On evaporation from the Indian Ocean

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ABSTRACT. The problem of evaporation from the oceans is briefly reviewed. Seasonal charts of evaporation are presented for the Indian Ocean area. Combining these charts with the precipitation charts of Jacobs, the main sources of moisture-supply to the atmosphere in this area are located. From a knowledge of Bowen's ratio, computations are made of the relative magnitudes of the sensible heat (enthalpy) and the total convective heat-energy lost by the sea-surface to the atmosphere. These computations confirm that the main mode of energy-transfer in the region of the trades is in the form of latent heat. A comparative discussion is given of the available estimates of the excess of evaporation over precipitation between the latitudes 40°N and 40°S. The question of how far quantitative determinations of the mass-circulation of the trade-wind cell and the toroidal and eddy-fluxes of water-vapour across its high-latitude boundaries can provide an independent aerological check on such estimates is examined.

1. Introduction

Water vapour, because of the energy of latent heat which it can supply to its environment when it condenses, plays a prominent part in the initiation and development of weather systems that range in size from a cumulus cloud to a cyclonic storm. Most of this water vapour comes to the atmosphere from the oceans. A study of evaporation from the oceans is thus of importance for a proper appraisal of atmospheric thermodynamics. Since in turn oceanic circulation is largely influenced by the magnitude and distribution of the energy transferred between sea and air, such a study is of equal significance in oceanography.

As early as 1686, Halley set out to determine "the quantity of vapour raised out of the sea by the warmth of the Sun". The first reasonable answer to Halley's problem appears to have been given by Brückner (1908). He extrapolated values from available observations on sea coasts and estimated that an average thickness of 106 cm was evaporated in a year from all the oceans. In comparison a figure of 140 cm quoted by Lutgens (1911) shortly after was excessively high.

In the years between 1890-1904, an extensive series of direct measurements of evaporation were made from pans placed on board of German expedition vessels. Discussion of these measurements by Schmidt (1916), Wust (1920, 1922) and Cherrubim (1931) conclusively proved that the evaporation from a pan was not representative of sea surface evaporation and that the exact magnitude of the pan co-efficient was uncertain.

The emphasis has therefore been, of late, to compute evaporation by indirect means. Two methods are generally adopted— (1) Computation from considerations of energy-balance, (2) Computation from mass-transfer equations derived from aerodynamical arguments.

In the energy-balance method, the rate at which energy is absorbed by the ocean is calculated as the difference between the energy of radiation (Q_s) received at the sea surface from the sun and sky, and that reflected by the sea surface back into space (Q_r). It is assumed that over a long period of time such as a year and for an enclosed body of the ocean the net amount of *stored* or advected energy is negligible and that the energy effectively absorbed is utilised either in evaporation or in direct conduction of sensible heat (enthalpy) to the atmosphere. If we represent the rates of loss by these two processes as Q_e and Q_h respectively, this assumption implies

$$
\begin{aligned}
Q_s - Q_r &= Q_e + Q_h \\
&= Q_e\,(1+R) \\
&= L.E\,(1+R)
\end{aligned}
$$

$$\text{or } E = \frac{Q_s - Q_r}{L(1+R)} \tag{1}$$

where E is the rate of evaporation and L the latent heat of vaporisation at the mean temperature of the sea. R, which has been written for Q_h/Q_e is known as Bowen's ratio. We shall subsequently refer to an expression derived for it by Bowen (1926); its order of magnitude was first given by Angström (1920) as 0·1.

Thus if we know average values of Q_s-Q_r for the different latitudinal belts, mean rates of evaporation can be calculated from equation (1). Various uncertainties are however involved in the computation of this quantity (Charnock 1951. Anderson 1952). Nevertheless the zonal distributions worked out by Schmidt (1915), Mosby (1936) and McEwen (1938) on independent but necessarily arbitrary assumptions come out to be fairly consistent though not satisfactorily so in equatorial regions. Accepting these values, Sverdrup (1951) has arrived at a figure of 99 ± 10 cm for the annually evaporated water from the region considered earlier by Bruckner. The corresponding estimates by Möller (1951) is 99·9 cm/year and by Riechel (1952) 95·5 cm/year.

As the simplified equation of energy-balance is derived by neglecting certain terms which is justifiable only on an extended scale of space and time, the method is clearly of limited utility for the determination of spatial or seasonal distribution of evaporation. For this purpose mass-transfer relationships which seek to express evaporative flux in terms of meteorological elements appear to hold more promise. Here, however, the converse question arises: Can such equations which give momentary and local fluxes be used to give time and space-averaged evaporation from climatic data? Pasquill (1949a) has shown how such an assumption leads to erroneous results over the land. In low latitudes over the sea where diurnal and seasonal temperature changes are small, the objection loses some of its force.

A mass-transfer equation developed by Sverdrup (1936) has been employed for computations of evaporation "along the Tusima warm current" by Miyazaki (1949) and from the Indo-Pacific region by Albrecht (1951). Miyazaki writes the equation as

$$E = \frac{0\cdot215\ (e_s-e_a)\ W}{8\cdot64 \log\left[\dfrac{a+z_0}{z_0}\right]^2+0\cdot16\ W}$$

Here E(mm/day) is the rate of evaporation and e_s (mm) the vapour pressure, corrected for salinity, at the sea-surface; e_a (mm) and W (cm sec^{-1}) are the vapour pressure and wind speed respectively at the observational level a cm above the sea surface. z_0(cm) is the 'roughness parameter' which represents a measure of the lack of smoothness of the boundary surface. In the case of the air-to-sea junction it is a complex function of wave height and wind speed.

A chief drawback to equations similar to those of Sverdrup arises from the restriction that in the absence of actual observations under normal oceanic conditions, various *ad hoc* assumptions have to be introduced about the dependence of the roughness parameter (z_0) on wind speed (W). Miyazaki and Albrecht adopted a formula due to Krummel* according to which z_0 increases *without limit* with wind speed. Rossby (1936) has however argued that at about a wind speed of 6-7 metres per second (Beaufort force 4) the sea surface changes from a hydrodynamically smooth to a rough surface and that above this speed the roughness parameter remains at a constant value of 0·6 cm. Munk (1947) has cited many independent physical processes taking place in the sea-to-air interface which seem to undergo an abrupt change at this wind speed and has identified it with the critical speed for instability required by the Helmholtz-Kelvin criterion. As against this, Marciano and Harbeck (1952) (see also Charnock 1951) have recently concluded from observations on Lake Hefner that a natural water surface is *always* rough with an average roughness

* According to this formula $z_0 = 0\cdot25$ cm when $W = 100$ cm sec^{-1} and $27\cdot0$ cm when $W = 2000$ cm sec^{-1}

parameter of about 0·6 cm under equilibrium conditions; they fail to find well-defined transitional velocity insisted on by Rossby and Munk.

The foregoing digression on the dependence of the roughness parameter on wind speed is relevant because the reliability of aerodynamical evaporation equations is linked closely with a satisfactory answer to this question. This has been clearly brought out by Sverdrup (1951) who has subjected to a critical comparison the various "models of evaporation" proposed by different workers (Miller 1937, Montgomery 1940, Sverdrup 1937, 1946, Norris 1948 and Bunker *et al* 1949). He has synthesized these models to yield the easily manipulable Daltonian equation

$$E = K\ (e_s - e_a)\ W_a \qquad (3)$$

where E is the evaporation in mm per day, e_s and e_a are the vapour pressures at the sea surface and in the air respectively in mb and W_a is the wind speed in m. sec^{-1} at the level of observation. K is designated by Sverdrup as the evaporation factor. It is related in a complicated way to the roughness parameter, wind speed etc. Its value differs especially at high wind speeds for different models due to difference of hypotheses about the roughness parameter and character of diffusivity imposed on these models*. This results in two extreme models giving evaporation values discrepant almost by a factor of four.

Such a wide divergence of results has led Jacobs (1942, 1943, 1949, 1950, 1951a, 1951b), in his regional study of evaporation over the N. Pacific and the N. Atlantic to discard all the theoretical models and arrive at a mean evaporation factor in a semi-empirical fashion. The following assumptions are implicit in Jacobs' work—

(1) Sverdrup's equation (3) can be written in terms of averaged quantities, *i.e.*,

$$\overline{E} = K\ \overline{(e_s - e_a)\ W_a} \qquad (4)$$

Due to limitation of the form in which climatic data are available further approximation may be forced as follows.

$$\overline{E} = K\ (\bar{e}_s - \bar{e}_a)\ \overline{W}_a \qquad (5)$$

(2) An average value for K can be calculated for a defined area of the ocean by stipulating that the energy-budget evaporation derived for this area from equation (1) can be substituted for E in (5).

(3) This value for K can be adopted for other adjacent areas and for all the seasons of the year.

Though the assumptions of Jacobs when worded as above are apparently indefensible, the gaps in our knowledge of the energy transformations in the inter-connected ocean-atmosphere system and the lack of accessibility of properly sorted climatic records probably justify his methods as the best that can be employed at present. The following study on the water-economics of the Indian Ocean which feeds a major portion of the Asiatic monsoon is accordingly carried out on lines similar to those of Jacobs. The evaporation factor derived here, however, is only two-thirds of the value adopted by Jacobs. This would suggest that Jacobs has over-estimated the supply of moisture from the equatorial zones of the oceans considered by him. We return to a discussion of this point in Section 2.

2. Sources of data and the method of computation of the evaporation factor, K

The only available source of relative humidity values over the sea suitable for evaporation computations is the "Atlas of Climatic Charts of the Oceans" (Macdonald 1938). This compilation contains average values of wet bulb depression $t_a - t_w$ for the following periods of the year—(1) December—January, (2) March—May, (3) June—August and (4) September—November. These periods roughly correspond to the cold weather, hot

* These models do not also take into account the increase of evaporation due to the increase in area of the sea surface under high wind conditions. According to Kazuhiko Terada (1954) this effect can almost double the evaporation rate in tropical storms

weather monsoon and post monsoon seasons of the Indian sub-continent. This compilation also contains averages of wind speed $\overline{W}_a$ for the same periods.

Charts of monthly mean isotherms of air temperature t_a and the sea surface temperature t_s have been published by the Marine Branch of the Meteorological Office, London, for the Indian Ocean region enclosed between the latitudes 30° N and 50°S. Average values of t_a and t_s were scaled from these charts for five-degree latitude-longitude "squares" (tessera) for the space and period specified by the availability of humidity data.

From the sea surface temperature t_s the vapour pressure e_s was obtained from the formula $e_s = 0{\cdot}98\ e_d$, where e_d is the saturated vapour pressure at t_s read out from Smithsonian Tables and 0·98 is a correction due to the presence of dissolved salt in sea water. The actual vapour pressure e_a at the air temperature t_a and the wet bulb depression $t_a - t_w$ was also read out from the same tables*.

The relevant details of the method by means of which the evaporation factor K has been determined for the region under study are summarised in Table 1. In this method an area of the ocean is selected for which the net advective transport of water by ocean currents is at a minimum. This is necessary to ensure that the heat carried away by these currents is negligible and the assumption in equation (1) that all the absorbed energy is utilised by the ocean either in evaporation or conduction is justified. We now equate an average rate of evaporation E' calculated from energy considerations and arrived at independently from ship measurements† to the rate in equation (5) so that we get

$$^{**}\ K = \frac{E'}{(\bar{e}_s - \bar{e}_a)\ \overline{W}_a} \tag{6}$$

The annual average of $(\bar{e}_s - \bar{e}_a)\ \overline{W}_a$ for the area is obtained from the climatological sources already quoted. K is then derived from equation (6). This K is used for calculation of evaporation for all the seasons.

The area defined by the latitudes 0° and 40°S and the longitudes 60° and 90°E was chosen for the Indian Ocean. This area is under the influence of the South Equatorial and the Eastern Counter currents and since both of these are mainly easterly drifts, the amount of energy taken by them across the latitude boundaries is on the average small. This statement may nevertheless need qualification for particular seasons since, for example, during the monsoons the advection of water from the S. hemisphere into N. latitudes can become quite important.

The values of the "radiation surplus" $(Q_s - Q_r)$ for the latitude belts 0—10°S, 10—20°S, 20—30°S and 30—40°S have been given by Schmidt as 218, 228, 229 and

*For a discussion of the errors involved in the determination of mean vapour pressures from average values of dry and wet bulb temperatures refer to Sumner and Tunnell (1949); also Brown (1953)

†As will be stated below, use has been made of Schmidt's latitudinal values of radiation surplus to calculate evaporation from energy considerations. Wust (1954) has recently pointed out that the evaporation figures given by Sverdrup using these values for the southern hemisphere are too large. We have here preferred to use the mean of the value obtained from energy balance and from ship measurements (Wust).

**It has been suggested by the referee that K could have been more appropriately derived from

$$\bar{E} = \frac{K}{4}\left\{\left[(\bar{e}_s - \bar{e}_a)\ \overline{W}_a\right]_1 + \left[(\bar{e}_s - \bar{e}_a)\ \overline{W}_a\right]_2 + \left[(\bar{e}_s - \bar{e}_a)\ \overline{W}_a\right]_3 + \left[(\bar{e}_s - \bar{e}_a)\ \overline{W}_a\right]_4\right\}$$

where $\bar{e}_s, \bar{e}_a, \overline{W}_a$ refer to seasona averages, and the subscripts 1, 2, 3, 4 refer to the four seasons. Since however, sea-surface and air-temperatures show little variation from season to season in the trade wind zone, much error is not involved in taking $(\bar{e}_s - \bar{e}_a)$ as a constant. The two expressions are then identical.

TABLE 1

Determination of the evaporation factor K for a selected area of the Indian Ocean

1. Latitude	0—40°S
2. Longitude	60—90°E
3. Schmidt's average value for $(Q_s—Q_r)$ cal/cm²/day	216·5
4. Evaporation from energy budget $E_1 = \frac{Q_s—Q_r}{L(1+R)}$ (mm/day)	3·71
5. Evaporation according to Wust's observations E_2 (mm/day)	3·38
6. Average E' (mm/day)	3·55
7. Computed $\overline{W}_a$ (m/sec)	6·7
8. Computed $(\bar{e}_s—\bar{e}_a)\overline{W}_a$	33·2
9. $K = \frac{E'}{(\bar{e}_s—\bar{e}_a)\overline{W}_a}$	0·107
10. Values of K (for the observed $\overline{W}_a$ of 6·7 m sec⁻¹) according to different theoretical models (*a*) Miller (1937) and Montgomery, (1940), (*b*) Bunker et al (1949), (*c*) Sverdrup (1936), (*d*) Sverdrup (1946), (*e*) Norris (1948)	(*a*) 0·054 (*b*) 0·107 (*c*) 0·154 (*d*) 0·208 (*e*) 0·231

191 cal cm^{-2} per day*. The mean of these has been taken as the average amount of energy that is totally available for evaporation and for direct heating of the atmosphere. It seems desirable to introduce a small correction to the values of the radiation surplus for the departure of the zonal amount of cloudiness from its average latitudinal distribution assumed by Schmidt. But this was not possible since Schmidt's exact figures of cloudiness were not easily available**.

Taking R in equation (1) to be equal to 0·1 and L at the average temperature of the sea as 584 cal gm^{-1}, the rate of evaporation works out to be 3·71 mm per day. Wust has given an observed value of 3·38 mm per day for the same area of the ocean. A value of 3·55 mm per day was therefore accepted as the average evaporation E'. The regional value of $(\bar{e}_s - \bar{e}_a)\overline{W}_a$ was obtained as 33·2 and the average wind speed was 6·7 m. sec^{-1}. Using these values we get $K = 0·107$ from equation (6).

The coincidence of this particular magnitude of the evaporation factor with that given by the theoretical model of Bunker (*see* Table 1) for the observed windspeed is doubtless fortuitous†. Jacobs, in his computations already mentioned, has adopted a mean value of 0·142†† for the N. Atlantic and the N. Pacific Oceans. Examining the meteorological observations in the Atlantic made on board the expedition vessel "Meteor" during the summer months, Sverdrup (1951) has, on the other hand, extracted a value of 0·080 as being more appropriate for the trade wind regions.

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* These values have been scaled from Fig. 1 of Sverdrup's (1951) article in the *Compendium of Meteorcl gy.*

** An examination of Brooks' (1930) data on the mean cloudiness over the earth, however, shows the error due to this to be negligible

† Bowden (1950) decided on an identical value for the Irish Sea

†† Jacobs (1951) obtained a value of 0·140 for a selected area lying between 20° and 25°N in the Pacific and 0·196 for a similar area in the Atlantic. He considers that while the unacceptably high value of the evaporation factor in the Atlantic might be the result of neglecting the pronounced inter-latitudinal heat transport by water currents on the western side of this ocean, the value for the Pacific is more reasonable as the effects of the current systems on its eastern and western sides tend to cancel each other

It is not to be discounted that the appropriate magnitude of the evaporation factor might change from region to region within the trades. As explained above, the value adopted for the present study has been derived for a section of the South Indian Ocean, a major portion of which lies within the trade wind zone. As such this value may not be strictly comparable with that of Jacobs; for available evidence (Flöhn 1950) justifies the expectation that noticeable differences between the two hemispheres might exist in the mean strengths of the trade wind cells, the average values of evaporation within their regimes and in the general pattern of (latent) heat transfer from low to high latitudes.

Kraus (1955) has drawn attention to the fact that over large areas of the trades the mean wind speed is marginal in the sense that it is very near the critical value of $6 \cdot 5 \text{ m sec}^{-1}$ at which the sea surface is generally supposed to change from a hydrodynamically smooth to a rough surface. This would mean that small changes in the mean speed might produce comparatively large changes in evaporation. It is possible to speculate that the secular changes in the evaporation-precipitation cycle in the tropics might be accounted for in this way.

3. Distribution of evaporation over the Indian Ocean

Fig. 1 abstracts the average seasonal and annual values of evaporation at different latitudes of the Indian Ocean. Figs. 3—7 illustrate its regional variation in greater details. A centre of maximum evaporation appears between the latitudes 15° and 20°S in all these figures. This centre is associated with the clear skies and dry subsiding air characteristic of the anti-cyclonic conditions which prevail in this region. Moreover, it appears to migrate with the subtropical high pressure field. Thus compared to its position in December—February it is shifted westward towards Madagascar Islands during March—May. The seasonal variation in this region of maximum evaporation is not very large. However, the supply of moisture is greater during June—August than during the other months. This is in agreement with the fact that whereas maximum evaporation occurs in summer over land it occurs in winter over the sea.

A second maximum of evaporation situated in the Arabian Sea off the coast of Africa can be seen in the charts for December—February and June—August. The former winter maximum is to be identified with the sub-tropical maximum finding its fullest development in similar latitudes over all the oceans. The maximum occurring during the monsoon months is however peculiar to the Indian Ocean and is attributable mainly to the high speeds of the overflowing air which has been partly depleted of its moisture content by precipitation during its journeys across the equatorial regions and undergone some mixing with dry continental air present at higher levels.

In all the seasons of the year, the lowest evaporation occurs in the equatorial zone or in the higher latitude regions of the S. Hemisphere. In the hot months of March—May, the Bay of Bengal and the Arabian Sea contribute some, though comparatively small, amount of water vapour to the atmosphere.

4. Variation of Bowen's ratio, R

It was supposed in equation (1) that R which is the ratio of Q_h given to the atmosphere from the ocean as sensible heat to Q_e which is given to it in the form of latent heat of vaporisation is nearly equal to $0 \cdot 1$. Bowen (1926) (*see* also Sverdrup *et al* 1942) has however obtained the following expression* for this ratio.

$$R = \frac{Q_h}{Q_e} = K \; \frac{t_s - t_a}{e_s - e_a} \cdot \frac{P}{1000} \qquad (7)$$

Here P is the sea surface pressure; t_s, t_a, e_s and e_a have the same significance as before

* This expression is derived on the assumption that the co-efficients of eddy transfer of heat and matter are identical. Contrary view points have been expressed by Pasquill (1949b); see also Frost (1953).

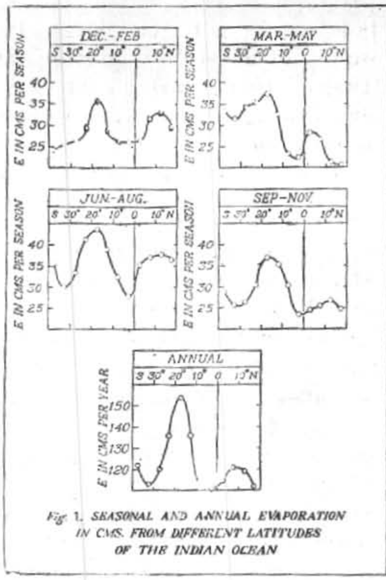

Fig. 1. SEASONAL AND ANNUAL EVAPORATION IN CMS. FROM DIFFERENT LATITUDES OF THE INDIAN OCEAN

Fig. 1

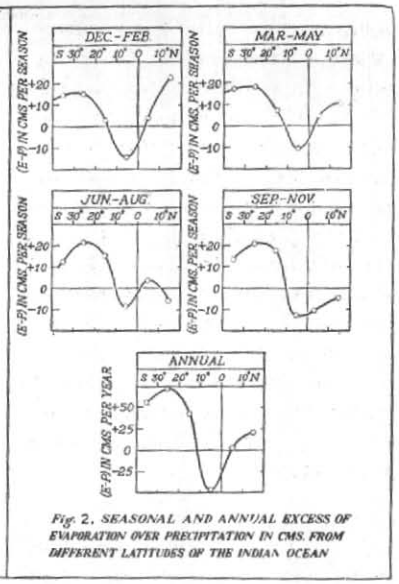

Fig. 2. SEASONAL AND ANNUAL EXCESS OF EVAPORATION OVER PRECIPITATION IN CMS. FROM DIFFERENT LATITUDES OF THE INDIAN OCEAN

Fig. 2

Fig. 3. Annual evaporation in cm over the Indian Ocean

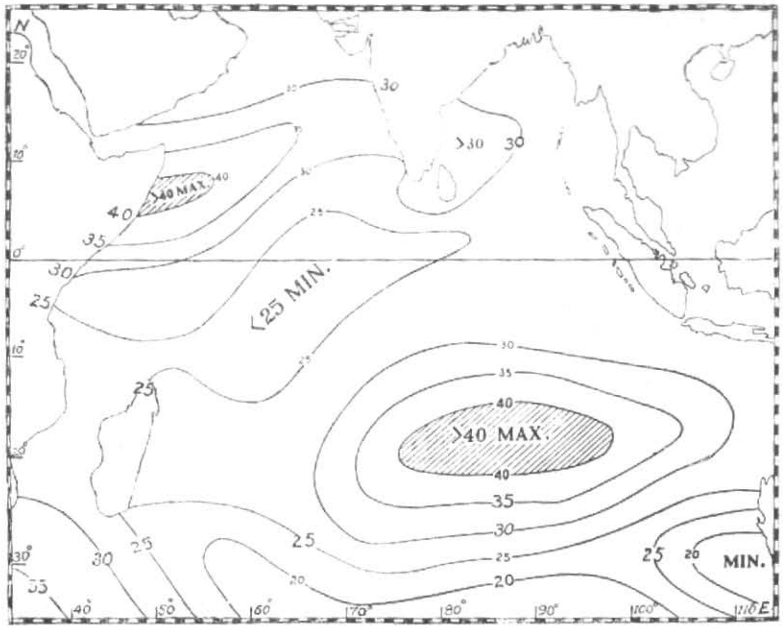

Fig. 4. Seasonal evaporation in cm over the Indian Ocean (December to February)

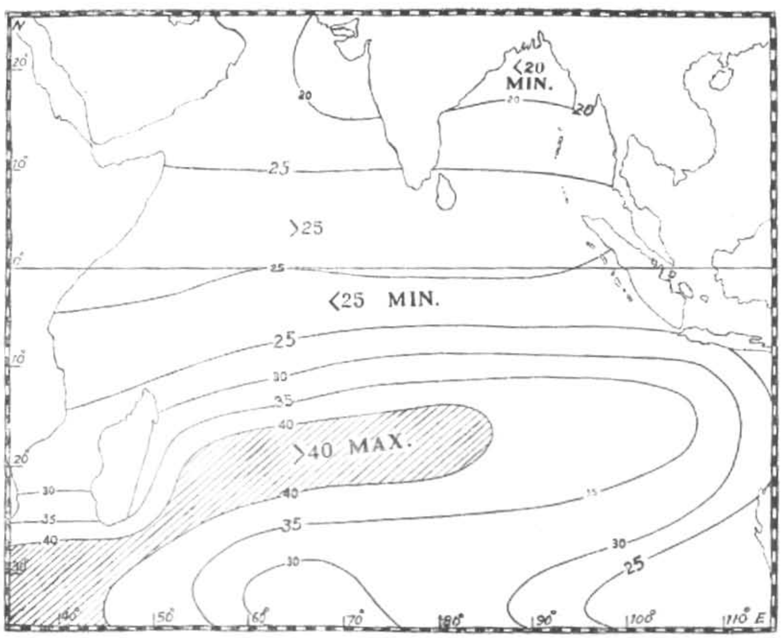

Fig. 5. Seasonal evaporation in cm over the Indian Ocean (March to May)

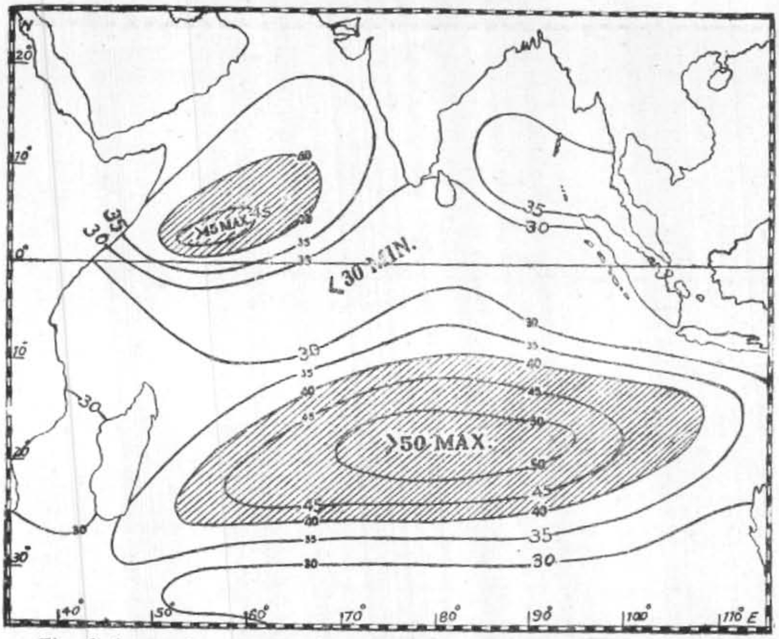

Fig. 6. Seasonal evaporation in cm over the Indian Ocean (June to August)

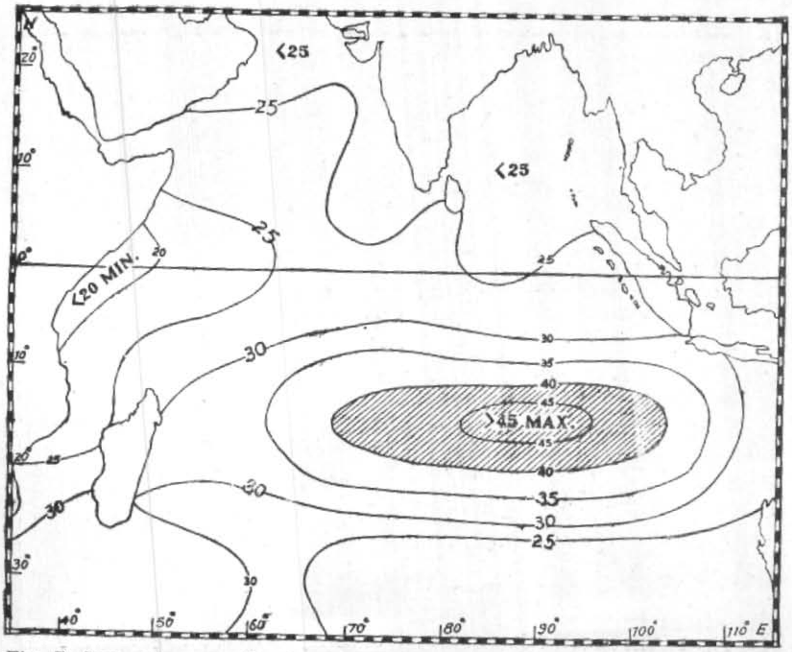

Fig. 7. Seasonal evaporation in cm over the Indian Ocean (September to November)

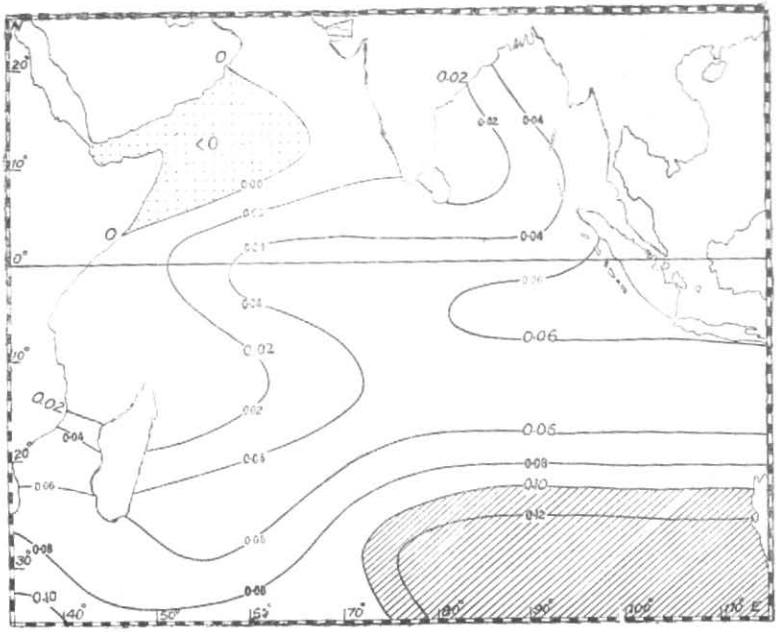

Fig. 8. Annual average of Bowen's ratio, R, over the Indian Ocean

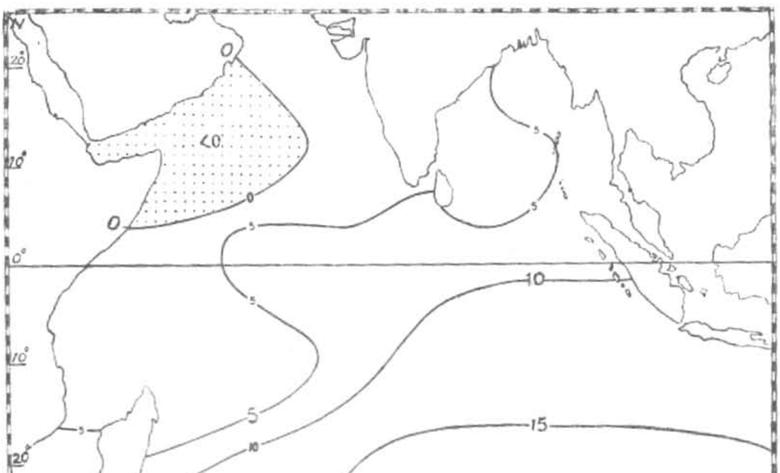

Fig. 9. The annual values of the rate of energy loss (cal cm^{-2} per day) by conduction (Q_h) over the Indian Ocean

TABLE 2

Seasonal variation of Bowen's ratio R over the Indian Ocean

Lat. range	Dec—Feb	Mar—May	Jun—Aug	Sep—Nov	Annual	Sverdrup*
20 — 10° N	0·03	0·00	—0·02	0·04	0·01	0·10
10 — 0° N	0·04	0·03	0·01	0·04	0·03	0·10
0 — 10° S	0·06	0·04	0·03	0·04	0·04	0·10
10 — 20° S	0·04	0·04	0·04	0·02	0·03	0·10
20 — 30° S	0·03	0·10	0·18	0·10	0·10	0·11
30 — 40° S	0·05	0·14	0·20	0·12	0·13	0·14

*Sverdrup's values for the N. Hemisphere are averages for the Atlantic and Pacific Oceans. His values for the S. Hemisphere are estimates

When temperatures are expressed in °C and the pressures in mb the constant K has a value of 0·67.

Fig. 8 shows the average annual distribution of Bowen's ratio for the Indian Ocean area calculated on the basis that pressure is 1000 mb everywhere over the sea. The seasonal averages for the various latitude belts are given in Table 2. It can be seen that the magnitude of this ratio in the equatorial zone is very small during all the seasons of the year. In the western portion of the Arabian Sea where air temperatures are high, there is even a tendency for it to be negative, which means that heat is conducted *from* the atmosphere *to* the ocean in this region. Though the seasonal variation is negligible the maximum value of R occurs during winter, *i.e.*, December—February in the N. Hemisphere and June—August in the S. Hemisphere.

Sverdrup (1951) has doubted whether the smallness of the value of R obtained for low latitudes is not due to errors made in the measurement of air temperature from within the heated body of the ship. His final estimates based partly on observations from the N. Atlantic and the N. Pacific have been included in Table 2 for comparison.

5. Amount of sensible heat Q_h and total convective energy $Q_a(=Q_e+Q_h)$ transferred from the Indian Ocean to the atmosphere

Knowing the distribution of evaporation E (and hence of Q_e) and the Bowen's ratio R, the amount of sensible heat Q_h conducted directly from the ocean to the atmosphere is computed from the relation $Q_h = RQ_e$ (Fig. 9). Since however the magnitude of R is very small, Q_h is small compared to Q_e . Thus whereas a loss of 280 cal cm^{-2} per day takes place from the sub-tropical regions of the S. Hemisphere, only 10 cal cm^{-2} per day are lost by direct conduction from the sea-surface*.

$Q_a = Q_e + Q_h$ represents the total loss of convective energy which, as far as the ocean is concerned, is entirely in the form of heat. Fig. 10 shows the annual distribution of Q_a over the Indian Ocean. Due to the smallness of the additive term Q_h the configuration of the isolines Q_e (or E) and Q_a are very similar.

$Q_{hp}=Q_h+Q_p$ gives the total amount of energy gained by the atmosphere by direct sea surface conduction (Q_h) and through the release of latest heat during precipitation (Q_p). Here again it can be seen from the relative magnitudes of Q_h and Q_p that the run of the Q_{hp} isopleths will be dominated more or less by the Q_p term. The general distribution of Q_{hp} over the oceans has been discussed at length by Jacobs (1951b).

The seasonal variation of Q_e , Q_h , Q_a , Q_p and Q_{hp} for the various latitude ranges of the Indian Ocean can be judged from Table 3. The values Q_e and Q_a have been tabulated for 5° ranges and those of Q_p (taken from Jacobs 1951a) and Q_{hp} for 10° ranges.

* Riehl (1954b) quotes Q_h to be approximately 5 per cent of Q_e in the trade wind zone

TABLE 3

Seasonal variation of Q_e, Q_h, Q_a, Q_p, and Q_{hp} over the Indian Ocean

Lat. range	Q_e	Q_h	Q_a	Q_p	Q_{hp}	Q_e	Q_h	Q_a	Q_p	Q_{hp}
	December—February					March—May				
20—15°N	194	6	200	60	66	134	0	134	69	69
15—10°	214	6	221			138	0	138		
10— 5°	207	8	215	164	171	172	5	177	149	154
5— 0°	172	7	179			180	5	186		
0— 5°S	171	10	181	266	276	144	7	151	219	226
5—10°	170	10	180			150	7	157		
10—15°	189	8	196	195	204	210	8	218	177	186
15—20°	232	9	242			239	10	249		
20—25°	190	6	196	82	87	225	23	248	106	129
25—30°	171	5	177			219	22	241		
30—35°	166	8	175	68	76	203	28	231	96	125
35—40°	158	8	166			213	30	243		
	June—August					September—November				
20—15° N	228	—5	223	274	269	159	7	165	192	199
15—10°	239	—5	235			172	7	179		
10— 5°	233	2	235	206	208	164	7	170	226	233
5— 0°	233	2	225			159	6	165		
0— 5°S	176	5	181	241	246	153	6	159	256	262
5—10°	202	6	208			195	8	202		
10—15°	245	10	255	163	173	227	5	231	120	125
15—20°	275	11	286			237	5	242		
20—25°	260	47	307	100	143	195	19	215	50	68
25—30°	215	39	253			168	17	185		
30—35°	192	38	231	132	174	166	20	186	91	112
35—40°	229	46	275			186	22	208		

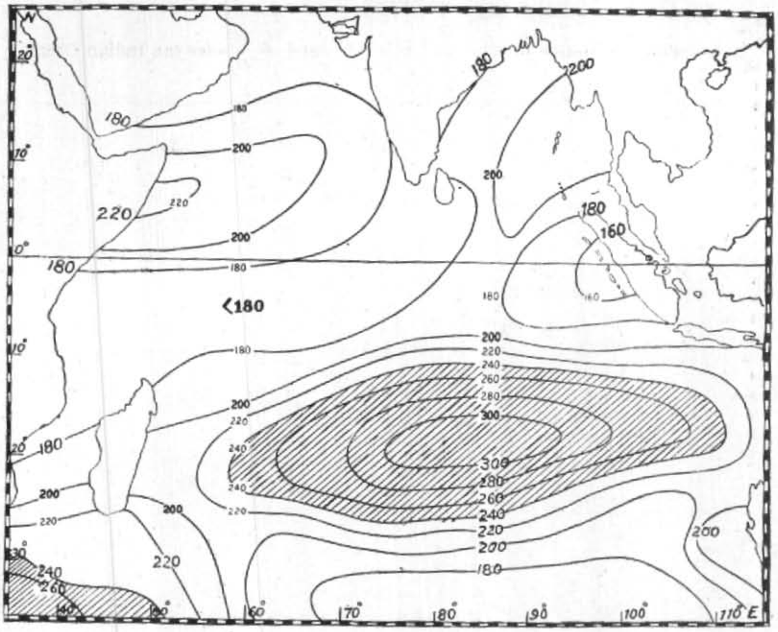

Fig. 10. The annual values of the total energy loss (cal cm^{-2} per day) by convection, $Q_a = Q_e + Q_h$, over the Indian Ocean

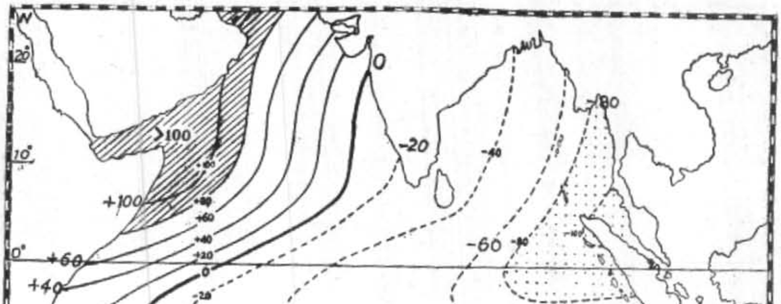

Fig. 11. Annual excess of evaporation over precipitation in cm over the Indian Ocean

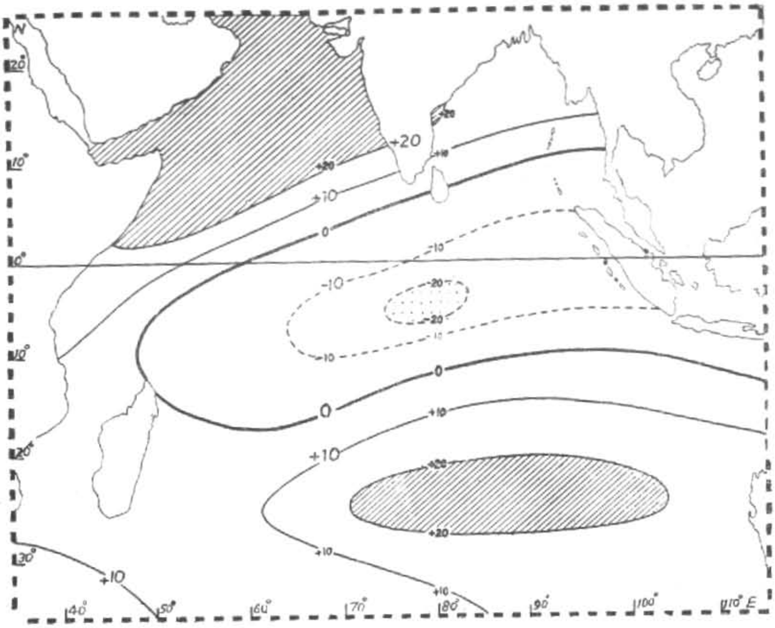

Fig. 12. Seasonal excess of evaporation over precipitation in cm over the Indian Ocean (December to February)

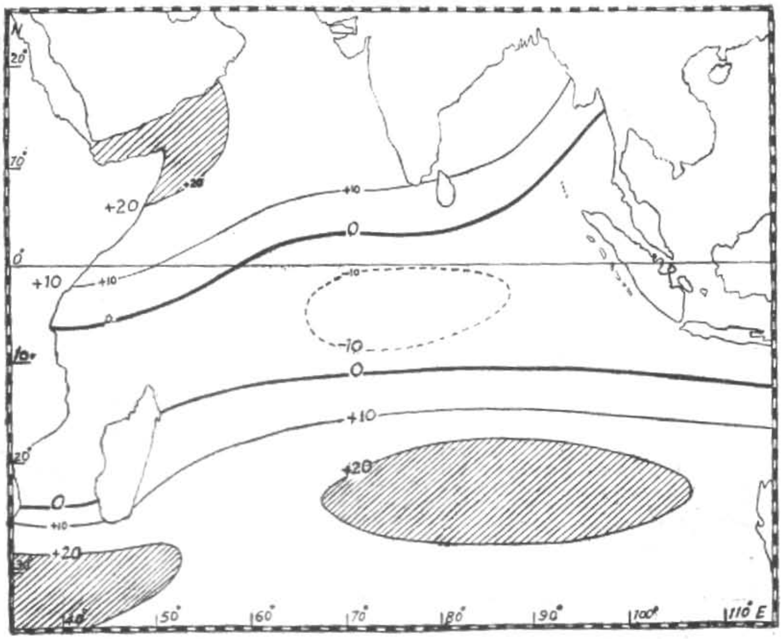

Fig. 13. Seasonal excess of evaporation over precipitation in cm over the Indian Ocean (March to May)

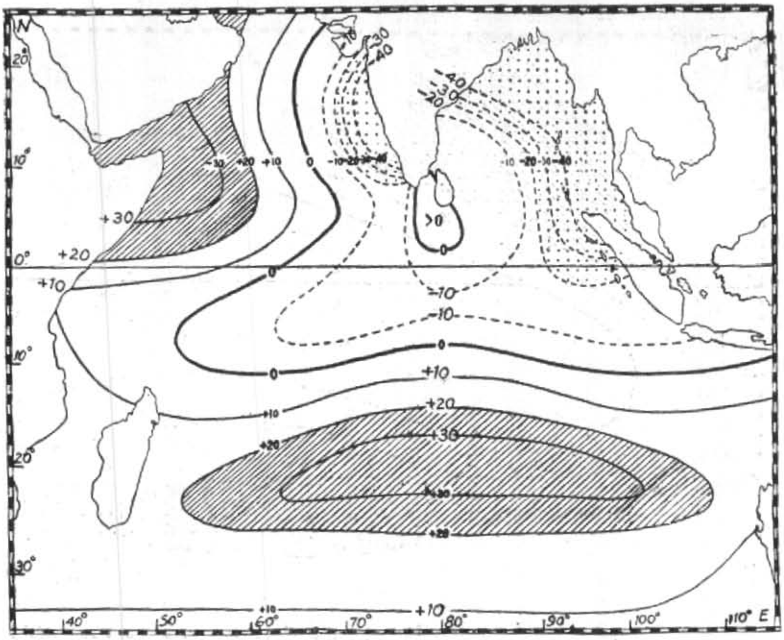

Fig. 14. Seasonal excess of evaporation over precipitation in cm over the Indian Ocean (June to August)

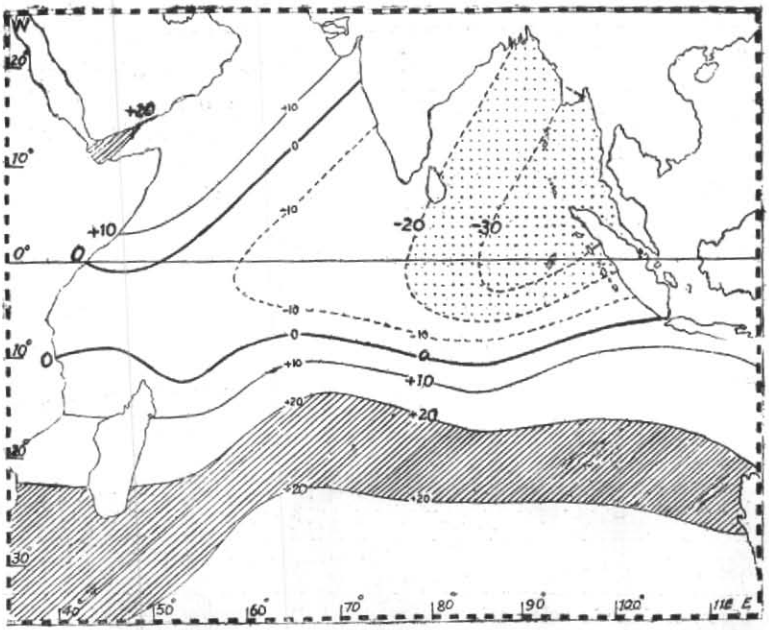

Fig. 15. Seasonal excess of evaporation over precipitation in cm over the Indian Ocean (September to November)

6. Distribution of the excess of evaporation over precipitation, (E—P), over the Indian Ocean

Comparative evaluation of the various estimates of precipitation over the oceans has been made by Jacobs (1951b) and Riechel (1952); (see also Riehl 1954). We have combined our evaporation charts with the precipitation charts of Jacobs and mapped out the excess of evaporation over precipitation in the Indian Ocean. The average latitudinal values of (E—P) thus obtained are given in Fig. 2 (p. 271) and the regional values in Figs. 11—15. These figures show that moisture for precipitation over land and over parts of the sea is mainly made available from two principal sources, namely, the sub-tropical areas in the S. Hemisphere and the western portion of the Arabian Sea, which are both regions of high evaporation and relatively small precipitation. High precipitation occurs in the west coast of Deccan and the Burma and Malayan coasts. Seasonal variation of precipitation and also of (E—P) is maximum in these regions.

7. The excess of evaporation over precipitation in the trade wind zone and its relation to the general circulation

Because of its usefulness as a "general circulation feature" for marking out the regions of sources and sinks of heat in the atmosphere, much attention has been devoted by geophysicists to the distribution of (E—P) over the oceans and over land. In Fig. 16 we have collected some of the reliable estimates of this quantity over the different oceans. The values of Wust (A) inserted in this figure have been obtained by him from "reduced" ship measurements of evaporation and from his own estimates of precipitation. For the N. Atlantic and the N. Pacific the "computed" values are those of Jacobs while for the Indian ocean they have been taken from Fig. 2*. The latitudinal means of these computed values weighted according to the areas occupied by the respective oceans have been marked in the curve for "all oceans". In this curve are also added (1) values worked out as differences between Sverdrup's estimates of evaporation from Schmidt's calculations of the radiation surplus, and Jacobs' estimates of precipitation and (2) Wust's (1954) values (Wust B) deduced from his recently modified empirical formulae which seek to relate (E—P) and the surface salinity of sea water S (expressed in parts per thousand by weight, of soluble matter) in a linear fashion. Wust's formulae are

$$E_n - P_n = 66 \cdot 7 \quad (S - 34 \cdot 47) \pm 7 \text{ cm/year for the N. Hemisphere} \tag{8}$$

and $$E_s - P_s = 80 \cdot 0 \, (S - 34 \cdot 92) \pm 8 \text{ cm/year for the S. Hemisphere} \tag{9}$$

Remembering that (E—P) is obtained as a difference of quantities neither of which can at present be directly measured, the agreement between the zonal averages for "all oceans" arrived at by independent methods must be considered as satisfactory at latitudes which are not very close to the equator. However in the important equatorial zone which is responsible for a major share of the atomospheric moisture supply† the disparity between the different estimates is seen to be admittedly large. The failure here of Wust's simple relationship between (E—P) and surface salinity is at once obvious. In drawing the mean curve for all oceans in Fig. 16 we have further adopted the view that Wust's (Wust A) results exaggerate the equatorial minimum of (E—P) in the N. Hemisphere and the "computed" values make a slight overestimate of the water evaporated from this region.

Coming to the individual oceans we find that the computed values disagree from Wust's (Wust A, *vide* Fig. 16) figures mainly for the N. Indian Ocean and for the subtropical zones of the N. Atlantic, while for the

* It has not been possible to include in this figure Albrecht's (1951) results for the Indo-Pacific region as no latitudinal values have been given by him

† According to Riehl (1954b) the belt between 25°N and 25°S furnish about 60 per cent of the global evaporation

TABLE 4

A draft estimate of the annual average of $(E—P)$ for the zone 0—30°N

Latitude	Average $E—P$ over the oceans (cm/year)	Estimated evaporation (cm/year)	Precipitation over land (cm/year)	$E—P$ over land (cm/year)	Average $E—P$ for land and ocean combined (cm/year)
0—10° N	−46·5	10·1	141·3	—131·2	—65·3
10—20° N	+49·5	11·3	82·9	—71·6	+30·3
20—30° N	+64·0	15·1	67·3	—52·2	+20·3

N. Pacific the mutual agreement is fairly close except near the equator. Both Wust's figures and the present computations indicate a residual annual excess of evaporation over precipitation for the N. Indian Ocean. This is discordant with Jacob's surmise that, for approximate hemispherical balance, oceanic precipitation must preponderate over evaporation in this area. From our charts it comes out that there is a well-marked maximum of evaporation off the African coast in winter as well as during the monsoons. The water intake into the atmosphere from this source more than compensates the large amounts of precipitation over the eastern portions of the Arabian Sea and the Bay of Bengal.

In any discussion of the water-economics of the entire region, however, we have to include in our reckoning the precipitation as well as the evaporation from the appropriate land areas. It is to be expected that particularly during summer a sizeable amount of water vapour is returned to the atmosphere by evaporation from land. This is supported by the charts of Albrecht (1951), according to which about 30 cm of water are evaporated from the Coromandal coast during the monsoon months June—August. Though no great accuracy can be claimed for the figures quoted above, it serves to underline that this source of moisture is not by any means to be neglected.

Another surprising feature brought out by Fig. 16 is that in spite of the abnormal northward transgressions of the equatorial trough during the monsoon months the mean annual minimum of $E—P$ occurs to the south of the equator in the case of the Indian Ocean whereas for the other oceans such a minimum is reached to the north of the equator.

We have given in Table 4 a 'draft' estimate of the annual average of $E—P$ for the zone 0—30°N, arrived at in the following way—Over the oceans values of $E—P$ for 10° latitude belts are taken from the mean curve given in Fig. 5. Over land it is assumed that evapotranspiration per unit area for the regions 0—10°, 10—20° and 20—30° are 100, 80 and 30 per cents of the corresponding values of evaporation given for the oceans, these percentage figures being put down in agreement with the tabulations of Wust reproduced in Hann-Suring's *Lehrbuch der Meteorologie* (p. 231). For precipitation over land we have used Meinardus' data provided in the same book on p. 468.

It can be argued from simple considerations that the excess of evaporation over precipitation within the trade wind zone 0—30°N calculated as above must tie up with the total poleward transport of water vapour across the northern boundary of this zone, if we discount on an annual basis any net interchange between the hemispheres. Thus according to Palmén (1954) the toroidal and eddy flux components of this transport add up to give $E—P$ as expressed by the relationship

$$m\,(q_1 - q_2) + W = E—P \qquad (10)$$

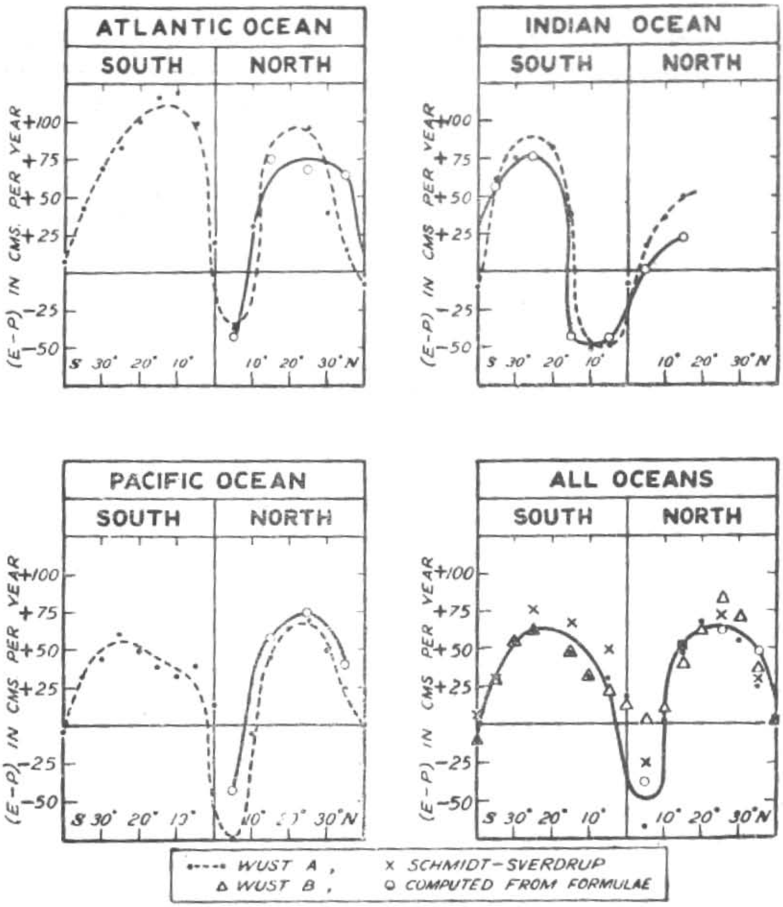

Fig. 16. Annual excess of evaporation over precipitation at different latitudes over the Oceans

where m is the mass circulation of the trade wind cell (assumed to extend upto 30°N), q_1 and q_2 are the average specific humidities of its northward and southward moving branches. W is the annual eddy-flux averaged over the latitude circle.

It is now possible to compute the left hand terms of equation (10) directly from aerological observations of wind and specific humidity. However, no reliable hemispherical estimates is as yet available (*see* Priestley 1951 and White 1951) due to the following reasons. Humidity measurements are not sufficiently accurate at higher levels. This by itself may not be very serious since it is known that the major part of the flux takes place in the lower troposphere below the 500-mb level. In this case its magnitude will be influenced strongly by topographical and local circulation factors. It would therefore become debatable as to how far averages drawn from scattered observation points can be considered as representative. Moreover, while it is recognised in general that $E—P$ over land and over the oceans will be of opposite sign in the region under consideration (except near the equator), only a vague assessment of the relation between their magnitudes can at present be given. This lack of knowledge arises no less from uncertainty of the magnitude of evaporation over land as from the unreliability of $E—P$ over the oceans. However equation (10) which expresses in an elegent manner the interlinkage between mass transport and energy in the circulation of the tropics collates the essential factors which a satisfactory solution of the problem will in future have to reconcile properly.

8. Acknowledgement

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