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**DEPARTMENT OF TRANSPORT  
METEOROLOGICAL BRANCH**

**THE DIABATIC CONTRIBUTION  
TO THICKNESS TENDENCY**

**BY  
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ABSTRACT

Important diabatic processes occurring in the atmosphere are briefly discussed and their contribution to thickness change is calculated for selected stations on the North Atlantic Ocean during a seven-day winter period. Results are compared with the diabatic thickness change determined by evaluating the terms in the thickness tendency equation. Although measurement errors are appreciable, values obtained by the two methods are comparable when the data have been suitably smoothed. The results indicate the flux of sensible heat and water vapour can be used to evaluate the diabatic contribution to thickness tendency when cold air advection is occurring over the sea.

CANADA - DEPARTMENT OF TRANSPORT - METEOROLOGICAL BRANCH

THE DIABATIC CONTRIBUTION TO THICKNESS TENDENCY

by

E. C. Jarvis

1. INTRODUCTION

It has been shown by Sutcliffe and Forsdyke (1950) that diabatic processes of heating and cooling can play a large role in altering the 1000-500-mb thickness. The modification of cold air masses as they pass over a warmer water surface has been discussed by Craddock (1951), Burke (1945), Klein (1945), and Burbidge (1951). Their studies indicate the increase in 1000-500-mb thickness due to heating from the underlying surface may occasionally be in excess of 800 ft in a 24-hour period. It has been noted by Clapp (1961) that error fields in mid-tropospheric contour forecasts are closely related to effective sources and sinks of heat at the surface of the earth. This has given fresh impetus to the question of the role of diabatic heating and cooling on synoptic-scale weather systems. The effect of heating from below on weather types has been discussed by Petterssen (1959). Although the relation between cloud and heat added to the air is not unique, he determined that the effect of heat sources and sinks on weather type can be very large.

The transfer processes occurring at the earth-air interface are not yet completely understood, however, simplified formulae that make use of customary synoptic data are available for computing the flux of heat and water vapour (Leese and Young, 1963; Duquet, 1961; and Petterssen, Bradbury and Pedersen, 1961). The effect of the release of latent heat of condensation due to large scale ascent on the heating of the atmosphere can be computed (Pedersen, 1961) and radiative fluxes can be estimated (Craddock, 1951). Some knowledge of the effect of these diabatic processes on the surface pressure field, upper-level wave patterns, and weather types could be of considerable assistance in weather forecasting. This study deals with the role of diabatic processes in altering the 1000-500-mb thickness field.

2. THE THICKNESS TENDENCY EQUATION

By introducing the hydrostatic assumption, the thermodynamic energy equation can be written (Reed, 1958):

$$\frac{\partial}{\partial t} \left( \frac{\partial Z}{\partial p} \right) = -\vec{v}_H \cdot \nabla \frac{\partial Z}{\partial p} + \frac{\alpha}{g\theta} \frac{\partial \theta}{\partial p} \omega - \frac{\alpha}{g\theta} \frac{d\theta}{dt}, \quad (1)$$

where  $Z$  is the height of the isobaric surface,  $\vec{v}_H$  is the horizontal wind vector at pressure  $p$ ,  $\omega$  is the vertical velocity in pressure coordinates,

$\alpha$  is the specific volume of air at pressure  $p$ ,  $g$  is the acceleration of gravity, and  $\theta$  is the potential temperature. When the thermodynamic energy equation is integrated from  $p_0 = 1000$  mb to  $p_5 = 500$  mb, it becomes

$$\frac{\partial \bar{h}}{\partial t} = A_h + \frac{\bar{\alpha}}{g\theta} \Delta \theta \bar{\omega} + H. \quad (2)$$

This is a form of the thickness tendency equation where bars denote mean values in the 1000-500-mb layer,  $A_h$  represents the advection of the 1000-500-mb thickness,  $h$ , with the mean wind  $\bar{h}$  in the layer, and  $H$  is the diabatic contribution to thickness tendency in the layer. The thickness tendency is therefore composed of an advective term,  $A_h$ , a diabatic term,  $H$ , and a dynamic term,  $D$ , where

$$D = \frac{\bar{\alpha}}{g\theta} \Delta \theta \bar{\omega}. \quad (3)$$

The advective term must always be accounted for in determining changes that take place in the thickness pattern. As pointed out by Sutcliffe and Forsdyke (1950), the whole basis of practical forecasting rests on the fact that changes take place largely by advection. The empirical relation that  $\partial h / \partial t \cong \frac{1}{2} A_h$  (Craddock, 1951) has had some use in forecasting, as a statistical relationship.

The dynamic term describes adiabatic changes in the thickness tendency that occur as a result of large-scale ascent and subsidence. In a dry atmosphere, subsidence produces warming which increases the thickness, and ascent results in cooling and a decrease in thickness. If a portion of the atmosphere becomes saturated due to large-scale ascent, the amount of cooling will be reduced because of the release of latent heat of condensation, and if the air is saturated and unstable, large-scale ascent will result in warming. That portion of the heat acquired or lost because of condensation and evaporation is part of the diabatic process. However, since it is most easily accounted for by adjusting the vertical velocity for initial unsaturation and release of latent heat (Petterssen et al, 1961) it is sometimes incorporated in the dynamic term of the thickness tendency equation.

### 3. THE DIABATIC PROCESSES

The diabatic term includes all processes in which heat is added to, or derived from, the atmosphere. The transfer of real (sensible) heat into the air from the earth's surface is the most readily observed of the diabatic processes, as indicated by the rapid modification of cold air masses as they pass over a warmer water surface. Leese and Young (1963), Petterssen, Bradbury and Pedersen (1961), and Duquet (1961) have expressed the flux of sensible heat in the form:

$$F_s = K_1 (T_s - T_a) V_a \quad (4)$$

where  $F_s$  is the flux of sensible heat, usually expressed in units  $\text{cal cm}^{-2} \text{min}^{-1}$ ,  $T_s$  is the sea surface temperature, and  $V_a$  is the wind at anemometer height.  $K_1$  is a constant of proportionality which assumes values of drag coefficient, roughness parameter, and Karman constant that are typical of neutral air mass stability and windy regimes at sea. This expression for the flux of sensible heat is regarded as being fairly reliable (Clapp, 1961), although it may not be entirely satisfactory for low winds and stable air masses. The equation is probably least reliable for cases where the air temperature is higher than the temperature of the sea and the flux of heat is downward (Haltiner and Wang, 1960). It is possible that a similar relation can be applied, in a statistical sense, to evaluate the heat flux over land. However, the lack of knowledge of soil temperatures and the variability of conductive heat capacities of typical substances composing the earth's surface prohibits a direct evaluation of the sensible heat flux in this manner. Synoptic experience suggests the heat transfer across an earth-air interface is much less than the transfer across a sea-air interface because, for practical purposes, the sea can be regarded as an infinite heat source, whereas the capacity of soil to conduct heat to or from its surface layer is much more limited. For this reason the heat flux between land and air is usually ignored.

Evaporation from the sea will occur when the sea surface temperature is higher than the dew-point temperature of the air near the surface. The latent heat of vaporization is acquired from the sea and will be realized by the air when condensation occurs. If a windy regime exists some water vapour will be evaporated from spray and the latent heat of vaporization will be derived from the air. However, this does not result in a net heat exchange since the heat will be returned to the air when condensation occurs. A similar argument holds for evaporation from falling rain. The water vapour content of the air is increased, but the latent heat of vaporization is derived from the air (Craddock, 1951).

As shown by Leese and Young (1963), the flux of water vapour from an underlying surface can be expressed in a form similar to the equation describing the flux of sensible heat:

$$F_{wv} = K_2 (e_s - e_a) V_a \tag{5}$$

where  $F_{wv}$  is the flux of water vapour,  $K_2$  is a constant of proportionality,  $e_s$  is the vapour pressure of the sea surface, and  $e_a$  is the vapour pressure of the air near the surface.

The warming of a cold air mass over a warmer sea always leads to convective cloud formation (Petterssen, 1959). This type of convective cloud will form over relatively small lakes, suggesting that some of the heat acquired by evaporation from the sea will be realized soon after, when convective cloud begins to form. Sensible heat is acquired by the air at the same time as the process of evaporation goes on, so not all the latent heat

acquired by evaporation will be released soon after it has been obtained, since the air can contain more water vapour at higher temperatures. Burke (1945) has assumed turbulent mixing causes the air to maintain a constant value of specific humidity from a height of about 15 meters to the top of the convective layer. If this assumption is made, the amount of water vapour acquired by evaporation that condenses and releases latent heat can be forecast. Reed (1958) and Klein (1945) have indicated that the relative humidity of a cold air mass over a sea surface is approximately 60 percent at a height of 15 meters above the ocean. For operational purposes, the specific humidity can be replaced by the mixing ratio, and the moisture distribution from a height of 15 meters to the top of the condensation layer can be readily determined to be 0.6 the mixing ratio that corresponds to the surface air temperature.

Figure 1 shows the alteration of an initial temperature curve by the addition of sensible heat at the sea-air interface, and the release of latent heat of condensation. The moisture distribution from 15 meters to the top of the condensation layer corresponds to mixing ratio  $r_a$  in figure 1. The mixing ratio  $r_i$ , in figure 1, represents the average initial mixing ratio to the top of the forecast convective layer, and  $r_f$ , the final mixing ratio to the top of the convective layer, is represented by curve DFC. So the gain in water by the air mass over the forecast period can be represented by  $r_a - r_i$ . Of this amount,  $r_a - r_f$  represents the amount of water vapour that has condensed. Therefore, the amount of water vapour acquired by evaporation from the sea, in the forecast period, which condenses is given by the ratio

$$\frac{r_a - r_f}{r_a - r_i} .$$

Hence,

$$F_L = F_{wv} \frac{r_a - r_f}{r_a - r_i} = K_2 \frac{r_a - r_f}{r_a - r_i} (e_s - e_a) V_a , \quad (6)$$

where  $F_L$  is the portion of water vapour flux,  $F_{wv}$ , that releases latent heat in cold air masses soon after the water has been acquired by evaporation from the sea. Equation (6) implies turbulent mixing is complete, and the horizontal diffusion of water vapour is negligible.

The ratio  $(r_a - r_f) / (r_a - r_i)$  was evaluated for 50 winter cases in which 24-hour trajectories terminated at Sable Island, or at weather ships Bravo, Cocoa, Delta or Echo.

Trajectories were determined by a grid method (Jarvis, 1964), and  $r_i$  was evaluated by averaging the mean initial mixing ratios in 50-mb layers to the top of the convective layer.  $r_f$  was determined by averaging the mean saturation mixing ratio of the forecast curve, curve DFC in figure 1, in 50-mb layers

to the top of the convective layer.  $r_a$  was assumed to be 0.6 the mixing ratio corresponding to the surface air temperature, at the 1000-mb pressure level, at the endpoint of the 24-hour trajectory.

For the 50 cases, the average value of the ratio  $(r_a - r_f) / (r_a - r_i)$  was found to be 0.57 with a standard deviation 0.18. The area of chief concern is where Arctic air masses move over a relatively cold sea. Because saturation vapour pressures are low in these conditions, the flux of water vapour is much less than the flux of sensible heat, so, in spite of the large standard deviation, for operational purposes,  $(r_a - r_f) / (r_a - r_i) \approx 0.5$ .

Then, equation (6) becomes

$$F_L = 0.5 K_2 V_a (e_s - e_a) \tag{7}$$

The release of latent heat of condensation that is brought about by large-scale ascent can be a significant diabatic process (Pedersen, 1961). As indicated in section 2, this effect can be accounted for by adjusting the vertical velocity for moisture content and the release of latent heat of condensation in the air. The detailed procedure, and assumptions involved, have been discussed by Harley (1963). In summary, an effective vertical velocity,  $\omega_E$ , is defined. This is the vertical velocity that would produce the same amount of precipitation in a saturated air mass in time  $\Delta t$ , as would the actual vertical velocity if the air was saturated for only a fraction of the time  $\Delta t$ . Ascending motion is required to produce precipitation. Since  $\omega_E$  and  $\omega_6$  are negative for ascent, the implication in the definition of  $\omega_E$  is that more ascent is required to produce a given amount of precipitation in a given time when the air is unsaturated than when it is saturated. Harley has shown that  $\omega_E$  can be approximated by

$$\omega_E = \omega_6 + 0.5 (T - T_d)_7 \tag{8}$$

where  $\omega_6$  is the vertical velocity at 600 mb, and  $(T - T_d)_7$  is the difference between temperature and dew point at 700 mb.

This equation pertains to a time interval,  $\Delta t$ , of 6 hours. It can be readily determined that, if the time interval is reduced to 3 hours, the equation for effective vertical velocity becomes:

$$\omega_E = \omega_6 + (T - T_d)_7 \tag{9}$$

Reducing the time interval places greater emphasis on the term  $(T - T_d)_7$ , and since it is subject to rather large local variations, constrains the usefulness of the concept of an effective vertical velocity. However,

since latent heat vertical velocity is not particularly sensitive to variations in effective vertical velocity, as shown by Table 1, and because tendency fields are usually adapted to a time interval of 3 hours, the effective vertical velocity given by equation (9) is used in this study.

Having determined the effective vertical velocity,  $\omega_E$ , the latent heat vertical velocity,  $\omega_H$ , can be determined from Table 1, where  $T_7$  is the 700-mb temperature.

Table 1

The relationship of latent heat vertical velocity,  $\omega_H$ , to 700-mb temperature,  $T_7$ , and effective vertical velocity,  $\omega_E$ .  $T_7$  is in degrees Celsius, and  $\omega_H$  and  $\omega_E$  are in units  $10^{-3}$  mb sec $^{-1}$ . (after Harley)

$T_7$	-14	-5	1	6
$\omega_H$	$0.5 \omega_E$	$\omega_E$	$1.5 \omega_E$	$2 \omega_E$

The thickness tendency due to large-scale vertical motion is given by equation (3).  $\bar{\omega}$ , the mean vertical velocity in the 1000-500-mb layer, can be replaced by the mean dry-adiabatic vertical velocity,  $\bar{\omega}_A$ , and a latent heat vertical velocity,  $\bar{\omega}_H$ , such that

$$\bar{\omega} = \bar{\omega}_A + \bar{\omega}_H \quad (10)$$

If a parabolic vertical velocity profile is assumed,  $\bar{\omega}_A$  will be approximately two thirds the vertical velocity determined by the Penner method (1963). The thickness tendency due to the release of latent heat by large-scale ascent becomes

$$\left(\frac{\partial h}{\partial t}\right)_L = \frac{\bar{\alpha}}{g \bar{\theta}} \Delta \theta \omega_H \quad (11)$$

It is not unusual to note large-scale subsidence to the rear of a developing cyclone over the oceans, as implied by strong cold air advection and anticyclonic, or weak cyclonic, mid-tropospheric vorticity advection (Penner, 1963). In this case no heat will be added to the air by condensation resulting from large-scale vertical velocities. However, the cold air mass, which was dry over land, may have moderate or heavy convective cloud and showers or thunderstorms when it moves over the sea. A large amount of water vapour must therefore have been acquired from the sea in a small interval of time. Convective currents, caused by the acquisition of heat and water vapour in the cold air, will have

supplied the release mechanism for substantial amounts of latent heat. Widespread local convection and associated large-scale subsidence are compatible. The effect of the large-scale subsidence is to speed up the stabilizing process, but not to inhibit local convection.

In addition, most of the moisture condensed, as a result of the large-scale vertical velocity, occurs in the warm air mass along active fronts or in the vicinity of a developing cyclone. Much of the water vapour in the warm air is acquired in the region of the subtropical highs and may have been transported hundreds of miles over a period of many days before the latent heat is realized by the air in middle latitudes.

Therefore, if attention is confined to areas of synoptic size, and to forecast periods of a few days or less, it appears necessary to account for latent heat released in the warm air mass by large-scale ascent, and latent heat acquired by cold air masses and then released by local convection, as two distinct processes.

The effect of radiative transfer is usually considered negligible over periods of a few days (Craddock, 1951; Burke, 1945; and Duquet, 1961) since the rate of heat exchange is relatively small and shows little horizontal variability other than latitudinal. Other diabatic processes are, however, closely linked with synoptic-scale weather systems. Also neglected for this period of time is the transformation of kinetic energy to heat in the friction layer. Petterssen (1959) has shown that this effect is an order of magnitude smaller than the flux of water vapour or sensible heat.

4. EVALUATION OF THE EFFECT OF DIABATIC PROCESSES ON THICKNESS TENDENCY

The flux of sensible heat is given by equation (4) where  $K_1 = 1.338 \times 10^{-3}$  when  $F_s$  is in  $\text{cal cm}^{-2} \text{min}^{-1}$ ,  $V_a$  is in knots, and  $T_s$  and  $T_a$  are in degrees Fahrenheit. Petterssen et al (1961) calculated the effect of  $F_s$  on the 1000-500-mb thickness,  $h$ . The flux of sensible heat was assumed to decrease linearly with pressure to a maximum depth of penetration of 700 mb. A heat flux of  $1 \text{ cal cm}^{-2} \text{min}^{-1}$  was then found to correspond to a thickness change of 82 feet in 3 hours, or 25 meters in 3 hours. This value was noted to be relatively insensitive to the depth of penetration. Using this relation, the thickness tendency that results from the transfer of sensible heat becomes

$$\left(\frac{\partial h}{\partial t}\right)_s = 0.01 V_a (T_s - T_a) \quad (12)$$

where  $V_a$  is in knots,  $T_s$  and  $T_a$  are in degrees Fahrenheit, and  $(\partial h / \partial t)_s$ , the thickness tendency due to sensible heat, is in feet per 3 hours.

The portion of the flux of water vapour that releases latent heat in cold air masses given by equation (7), where  $K_2 = 0.739 \times 10^{-3}$ ,  $F_L$

is in  $\text{cal cm}^{-2} \text{min}^{-1}$ ,  $V_a$  is in knots and  $e_s$  and  $e_a$  are in millibars. If the condensation level is assumed to be 900 mb, the forecast tephigram intersects the initial curve close to 600 mb when the curve forecast using only the sensible heat term intersected the original curve at 700 mb (figure 1). As shown by figure 2, the decrease in saturation mixing ratio with decreasing pressure to 600 mb is nearly linear over the range of predicted temperatures which is of interest. Then the same criterion can be applied to the flux of water vapour as was assumed for the flux of sensible heat, and the thickness tendency resulting from the flux of water vapour in cold air,  $(\partial h / \partial t)_v$ , becomes

$$\left(\frac{\partial h}{\partial t}\right)_v = 0.03(e_s - e_a) V_a \quad (13)$$

where  $(\partial h / \partial t)_v$  is in units feet per 3 hours,  $V_a$  is in knots, and  $e_s$  and  $e_a$  are in millibars.

The thickness tendency resulting from condensation due to large-scale vertical velocities,  $(\partial h / \partial t)_L$ , was found in section 3 and can be written

$$\left(\frac{\partial h}{\partial t}\right)_L \approx 360 \frac{\Delta \theta}{\bar{\theta}} \omega_H \quad (14)$$

$(\partial h / \partial t)_L$  is in units feet per 3 hours,  $\omega_H$  is in units  $10^{-3} \text{mb sec}^{-1}$ ,  $\bar{\theta}$  is the mean potential temperature in the 1000-500-mb layer in degrees Kelvin, and  $\Delta \theta$  is the change in potential temperature between 1000 mb and 500 mb in degrees Kelvin.

$(\partial h / \partial t)_s$  and  $(\partial h / \partial t)_v$  were evaluated every 3 hours at Sable Island and at weather ships Alfa, Bravo, Cocoa, Delta and Echo for the 7-day period March 1 to 7, 1961.  $(\partial h / \partial t)_L$  was evaluated every 12 hours for these stations during the 7-day period, using the method described by Harley (1963) to determine  $\omega_H$ .  $\bar{\theta}$  was evaluated by averaging the potential temperature in 100-mb layers. It has been pointed out by Petterssen et al (1961), and by Leese and Young (1963) that smoothing of the data is necessary. To avoid large measurement errors, the thickness tendency was evaluated at only selected stations which had both surface and radiosonde observing stations. Because of the great distance separating stations, areal smoothing was not possible, so the data for each station was time-smoothed over the seven-day period. The result of the computations is shown in Table 2.

Table 2

Seven-day average values of thickness tendency in feet per 3 hours resulting from sensible heat flux,  $(\partial h/\partial t)_s$ , flux of water vapour,  $(\partial h/\partial t)_v$ , and heat acquired by condensation due to large-scale ascent,  $(\partial h/\partial t)_L$ . The thickness tendency due to the sum of these diabatic processes is F. <sup>L</sup>

STATION	$(\partial h/\partial t)_s$	$(\partial h/\partial t)_v$	$(\partial h/\partial t)_L$	F
A	29	2	2	33
B	23	3	0	26
C	1	1	2	4
D	18	4	14	36
E	5	4	25	34
SA	3	1	14	18

The diabatic term of the thickness tendency equation was evaluated from

$$H^* = -A_h + \frac{\partial h}{\partial t} - D_A$$

where  $H^* = H + D_L$ ,  $D_L = 360 (\Delta \theta/\bar{\theta}) \omega_H$  and  $D_A = 360 (\Delta \theta/\bar{\theta}) \omega_A$ .  $A_h$  is the advection of the 1000-500-mb thickness with the 500-mb wind,  $(\partial h/\partial t)$  is the 3-hour thickness tendency interpolated from the observed 12-hour thickness change,  $D_A$  is the dynamic change resulting from the mean dry adiabatic vertical velocity,  $\omega_A$ , and  $D_L$  is the change resulting from latent heat vertical velocity,  $\omega_H$ . Table 3 shows the contribution of the various terms of the thickness tendency equation for weather ships, Alfa, Bravo, Cocoa, Delta and Echo, and for Sable Island for the 7-day period.

Table 3

Seven-day average values of the terms of the thickness tendency equation, in units feet per 3 hours. The thickness tendency, advective, and dry-adiabatic dynamic terms are added to get  $H^*$ , the total diabatic contribution to thickness change. F is the thickness tendency due to the sum of the diabatic processes shown in Table 2.

STATION	$(\partial h/\partial t)$	$A_h$	$D_A$	$H^*$	F
A	+4	-43	13	34	33
B	-4	-55	23	28	26
C	-8	-26	15	3	4
D	-5	-36	7	38	36
E	-8	-21	6	32	34
SA	+6	-55	42	19	18

5. DISCUSSION

The map series selected for this study was considered typical of weather patterns on the North Atlantic Ocean in winter. Cold air advection dominated the region during the seven-day period, but was broken in the south by several frontal cyclones, some of which reached appreciable intensities, that passed from the American seaboard to near ship Delta and then to the south of ship Cocoa.

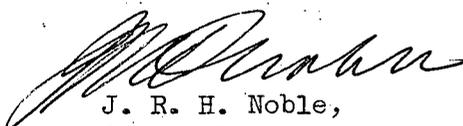
Less reliance can be placed on the values in Tables 2 and 3 than if unsmoothed data had been used. However, it is expected that the measurement errors, although appreciable, would have a random distribution, and the time-smoothing employed would reduce their magnitude.

Table 2 shows the influence of the various types of weather regimes on the diabatic processes. The effect of sensible heat on thickness tendency was dominant at weather ships Alfa and Bravo. Although cumulus cloud was observed at these stations for most of the seven-day period, the latent heat effect that resulted by evaporation from the sea was very small. This was because sea and air temperatures were too low to support much water vapour. Because subsiding air was associated with continued cold advection at these stations, the contribution to thickness tendency resulting from condensation due to large-scale vertical velocities was small. However, this term becomes dominant at stations lying close to the cyclone's tracks. Although the influence of evaporation was small at all stations, where high water temperatures existed, for example at ship Echo, this term and the sensible heat term were comparable.

Values showing the mean contribution of the individual terms of the thickness tendency equation, given in table 3, must be regarded as being fairly reliable because of the close agreement between F in table 2 and H\* in table 3. The table indicates that the diabatic contribution to thickness tendency can be as large, though of opposite sign, as the advection term when cold advection exists. It is doubtful that the diabatic contribution would be as large as this when warm air advection occurred, since the diabatic processes are suppressed by pronounced stability.

Although the analysis undertaken in this study leaves much to be desired, the results suggest that diabatic processes must be considered if prognostic pressure and contour fields are produced over large water bodies in winter when cold air advection exists. It also suggests that when these conditions occur, the flux of sensible heat and water vapour can be used to estimate the magnitude of the diabatic contribution to thickness tendency.

APPROVED,



J. R. H. Noble,  
Acting Director.

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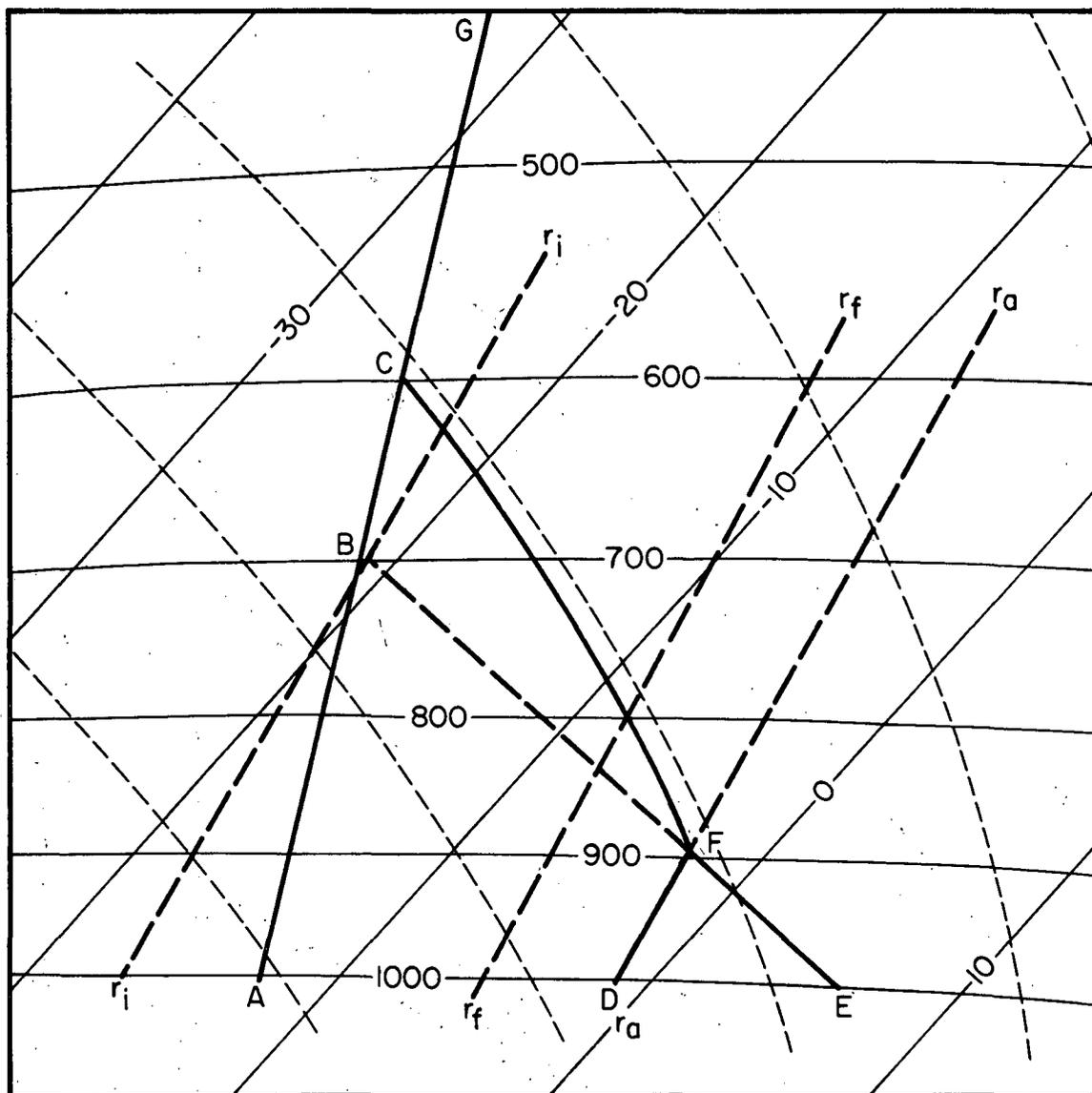


Fig. 1  
 A Schematic Tephigram Showing the Alteration of an Initial Temperature Curve, ABCG, to Curve EFBCG by the Addition of Sensible Heat at the Sea-air Interface, and to Curve EFCG by the Addition of Sensible Heat and Realization of Latent Heat of Condensation. The Top of the Condensation Level is 600 mb.

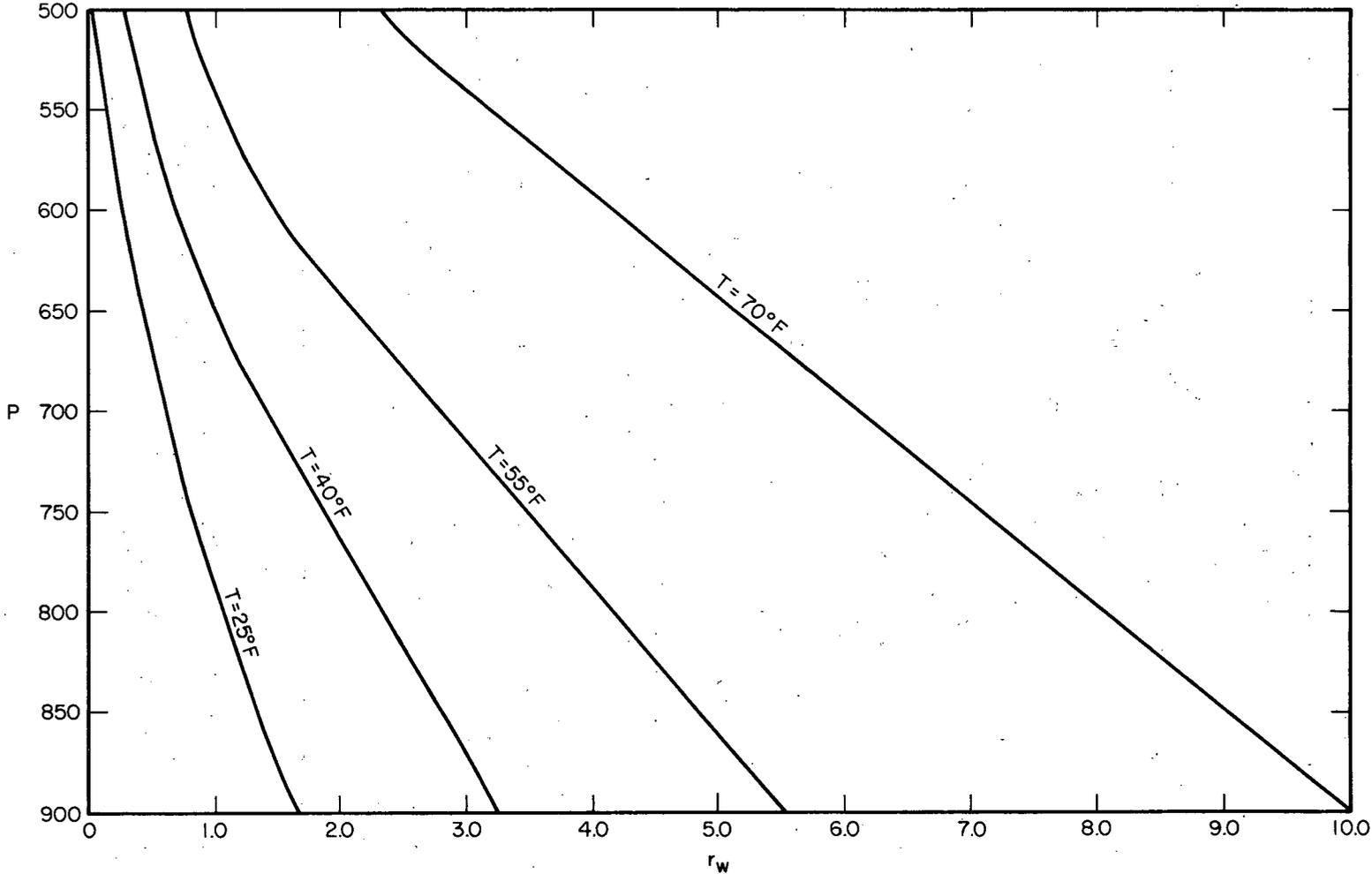


Fig. 2  
The Variation of Saturation Mixing Ratio,  $r_w$ , With Pressure, p,  
Along the Pseudo-adiabat Corresponding to the Surface Temperature, T.

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TEC-521 : 551.510.522  
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10 pps. 2 figs. 16 refs. 3 tables. 14 eqns.

Subject reference: 1. Heat Exchange Sea-atmosphere.  
2. Thermodynamic Energy Equation.  
3. North Atlantic Ocean.

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ABSTRACT: Important diabatic processes occurring in the atmosphere are briefly discussed and their contribution to thickness change is calculated for selected stations on the North Atlantic Ocean during a seven-day winter period. Results are compared with the diabatic thickness change determined by evaluating the terms in the thickness tendency equation. Although measurement errors are appreciable, values obtained by the two methods are comparable when the data have been suitably smoothed. The results indicate the flux of sensible heat and water vapour can be used to evaluate the diabatic contribution to thickness tendency when cold air advection is occurring over the sea.

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