A Test -study on the Water Vapour Balance in the Atmosphere over the North Atlantic Ocean

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Abstract

In this study the data of 10 days of Jan. 1957 were analysed to examine the validity of water vapour balance in the atmosphere. It is demonstrated that precipitation over the North Atlantic Ocean, surrounded by the continents with the sufficient number of observation stations along their coastal line, can be estimated from the consideration of the energy balance among evaporation from the sea surface, transport of the atmospheric water vapour, and the precipitation. The estimated precipitation distribution was compared with the low cloud distribution which may be considered to be a fairly good indicator of precipitation. And also the relationship between precipitation and the mean vertical velocity on the 850 mb, 700 mb and 500 mb was examined. It is concluded that the precipitation has a close relationship with the low cloud amount, but a meaningful relation between precipitation and the mean vertical velocity was not detected.

1. Introduction

Benton and Estoque[1] (1954) studied the balance between evapotranspiration, transport of water vapour, precipitation, and discharges from rivers for the case of the atmosphere over the North American and Canadian continent and evaluated the evapotranspiration from the measurable quantities, that is, transport of water vapour, precipitation and discharges from the rivers.

Weather observations are sparse over the oceanic surfaces, and, moreover, the major source of data over the ocean are reports from the merchant vessels, which do not observe precipitation. This is one of the reasons why the study of the energy problem over the ocean has been less promoted than over the lands.

Wüst[2] (1936) firstly related the difference between evaporation and precipitation to the salinity, based on which he obtained precipitation by measuring evaporation and salinity. Later, Jacobs[3] (1951) constructed the chart of annual and seasonal precipitation with use of the results given by Wüst and the Soviet World Atlas. On the other hand, G. B. Tucker[4] 1961 recently drew the map of seasonal precipitation over the Atlantic ocean. In his determination of precipitation the correlation between the present weather and the precipitation which was obtained from the land observations was applied for the data of the present weather and precipitation reported by the weather ships over the Atlantic ocean.

Present author shows how it is possible to estimate precipitation from the view of the balance among evaporation from the surface, transport of the atmospheric water vapour and precipitation, over the ocean area bounded by the east coastal lines of America, the Cuban islands, the west part of Africa, and the southern Europe and

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2. Data and equations for computations

The data of lst through 10th Jan., 1957, in the Northern Hemisphere Data Tabulation Daily Bulletin issued from the U. S. Weather Bureau were used. The observation stations were listed in the Table 1 (1), and locations of the main stations and ships were shown in Fig. 1. For the analysis of the upper atmosphere, we used the coastal

Table 1. Observation stations and ships

(1) Observation Stations

(72)	202	Miami
	206	Jacksonville
	208	Charleston, S. C.
	304	Hatteras, N. C.
	405	Washington, D. C.
	506	Nantucket, Mass.
	606	Portland, Maine
	807	Argentina
(04)	018	Keflarik
	270	Narssarssuaq
	509	Lajes (Acores)
(78)	016	Kindley-Berm
	367	Guantanamo
	467	Sabana de la Mar
	526	San Juan
	897	Raizet, Guandeloupe
	089	Bone fish Bay
(03)	005	Lerwick
	026	Stornoway
	808	Cambornel
	917	Alderqrove
	953	Valentia
(07)	110	Brest
(08)	521	Funchal
	536	Lisboa/Portela
	594	Jal
(61)	401	Ft-trinqvet

(2) Weather Ships

02	B	O	565	510
03	C	O	525	355
04	D	O	440	410
05	E	O	350	480
041	K	O	450	160
520	J	O	523	201
061	I	O	589	192
063	A	O	620	330

(3) Merchant Ships

N	W
60—65	0—10
40—50	60—70
40—50	50—60
40—50	40—50
40—50	30—40
40—50	20—30
40—50	10—20
40—50	0—10
30—40	70—80
30—40	60—70
30—40	50—60
30—40	40—50
30—40	30—40
30—40	20—30
30—40	10—20
20—30	80—90
20—30	70—80
20—30	60—70
20—30	50—60
20—30	40—50
10—20	70—80
10—20	60—70
10—30	30—40
10—30	10—30
10—20	40—60

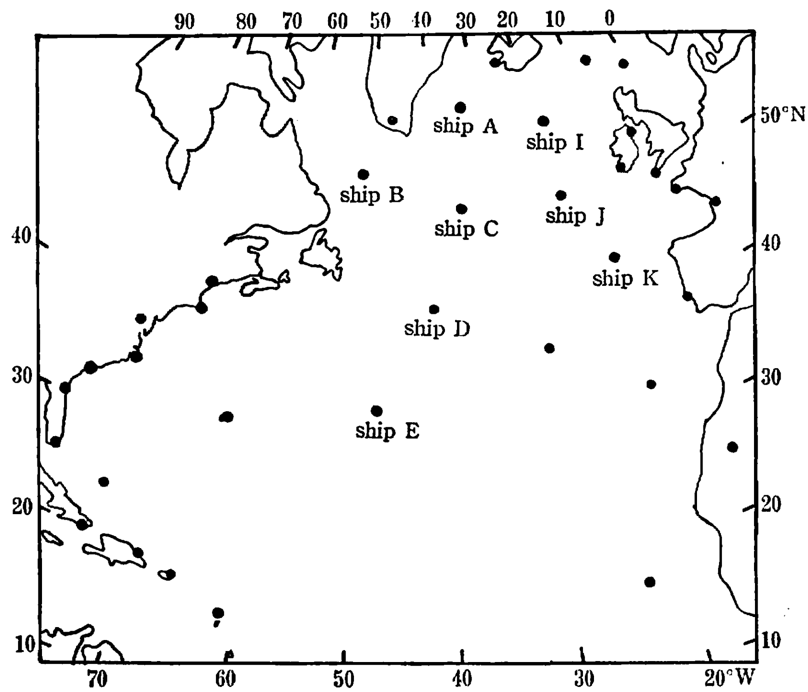

Fig. 1. Position of observation stations over and near the North Atlantic Ocean

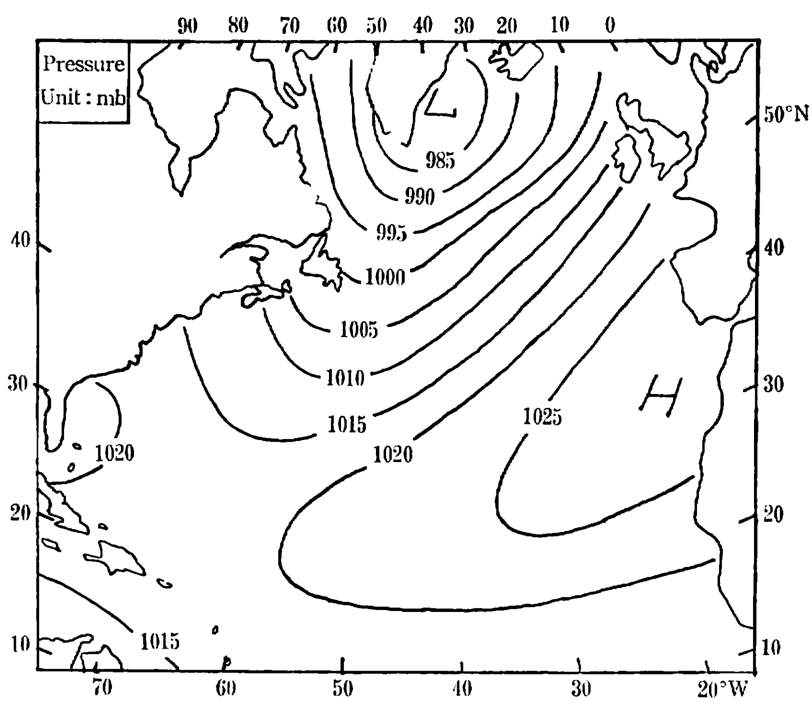

Fig. 2. Mean surface pressure distribution during the period 1st through 10th of January 1957

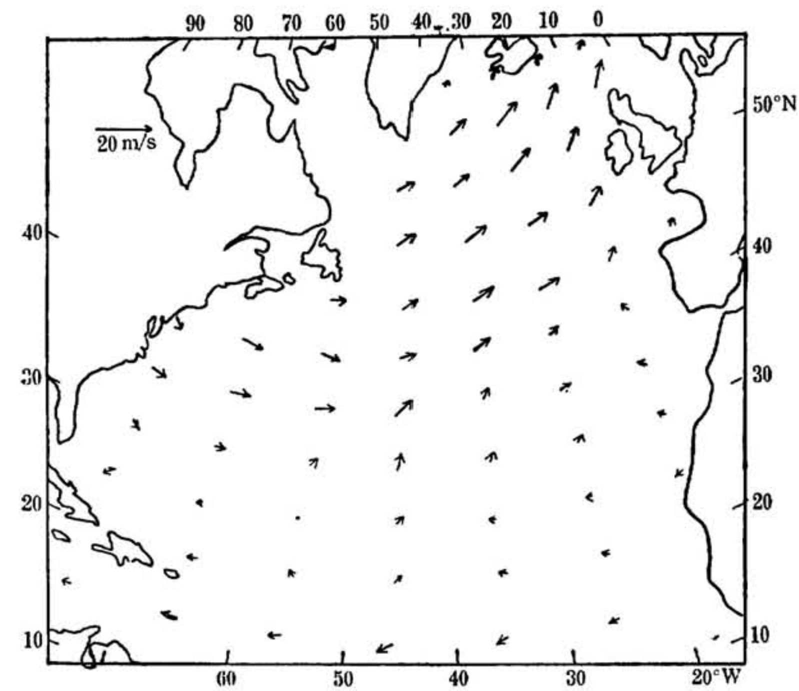

Fig. 3. Surface level wind vectors

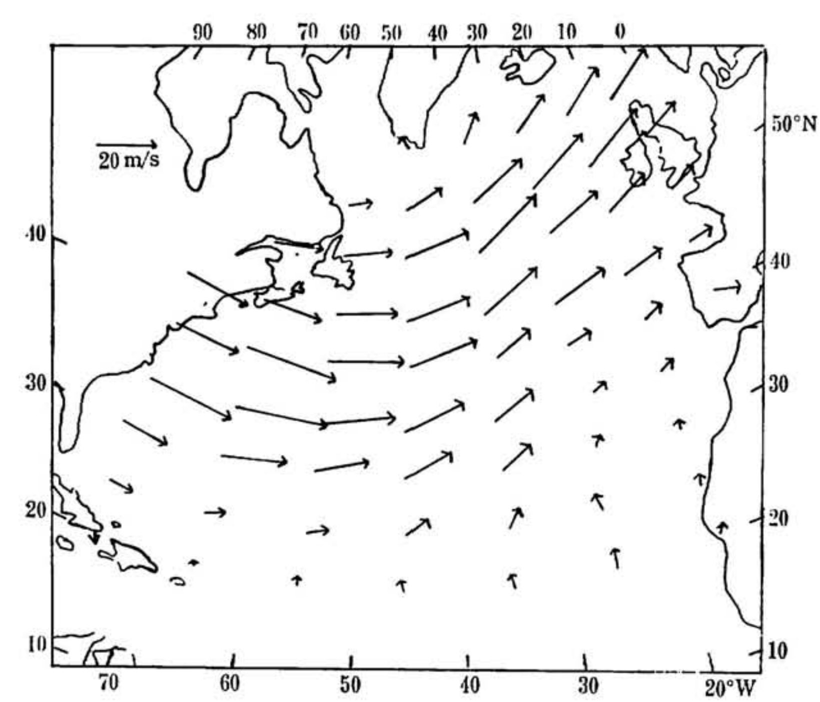

Fig. 4. 500 mb level wind vectors

land observations and oceanic weather ships (Table 1 (2)), whereas more emphasis was put on the ships (Table 1 (3)) for the surface analysis, because the land observations appear to be influenced by orographic effects and do not represent their neighbouring areas.

We copied the data of velocity, temperature, humidity twice a day for the upper atmosphere, and those of surface pressure, velocity, temperature, dew point, and cloud amounts for the surface of the set up regions. During the period of the concerned 10 days the low existed along the coasts of the American continent which sometimes moved towards the northern part of the ocean near Greenland, and the anticyclone in the subtropical region extending from the central Atlantic ocean to the south of the Iberian peninsula. The mean map of the surface pressure over the above 10 days is shown in Fig. 2 and those of the surface level and 500 mb velocity fields are in Fig. 3and 4.

The essence of the water vapour balance is explained in the following. The increment of the water in the atmosphere per unit time and per unit mass of the atmosphere is,

$$-\frac{dc}{dt}=\frac{\partial q}{\partial t}+u\frac{\partial q}{\partial x}+v\frac{\partial q}{\partial y}+\omega\frac{\partial q}{\partial p} \tag{1}$$

in the p coordinate. Specific humidity is denoted by q, water concentration by c and vertical p velocity by $\omega\left(=\frac{dp}{dt}\right)$. If we consider the period long enough, the first term in the above equation becomes relatively small and negligible. The equation (1) can be rewritten with the aid of the continuity equation and takes the form of,

$$-\frac{dc}{dt}=\frac{\partial qu}{\partial x}+\frac{\partial qv}{\partial y}+\frac{\partial \omega q}{\partial p}$$

By integrating the above equation with p from the top of atmosphere (P_∞) to the surface level (P_S), we get,

$$-\int_{P_\infty}^{P_S}\frac{dc}{dt}dp=\int_{P_\infty}^{P_S}\left(\frac{\partial qu}{\partial x}+\frac{\partial qv}{\partial y}\right)dp+\int_{P_\infty}^{P_S}\frac{\partial \omega q}{\partial p}dp$$

Setting the boundary condition $\omega q=0$ at $P=P_\infty$ that was assumed to be at 400 mb level in the present analysis, the following equation is obtained,

$$P_r=E-M \tag{2}$$

where we defined the three quantities as follows;

$$P_r\equiv\frac{1}{g}\int_{P_\infty}^{P_S}\frac{dc}{dt}dp$$

(the condensation amount occuring above P_S level, that is, precipitation observed at the surface level)

$$E\equiv-\frac{1}{g}(\omega q)_{P_S}$$

(the water vapour amount transported upward through the P_S level, that is, evaporation from the sea surface)

$$M \equiv \frac{1}{g} \int_{P_\infty}^{P_S} \left(\frac{\partial qu}{\partial x} + \frac{\partial qv}{\partial y} \right) dp$$

(the divergence field of the water vapour transport within the region bounded by the P_∞ and P_S level)

As is clear from the equations (2), evaporation becomes larger than the precipitation where $M > O$, while the opposite relation holds where $M < O$. The excess of the water vapour (evaporation over precipitation) in a setup region will be transported to the neighbouring continents or parts out of the region and supply energy to the atmosphere there when it comes to condense.

Evaporation is estimated by the evaporation chart constructed by Kondo[5], [6] (1962). The process of analysis of the quantity M is, (1) to obtain the components u and v, (2) to compute the quantities qu and qv, (3), to take the mean of these quantities for 10 days and integrate with regard to P from the surface to 400 mb, and (4), to compute the divergence by performing differentiation with horizontal axes x and y. Although we are provided with the actual data until 500 mb level, the interpolation from the values at the surface, 850. 700 and 500 levels makes it possible to estimate the value at 400 mb. Examples of the distribution of qu and qv with height is shown in Fig. 5.

3. Results and Discussions

In winter cold air masses which have passed over the American continent, comes into the western Atlantic ocean, and becomes unstable because of the underlying warm Gulf Stream. Moreover, the westerly is strong in the middle latitudes. These two effects contribute in creating strong evaporation as shown in Fig. 6. The intense evaporation area spans from the western Atlantic off the east coast of the U. S. to the north-western part of the ocean south of Greenland. The center of this area is located north of Bermuda, and evaporation there amounts to more than 18mm/day. In the subtropical region where the high is located, evaporation amounts only to 3 mm/day because of the calm wind. The general tendency of the obtained evaporation distribution is similar to that of Jacobs (1951). The averaged value of evaporation over the whole Atlantic ocean (20−10°N) is 6 mm/day, while 4 mm/day evaluated from the results of the seasonal distribution given by Jacobs.

Our estimate is larger than Jacobs'. We can attribute this discrepancy partially to the difference of the length of the periods of two studies. However, the essential reason is found in the difference between the Kondo's evaporation chart and Jacobs'[7] methods (1942) in estimating evaporation. The evaporation chart method which owes to the theory of turbulence given by Yamamoto[8] (1959) considers not only the stability of the atmosphere but also the roughness of the sea surfaces caused by the wind. Jacobs' method was built on the simple basis that evaporation (E) is proportional to the product of the difference of the saturation water vapour pressure (e_S) right above the surface and water vapour pressure (e) in the atmosphere slightly

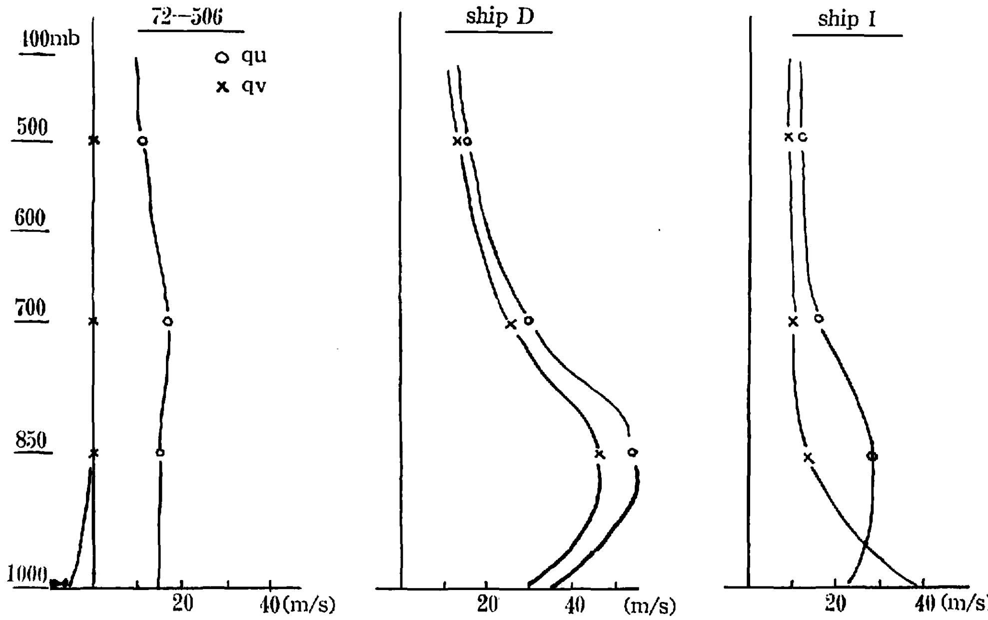

Fig. 5. Examples of vertical distribution of qu and qv

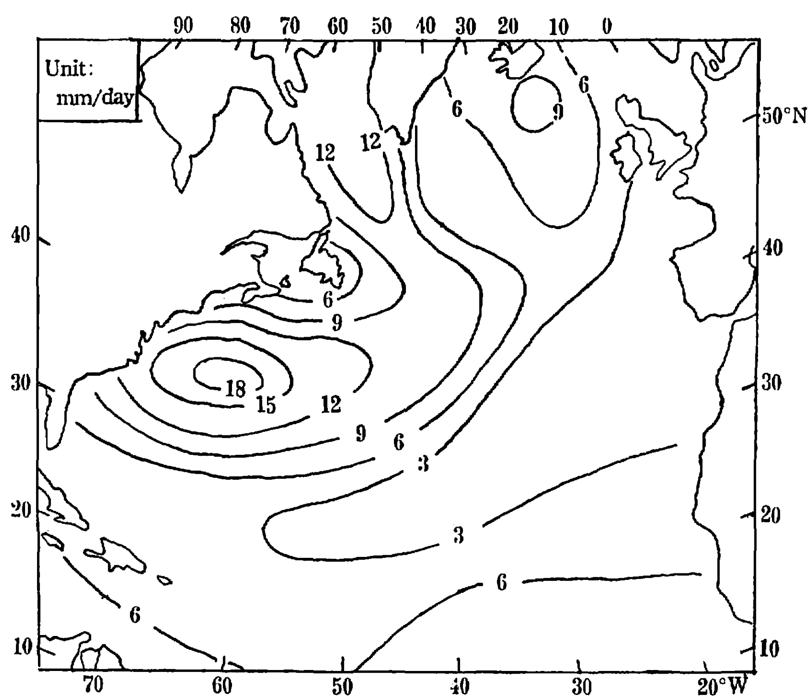

Fig. 6. Distribution of evaporation from the North Atlantic Ocean

high above the surface, and the wind velocity (V), as formulated by $E=k(e_s-e)V$, where k was empirically determined using the evaporation obtained from the heat budget theory for particular regions. This simplification in estimating evaporation results in the fact that Jacobs' equation gives the large value for the weak wind and small value for the strongwind as in the middle latitude in winter. In addition, elimination of the stability consideration causes the underestimation in the Jacobs' method.

The divergence field of water vapour transport is shown in Fig. 7. The western, central and southern parts of the Atlantic ocean experience the positive M with the most intense areas over the east coast of the U. S. and over the 10−20 N region near the equator. The negative value of M is found in the higher latitude (40−60 N)zone and in the south of the Iberian peninsula. The water vapour should come into the negative region of M from the positive region. The averaged value of M over 20−60 N region is about $2\cdot10^{-1}$ gr/day cm^2 (2 mm/day) which is roughly equal to the corresponding value estimated from Jacobs' results of E-P. According to the Benton and Estoque' s estimation (1954), −0.8 mm/day of M exists over the American continent in winter. It is understandable that M takes negative values over any continent since it acts as a water vapour sink, and this negative must be somehow compensated by the positive value M over the oceans acting as a water vapour source.

Precipitation computed from the equation (2) is shown in Fig. 8. The general picture of precipitation is similar to the evaporation field, but some differences are

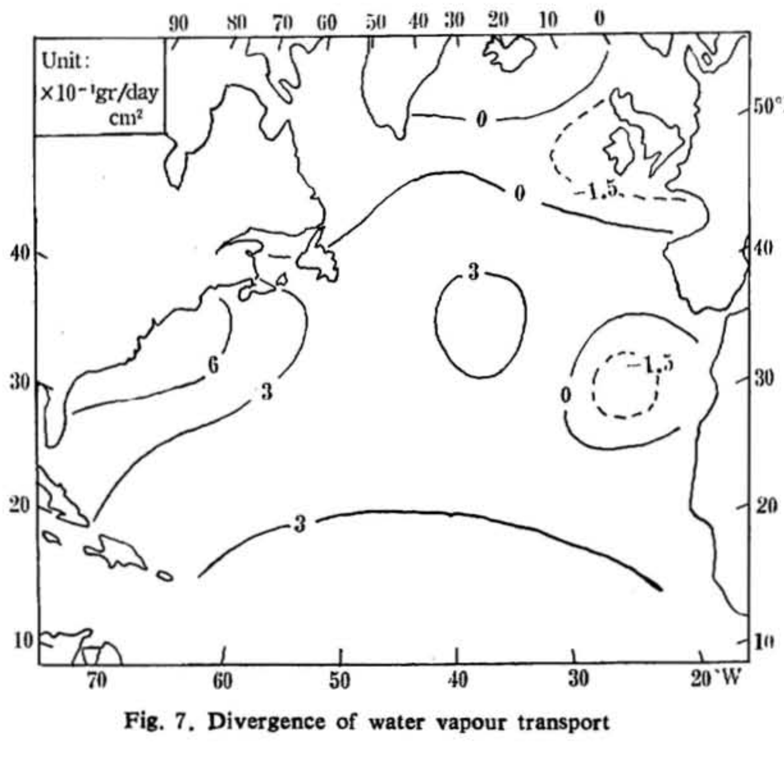

Fig. 7. Divergence of water vapour transport

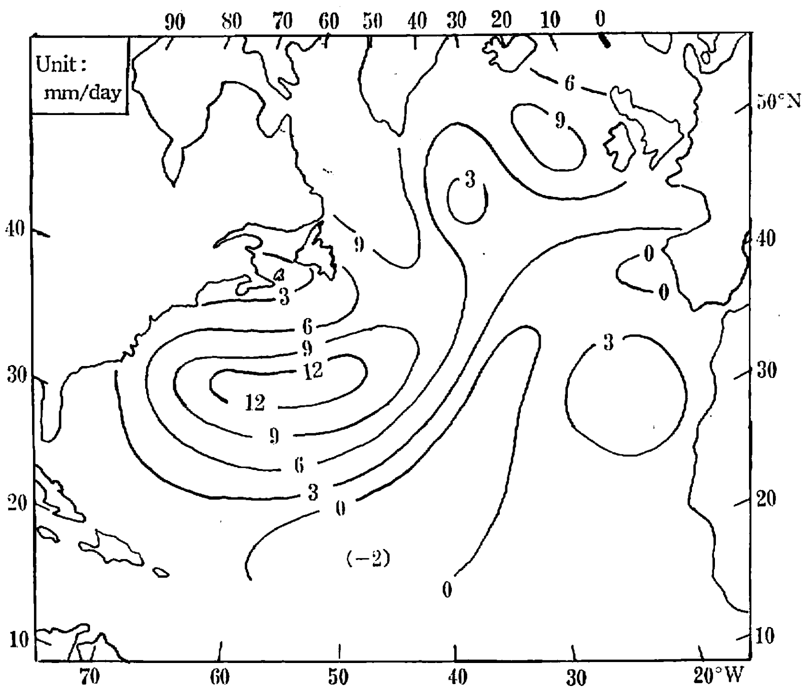

Fig. 8. Computed precipitation distribution

seen in that the center of the strong precipitation area is located slightly more eastward and generally precipitation is weaker in intensity than evaporation. The intense area discovered over the Madiera Canary islands seems to be caused by the cyclone which occassionally passed over this area, but it would have been eliminated if smoothing out the chart by taking a longer period. The negative area of precipitation found in this figure must be the error in the computation. The distribution of the cloud amount of low clouds below 2500ft shown in Fig. 9 is in fairly good agreenent with the precipitation field. It is understood that cloud amount should not represent the precipitation exactly, but it can be of help to guess precipitation.

The averaged value of precipitation over 20−60 N is 4.0 mm/day. Both estimates given by Tucker (1961) and by Jacobs (1951) amount to 2∼3 mm/day which is less than the present result. It is because there might be much rain during the period of lst through 10th Jan., 1957. The better comparison can not be made until we try to find the result for the period as long as Jacobs or Tucker did.

The mean vertical p-velocity were analysed for 850 mb, 700 mb (not shown) and 500 mb, shown in Fig. 10 and Fig. 11. They were obtained from integrating the equation of continuity written in the p-coordinate with the lower boundary condition that the mean vertical p-velocity is zero ($\omega = 0$) at the surface.

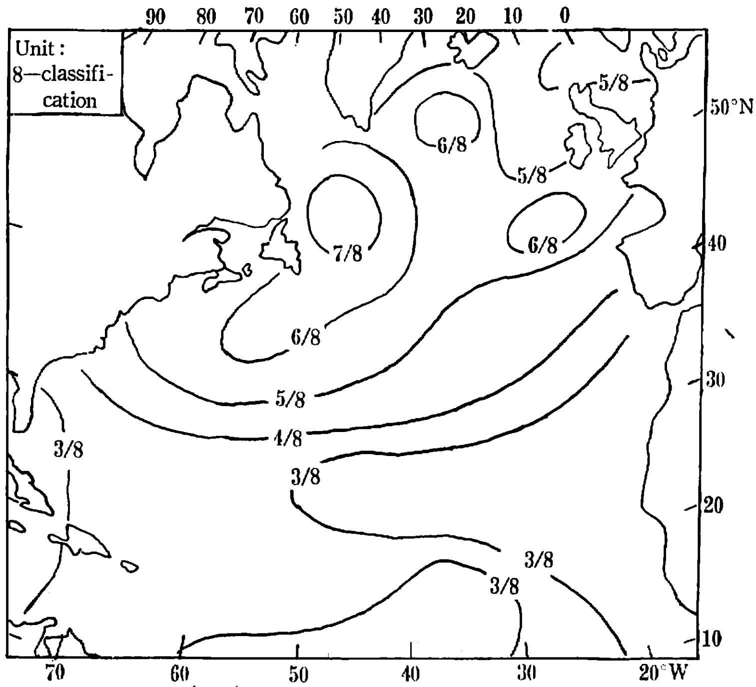

Fig. 9. Cloud amount distribution of low cloud

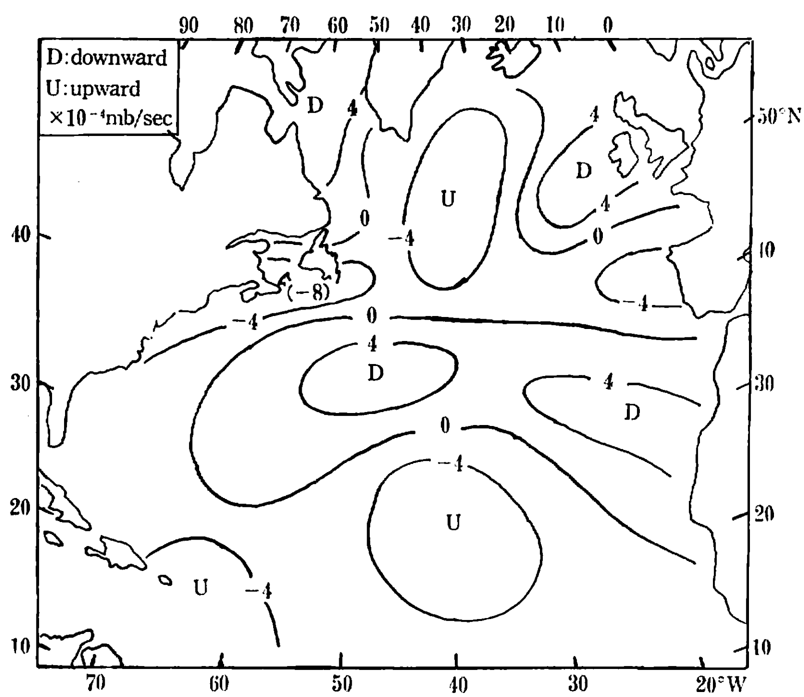

Fig. 10. Distribution of ω at 850 mb level

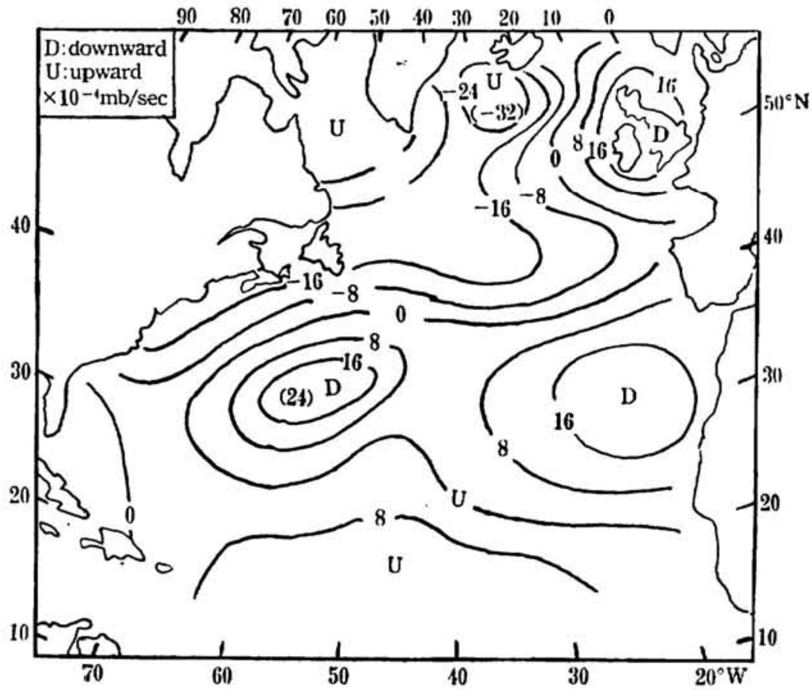

Fig. 11. Distribution of ω at 500 mb level

We obtained 4×10^{-4}, 8×10^{-4} and 16×10^{-4} mb/sec for the value of ω at the levels of 850, 700 and 500 mb, respectively. In the subtropical region where an anticyclone exists, the descending motion dominates, as shown in the maps. The belt of the ascending motion extends from the east coasts of the U. S. and Canadian continents to the south of Greenland and Ireland. The pattern of the mean vertical p-velocity is not in agreement with that of precipitation, especially over the east of Bermuda and the west of Iberian peninsula. The vertical velocity (or vertical p velocity) is often used as a representative of precipitation on the basis of the positive correlation between the two quantities. However, this is significant only when the argument is concerned with a small area. As is seen in the present analysis, there is little correlation between the precipitation and vertical velocity when they are treated over the wide area. This absence of the relationship of the precipitation with the velocity (or vertical p-velocity, can be attributed to disappearance of the perturbation of vertical motion which directly contributes to precipitation in the stage of taking average of the vertical velocity over the wide area.

4. Conclusions

It is impossible to get actual distribution of precipitation because of no reports of precipitation from the oceanic weather ships. The present author showed how it was possible to get the information of precipitation by means of the water vapour balance indirectly.

The most intense evaporation is located over the east of the American continent north of Bermuda, while that of precipitation is little more eastward. Neither evaporation nor precipitation is strong over the subtropical area. Generally speaking, there are strong precipitation and evaporation in the cyclonic area.

The averaged values over the concerned ocean area for evaporation, precipitation and the divergence of the water vapour transport are about 6, 4, 2 mm/day, respectively. The positive amount 2 mm/day of M $(=E-P)$ means that this excess amount of the water vapour is being transported to the other region of the ocean or to the American and European continents.

The comparison between the vertical velocity and the precipitation showed no relation between them, objecting to the ordinary conception of the vertical velocity. It became clear in the present paper that it is not necessary for area of the ascending motion to have precipitation, as in the spots of the east of Bermuda, west of England and south of Iceland. We came to think that the eddies of vertical velocity greatly contributes in precipitation so that we can not relate the precipitation directly to the mean vertical velocity.

References

(1) Benton, G. S. and M. A. Estoque, 1954 : Water vapour transfer over the North American continent. J. Met., Vol. 11, 462-477.

(2) W st, G., 1936: Oberfl chensalzgehalt, Verdunstung, und Niederschlag auf dem Weltmeere., L nderkundliche Forshung, Festschrift Norbert Krebs, SS. 347-359.

(3) Jacobs, W. C., 1951: Largescale aspects of energy transformation over the oceans. Compendium of Met. Ame. Met. Soc., 1057—1070.

(4) Tucker, C. B., 1961: Precipitation over the North Atlantic Ocean. Quart. J. R. M. Soc., Vol. 87, 147—158.

(5) Kondo, J., 1962: Evaporation from extensive surfaces of water. Sci. Re. Tohoku Univ., Ser. 5, Geophys., Vol. 14, 107—119.

(6) Kondo, J., 1963: Evaporation from the Japan Sea in the winter monsoon season. Sci. Rep. Tohoku Univ., Ser. 5, Geophys., Vol. 15, 67—75.

(7) Jacobs, W. C., 1942: On the energy exchange between sea and atmosphere. J. Marine Res., Vol. 5, 37—66.

(8) Yamamoto, G., 1959: Theory of Turbulent Transfer in Non-Neutral Conditions. J. Met. Soc. Japan, Vol. 37, No. 2, 60—69.