Interactions between North Atlantic Clouds and the Large-Scale Environment

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ABSTRACT

This paper addresses the problem of understanding and predicting the presence of clouds and their effects on the atmosphere in the midlatitudes of the North Atlantic Ocean. The European Centre for Medium Range Weather Forecasting initialized analyses and the U.S. Air Force Three-Dimensional Nephanalysis are employed to construct a joint time series of gridpoint values of cloudiness and large-scale meteorological fields, including heat and moisture budgets, for January 1979. Interpretation of cloud in the context of the large-scale flow is given for the monthly average situation, disturbed and undisturbed conditions, and a longitudinal cross section through a baroclinic wave. In general, middle clouds are formed primarily due to the benefit from large-scale three-dimensional moisture convergence; and low cloud formation depends on surface moisture flux and the static stability. Upper-level moisture was deemed to be sufficiently unreliable so that no inferences regarding high clouds could be made. Comparison of the relative humidity field with cloud cover in a cross section of a baroclinic wave shows that peak cloud fractions are displaced approximately 3.5° latitude to the east of the peak relative humidities. From concurrent examination of the residual heat and moisture sources, it is suggested that clouds do not respond instantaneously to the large-scale relative humidity field, but take a period of time on the order of hours to adjust in terms of evaporation and condensation.

The relationship of cloud fraction to the large-scale humidity field is examined, along with several diagnostic parameterizations of cloud fraction currently employed in general circulation models. The grid-scale threshold relative humidity below which cloud, on average, does not occur was determined to show a strong decrease with height. It was shown that appropriate “tuning” of a diagnostic relative humidity-based parameterization can result in accurate parameterized mean monthly total cloud amount for the region, and layer cloud fractions to within 5% of observed layer cloud fractions. However, this type of cloud fraction parameterization appears to be unable to diagnose layer cloud fraction on the smaller time and space scales that are undoubtedly required for obtaining the correct local cloud radiative and hydrological feedback with the dynamics.

1. Introduction

This paper focuses on the problem of understanding and predicting the presence of clouds and their effects on the atmosphere in one of the most persistently cloudy regions of the Northern Hemisphere, namely, the midlatitudes of the North Atlantic Ocean. There now exists a great volume of literature confirming the vital role of cloudiness in both weather and climate prediction. The importance of clouds arises through the coupling of radiative, dynamical, and hydrological processes in the atmosphere through the reflection, absorption, and emission of radiation, the release of latent heat, and the redistribution of sensible and latent heat and momentum. While a substantial effort has been spent in attempting to understand the interaction of convection, the planetary boundary layer, and the large-scale flow in the tropics and subtropics, there has been no comparable effort to understand the interactions of cloudiness with the large-scale flow in midlatitude maritime regions dominated by frontal systems.

The role of clouds in climate is well understood in terms of the albedo and greenhouse effects. Arking (1991) has provided a useful review of research on the impact of clouds on the earth’s radiation balance. All recent estimates agree that the annual, global mean effect of clouds on the net flux of radiation at the top of the atmosphere is cooling when compared with a clear-sky atmosphere (dominance of the albedo effect). Any long-term change in cloudiness that results in a global change of albedo and greenhouse warming can eventually lead to a global climate change. The sensitivity of atmospheric general circulation models (GCMs) to cloudiness has been shown by Manabe and Stouffer (1980), Ramanathan et al. (1983), Roeckner and Schlese (1987), Wetherald and Manabe (1980), Cess et al. (1989), and Slingo (1990), among others.

In addition to the importance of clouds in the earth’s radiation balance, diabatic processes associated with clouds substantially modify the baroclinic waves of extratropical latitudes. The influence of clouds on short-term weather patterns arises primarily from the effect of latent heat release on the development of synoptic...
systems. In the climatological mean, latent heat release due to cloud formation over the North Atlantic during winter is about 1°–2°C day⁻¹ (Geller and Avery 1978; Schubert and Herman 1981). For comparison, it is only in the equatorial tropics that larger latent heating rates are found. Clouds thus provide a strong source of mid- and upper-tropospheric latent heating over the North Atlantic cyclone tracks. Studies of extratropical cyclones by Smith et al. (1984), Pauley and Smith (1988), and Zimmerman et al. (1989) show that latent heat release can have direct and indirect effects on the enhancement of mechanisms that lead to the further development of cyclones. Diagnostic studies (e.g., Gyakum 1983) and numerical experiments (e.g., Anthes et al. 1983; Kuo and Low-Nam 1990) point to latent heat release as a key ingredient in “explosive marine cyclogenesis.” Hoskins (1980) has shown theoretically the mechanisms through which heating could enhance baroclinic development. It has also been suggested that latent heating plays a role in the development of smaller-scale baroclinic systems (polar lows), which frequently develop to the south and east of Greenland (e.g., Rasmussen 1979).

An understanding of large-scale cloud diabatic processes and accurate representation of diabatic processes in GCMs is thus a prerequisite for accurate weather and climate predictions. The proper formulation of cloud properties and the hydrological cycle in GCMs unfortunately remains an unsolved problem. Difficulty in predicting cloud amount, height, and depth arises from lack of knowledge of the processes resulting in cloud formation and dissipation on the large scale. Since most clouds are either horizontally or vertically subgrid scale, the complicated interactions between moist convective turbulence, radiation, cloud microphysical processes, and the large-scale dynamics that result in cloud formation and dissipation must be somehow parameterized in terms of the grid-scale variables.

In this study the European Centre for Medium Range Weather Forecasts (ECMWF) initialized analyses and the U.S. Air Force 3D Nephanalysis (3DNEPH) are employed to construct a joint time series of gridpoint values of cloudiness and large-scale meteorological fields for January 1979 in the North Atlantic. Heat and moisture budgets are determined following the techniques of Yanai et al. (1973). This dataset provides a basis for the phenomenological description of the cloudiness in terms of the atmospheric forcing parameters, thus elucidating the processes contributing to the formation, maintenance, and dissipation of the clouds, and also determining such fundamental diagnostics as the vertical distribution of latent heating. Interpretation of cloud in the context of the large-scale flow is given for the monthly average situation, disturbed and undisturbed conditions, and a longitudinal cross section of a baroclinic wave. The relationship of cloud fraction to the large-scale humidity field is examined, along with several diagnostic parameterizations of cloud fraction currently employed in GCMs.

2. Data

The time period under consideration is January 1979, corresponding to the First GARP Global Experiment (FGGE) Special Observing Period I. The specific area in the North Atlantic that is examined extends from 40° to 60°N and from 10° to 50°W. Datasets employed in the analysis include the ECMWF FGGE IIIb dataset, 3DNEPH, the Nimbus-7 Scanning Multichannel Microwave Radiometer (SMMR) data, and the High-Resolution Infrared Sounder 2–Microwave Sounding Unit (HIRS2–MSU) radiances. These data were averaged daily on a 1.875° latitude and 1.875° longitude grid at 0000 and 1200 UTC, coincident with the ECMWF analyses.

The region and period that we have chosen to examine has probably the best conventional data coverage of any oceanic region in the world (Bjorheim et al. 1981). Analysis of the distribution of data used in the ECMWF analyses gives an average for the region of 82 and 100 surface ship observations at 0000 UTC and 1200 UTC, respectively (Bjorheim et al. 1981). Aircraft reports (utilized in both the 3DNEPH and ECMWF) are abundant, but appear to vary with the day of the week (Bjorheim et al. 1981). There are six radiosondes from stationary weather ships and from islands in this region.

a. Atmospheric dynamic and thermodynamic data

The ECMWF four-dimensional data assimilation scheme utilized in the “main” (1980/81) FGGE level IIIb analyses is described by Bengtsson et al. (1982). It is an intermittent data assimilation system consisting of a multivariate optimum interpolation analysis, a nonlinear normal-mode initialization, and a high-resolution model, which produces a first-guess forecast for the subsequent analysis. The analysis consists of two parts: one for the simultaneous analysis of surface pressure, geopotential height, and horizontal wind, and another part for analysis of humidity. Hydrostatic balance is achieved through conversion of temperature observations into thicknesses prior to the assimilation. The level IIIb dataset contains both basic analysis parameters as well as derived parameters. The basic parameters are uninitialized and consist of geopotential height, sea level pressure, and horizontal wind components. The derived parameters are temperature, relative humidity, and vertical velocity. The relative humidity is determined from the precipitable water in each analysis layer (the basic analysis parameter for humidity) and the temperature. The vertical velocity is calculated from initialized divergences.
The "final" (1985/86) ECMWF FGGE analyses, which are utilized in this study, contained numerous changes to the assimilation scheme and also contained corrected and additional data (Pailleaux 1986). The final dataset contained additional data from the regional experiments conducted during FGGE, additional ship data, higher-density satellite temperature and humidity retrievals, and cloud-track winds. Changes in the mass and wind analysis algorithm are described by Shaw et al. (1987). The main changes to the humidity analyses are discussed by Illari (1985) and Pasch and Illari (1985). The incorporation of diabatic effects in the normal-mode initialization are discussed in Wergen (1988). The assimilating model was a T63 spectral model (Girard and Jarraud 1982), and the physical parameterizations are described by Tiedke et al. (1988). The Arpe (1985a,b) showed that the final analyses agreed better with observational data than the main analyses.

Both the initialized and uninitialized versions of the final ECMWF FGGE IIIb analyses were examined. Although the initialized analyses may possess dampened divergences, a number of other factors are more attractive in the initialized analyses when compared with the uninitialized analyses. A comparison was therefore made between the final initialized and uninitialized analyses on a gridpoint-by-gridpoint basis for the region and time period under consideration.

For horizontal velocity components, the average absolute difference between the initialized and uninitialized gridpoint values was 0.5 m s\(^{-1}\), although there was no systematic difference. Slightly higher differences were found at the highest levels, with the maximum difference being 4 m s\(^{-1}\). Derived vorticity values differed by an average of less than 10%. Differences in the derived divergences were larger, averaging 30%, but it did not appear that the initialized divergences were systematically smaller in magnitude. The divergences of the two datasets had a correlation coefficient of 0.9.

Vertical velocities for the uninitialized analyses were derived using the method of O'Brien (1970), while vertical velocities for the initialized analyses were a byproduct of the diabatic nonlinear normal-mode initialization procedure. A comparison of the derived uninitialized vertical velocities with the initialized values showed that the uninitialized vertical velocities were systematically smaller than the initialized values, with little correlation and substantially different vertical-velocity profiles.

The uninitialized temperatures were derived from the geopotential fields using the hydrostatic equation. The initialized temperatures were systematically higher than the uninitialized temperatures. The average absolute difference was 1°C and the maximum difference was 5°C, the largest differences occurring at lower levels. The uninitialized mixing ratios were obtained from the layer precipitable water values after Lorenc and Tibaldi (1980). A comparison of the initialized with the uninitialized values showed that the 1000- and 850-mb initialized values were systematically ~20% higher than the uninitialized values. Above 500 mb the initialized values were systematically ~20% lower than the uninitialized values. The two datasets therefore have markedly different vertical profiles of humidity. It is presumed that the use of the vertical structure functions (Uppala 1986) in the derivation of the initialized mixing ratios from the precipitable water values accounts for this difference.

In addition to the primary meteorological quantities, higher-order terms from the heat and moisture budgets (see section 3) were compared. The largest differences were associated with those budget terms that contained the vertical velocity. In spite of rather large random differences, the systematic differences for the mean monthly budget residuals were less than 10%.

In light of the aforementioned comparisons, the initialized analyses have been chosen for this study. Specific reasons for choosing the initialized analyses include the following: improved vertical structure of the temperature and humidity fields, while spurious divergences have been eliminated, the initialized divergences do not appear to be damped; the initialization is generally believed to produce more realistic extra-tropical vertical velocities (e.g., Savijärvi 1983); and the initialized analysis fulfills the integral constraints and provides a dynamically consistent dataset. However, the overall quality of the humidity fields remains uncertain.

The fields obtained from the ECMWF initialized analyses and employed in this study are temperature \(T\), absolute humidity \(q\), height \(z\), zonal velocity \(u\), meridional velocity \(v\), and vertical \(p\) velocity \(\omega\), at levels 1000, 850, 700, 500, 400, 300, and 200 mb, with a horizontal resolution of 1.875° in both latitude and longitude.

b. Cloud-cover data

The cloud-cover data used in this study are the U.S. Air Force 3DNEPH, which is described by Fye (1978). The 3DNEPH provides total cloud cover, cloud amount at 15 levels, heights of the lowest bases and highest tops, cloud types, and significant weather, with a horizontal spatial resolution of about 44 km at 60°N and a nominal temporal resolution of 3 h. The 3DNEPH is a blend of satellite (threshold analysis) and conventional observations; the conventional observations are obtained from surface, aircraft, and radiosonde reports. An assessment of the quality of the 3DNEPH for this region and period is given by Tian and Curry (1989). Because of the density of conventional observations in this region, the 3DNEPH provides a reasonable analysis of layer cloud amounts.

The 15 layers of 3DNEPH were assigned to the nearest layers of the ECMWF analyses, which contains 7 layers. Each of the lowest four ECMWF layers con-
sists of at least two 3DNEPH layers. An algorithm based on random and maximum overlap assumption is employed to determine the cloud fraction for each of these ECMWF layers (Tian and Curry 1989).

c. Precipitation data

The microwave brightness temperatures measured by the Nimbus-7 SMMR, described by Gloersen and Barath (1977), are employed in the determination of precipitation rates. Because of the relatively narrow swath of observations and due to the fact that the instrument was turned off every other day, there is not full data coverage for the entire region and period under analysis. Precipitation rates were determined after Curry et al. (1990), which employed the 37-GHz vertically and horizontally polarized brightness temperatures.

d. Sea surface temperatures

Sea surface temperatures were determined after Susskind and Reuter (1985) using the HIRS2-MSU satellite data. The temperatures were determined on a horizontal resolution of 2.5° latitude × 2.5° longitude; values were interpolated onto the ECMWF grid spacing. Mean monthly temperatures at each grid point were used.

3. Method of analysis

The sensible heat and moisture budgets per unit mass of air are derived following the bulk diagnostic method proposed by Yanai et al. (1973), with the exception that the heat budget equation is written in terms of enthalpy \( c_p T \) instead of dry static energy:

\[
\nabla \cdot \mathbf{q} + \frac{\partial \dot{\omega}}{\partial p} = 0
\]

\[
\frac{\partial}{\partial t} (c_p \bar{T}) + \nabla \cdot \left( c_p \mathbf{v} \bar{T} \right) + \left( \frac{\partial}{\partial p} - \frac{R}{c_p p} \right) (c_p \bar{\mathbf{v}} T) = Q_R + L(c - e) \tag{2}
\]

\[
\frac{\partial \bar{q}}{\partial t} + \nabla \cdot \bar{\mathbf{v}} q + \frac{\partial \bar{\omega} q}{\partial p} = e - c. \tag{3}
\]

Most of the notations bear the conventional meanings, with \( Q_R \) representing the net radiative heating, \( c \) the rate of condensation per unit mass, \( e \) the evaporation rate, \( p \) pressure, \( L \) latent heat of condensation, \( c_p \) specific heat at constant pressure, and \( \nabla \cdot \mathbf{v} \) the horizontal divergence operator on a constant-pressure surface. The overbar denotes the horizontal average over the grid box, for example, \( \bar{\cdot} = A^{-1} \int \int \left( \cdot \right) a^2 \cos \phi d\lambda d\phi \), where \( A \) is the grid-box area, \( a \) the radius of the earth, \( \lambda \) longitude, and \( \phi \) latitude. All of the terms on the left-hand side of (1)–(3) can be evaluated using the gridded fields from the ECMWF dataset.

Terms on the right-hand side of (2)–(3) represent the source terms, which cannot be directly evaluated using the gridded fields from the ECMWF IIb dataset. Equations (1)–(3) may be rearranged to give

\[
Q_1 = \frac{\partial}{\partial t} (c_p \bar{T}) + \nabla \cdot \left( c_p \bar{\mathbf{v}} T \right) + \left( \frac{\partial}{\partial p} - \frac{R}{c_p p} \right) (c_p \bar{\omega} T) - \nabla \cdot \frac{\partial}{\partial p} (c_p \bar{\omega} T) + Q_R + L(c - e) \tag{4}
\]

\[
Q_2 = \frac{\partial \bar{q}}{\partial t} + \nabla \cdot \bar{\mathbf{v}} q + \frac{\partial \bar{\omega} q}{\partial p} = -\nabla \cdot \bar{\omega} q' - \frac{\partial}{\partial p} \bar{\omega} q' + e - c. \tag{5}
\]

The cross-correlation terms in (4) and (5) represent the turbulent eddy fluxes ranging in scale from convective and cloud-induced eddies to other mesoscale circulations. The terms \( Q_1 \) and \( Q_2 \) denote the apparent sensible heat moisture sources, respectively, and can be determined as the residual from the grid-scale variables. The residual heating is due to radiation, latent heat release, and the divergence of the eddy heat flux; the residual moisture source is due to condensation, and to the divergence of the vertical eddy moisture flux. The divergences of the horizontal eddy heat and moisture fluxes are small (e.g., Pedigo and Vincent 1990). In evaluating \( Q_1 \) and \( Q_2 \) as residuals, it must be kept in mind that the residuals may also contain contributions from errors that are associated with the analysis or with the finite-differencing schemes.

Assuming that the horizontal subgrid-scale eddy flux is small (after Pedigo and Vincent 1990), the vertical eddy flux of total heat \( F \) is obtained by combining (4) and (5):

\[
F = \frac{1}{g} \int_{p_T}^{p_0} (Q_1 + LQ_2 - Q_R) dp. \tag{6}
\]

By integrating (4) and (5) from the surface \( (p_0) \) to 200 mb \( (p_T) \), we obtain (after Yanai et al. 1973)

\[
\frac{1}{g} \int_{p_T}^{p_0} (Q_1 - Q_R) dp = LP_0 + S_0 \tag{7}
\]

\[
\frac{1}{g} \int_{p_T}^{p_0} LQ_2 dp = L(P_0 - E_0) \tag{8}
\]

\[
\frac{1}{g} \int_{p_T}^{p_0} (Q_1 + LQ_2 - Q_R) dp = LE_0 + S_0 \tag{9}
\]

where \( S_0 \) is the supply of sensible heat from the surface, \( E_0 \) is the rate of evaporation from the sea surface, and \( P_0 \) is the amount of precipitation received at the sea surface. Therefore, the vertical integrals of the large-
scale heat and moisture equations yield information on the surface precipitation, evaporation, and sensible heat flux as a lower boundary condition. When compared with independent observations or estimates of \( S_0, E_0, \) and \( P_0, \) (7)–(9) can be used to check the accuracy of the estimates of \( Q_1, Q_2, \) and \( Q_R. \)

### a. Radiative heating rates

The net radiative heating rate \( Q_R \) is the sum of the solar (SW) heating rate and longwave (LW) cooling rates. Both the LW and SW heating rates are calculated after Harshvardhan et al. (1987). For SW, ozone and water vapor absorption bands are considered, and Rayleigh scattering is included. Multiple scattering from clouds is treated using a two-stream delta-Ed-dington approximation to provide fluxes for a range of conditions varying continuously from a clear sky to cloudy sky of arbitrary optical depth. Longwave radiation is calculated using an emissivity approach over six water vapor bands, three carbon dioxide bands, and a single ozone band; water vapor continuum absorption is also included. The parameterization allows for fractional cloud cover and random or maximum overlap. A computational scheme utilizing the probability of a clear line of sight between each layer and all other layers, the ground, and the top of the atmosphere is used to treat the radiative transfer in the presence of clouds. The only assumptions required are that clouds do not reflect in the longwave, and that they fill the model layer in the vertical. Nonblack clouds are assigned a random overlap cloud fraction equal to the emissivity.

Radiative heating rates are calculated for each grid point and time using the ECMWF temperatures and humidities and the 3DNEPH cloudiness. Above the 200-mb level, cloud fraction is assumed zero, and temperature and water vapor mixing ratio are taken from McClatchey et al. (1972) for midlatitude winter, which also supplies the vertical distribution of ozone concentration for the model.

### b. Precipitation rates

The SMMR-derived precipitation rates are retrieved using the algorithm described by Curry et al. (1990) in which the horizontal and vertical brightness temperatures at 37 GHz are used. The algorithm requires that a threshold brightness temperature be chosen for the onset of rain. The threshold value is determined according to the criterion \( P^* < 0.8, \) where \( P^* \) is defined by Petty and Katsaros (1990) as

\[
P^* = \frac{T_V - T_H}{T_{V_{CLR}} - T_{H_{CLR}}},
\]

where \( T_V \) and \( T_H \) are vertically and horizontally polarized brightness temperatures at 37 GHz, respectively, and the subscript CLR refers to clear sky conditions.

The rain rates \( P_0 \) are then calculated using the following algorithm:

\[
P_0 \text{ (mm h}^{-1}\text{)} = \left( \frac{T_H - C}{3.3} \right)
\]

where \( C \) is determined to be

\[
C = T_V - 0.8 \sigma (T_{V_{CLR}} - T_{V_{CLR}}).
\]

### c. Surface heat fluxes

Surface latent and sensible heat fluxes are obtained by using bulk aerodynamic properties. The parameterization scheme introduced by Louis (1979) relates Monin–Obukhov scale height to bulk Richardson number. The surface sensible heat flux is evaluated using the following expression:

\[
\theta_s = \frac{a^2}{R} u \Delta \theta F \left( \frac{z}{z_0}, R_i \right)
\]

where \( R \) is the ratio of the drag coefficients for momentum and heat in the neutral limit, \( a \) is the drag coefficient in neutral conditions, \( u \) is the wind speed, \( z_0 \) is the surface roughness length, and \( R_i \) is the bulk Richardson number. Difficulty is encountered when dealing with cases of \( u = 0, \) since \( R_i \) becomes infinite. For stable cases, the vertical heat flux approaches zero when wind speed is zero. For unstable cases (\( \Delta \theta > 0 \)), the heat flux can be obtained from

\[
\theta_s = \frac{a^2 b}{R c} \left( \frac{g z}{\theta} \right) \left( \Delta \theta \right) \left| \Delta \theta \right|^{1/2}
\]

where \( b \) and \( c \) are constants (see Louis 1979). Sensible heat flux can be obtained by multiplying (13) or (14) by \( -\rho c_p T \), where \( \rho \) and \( c_p \) represent air density and specific heat capacity at constant pressure, respectively. An analogous approach is employed to derive the surface moisture flux.

### 4. Monthly mean conditions

During this period, a subtropical high with small variation in position dominated the southern portion of this region. To the north was a region of strong baroclinic instability, characterized by a succession of frontal systems.

The mean cloud characteristics for this region and period are described by Tian (1988), Tian and Curry (1989), and Curry et al. (1990). The average total cloud cover of this region and period was 64%.

The period of our analysis was dominated by layered clouds, with cumuliform cloud types being reported for 10% of the observations and contributing little to the total cloud cover. The most frequent cloud types occurring in this region were stratus and stratocumulus, with a
combined frequency occurrence of 66%. The frequency of cases with completely clear sky was about 10%.

Figure 1 shows the vertical profiles of the areal-averaged monthly mean quantities obtained from the ECMWF analyses for January 1979 in the region of interest. The monthly mean temperature and absolute humidity profiles show typical decreases with height. Relative humidity is highest at the low levels, with a secondary maximum at 400 mb.

The areally averaged mean monthly \( U \)-component wind velocity is westerly, the magnitude increasing nearly linearly with a decrease in pressure from a nearly zero value at 1000 mb. The mean monthly \( V \)-component wind velocity is nearly zero at all levels, although the instantaneous \( V \) velocity is not necessarily small, the average of positive and negative values being responsible for the nearly zero mean values. The monthly averaged vertical \( p \)-velocity profile shows downward motion at each level, with maximum velocity occurring at the 400-mb level.

\[ a. \textbf{Heat and moisture budgets} \]

Figure 2 shows the mean monthly heat budget, averaged over the entire region. The terms in the heat budget equation (2) are expressed in units of degrees Celsius per day. The most significant term in the heat budget at lower levels is the horizontal cold-air advection, which is nearly balanced by the large-scale residual heat source. Above 500 mb, the adiabatic warming and vertical transport terms become increasingly dominant, and the horizontal transport becomes less significant. The residual term \( Q_1 \) (which includes radiative heating, latent heating, and vertical eddy heat flux divergence) becomes negative above 500 mb. The positive value of the mean local rate of change at each level signals a warming trend during this season, although the magnitude is small compared to most of the other terms.

Figure 3 shows the monthly mean moisture budget averaged over the entire region. The terms in the moisture budget equation (3) are expressed in units of grams
of water vapor per cubic meter of air per day. Analogous to the heat budget, the horizontal moisture advection term is dominant at lower levels and is nearly balanced by a large-scale residual moisture source. The moisture source \(Q_2\) is associated with evaporation of cloud drops and the vertical eddy moisture flux divergence. Positive vertical moisture advection dominates at the upper levels and is offset by the large-scale residual moisture sink.

It is seen from Fig. 4 that the net radiative heating \(Q_R\) is dominated by the longwave cooling, resulting in net radiative cooling at each level. The local maxima at 400 mb in both shortwave and longwave heating reflect the relatively sharp decrease in cloud fraction (Fig. 7) and cloud optical thickness between 500 and 400 mb. Comparison of monthly mean vertical profiles of \((1/c_p)(Q_1 - Q_R)\) and \((L/c_p)Q_2\) is shown in Fig.

5. At low levels (below 800 mb) there is a positive heat and moisture source, associated with the transport of sensible and latent heat from the surface. Since the average vertical velocity at these levels is downward, it appears that the contributions to the \(Q_1\) and \(Q_2\) budgets are dominated by the periods of cloud formation, which presumably are principally associated with episodes of rising motion. The negative moisture source and positive value of \((1/c_p)(Q_1 - Q_R)\) at levels between 800
and 300 mb are consistent with large-scale condensation occurring at these levels. At 400 mb, it is not clear why \((1/c_p)(Q_1 - Q_R)\) is not balanced by the moisture sink, as would be expected if condensation is the only contributing physical process. Above the 400-mb level the heat and moisture sources are nearly zero.

A comparison of these heat and moisture budgets can be made with those determined for other geographical regions, specifically the summertime Arctic Ocean (Curry and Herman 1985) and the tropical ocean (e.g., Yanai et al. 1973). Over the Arctic Ocean the dominant features are warm-air advection and a residual heat sink at all levels. These features represent warm-air advection over a cold surface. There is little moisture from the surface, the moisture sink at all levels associated with condensation. The Arctic Ocean heat and moisture budgets are clearly different from the situation in this study, where the underlying surface is warm and the low levels are dominated by cold-air advection. The tropical ocean heat and moisture budgets, as typified by the study of Yanai et al. (1973), also present different characteristics in the heat and moisture budgets. Strong convective activity in tropical regions results in a heat source and a moisture sink associated with condensation, having a maximum between 500 and 800 mb. In the present study, the maximum values of \(Q_1\) and \(Q_2\) occurred lower in the atmosphere, near the surface.

According to Eq. (6), the vertical eddy heat transport at each level can be obtained by integrating the quantity \(Q_1 + LQ_2 - Q_R\) from pressure \(p\) to the top of the model, assuming the vertical eddy heat transport at 200 mb is zero. The vertical profile of the mean monthly vertical eddy transport shown in Fig. 6 is seen to decrease sharply with height. This is in contrast to the result of Yanai et al. (1973) for the tropics, which shows a strong secondary maxima at 700 mb. Notice that the value of \(F_0\) in the present study is even slightly larger than the tropical value determined by Yanai et al. (1973).

### Table 1. Comparison of surface sensible and latent heat fluxes and precipitation determined from the heat and moisture budgets with independently determined values. Also shown are the correlation coefficients between the budget-derived and independently derived values, and the values determined by Yanai et al. (1973) for the tropics.

<table>
<thead>
<tr>
<th></th>
<th>Budget</th>
<th>Independent</th>
<th>Correlation</th>
<th>Yanai</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sensible heat (W m(^{-2}))</td>
<td>60</td>
<td>62</td>
<td>0.33</td>
<td>14</td>
</tr>
<tr>
<td>Latent heat (W m(^{-2}))</td>
<td>104</td>
<td>108</td>
<td>0.30</td>
<td>188</td>
</tr>
<tr>
<td>Precipitation (W m(^{-2}))</td>
<td>107</td>
<td>73</td>
<td>0.41</td>
<td>526</td>
</tr>
</tbody>
</table>

**b. Calibration of the large-scale budgets**

Equations (7) and (8) can be used to derive values of the surface sensible heat flux and the surface precipitation, provided that a value is available for the Bowen ratio, which is defined to be the ratio of the sensible to latent heat flux. The Bowen ratio was determined to be 0.58 from these data. Independent values of the surface latent and sensible heat fluxes may also be obtained from (13) or (14), and precipitation may be determined from the SMMR data using (11). By comparing the budget-derived values of \(S_0\), \(LE_0\), and \(LP_0\) with the independently derived values, an assessment of the consistency of the heat and moisture budgets can be made.

Table 1 compares the budget-derived values of \(S_0\), \(LE_0\), and \(LP_0\) with the independently derived values (note that the budget-derived value of \(E_0\) has been implicitly determined by the assumed Bowen ratio). Comparison of the monthly averaged values of \(LE_0\) and \(S_0\) determined by the two different methods agrees within 4%. Mean monthly precipitation values are seen to differ by 30%. Although the relative errors of budget-derived and independently derived values are uncertain, this comparison indicates that \((Q_1 - Q_R)\) and \(Q_2\) show a systematic error of less than a factor of 1.5.

Comparison of \(S_0\), \(LE_0\), and \(LP_0\) for the North Atlantic with values presented by Yanai et al. (1973) for the tropics shows several differences. The Bowen ratio for the tropics was determined to be 0.076 by Yanai et al., with the sensible heat flux being substantially smaller for the tropics than for the midlatitudes. In the tropics, \(S_0\) can be neglected with respect to \(LP_0\) and \(LE_0\); this is clearly not the case in the midlatitudes. In
the tropics, $L(P_0 - E_0)$ is 338 W m$^{-2}$, while the corresponding value in the midlatitudes is nearly zero.

Also shown in Table 1 is the correlation coefficient between the budget values and independent values of $S_0$, $LE_0$, and $LP_0$ for the twice-daily gridpoint values. The correlations are not seen to be particularly high, although all are significantly different from zero at the 99% confidence level. This indicates that while the mean monthly budgets show reasonable consistency, there is a larger random error for the instantaneous gridpoint values.

c. Interpretation of mean cloud cover

The vertical profile of monthly mean cloud cover shown in Fig. 7 has a maximum of 40% at the 850-mb level, and the layer cloud cover decreases to below 10% above 400 mb. A comparison of Fig. 7 with the vertical profile of relative humidity shown in Fig. 1 indicates that there is no simple diagnostic relationship between mean monthly cloud fraction for a layer and the mean layer relative humidity. An interpretation of the vertical profile of mean monthly cloud cover can be made by concurrent examination of the vertical profiles of the heat and moisture budgets, shown in Figs. 2 and 3.

While the three-dimensional moisture convergence drops to zero rapidly at high levels, the three-dimensional heat convergence becomes relatively large and contributes to three-dimensional relative humidity divergence (see Curry and Herman 1985). The majority of heat convergence at high levels comes from adiabatic warming, which results in the dissipation of clouds. Between 800 and 500 mb, there is three-dimensional heat flux divergence plus three-dimensional moisture convergence, resulting in relative humidity convergence in this layer; this is a layer of relatively high cloudiness.

Below 800 mb there is little large-scale relative humidity convergence, the cloudiness associated principally with turbulent transport of moisture from the surface. Despite the fact that maximum value of monthly mean relative humidity occurs at 1000 mb, the monthly mean cloudiness peaks at the 850-mb level. The failure to account for the high cloudiness at the 850-mb level by relative humidity can be attributed to the role of other boundary-layer characteristics, particularly the static stability. The more stable the layer, the more likely it is that the moisture released from the surface will be trapped in the layer. The stability parameter $\Theta / \Theta_p$ is calculated for both the 1000- and 850-mb levels, with resulting values of $-0.0228$ and $-0.0420$ K mb$^{-1}$, respectively. The static stability at 850 mb is seen to be almost twice as large as at 1000 mb. The moisture transported out of 1000 mb is apparently trapped in the 850-mb layer, so that the mean cloud fraction is larger at 850 mb than at 1000 mb.

In general, middle clouds are formed primarily due to the benefit from large-scale three-dimensional moisture convergence; and low-cloud formation depends on surface moisture flux and the static stability.

5. Relation of budgets and cloud characteristics to the synoptic situation

Analysis of the mean monthly budgets concurrently with cloud observations and surface precipitation, latent and sensible heat fluxes allows the gross consistency of the monthly heat and moisture budgets to be assessed and an interpretation of the mean monthly cloud cover to be made. However, hidden in the mean values are a series of baroclinic waves with vastly different heat and moisture budget characteristics in different portions of the wave. In this section, the heat and moisture budgets for different synoptic situations are examined.

Figure 8 compares the mean heat and moisture budgets for surface anticyclonic and cyclonic conditions. The criterion for selecting a grid point as corresponding to a surface anticyclone/cyclone was simply that the 1000-mb vorticity be in the lowest/top 20%. Shown in Fig. 9 are the vertical profiles of mean layer cloud cover for the corresponding anticyclonic–cyclonic conditions. It is seen that cloud fraction is substantially greater under cyclonic conditions at all levels except at 200 mb, where cloud fraction is very low in both situations.

Comparison of the heat and moisture budgets for the anticyclonic and cyclonic situations in Fig. 8 shows
that all of the budget terms are of opposite sign at nearly all model levels. In comparing the heat budgets, it is seen that the three-dimensional heat flux convergence term is positive for the anticyclone and negative for the cyclone, the adiabatic warming term being the individual term of largest magnitude in both cases. The three-dimensional moisture flux convergence term is positive at all levels for the cyclonic cases, while for the anticyclonic cases the three-dimensional moisture flux convergence is negative below 500 mb with small, slightly higher values at higher levels. For the cyclonic situations the residual heat source is positive at all levels except 400 mb (reflecting large radiative cooling rate at this level), and there is a moisture sink at all levels, the maxima of both terms occurring at 850 mb. These residual terms are consistent with large-scale condensation occurring at all levels during disturbed conditions. By contrast, in the anticyclonic cases, $Q_2$ is positive at all levels, and only at 1000 and 850 mb is there evidence of large-scale condensation, accompanied by a moisture flux from the surface.

It is also of interest to examine the heat and moisture budgets in a longitudinal cross section through a baroclinic wave. Figure 10 shows the surface pressure field and the 500-mb heights for 1200 UTC 3 January, the region of interest lying within the box. Figure 11 represents a longitudinal cross section at 46.875°N of the meteorological fields across the baroclinic wave. The relative humidity field is closely tied to the vertical velocities, the driest regions associated with the strongest sinking motion to the west of the upper-air trough, and in the region of the surface cold-air advection. Comparison of the relative humidity field with the cloud cover shows that peak cloud fractions are displaced approximately 3.5° longitude to the west of the peak relative humidities. The highest cloud amounts, particularly near −19°, correspond to low local relative humidities, particularly above 800 mb. It is suggested from examining this cross section that the clouds do not respond instantaneously to the large-scale relative humidity field, but take a period of time on the order of hours to adjust to relative humidity field in terms
of evaporation and condensation. This is further supported by examining the $Q_1$ and $Q_2$ fields shown in Fig. 11. As the cloud fields progress from west to east, there is a moisture source as the cloud evaporates on the westward side and a moisture sink as condensation occurs on the eastward side.

6. Cloud diagnostic relationships

Central to all cloud parameterizations in GCMs, be they diagnostic or prognostic, is the determination of a grid-scale threshold relative humidity below which cloud does not occur. It is observed that cloud formation occurs well before the large-scale relative humidity reaches 100%, which is believed to be associated with subgrid-scale fluctuations in relative humidity. Values for the threshold relative humidity $\text{RH}_c$ that have been employed in GCMs include: 100% (Somerville et al. 1974); 97% (Wetherald and Manabe 1980); and 80% (Sundqvist 1978). More complex specifications have been made, whereby the value of $\text{RH}_c$ varies with height in the model: 80% for low and high clouds; 65% for midclouds (Slingo 1987), from 38% to 56% (Mitchell and Hahn 1990); 92.5% for the lowest two model levels and 85% for higher levels (Smith 1990); and polynomial function of pressure, with $\text{RH}_c$ varying from 97% at 1000 mb, reaching a minima of 43% at 700 mb, and increasing 85% to 200 mb (Geleyn 1981). Mitchell and Hahn (1990) have pointed out the dependence of $\text{RH}_c$ on the characteristics of the model initialization and physics, since significant errors seem possible in model moisture fields.

To determine $\text{RH}_c$, a scatter diagram of observed cloudiness against analyzed relative humidity can, in principle, be employed. It is, however, difficult to determine $\text{RH}_c$ from Fig. 12 (note that relative humidities at levels 500 mb and above are determined with respect to ice, since temperatures at these levels are generally below $-30^\circ\text{C}$). The relationship between fractional cloud cover and model relative humidity deteriorates with height; at 200 and 300 mb there is in fact a negative correlation between cloud fraction and relative humidity, and at 200 mb no clouds occur at relative humidities (with respect to ice) exceeding 65%. It would

![Fig. 9. Comparison of cloud cover for cyclonic (solid) with anticyclonic (dash) situations.](image)

![Fig. 10. Surface pressure field and 500-mb heights for 1200 UTC 3 January. Region analyzed here is indicated by the box (40°–60°N, 10°–50°W).](image)
appear that this relationship at 1000 mb is so good because of the fairly large number of surface (ship) observations in this region. At higher levels, the analyzed moisture field becomes increasingly dependent on the model vertical transport of moisture (i.e., the convective parameterization). From this analysis it
would appear that the model is not transporting sufficient moisture to high levels.

An approach to determine the threshold relative humidity, taking into account inadequacies in the analyzed relative humidity field, was recently proposed by Mitchell and Hahn (1990) in which the cumulative frequencies of observed cloud fraction and analyzed grid-scale relative humidity are used. The point of using
Table 2. Threshold relative humidity determined after Mitchell and Hahn (1990) and the exponent in cloud fraction prediction equation (15).

<table>
<thead>
<tr>
<th>p (mb)</th>
<th>Threshold relative humidity</th>
<th>Exponent</th>
</tr>
</thead>
<tbody>
<tr>
<td>1000</td>
<td>73</td>
<td>2.8</td>
</tr>
<tr>
<td>850</td>
<td>42</td>
<td>3.6</td>
</tr>
<tr>
<td>700</td>
<td>29</td>
<td>2.2</td>
</tr>
<tr>
<td>500</td>
<td>51*</td>
<td>3.0</td>
</tr>
<tr>
<td>400</td>
<td>83*</td>
<td>2.8</td>
</tr>
<tr>
<td>300</td>
<td>46*</td>
<td>1.5</td>
</tr>
<tr>
<td>200</td>
<td>22*</td>
<td>2.3</td>
</tr>
</tbody>
</table>

* Asterisks refer to relative humidity with respect to ice.

tained from the projection of the frequency distribution of the cloud analysis onto that of the RH analysis. The best-fit exponent is determined to lie between 1.5 and 3.6 at different levels, as shown in Table 2.

Smith (1990; appendix C) developed an expression for cloud fraction in terms of relative humidity using the cloud distribution concepts developed by Sommeria and Deardorff (1977) and Mellor (1977). The parameterization takes the following form:

$$ CF = 4 \cos^2 \left( \frac{\Pi}{3} + \frac{\vartheta}{3} \right) $$

where

$$ \vartheta = \cos^{-1} \left[ \frac{3}{2\sqrt{2}} \left( \frac{RH - RH_c}{1 - RH_c} \right) \right] $$

for $RH_c < RH < (5 + RH_c)/6$.

$$ CF = 1 - \left[ \frac{3}{2\sqrt{2}} \left( \frac{1 - RH}{1 - RH_c} \right) \right]^{2/3} $$

for $(5 + RH_c)/6 \leq RH \leq 1$. (16)

Table 3 gives a comparison of the observed mean monthly layer cloudiness and the predicted cloud amount at each level, using the Slingo (1987), Mitchell and Hahn (1990), and Smith (1990) parameterizations. In calculations using the Mitchell and Hahn (1990) and Smith (1990) parameterizations, the values of RHc in Table 2 were employed; for the Slingo (1987) parameterization, the values of RHc reported in that study were used (note that at 1000 mb the stability adjustment is not employed here in using the Slingo parameterization). Total cloud amount was determined from the parameterizations by assuming maximum overlap for adjacent layers, and random overlap for cloud layers separated by a clear layer. In parentheses is shown the coefficient of correlation between the parameterized and observed cloud fractions, using twice-daily gridpoint values. In terms of mean monthly layer cloud fraction, the Mitchell and Hahn (1990) and Smith (1990) values agree significantly better with

Table 3. Comparison of observed and predicted monthly mean cloud fractions (correlation coefficients of observed gridpoint values with predicted values are given in parentheses; * indicates correlation is not significantly different from zero at the 95% confidence level).

<table>
<thead>
<tr>
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<tbody>
<tr>
<td>1000</td>
<td>26</td>
<td>22 (0.032)</td>
<td>23 (0.031)</td>
<td>26 (0.037)</td>
</tr>
<tr>
<td>850</td>
<td>40</td>
<td>15 (0.183)</td>
<td>36 (0.222)</td>
<td>37 (0.223)</td>
</tr>
<tr>
<td>700</td>
<td>29</td>
<td>12 (0.311)</td>
<td>30 (0.399)</td>
<td>21 (0.400)</td>
</tr>
<tr>
<td>500</td>
<td>18</td>
<td>15 (0.370)</td>
<td>22 (0.415)</td>
<td>21 (0.413)</td>
</tr>
<tr>
<td>400</td>
<td>7</td>
<td>15 (0.236)</td>
<td>11 (0.209)</td>
<td>9 (0.200)</td>
</tr>
<tr>
<td>300</td>
<td>5</td>
<td>0 (0.113)</td>
<td>5 (0.221)</td>
<td>2 (0.212)</td>
</tr>
<tr>
<td>200</td>
<td>2</td>
<td>0 (0.000)*</td>
<td>2 (0.044)</td>
<td>1 (0.046)</td>
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</table>

Total cloud cover

<table>
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<tr>
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<tbody>
<tr>
<td>65</td>
<td>55 (0.350)</td>
<td>66 (0.378)</td>
<td>62 (0.382)</td>
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observations than does the Slingo (1987) parameterization, particularly at the 850- and 700-mb levels. Total cloud fraction is best parameterized by Mitchell and Hahn, with the Smith parameterization performing adequately. The correlation of the observed with the parameterized layer cloud fractions for twice-daily gridpoint values is generally poor, particularly at 1000 mb, although the correlations using the parameterizations of Mitchell and Hahn and Smith are slightly larger than those using the Slingo parameterization.

From these results, it is seen that appropriate tuning of a diagnostic relative humidity–based parameterization can result in accurate parameterized mean monthly total cloud amount for the region, and layer cloud fractions to within 5% of observed layer cloud fractions. However, this type of cloud fraction parameterization, as indicated by the low correlations, appears to be unable to diagnose layer cloud fraction on the smaller time and space scales that are undoubtedly required for obtaining the correct local cloud radiative and hydrological feedbacks with the dynamics.

b. Other diagnostic relationships

In the previous discussion, it was shown that diagnostic parameterizations based solely on relative humidity are not very successful for daily gridpoint values. Since large-scale layer clouds can also be related to dynamical processes, we examine here whether gridpoint layer cloudiness is diagnostically related to vertical velocity, moisture and relative humidity transport, height tendency, static stability, and surface heat and moisture flux for low clouds.

Table 4 gives the correlation of layer cloud fraction with vertical velocity. It is seen from the table that the correlation of cloud fraction with vertical velocity is higher even than those for relative humidity as shown in Table 4, particularly for levels between 700 and 300 mb. Correlations of cloud fraction with height tendency, horizontal moisture advection, and three-dimensional relative humidity convergence are all insignificantly different from zero at the 99% confidence level and are not shown here.

Table 5 gives the correlation of the 1000- and 850-mb layer cloud fraction with static stability and surface sensible and latent heat fluxes. Particularly at 1000 mb, correlations with the sensible and latent heat fluxes are seen to be larger than any other parameter that has been tested. All correlations in Table 5 are significantly different from zero at the 99% confidence level.

7. Summary and conclusions

This study has examined the relationships between gridscale cloudiness and the large-scale environment in the North Atlantic. The ECMWF initialized analyses and the U.S. Air Force 3DNEPH were employed to construct a joint time series of gridpoint values of cloudiness and large-scale meteorological fields for January 1979. Heat and moisture budgets were determined following the techniques of Yanai et al. (1973). Observations of precipitation using satellite microwave measurements and calculations of surface sensible and latent heat fluxes by bulk aerodynamic method provided a check on the accuracy of the heat and moisture budgets. It was determined that the mean monthly values of $Q_s$ and $Q_r$ were correct to within a factor of 1.5, although individual gridpoint values may have a larger error. The Bowen ratio and value of $L(P_b - E_b)$ determined for this North Atlantic dataset showed marked differences to values determined for the tropics (e.g., Yanai et al. 1973). The North Atlantic Bowen ratio was determined here to be 0.58, compared to a tropical value of 0.076. In the tropics, $L(P_b - E_b)$ was 338 W m$^{-2}$, compared to a value of nearly zero for this dataset.

The most significant term in the mean monthly heat budget at lower levels was the horizontal cold-air advection, which was nearly balanced by the large-scale residual heat source. Above 500 mb, the adiabatic warming and vertical transport terms became increasingly dominant, and the horizontal transport less significant. The residual term became negative above 500 mb. Analogous to the heat budget, the horizontal moisture advection term was dominant at lower levels and nearly balanced by a large-scale residual moisture source. Positive vertical moisture transport dominated at the upper levels and was offset by the large-scale residual moisture sink. At levels below 800 mb, the positive heat and moisture source was associated with the transport of sensible and latent heat from the surface. The negative moisture source and positive value of $(Q_s - Q_R)$ at levels between 800 and 300 mb were consistent with large-scale condensation occurring at these levels. Interpretation of the mean monthly cloud

<table>
<thead>
<tr>
<th>$p$ (mb)</th>
<th>Correlation</th>
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<tbody>
<tr>
<td>1000</td>
<td>-0.0021*</td>
</tr>
<tr>
<td>850</td>
<td>-0.1851</td>
</tr>
<tr>
<td>700</td>
<td>-0.3733</td>
</tr>
<tr>
<td>500</td>
<td>-0.4436</td>
</tr>
<tr>
<td>400</td>
<td>-0.3822</td>
</tr>
<tr>
<td>300</td>
<td>-0.3019</td>
</tr>
<tr>
<td>200</td>
<td>-0.0188*</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>$p$ (mb)</th>
<th>Sensible heat</th>
<th>Latent heat</th>
<th>Static stability</th>
</tr>
</thead>
<tbody>
<tr>
<td>1000</td>
<td>0.1604</td>
<td>0.1312</td>
<td>0.1648</td>
</tr>
<tr>
<td>850</td>
<td>0.1096</td>
<td>0.0633</td>
<td>0.0945</td>
</tr>
</tbody>
</table>
cover in the context of the heat and moisture budgets shows that middle clouds are formed primarily due to the benefit from large-scale three-dimensional moisture convergence; and low-cloud formation depends on surface moisture flux and the static stability. The quality of the upper-level moisture analyses was deemed to be too poor to make inferences about high clouds.

Heat and moisture budgets were also compared for anticyclonic and cyclonic situations. Cloud fraction is substantially greater under cyclonic conditions at all levels except at 200 mb, where cloud fraction is very low in both situations. Comparison of the heat and moisture budgets for the anticyclonic and cyclonic situations shows that all of the budget terms are of opposite sign at nearly all model levels. For the cyclonic situations the residual heat source is positive at all levels except 400 mb (reflecting large radiative cooling rate at this level), and there is a moisture sink at all levels, the maxima of both terms occurring at 850 mb. These residual terms are consistent with large-scale condensation occurring at all levels during disturbed conditions. By contrast, in the anticyclonic cases, \( Q_2 \) is positive at all levels, and only at 1000 and 850 mb is there evidence of large-scale condensation, accompanied by a moisture flux from the surface.

Heat and moisture budgets in a longitudinal cross section through a baroclinic wave were also examined. The relative humidity field is closely tied to the vertical velocities, the driest regions associated with the strongest sinking motion to the west of the upper-air trough and the surface cold-air advection from the south. Comparison of the relative humidity field with the cloud cover shows that peak cloud fractions are displaced approximately 3.5° latitude to the east of the peak relative humidities. It is suggested from examining this cross section that the clouds do not respond instantaneously to the relative humidity field, but take a period of time on the order of hours to adjust to the large-scale relative humidity field in terms of evaporation and condensation. This is further supported by examining \( Q_1 \) and \( Q_2 \). As the cloud fields progress from west to east, there is a moisture source as the cloud evaporates on the westward side and a moisture source as condensation occurs on the eastward side.

Determination of the grid-scale threshold relative humidity below which cloud, on average, does not occur was determined after Mitchell and Hahn (1990) to vary with height, ranging from 22% to 83%. Mitchell and Hahn (1990) have pointed out that the threshold relative humidity should reflect the statistical properties of the model's forecast relative humidity fields, so the actual values of the threshold relative humidity and their variation with height will be model dependent. The relationship between fractional cloud cover and model analyzed relative humidity was determined to deteriorate with height; at 200 and 300 mb there was, in fact, a negative correlation between cloud fraction and relative humidity, and at 200 mb no clouds occur at relative humidities (with respect to ice) exceeding 65%. From this analysis it is suggested that the model is not transporting sufficient moisture to high levels.

Comparison of the observed mean monthly layer cloudiness and the predicted cloud amount at each level using several relative humidity–based diagnostic cloud parameterizations. It was shown that appropriate tuning of a diagnostic relative humidity–based parameterization can result in accurate parameterized mean monthly total cloud amount for the region, and layer cloud fractions to within 5% of observed layer cloud fractions. However, this type of cloud fraction parameterization, as indicated by the low correlations, appears to be unable to diagnose layer cloud fraction on the smaller time and space scales that are undoubtedly required for obtaining the correct local cloud radiative and hydrological feedbacks with the dynamics. The correlation of cloud fraction with other meteorological parameters was investigated; correlation of cloud fraction with vertical velocity was significant, as was correlation of low-cloud fraction with static stability and surface heat fluxes. The deficiencies of purely diagnostic techniques emphasize the need to develop more physical models of the mechanisms controlling cloud cover, including additional prognostic equations for cloud processes.

From the results of this study it is inferred that mean monthly cloud amounts and heat and moisture budgets can be determined accurately from the ECMWF analyses, although the instantaneous gridpoint values can show large errors. This may reflect deficiencies in our understanding of the diabatic processes associated with large-scale processes and may also reflect problems with the analyzed humidity fields. We cannot expect a realistic parameterization of cloud in terms of grid-scale moisture until we are confident of the grid-scale moisture fields.

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REFERENCES


