4 pressure levels). An example is given in Fig. 6. The rms-mass flux divergence calculated from Fig. 6 over all boxes of the Northern Hemisphere for each season is of the order 100×10^{-5} kg $m^{-2} s^{-1}$ which corresponds to some 10^{-7} s^{-1}. This is spurious indeed and demonstrates the high quality of Oort's general circulation statistics. It nevertheless causes errors in the box heat flux divergence of the order 300 W m^{-2} which is not spurious but swamps the heat budget.

It follows that the corrections necessary to bring the mass flux into balance are small. As Tab. 1 indicates they are between 0.31 - 0.99 m s^{-1} and smaller than the principal interpolation errors estimated by Oort (1978). The impact upon the heat budget is shown in Tab. 2. The mass correction reduces the heat imbalance by a factor of about 15.

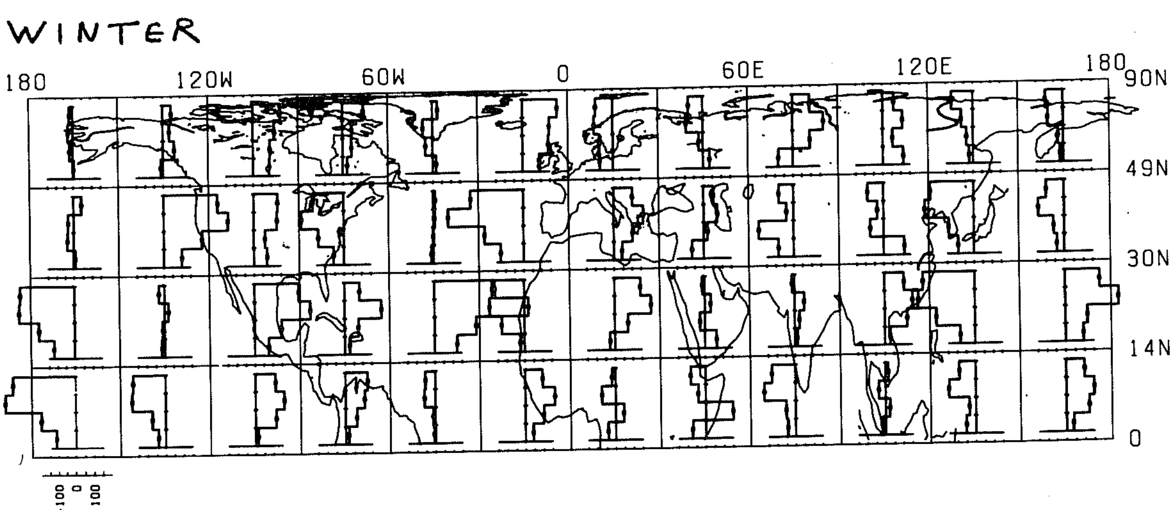

Fig. 6 3 D-mass flux divergence in atmospheric boxes. Histograms in each region show vertical profiles of box-integrated $\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}+\frac{\partial \omega}{\partial p}$ which should, but does not, vanish. Layers: 0-250, 250-500, 500-750, 750-1000 mb. u, v, ω fields from Oort and Rasmusson (1971). Unit 10^{-5} kg m^{-2} s^{-1}.

Table 1 rms-corrections for mass flux field.
Italicized numbers: rms-errors due to inhomogeneities of radiosonde data according to Oort (1978).

	Winter	Spring	Summer	Fall
$\bar{u}$ (m s^{-1})	0.99 (*3.25*)	0.72	0.38 (*3.87*)	0.53
$\bar{v}$ (m s^{-1})	0.37 (*1.39*)	0.38	0.31 (*1.37*)	0.43
$\bar{\omega}$ (10^{-2}Pa s^{-1})	0.23	0.19	0.12	0.16

Table 2 rms-imbalance for h-budget with/without correction of mass budget in W m^{-2}; average over 192 equal-mass boxes of Northern Hemisphere.

	without correction	with correction
Winter	406	26
Spring	357	25
Summer	262	17
Fall	376	19

6. NORTHERN HEMISPHERE HEAT BUDGETS WITH ERROR ESTIMATE

Since the mass flux correction is usually quite small the corrected heat transport components are not much different from the observations. This is demonstrated in Fig. 7 for one specific case in the vertical mean (no $\overline{\omega' h'}$ components shown). The horizontal transports of $c_p T$, gz, Lq are due to cell plus eddy fluxes according to the convention of eq. (3.3).

Fig. 7 is part of a complete Northern Hemisphere h-budget of Hacker (1980) with main emphasis on the vertical profiles of $\overline{\omega' h'}$. The result for the solstice seasons is shown in Fig. 8. We shall not dwell on too many details but limit ourselves to the most prominent features.

a) $\overline{\omega' h'}$ is in most geographical regions and in most atmospheric levels upward, consistent with the zonal mean (Fig. 3).

b) Its order of magnitude is 100-300 W m^{-2} which indicates the significance of this heat transport component.

c) The error is equal in the vertical, due to the method employed. The error is different in the horizontal from one geographical region to another. It varies between zero to more than 100 W m^{-2} with an average of less than 26 W m^{-2} (Tab. 2).

d) The error is highest in regions with high mass flux correction. This is the case, for example, in the region off the coast of Spain (compare Figs. 6, 8).

e) $\overline{\omega' h'}$ is high in the ITCZ, notably in summer, and very high over the midlatitude western oceans in winter (Gulf Stream, Kuroshio).

f) $\overline{\omega' h'}$ is small when convection is suppressed, particularly over the continental subtropics in winter. Note the difference over India between winter and summer.

g) $\overline{\omega' h'}$ is positive (downward) in higher levels off the west coast of North America (particularly in winter), seemingly due to subsidence. It is also positive over Northeast Siberia in both seasons.

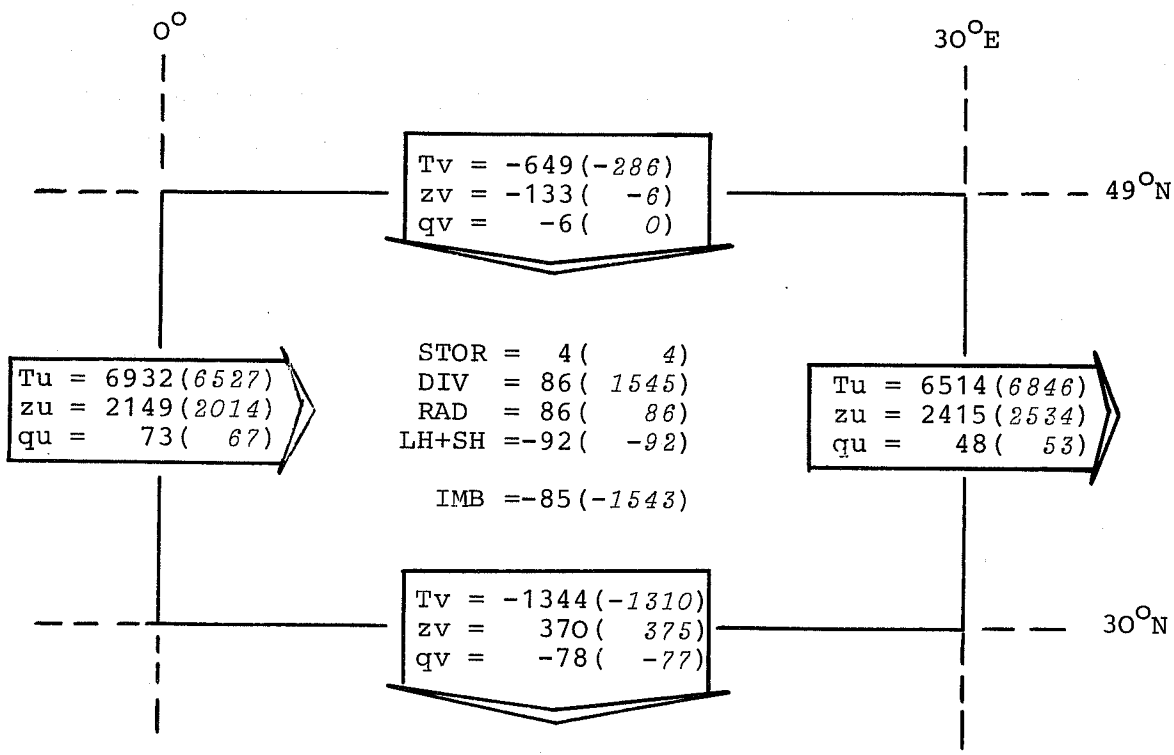

Fig. 7 Box heat budget over the Mediterranean in summer.
Vertical integral of h-budget ($h \equiv c_p T+gz+Lq$) between 0°-30°E, 30°-49°N. Numbers in W m^{-2}. Fluxes positive downward, eastward and northward. Italicized numbers: mass budget spoiled with spurious divergence.

Tu, Tv: horizontal transports of sensible heat
zu, zv: horizontal transports of geopotential
qu, qv: horizontal transports of latent heat
STOR: storage of h
DIV: divergence of horizontal flux of h (data from Oort)
RAD_{TOP}: 57 W m^{-2} (data from Winston)
RAD_{BOTTOM}: 143 W m^{-2} (data from Oort)
LH = - 43 W m^{-2} (data from Budyko)
SH = - 49 W m^{-2} (data from Budyko)
STOR+DIV+RAD+LH+SH+IMB = 0

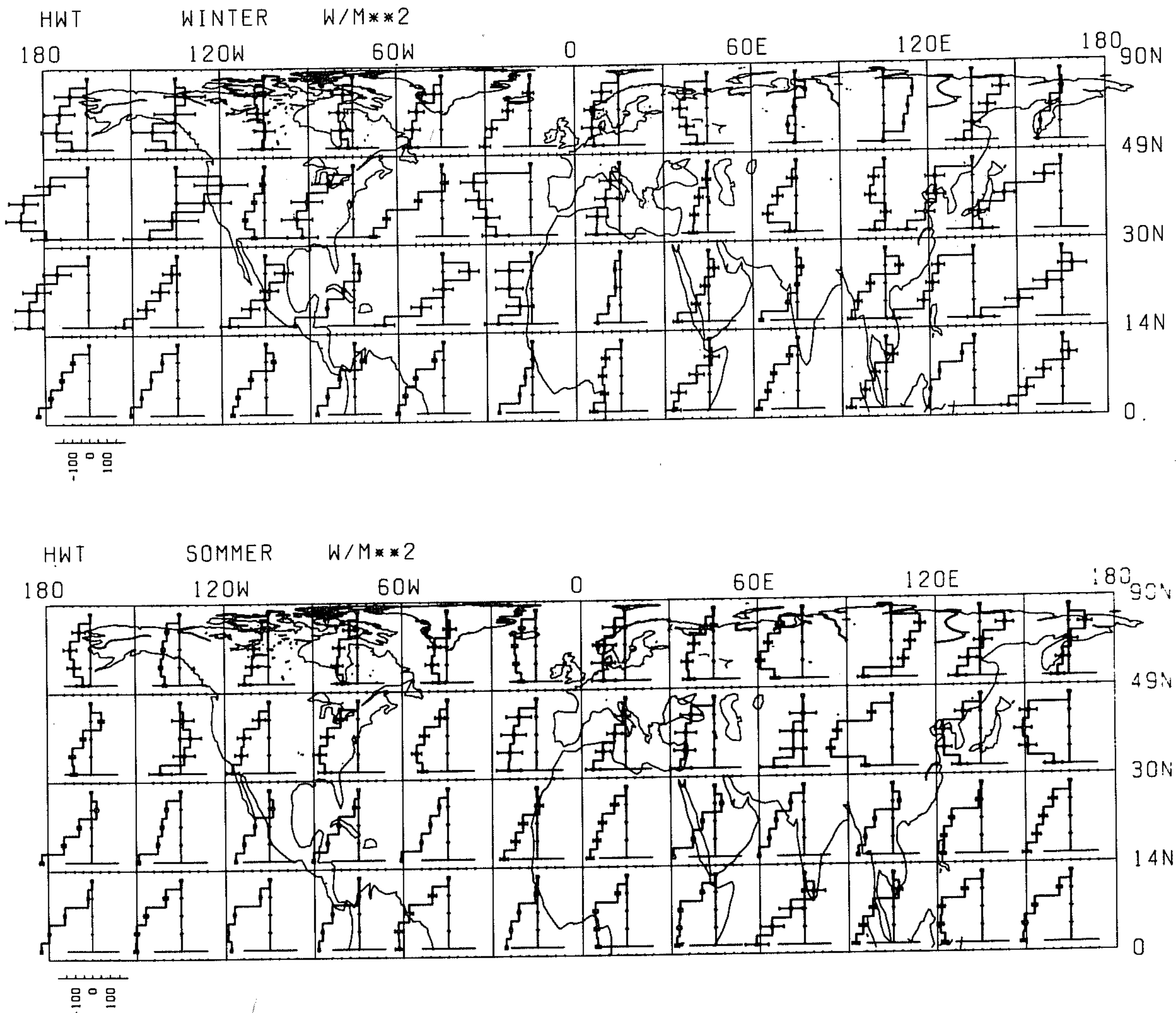

Fig. 8 $\frac{1}{g}\,\overline{\omega' h'}$ due to eddies of all scales according to Hacker (1980).

$h \equiv c_p T + gz + Lq$ = moist static energy

Key for each of the 4x12 profiles:

Abscissa: flux in units W m^{-2}, one tick = 50 W m^{-2}, negative upward.

Ordinate: pressure levels: 0 kPa (flux zero), 25 kPa, 50 kPa, 75 kPa, 100 kPa (flux = LH + SH, after Budyko, 1963).

Error bars: equal to IMB following from overspecified input data.

7. OUTLOOK

The results presented are consistent with general climatological experience. The present method seems to yield objective estimates for $\overline{\omega'h'}$ which is an important component of the heat budget in flux form and of diabatic heating.

Similar estimates for the other components of the atmospheric heat budget, notably precipitation flux and vertical eddy moisture flux, are in preparation. Also in preparation are regional estimates for the ALPEX area, similar to recent work of Behr and Speth (1977) for the Baltic Sea region. These should, on the basis of the FGGE data, be carried out for typical synoptic periods.

One shortcoming of the present box budgets should not be overlooked: It is not possible to split $\overline{\omega'h'}$ into contributions from different scales. In order to do that one would need more refined methods than the present crude budgeting approach.

The presented diagnostic method has been dictated by the need to obtain estimates of diabatic heating and its flux components as close to observations as possible. No resort to modelling or parameterization has been made. We conclude that the indirect budgeting techniques described here may continue to yield reasonable quasi-observed fluxes until direct area-representative measurements of subsynoptic fluxes (rain, $\overline{\omega'T'}$, $\overline{\omega'q'}$) will become available.

8. ACKNOWLEDGEMENTS

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