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**Contents**



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# Turbulent Fluxes over Lake Ontario During a Cold Frontal Passage

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ABSTRACT

The turbulent fluxes of momentum, sensible heat and latent heat associated with a cold frontal passage of October 8, 1972 over Lake Ontario are examined. There are marked increases in the momentum and moisture fluxes after the frontal passage but the effects on the sensible heat flux are more complicated. The effects on the lake surface temperature are discussed.

## 1 Introduction

Large water masses play an important role in modifying the weather of their neighbouring land area. Lake Ontario is no exception. The lake-effect winter storms that pile snow on the leeward shores are one example. Another is modification of the Toronto heat island by on-shore winds (Munn et al, 1969). These effects result from the interaction of the atmosphere and the lake. The primary interactions are the exchanges of momentum, sensible heat and mass (water) across the interface between the atmosphere and the lake. The water surface exerts a drag on the underside of the moving air mass and tends to slow it down. To maintain the flow near the surface there is a downward transfer of momentum in the air. Since the water surface, unlike the land surface, can respond to air motions, momentum is transferred into the water resulting in a mean drift (the currents) and in surface water waves. If the air temperature is different from the water temperature, then there will be a transfer of sensible heat between the water and the air. Similarly, if the specific humidity of the air is different from the effective saturated specific humidity of the water, then there will be a transfer of moisture and latent heat between the water and the air. Heat transfer will also result from the imbalance of incoming and outgoing radiation.

The details of these transfers have been the subject of numerous studies (e.g. Kraus, 1972). An end product of these studies will be a model or parameterization technique to deduce the fluxes from synoptic scale meteorological measurements. Although there is still much investigation to be done, the bulk transfer equations that have been used over the years give useful approximate values for these fluxes. The equations for momentum, heat and moisture fluxes are:

$$\tau = \rho C_D(\bar{u}_{10})^2$$

$$H = \rho c_p C_H \bar{u}_{10}(T_w - T_a)$$

$$E = \rho C_E \bar{u}_{10}(q_w - q_a)$$

where:

$\tau$ — the downward momentum transfer which equals the surface stress
$H$ — the upward sensible heat flux
$E$ — the upward moisture flux or evaporation ($LE$ is the latent heat flux where $L$ is the latent heat of vaporization)
$\rho$ — the density of the air
$c_p$ — the specific heat at constant pressure
$\bar{u}_{10}$ — the mean wind speed at 10m
$T_w$ — the surface water temperature
$q_w$ — the saturated specific humidity corresponding to the surface water temperature. This assumes that the vapour pressure of the air immediately adjacent to the water is in vapour pressure equilibrium with it.
$T_a$ — the temperature of the air at 10m
$q_a$ — the specific humidity of the air at 10m

$C_D$, $C_H$, $C_E$ are the empirical coefficients required to make these equations valid. $C_D$ is usually called the drag coefficient.

The appropriate values for the coefficients $C_D$, $C_H$, $C_E$ have been the objective of numerous studies (e.g. Pond et al, 1971). For this work the result of Smith (1973) will be used for $C_D$ and it will be further assumed that $C_H = C_E = C_D$. This assumption is questionable but the available data are not conclusive as to how to assign values for $C_H$ *and* $C_E$. Smith's result is

$$C_D = (0.58 + 0.068\bar{u}_{10}) \times 10^{-3}, \ \bar{u}_{10} \text{ in m s}^{-1},$$

so that the $C_D$ used is dependent on wind speed but not on atmospheric stability. For $\bar{u} = 10$ m s$^{-1}$, $C_D = 1.26 \times 10^{-3}$. More recent results by Smith (1974) for direct measurements of $C_D$, $C_H$ and $C_E$ over Lake Ontario agree with these results.

During the International Field Year on the Great Lakes (IFYGL) several arrays of meteorological buoys were deployed on Lake Ontario. For the present study the data from the buoys of the Canada Centre for Inland Waters, Burlington, Ontario, were used. Wind speed and direction, air temperature and vapour pressure at 4 m were measured using a cup anemometer and vane, thermistor and lithium choloride humidiometer, respectively (Elder and Brady, 1972). Surface water temperature was measured with a thermistor mounted on the side of the buoy so as to measure the temperature of the upper

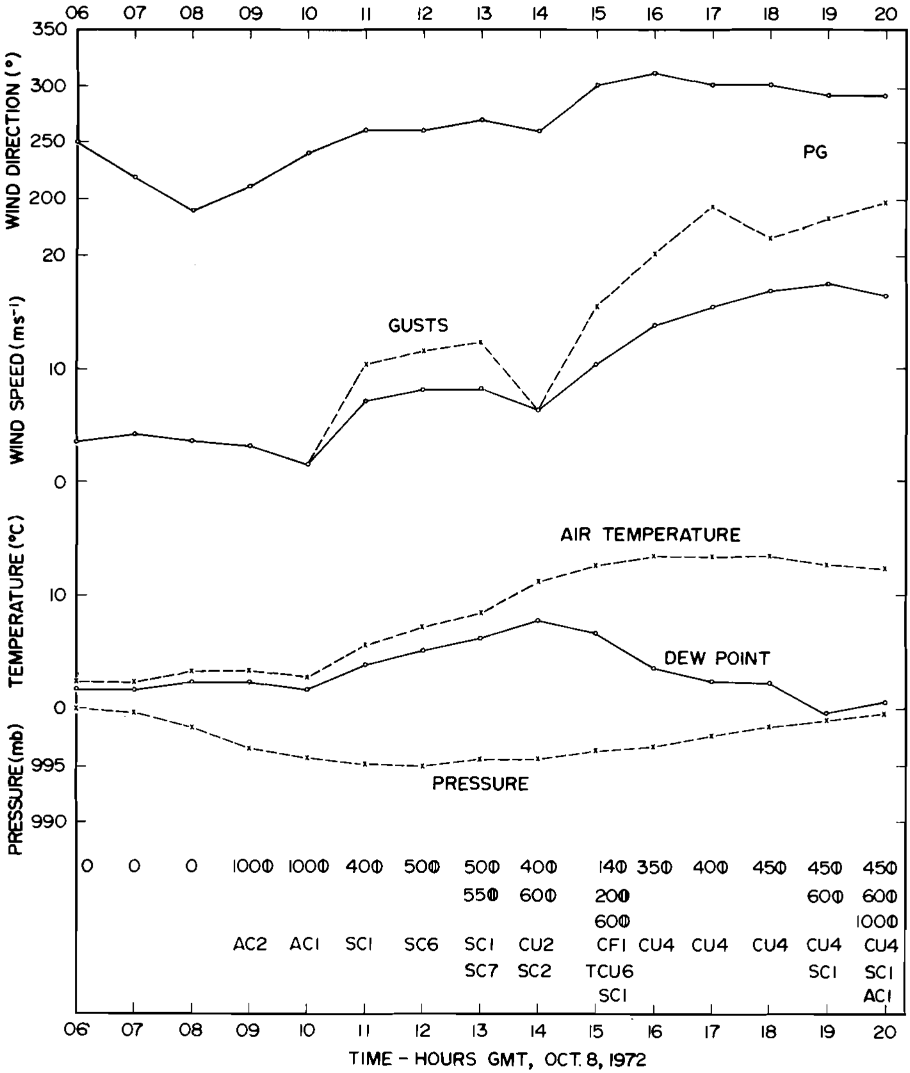



Fig. 1. Hourly weather observations at Toronto International Airport (YZ) for the period 0600–2000Z, October 8, 1972. Wind direction, wind speed, temperature, dewpoint, pressure and cloud amounts and types are given. The cloud heights are in hundreds of feet as reported in the hourly observations. PG in the upper right is the peak gust reported.

few centimetres of water. From these data the surface fluxes were calculated by using the above formulae.

The objective of this paper is to examine the time and space variations of the turbulent fluxes as a cold front passes over Lake Ontario. First the synoptic situation will be described and then the variation of the fluxes with time at one point and with space relative to the front will be studied.

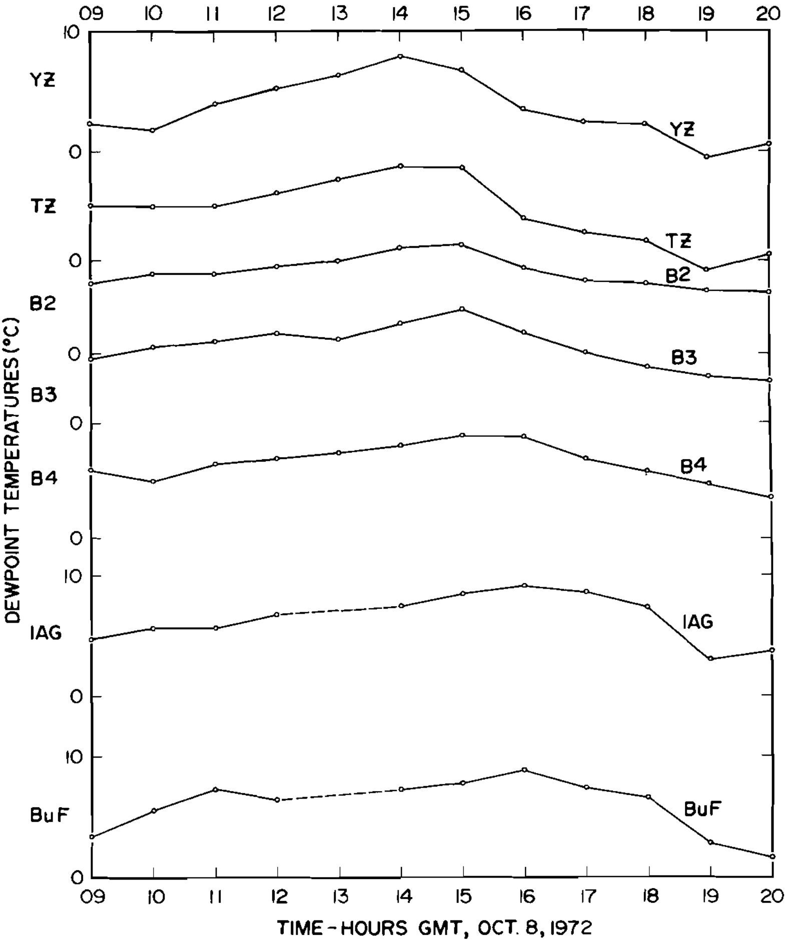



Fig. 2. Hourly dewpoint temperatures for the period 0900–2000Z, October 8, 1972. Stations are Toronto International Airport (YZ), Toronto Island Airport (TZ), three IFYGL buoys (B2, B3, B4, see Fig. 3), Niagara Airport (IAG) and Buffalo Airport (BUF).

## 2 Synoptic Situation

For the first five days of the month of October a large high pressure area covered most of eastern Canada and the United States. A weak quasi-stationary front moved over Lake Ontario on October 7th and was followed by a ridge line early on the 8th. At the same time a deep low pressure centre (986 mb) moved into northwestern Ontario. At 00Z on October 8th a cold front associated with this low was west of Lake Michigan and moving rapidly east-south-east. The details of the passage of this front over Lake Ontario are the subject of this paper.

As is usually the case upon detailed examination, this cold front was not a simple, easily-defined feature. A look at the hourly weather observations for Toronto International Airport (Fig. 1) will demonstrate this. The first event to be noted is the sudden wind speed increase at 1100Z, the wind being 1.5 m $s^{-1}$ at 1000Z and 7.2 m $s^{-1}$ with gusts to 10.3 m $s^{-1}$ at 1100Z. The wind direction shifted from 240° to 260°. This sudden increase in wind speed can be followed along the lake shore, occurring at Peterborough at 1300Z, at Trenton at 1500Z, and at Rochester about 1700Z. This wind change line was not closely related to the pressure minimum line which was 1–2 hours behind it on the north shore but preceded it on the south shore. A second marked wind change occurred at Toronto International Airport between 1400Z and 1500Z and this accompanied the beginning of a rapid decrease in dewpoint, which dropped 8.4°C between 1400Z and 1900Z. This second feature has been identified as the cold frontal passage. The change of slope of the dewpoint trace is the most easily identifiable feature and it can be traced through a sequence of stations. In Fig. 2 the hourly dewpoint values for Toronto International Airport, Toronto Island airport, 3 buoy stations, Niagara Falls and Buffalo are shown. These stations are in a line approximately perpendicular to frontal motion with orientation about 325°. These results indicate a frontal speed over western Lake Ontario of about 8 m $s^{-1}$.

In Fig. 3 the synoptic situation for 1500Z on October 8, 1972 is shown. The wind shift line is on a line Trenton to west of Rochester while the front is along a Peterborough-Toronto line. By 1800Z the cold front had passed eastward over most of Lake Ontario and was south of Buffalo. The main effect of the cold frontal passage was to cause a large change in the atmospheric moisture content. This is illustrated in Fig. 4 which shows three upper-air soundings from Lakeside, a special IFYGL radiosonde station northwest of Rochester. At 1200Z the lowest 300 m is stably stratified (temperature 8–10°C) with a dewpoint temperature of about 4°C. By 1500Z the lowest 700 m had an adiabatic lapse rate and the dewpoints have increased to about 6°C near the surface to saturated values of about 8°C at 600 m. By 1800Z, after the frontal passage the air temperature lapse rate is still dry adiabatic and slightly warmer than at 1500Z while the average dewpoint of the lowest 600 m has dropped by about 7°C.

## 3 The Turbulent Fluxes

In Fig. 5 the meteorological parameters and the fluxes from buoy 7 are plotted. Buoy 7 has a location almost in the centre of the lake. The wind shift line is not well marked but may have passed by 1200Z. The front is better defined with frontal passage between 1500 and 1600Z. Since the wind stress is proportional to between the square and cube of the wind speed it responds more dramatically to wind speed changes. There is a rapid increase in stress until 1200Z and then approximately constant values until frontal passage when the value increases by almost 50%. The kinematic stress remained about 0.25 $m^2$ $s^{-2}$ until about 0600Z on October 9th when it decreased to less than one-half that value. The wind speed decrease at that time was likely due to the combined effects of the decreasing free atmosphere wind speed and a diurnally caused stable layer inhibiting momentum transfer downward from the free atmosphere.

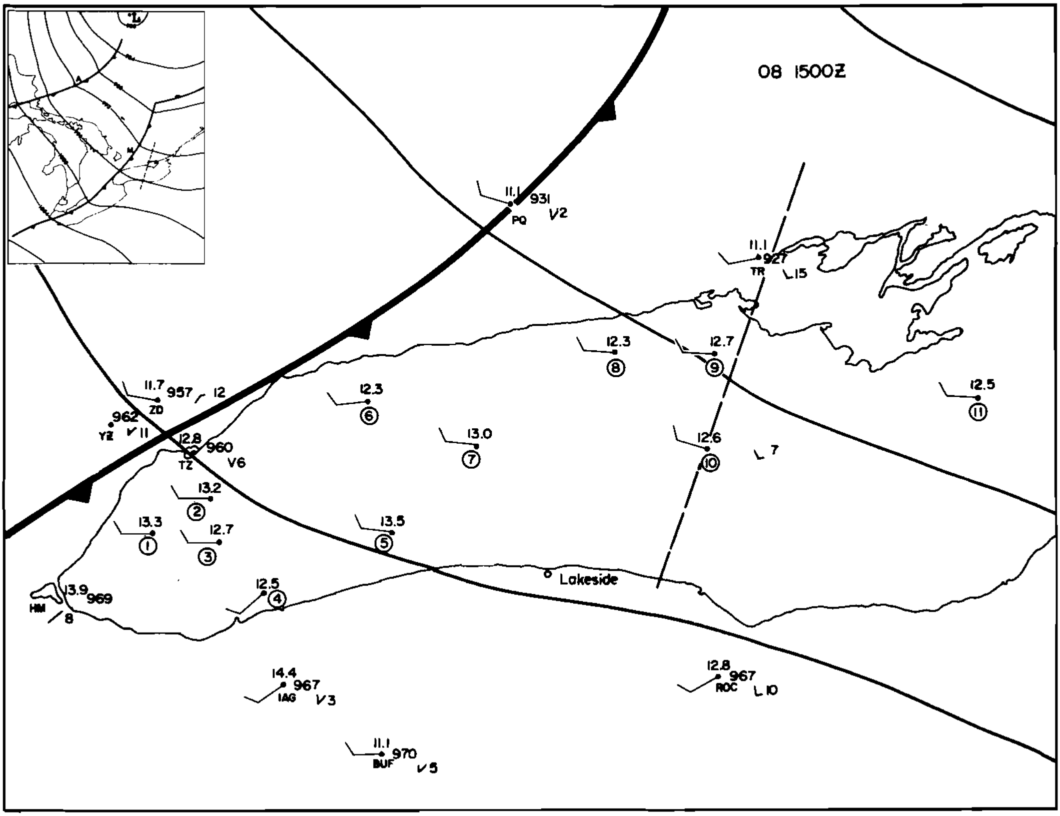



Fig. 3. Synoptic weather situation for 1500z, October 8, 1972. The wind speed (m $s^{-1}$) and direction are indicated by lines with barbs as usual. The numbers are air temperature (C) (above), pressure (tenths of millibars, omitting the hundreds and thousands) (above right), pressure tendency (tenths of millibars in 3 hours) (right) and station identifier (designator or circled number for buoy stations) (below). The 994 and 996 mb isobars, the cold front and the wind shift line (dashed) are indicated. The insert shows the synoptic situation for a larger area.

The evaporation and hence the latent heat flux increased (18 W $m^{-2}$/6 hr) then decreased just before frontal passage. When the dry air behind the front reached the buoy the flux increased rapidly and averaged about 270 W $m^{-2}$ for the next 12 hours. The sensible heat flux variation was quite different. The values were near 50 W $m^{-2}$ until 1400Z and then decreased to near zero by 1600Z. This was followed by a gradual increase (80 W $m^{-2}$/11 hr) for 18 hours, with the flux values then levelling off at about 100 W $m^{-2}$. These variations in the sensible heat flux are due to conflicting trends. First, the air temperature was rising due to both warm air advection and heating from the water surface below. In addition, the solar radiation input would tend to create a diurnal variation. This would cause a further temperature increase. After the frontal

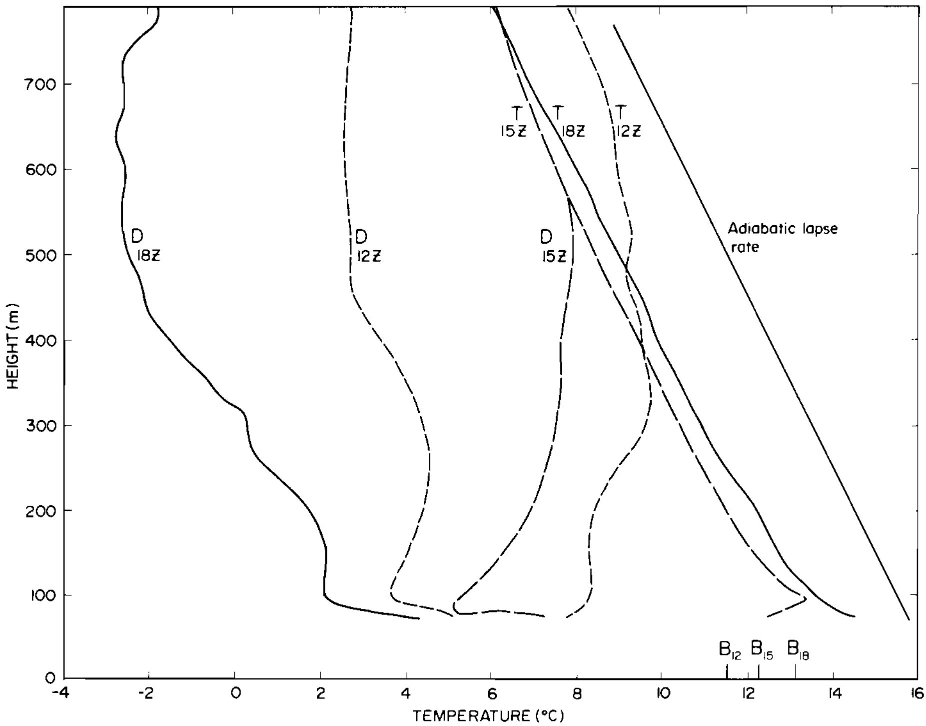



Fig. 4. Radiosonde data from Lakeside (see Fig. 3). T and D are the temperature and dewpoint temperature, respectively, for the times indicated. The B's are the air temperatures at buoy 5 for the times indicated.

passage (by 1600Z, 1100 EST) the air temperature started to decrease but only very slowly. The air immediately behind the front had been warmed by its passage over the warm land (note the diurnal temperature variation at Toronto International Airport [Figure 1]). The maximum air temperature at the buoy was 13.5°C at 1600Z (1100 EST). The maximum on October 9th (the following day) was 8.4°C at 2200Z (1700 EST). The frontal passage distorted the diurnal temperature variation and reduced the peak by 5°C from one day to the next. The second effect was the decrease in water surface temperature which dropped 0.5°C between 1500Z and 1600Z and continued to decrease, but more slowly, after that. The decrease in water surface temperature was due to increased heat (both latent and sensible) loss and due to increased mixing of the cooler subsurface water to the surface associated with the windspeed increase. The net result of these effects was to reduce the air-water temperature difference and, hence, reduce the sensible heat flux.

The space-time variation of the fluxes over the lake was first examined by plotting the fluxes and the front for each hour during the interval 1400–1800Z of October 8th. It was found that the frontal effects on the fluxes did not show

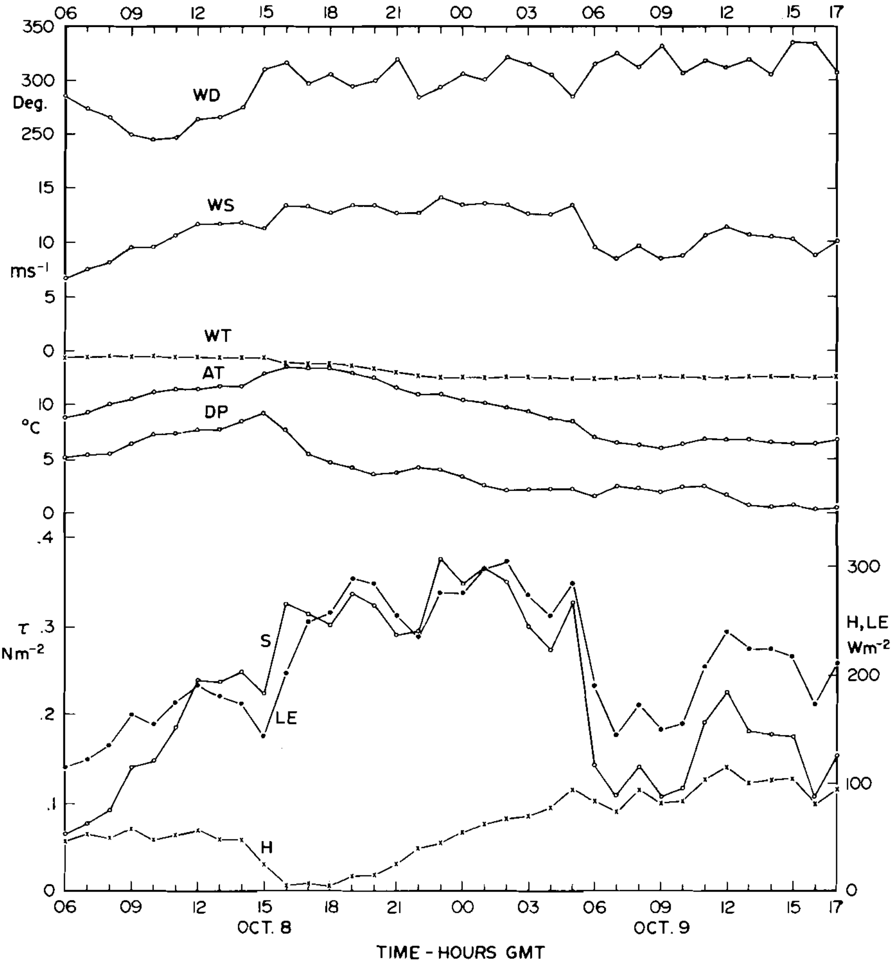



Fig. 5. Hourly weather observations and computed fluxes at buoy 7 for period 0600z, 8 October to 1700z, 9 October, 1972. From top are wind direction (WD), wind speed (WS), water temperature (WT), air temperature (AT), dewpoint (DP), stress (S, $\tau$), heat flux (H), latent heat flux (LE).

up particularly strongly. There are several possible reasons for this. First, the local topographical and other anomalies in the lake, such as the proximity to shore and the Niagara River inflow, distort the flux pattern associated with the front, and, second, any errors in the measurements at any one buoy would also distort the flux patterns. Third, there is a lack of spatial resolution in that there are only 11 flux estimates. To overcome these deficiencies the following procedure was adopted. For each buoy location the flux values for each hour were divided by the maximum flux at that buoy that occurred during the 24-hour period of October 8th. This should reduce the effects of lake anomalies and sensor errors. The 24-hour period is somewhat arbitrary but slightly altering it would not affect the general results. It was then assumed that the variations in time of the fluxes at a point fixed relative to the front were smaller than the variations in space relative to the front, and that a time-space transformation would be meaningful. It was assumed that the front remained constant in shape during its passage over the lake and the flux values were plotted relative to a point on the front which was assumed to follow a streamline along the geostrophic wind. The period 1400 to 1800Z was used so that the end results (Fig. 6) are flux values at 55 points relative to the front.

The wind stress pattern, $\tau/\tau_{max}$, Fig 6a, shows a maximum area behind the front with a minimum in front. A secondary maximum is in advance of the front and appears to be associated with the wind shift line. The low values to the north-west of the front are due to buoys 8 and 9 which both measured strong winds at 2000Z (20% higher than before and after 2000Z). The evaporation flux pattern, $E/E_{max}$, (Fig. 6b) shows the clearest frontal effects. Ahead of the front, values are almost all less than 0.5 whereas behind the front the values are usually greater than 0.8. (Because the humidity sensor on buoy 8 was not working there is no data on the evaporation flux for it). Typical values of $LE$ behind the front were about 300 W m$^{-2}$, with one value as high as 450 W m$^{-2}$. The mean latent heat flux for the month of October was 103 W m$^{-2}$ (Pinsak and Rodgers, 1974). Heat fluxes during cold, dry air outbreaks behind cold fronts, therefore, are very important contributors to the net energy transfer.

The sensible heat flux, $H/H_{max}$, (Fig 6c) does not show the patterns that would have been expected. As explained previously, these results are due to the interaction of several influences on the air-water temperature difference. Since time of day generally increases from right to left in Fig. 6c an increase of air temperature during the day would show up as a decrease in heat flux from right to left. This is what is evident in Fig. 6c with the lowest values behind the front and the largest values ahead of it. The maximum to the southeast of the front is associated with the wind maximum near the wind shift line. According to Pinsak and Rodgers (1974), the average sensible heat flux for October was 33 W m$^{-2}$. Flux values in advance of the cold front (i.e., early on October 8, 1972) were typically 50–80 W m$^{-2}$ and, after decreasing initially after the frontal passage, were 100–120 W m$^{-2}$ over most of the lake by about 1200Z on October 9th.

## 4 The Effects

The effects of a cold frontal passage can be best illustrated by examining the airborne radiation thermometer (ART) water surface temperature distributions of October 5th and 10th, 1972 (Fig. 7). On October 5th the lake temperature was fairly uniform with most values between 14 and 16°C. The mean was 15.3°C. By October 10th, after the frontal passage, the situation was drastically different. The mean temperature was down to 11.5°C with some water near the north shore below 5°C. The cold water on the north side is due to

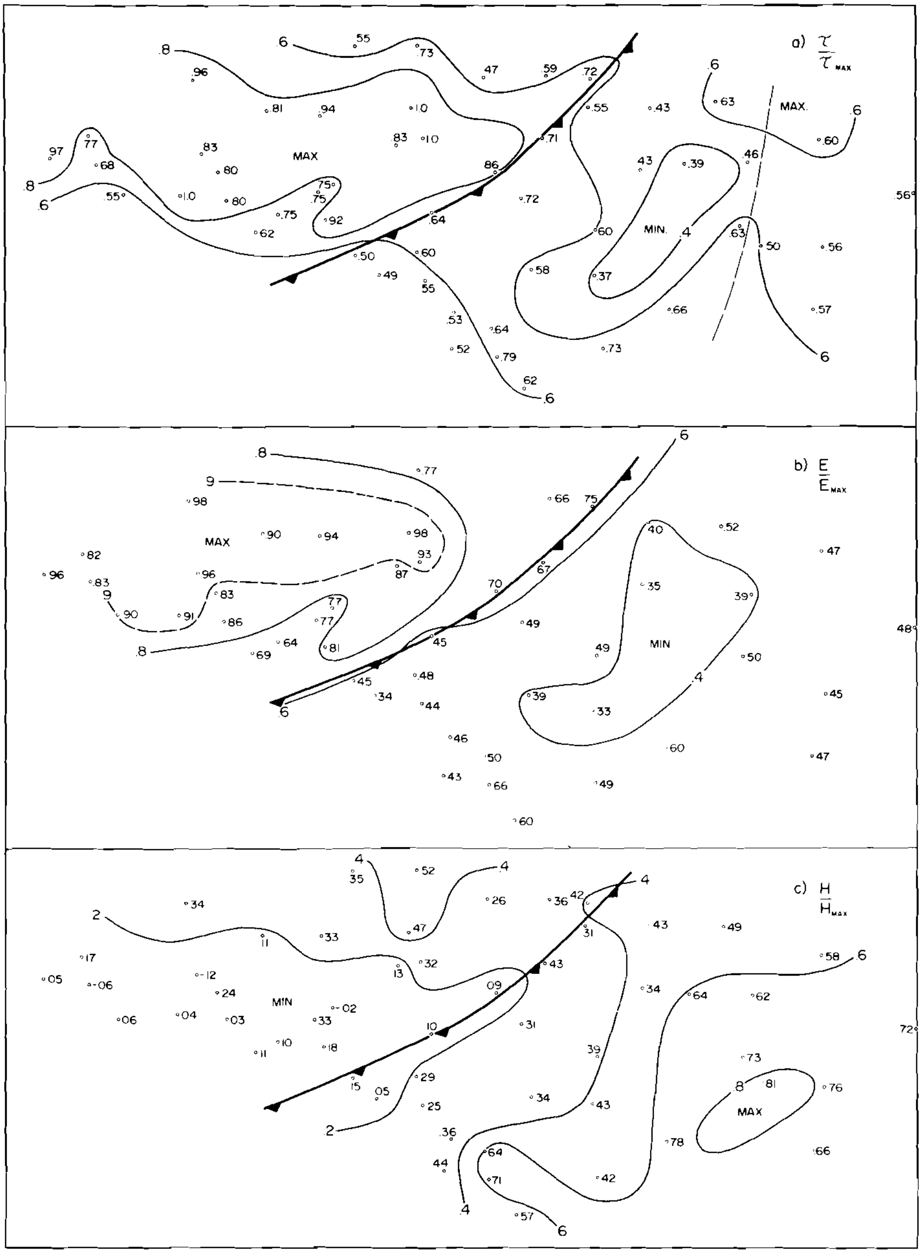
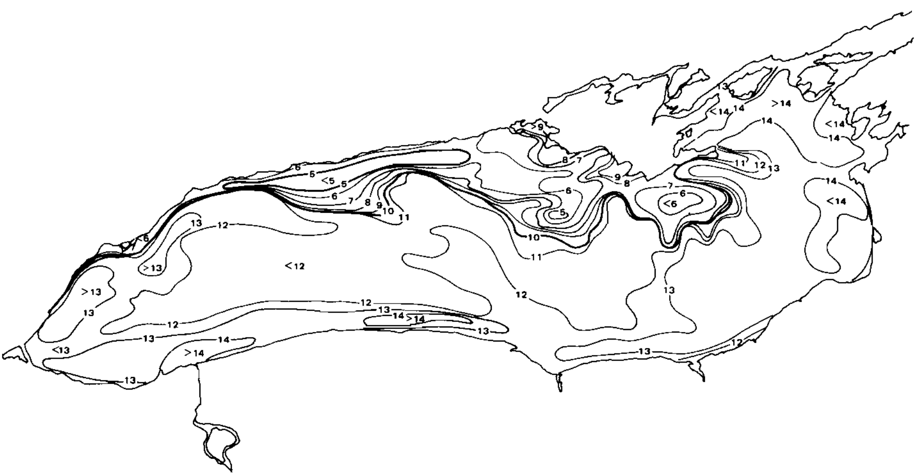



Fig. 6. Ratios of fluxes to the 24 hr maximum and relative to the cold front for a) stress, b) evaporation and c) sensible heat.

ART SURVEY OCT 5/72 MEAN TEMP. 15.3°



ART SURVEY OCT 10/72 MEAN TEMP. 11.5°

Fig. 7. Surface water temperature from airborne radiation thermometer (ART) surveys of October 5 and October 10, 1972, Isotherms in degrees C. Data provided by Hydrometerological and Marine Applications Divisions, Atmospheric Environment Service.

colder subsurface water having upwelled because of the wind-induced surface water transport to the south. Over the southern two-thirds of the lake the water temperatures are 11 to 13°C, representing a 3°C decrease from October 5th. Part of this decrease would be due to the heat loss due to latent and sensible heat fluxes. A total flux of 400 W $m^{-2} \simeq 800$ cal. $day^{-1}$ $cm^{-2}$, would cool by 1°C in two days a layer of water about 16 m deep which is the order of the depth of the mixed layer. The remaining cooling would be due mainly to further vertical mixing and some heat losses on the days before the frontal passage.

The importance of frontal passages similar to the one described can be illustrated by considering the two month period September 21 to November 21, 1972 during which 11 ART surveys were made. The mean lake surface temperature dropped 10.7°C (16.9°C to 6.2°C) during this period but the rate of decrease was very non-uniform. Between September 21 and October 4 there were several weak frontal passages but no sustained strong winds and the mean temperature dropped only 1.6°C. Between October 5 and 18 the drop was 6.7°C as a result of the frontal passage described and other ones on October 12, 14 and 17. The frontal passage on October 17 was the most similar to that of October 8. Between October 18 and November 21 the decrease was a further 2.4°C. Thus the periods of frontal passage followed by strong northwesterly winds were responsible for over 60% of the temperature decrease although occupying less than 15% of the time period.

It is interesting to note in conclusion the tremendous amounts of energy involved in these lake-to-atmosphere latent and sensible heat transfers. During the month of October the total energy transfer through the surface of the lake would be about $7 \times 10^{25}$ ergs (using Pinsak and Rodgers, (1974) mean fluxes) which is equivalent to almost 2000 megatons of T.N.T. or twenty 100 megaton nuclear bombs.

**Acknowledgements**

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Thanks are also due to R.D. Paterson, V.S. Derco and G. den Hartog for comments on the paper.

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# Une Methode d'Etude Experimentale "in situ" de la Zone d'Echange entre les Nuages Convectifs et leur Environnement

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RESUMÉ

Le développement d'un projet de recherche consacré à l'étude des nuages convectifs a amené les auteurs à concevoir et à élaborer un système complet de mesures adapté à cette étude. Les grandeurs thermodynamiques classiques (Température, Pression, Humidité) ainsi que les vitesses d'ascendance autour et à l'intérieur des nuages sont mesurées grâce à l'utilisation d'un avion Cessna 206 équipé de ce système. Une première étape consiste à mettre en évidence les phénomènes relatifs aux échanges entre le nuage et le milieu environnant.

## 1 Introduction

L'étude de la dynamique des cellules convectives a déjà fait l'objet de nombreux travaux de recherches très souvent théoriques, quelque fois réalisés en laboratoire (Squires et Turner, Morton, Priestley, etc. ...) et moins fréquemment de mesures "in situ" au niveau du nuage lui-même (Warner, Musil, Schleusener ...).

Les travaux théoriques ont pris une grande importance lorsqu'il a été possible de réaliser des simulations de phénomènes convectifs sur ordinateur. On a pu ainsi progresser dans la compréhension des mécanismes d'évolution des cellules convectives ou dans leur modification (Simpson, Weinstein, Dennis, etc...).

L'efficacité de ce type de recherche reste liée à l'amélioration des modèles théoriques, qui dépend alors d'une meilleure connaissance des phénomènes naturels. La majorité des résultats donnés par ces modèles sont, en effet, confrontés aux observations effectuées à partir du sol grâce au radar. Or l'établissement de relations rigoureuses entre les paramètres physiques de l'atmosphère et les données radar comporte de grosses difficultés. Il est donc fondamental de procéder à de nombreuses mesures in situ.

Des auteurs comme Bjerknes, Asai et Kasahara, ont montré l'intérêt que présente l'étude des mouvements d'air de compensation qui accompagnent le développement d'un nuage. Du fait de la complexité et de la diversité des phénomènes qui entrent en jeu, notre approche s'est momentanément limitée à l'observation des échanges entre le nuage et le milieu environnant.

Afin de disposer d'un ensemble de mesure pour les expérimentations envisagées nous avons été amenés à réaliser au laboratoire une grande partie des capteurs utilisés ainsi que toute la chaîne d'acquisition des données. Cet équipement dont le principe est déjà connu, constitue cependant un ensemble bien adapté aux enregistreurs magnétiques analogiques utilisés, d'un type classique, ce qui réduit le coût global du système.

L'ensemble des capteurs, des dispositifs électroniques associés ainsi que des enregistreurs magnétiques occupe un volume de 0,150 $m^3$ et pèse environ 25 kilogrammes. Alimenté sous 28 volts continus, le système complet nécessite une puissance électrique d'environ 60 watts. La majeure partie des mesures a été effectuée à bord d'un Cessna 206 turbochargé.

**2 Les équipements**

Deux types de paramètres sont mesurés ; d'une part les grandeurs thermodynamiques classiques (pression, température, degré hygrométrique), d'autre part, les divers paramètres permettant de connaître à tout instant la valeur de la composante verticale de la vitesse des courants aériens.

**a** *Les Grandeurs Thermodynamiques*

*Les Capteurs* (*Fig. 1*)

Certains des capteurs utilisés sont commercialisées, d'autres ont été conçus et réalisés au laboratoire.

La pression atmosphérique est mesurée par un capteur fabriqué par la SFIM.[1] Il s'agit du modèle H335 constitué d'un empilement de capsules anéroïdes dont une extrémité est fixe et l'autre entraîne dans son déplacement le curseur d'un potentiomètre de valeur ohmique totale de 500 ohms. L'entrée du capteur est branchée sur l'une des prises statiques de pression.

Une thermistance K19 fabriquée par Siemens[2] mesure la température. Ses dimensions très réduites lui confèrent un temps de réponse de l'ordre du dixième de seconde et sa valeur ohmique à 20°C est de 10000 ohms. Elle est disposée à l'interieur d'un tube métallique dont le comportement aérodynamique a été étudié avec soin. Des ouvertures pratiquées en couronne sur le tube et à l'arrière du support assurent une ventilation de l'ordre de 3 m/s lorsque l'avion vole à sa vitesse de travail. L'intérieur du tube a été noirci et l'extérieur anodisé afin d'affranchir le capteur des effets du rayonnement solaire.

Le capteur d'humidité est constitué par une plaquette à film de carbone fabriquée par Bendix International sous la nomination Ml 476 Humidity Sensor et dont la résistance électrique varie avec l'humidité relative de l'air. Afin d'éviter la destruction du film de carbone par l'eau liquide, cette plaquette est montée à l'intérieur d'un tube dont la structure permet au capteur d'effectuer des mesures jusqu'à l'intérieur des nuages. Pour cela une série de trois chicanes, en forme d'entonnoir, a pour mission d'éliminer les gouttes d'eau présentes dans le flux d'air qui traverse le tube. Les deux capteurs de

[1]Société de Fabrication d'Instruments de Mesure, 13, Avenue de Ramolfo-Garnier, 91 Massy, France.

[2]Agence canadienne: Siemens Canada Limited, 407 McGill Street, Montreal 1, P.Q.

Fig. 1 Les capteurs de température et d'humidité : à gauche, deux capteurs de température (tubes métallisés) à droite, le capteur d'humidité (tube plastique transparent).

température et d'humidité sont disposés sous l'aile de l'avion, à une distance telle que les remous crées par l'écoulement turbulent ne perturbent pas les mesures.

*Acquisition des données (Fig. 2 et 3)*

Tous les paramètres mesurés sont transformés, par conversion électronique analogique, en signaux à fréquence variable. Ces signaux sont ensuite composés deux à deux par un système de multiplex fréquentiel puis enregistrés sur bande magnétique. Chaque signal composite est donc la somme de deux tensions dont les fréquences appartiennent l'une au canal 200–1 000 Hz, l'autre au canal 2 000–10 000 Hz. L'addition s'effectue dans un mélangeur basse fréquence à deux voies dont le signal de sortie est enregistré sur l'une des deux voies d'un magnétophone Uher[1] type 4 400, ($E_1$).

[1] 147, Rue Jean Pierre Timbaud, 92 400 Courbevoie, France

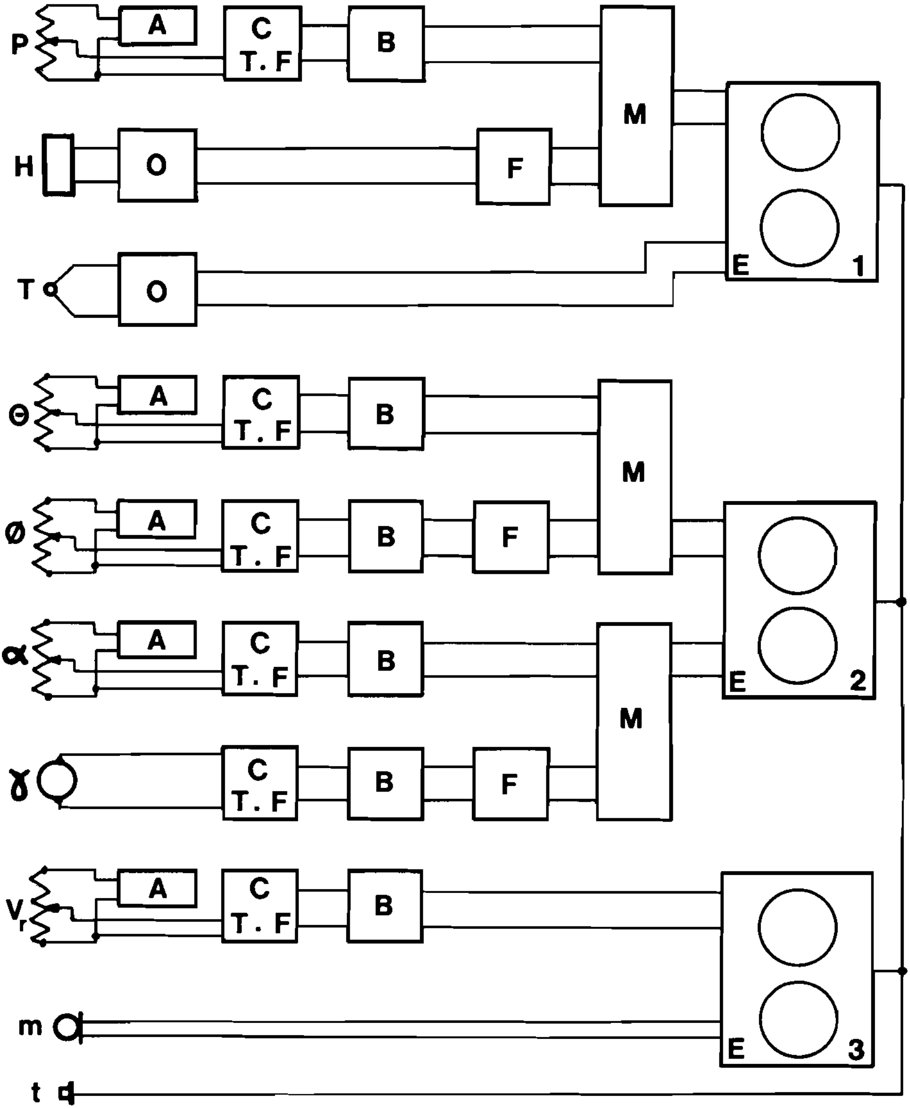


Fig. 2 Traitement et enregistrement des données : A : Alimentation stabilisée, O : Oscillateur, C.T.F. : Convertisseur tension-fréquence, B : Circuit bistable, F : Filtre passe-bas, M : Mélangeur, E : Enregistreur.

Dans le cas de la mesure de la température, la thermistance (comme la plaquette de mesure de l'humidité) est intégrée dans un oscillateur à unijonction qui délivre une tension alternative dont la fréquence varie avec la résistance de celle-ci. L'utilisation d'un tel oscillateur possède l'avantage de soumettre la thermistance au passage d'un courant suffisamment faible pour qu'aucun effet anémométrique ne se manifeste. Afin d'obtenir une meilleure sensibilité finale, on n'a pas limité ici la fréquence aux bornes d'un canal, elle

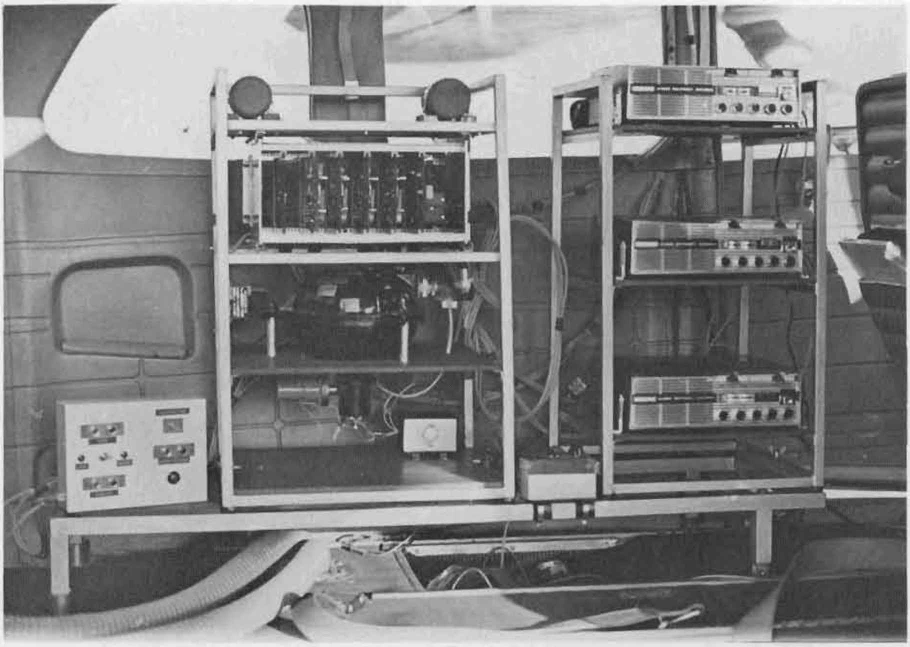

Fig. 3 L'ensemble du système aéroporté : électronique et magnétophones, à l'intérieur d'un Cessna 206.

peut varier entre 500 Hz et 6000 Hz environ. Le signal "température" n'est donc pas mélangé à un autre mais enregistré directement sur la voie 2 du même magnétophone Uher que précédemment, ($E_1$).

**b** *Paramètres d'Ascendance*

Afin de pouvoir déterminer la vitesse des courants verticaux pour toutes les attitudes de vol possibles de l'avion, il s'avère indispensable de connaître simultanément les valeurs de cinq grandeurs différentes :

L'intensité $V_r$ de la vitesse de l'air par rapport à l'avion

La direction $\alpha$ de la vitesse de l'air par rapport à l'avion

Les 2 angles $\Phi$ et $\Theta$ caractérisant l'assiette de l'avion par rapport au plan horizontal.

L'accélération normale $\gamma$ subie pa l'avion.

Un calcul simple montre que la composante verticale $w$ de la vitesse absolue d'ascendance a pour expression :

$$w = \frac{V_r(\alpha - \Theta)}{\cos \Phi} + \int_{t_0}^{t_1} \left( \gamma \frac{\cos \Phi}{\cos \Theta} - g \right) . \mathrm{d}t$$

*Les Capteurs*

L'intensité $V_r$ de la vitesse de l'air par rapport à l'avion est mesurée par un tube de Pitot relié à un variomètre différentiel du type H330 de la SFIM. Son fonctionnement est identique à celui que l'on utilise pour la mesure de la

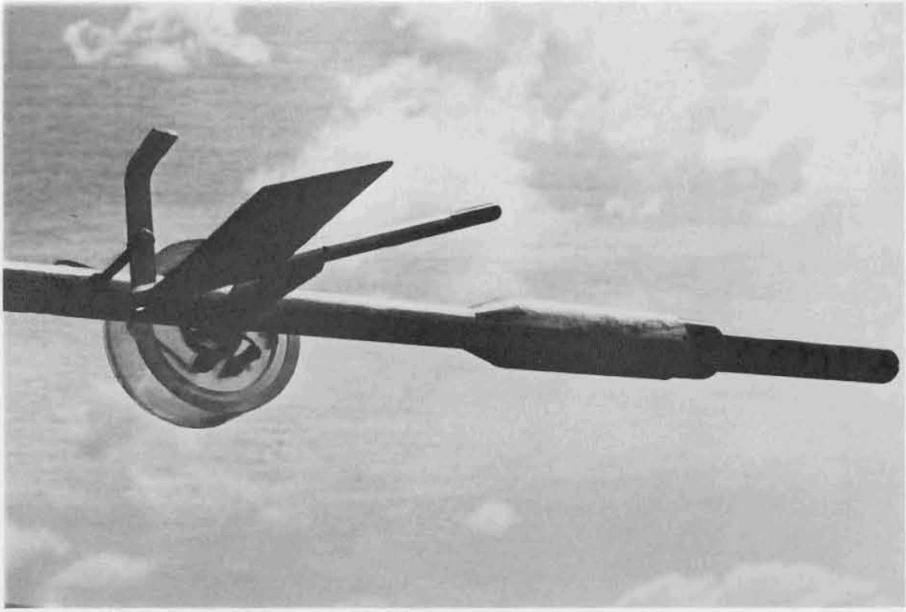

Fig. 4 Le tube de Pitot et l'ailette utilisés pour les mesures d'ascendances.

pression atmosphérique. Les variations de vitesse relative provoquent le déplacement du curseur d'un potentiomètre de 500 ohms.

La direction $\alpha$ de l'air par rapport à l'avion est donnée par une ailette fixée à l'extrémité d'une perche métallique, et qui s'oriente dans le vent relatif. Elle entraîne dans sa rotation, le curseur d'un potentiomètre dont les frottements ont été réduits au minimum (Fig. 4).

Les angles $\Theta$ et $\Phi$ caractérisant l'assiette de l'avion par rapport au plan horizontal sont déterminés à l'aide d'un dispositif gyroscopique dont l'axe de référence reste vertical. Chacun des deux axes de rotation de la suspension à la cardan du gyroscope est solidaire du curseur d'un potentiomètre. Les amplitudes maximales choisies sont de $+60°$ et $-60°$ pour l'angle de roulis $\Phi$, et de $+20°$ à $-20°$ pour l'angle de tangage $\Theta$.

Enfin l'accélération $\gamma$ subie par l'avion est mesurée par un accéléromètre SFIM modéle JT 2041S à sortie analogique dont la gamme de mesure s'étend de $-10$ à $+30$ m/s² (c'est à dire symétrique autour de la valeur de g). Cet accéléromètre est fixé sur un chassis perpendiculairement au plancher de l'avion.

*Acquisition des données (Fig. 2 et 3)*

Roulis $\Phi$ et tangage $\Theta$: chacun des potentiomètres est soumis à la tension délivrée par une alimentation stabilisée à circuit intégré. Chacune des deux tensions variables est alors appliquée à l'entrée d'un convertisseur Tension-Fréquence TBO5. Après la mise en forme, le signal relatif à l'angle $\Phi$

occupe la gamme de fréquence 200–1000 Hz et le signal relatif à l'angle $\Theta$, la gamme 2 000–10 000 Hz. La tension à fréquence inférieure à 1 000 Hz est filtrée puis les deux signaux sont mélangés avant d'être enregistrés sur la voie 1 d'un second magnétophone stéréophonique ($E_2$).

Angle $\alpha$ : le signal relatif à cet angle est obtenu de la même manière que précédemment. Il occupe la gamme de fréquences 2 000–10 000 Hz et est appliqué à l'entrée de la voie 1 d'un mélangeur.

Accélération $\gamma$ : la tension délivrée par l'accéléromètre est transformée par un convertisseur Tension-Fréquence en signal de fréquence variable qui, après mise en forme, occupe la gamme de fréquence 200–1 000 Hz. Il est filtré puis appliqué à l'entrée de la voie 2 du même mélangeur que précédemment. Le signal composite est alors enregistré sur la voie 2 du second magnétophone $E_2$.

Vitesse $V_r$ : à partir d'un potentiomètre intégré au manomètre différentiel, on élabore le signal correspondant de la même manière que pour l'angle $\alpha$. Ce signal est enregistré sans être mélangé à un autre, sur la voie 1 d'un troisième magnétophone stéréophonique ($E_3$).

c *Commentaire – Synchronisme*

La deuxième voie, *m*, du troisième magnétophone est consacrée à l'enregistrement du commentaire parlé du chef d'expérimentation. Ce commentaire contient toutes les informations précises concernant le déroulement de chaque expérience (appréciations qualitatives, désignations des instants de pénétration dans les cellules ainsi que des sorties, valeurs du cap suivi par l'avion etc ...).

D'autre part, une bonne simultanéité dans les enregistrements de tous les paramètres étant indispensable, il est procédé au marquage périodique et simultané des trois bandes magnétiques par un signal caractéristique dit de "topage". Tout glissement intempestif d'une bande par rapport à la bande de référence contenant le commentaire peut ainsi être évité.

**3 Méthodes de mesures**

Afin de contrôler étroitement le déroulement de chaque expérimentation, l'avion de mesures travaille en relation permanente avec un radar 3 cm. Ce radar localise avec précision la ou les cellules étudiées. Les relevés effectués sur le scope radar sont ensuite comparés aux mesures "in situ".

Avec un seul avion équipé de capteurs, il est possible de sonder des cumulus congestus dont le développement vertical atteint 5 000 mètres environ. Notons toutefois que la plupart des cellules étudiées avaient un développement de 2 000 mètres. La cellule est sondée à partir du moment où son développement vertical dépasse 500 mètres. L'avion poursuit le sondage pendant son évolution, soit jusqu'à sa disparition, soit jusqu'au moment où sa taille rend impossible la continuation des mesures. Avant d'approcher la cellule elle-même, l'avion effectue localement un sondage vertical de température et d'humidité depuis le sol jusqu'au niveau de son sommet.

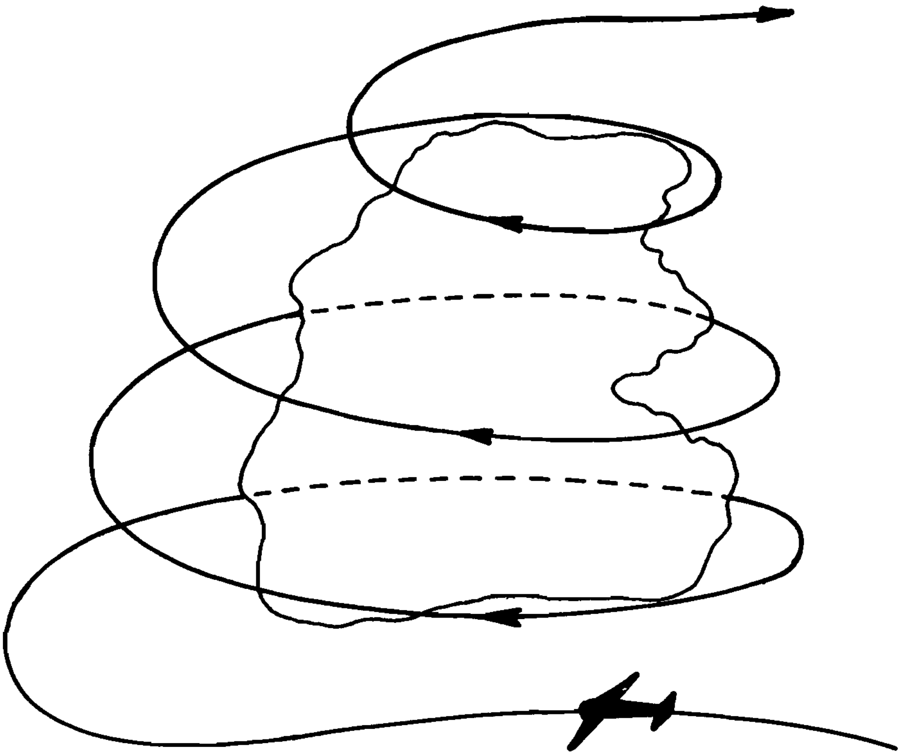

Fig. 5 Trajectoire de l'avion de mesure au cours du sondage d'un cumulus. La plupart des cellules sondées ont un développement vertical d'environ 2000m et un diamètre allant de 500 à 2 000 m.

Ensuite, chaque cellule isolée est étudiée de la manière suivante :
— passage à la base de la cellule,
— palier horizontal autour de la cellule, aussi proche que possible de celle-ci (quelques mètres), à 200 mètres environ au-dessus de la base et, éventuellement, traversées à la même altitude
— Palier et traversées au niveau du centre de la cellule
— palier et traversées à 200 mètres au-dessous du sommet
— passage au niveau du sommet.

Lorsque l'activité de la cellule sondée le permet, les traversées sont effectuées suivant l'axe de déplacement de la cellule et la direction perpendiculaire (Fig. 5).

## 4 Traitement des données

### a *Précision des mesures*

Il est indispensable de savoir dans quelle mesure les données enregistrées sont significatives. Deux éléments entrent en jeu dans la détermination de la précision des mesures : la réponse statique de l'ensemble capteur-enregistreur d'une part, sa réponse dynamique d'autre part.

La réponse statique concerne la précision absolue d'une mesure ponctuelle donnée. Compte tenu de l'élément capteur et du taux de pleurage des enregistreurs, l'évaluation de la sensibilité pour chaque paramètre mesuré a donné les résultats suivants :

| | |
|---|---|
| Température | : $^1/_{10}$ de degré Celsius |
| Humidité | : 1% d'humidité relative |
| Pression | : 4 millibars. |

En ce qui concerne les mesures relatives à la composante verticale des courants aériens l'évaluation de l'incertitude a donné comme résultat un maximum de 10% dans le cas le plus défavorable.

La réponse dynamique fait intervenir la notion fondamentale de temps de réponse du capteur associé à la vitesse de croisière de l'avion. Ce sont en fait ces éléments qui contribuent le plus à l'incertitude de la mesure. On peut émettre à ce sujet plusieurs remarques.

Tout d'abord, le temps mis par l'avion pour l'exploration d'un tour de cellule doit être petit devant le temps d'évolution de cette cellule. L'expérience a montré qu'il l'était.

D'autre part, la vitesse de croisière de l'avion n'est pas nécessairement constante. Toutefois elle doit rester inférieure à une certaine valeur $v_0$. Cette vitesse associée au temps de réponse $t_0$ d'un capteur définit l'intervalle de variabilité $i_0$ susceptible d'être enregistré, dans l'hypothèse habituellement admise de "l'atmosphère figée" :

$$i_0 = v_0 \times t_0$$

Le capteur possédant le temps de réponse le plus élevé est le capteur d'humidité : $t_0 = 1$ seconde.

La vitesse maximum d'exploration $v_0$ étant fixée à environ 40 m/s, l'intervalle maximum de définition spatiale est de l'ordre de 40 mètres.

D'autre part, le capteur de pression possède une incertitude de mesure équivalant à 40 mètres environ (le temps de réponse n'intervenant pas ici puisque la pression varie toujours très lentement au cours de l'expérimentation).

Toutes ces considérations nous permettent de déterminer le volume élémentaire ou la "maille" de mesure qui peut donc être assimilée à un cube de 40 mètres de côté dans le cas le plus défavorable (humidité).

**b** *Exploitation des paramètres atmosphériques*

Le dépouillement des bandes magnétiques s'effectue sur un ensemble constitué de filtres (passe-bas et passe bande) et de convertisseurs Fréquence-Tension. L'ensemble des mesures est alors transcrit sur papier par enregistrement graphique. On sélectionne ensuite les fractions de mesure qui présentent un intérêt particulier. On sélectionne ainsi une série "d'événements" qui sont exploités d'une manière approfondie. Ce sont en

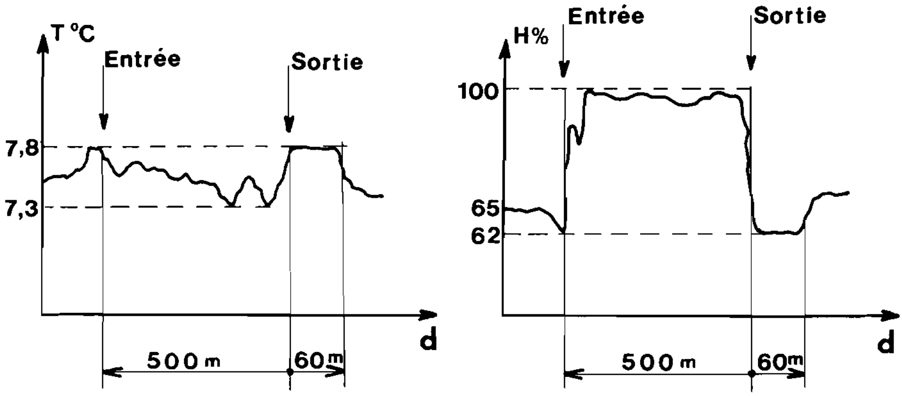


Fig. 6 Variations de la température et de l'humidité au cours de la traversée d'un cumulus.

particulier les paliers horizontaux constituant un tour de cellule et les traversées (Fig. 6).

Les paliers horizontaux permettent de tracer quelques indicatrices de température et de degré hydrométrique. Une telle présentation des résultats possède l'avantage de mettre en évidence les anisotropies existant autour des cellules pour ces deux paramètres (Fig. 7).

Quant aux traversées, leur tracé présente un grand intérêt puisqu'ils permettent une étude approfondie des zones frontières où s'effectue une partie des échanges entre le nuage et son environnement.

L'analyse spectrale constitue l'un des moyens les mieux adaptés à l'étude des phénomènes fluctuants. Dans le problème qui nous occupe, cette analyse peut permettre de mieux caractériser la turbulence associée au développement d'une cellule.

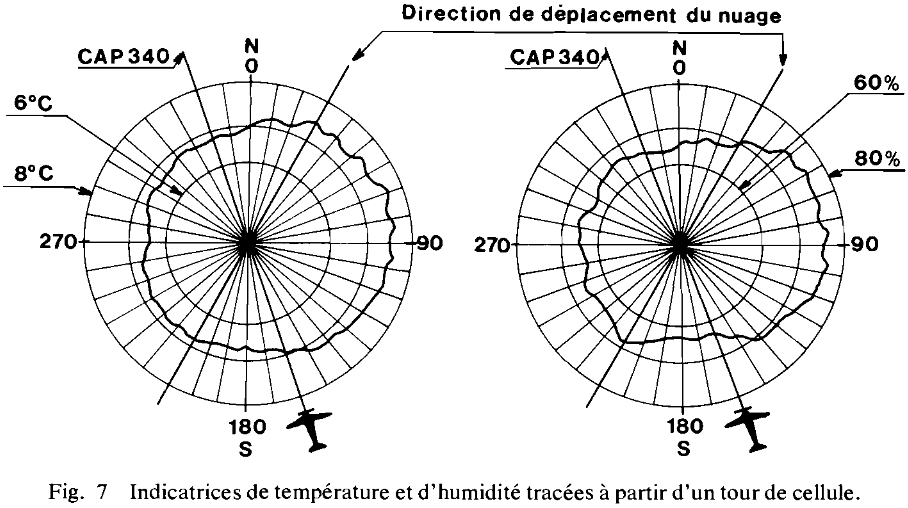


Fig. 7 Indicatrices de température et d'humidité tracées à partir d'un tour de cellule.

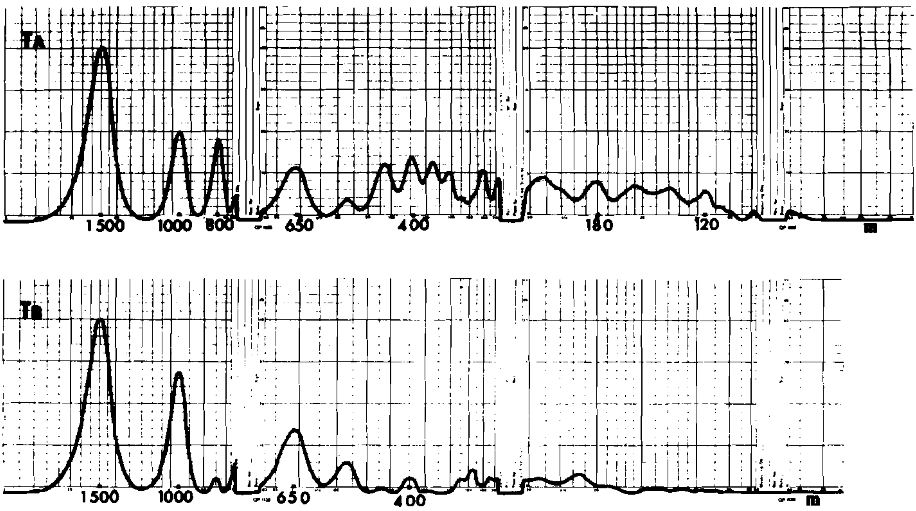


Fig. 8 Deux exemples d'analyse spectrale de la température autour des cellules. L'échelle des abscisses, primitivement graduée en hertz, a été convertie en mètres en tenant compte de la vitesse de vol.

En effet les fluctuations de la température et de l'humidité sont reliées physiquement à la turbulence du milieu. L'analyse spectrale peut alors indiquer l'intensité et la dimension de cette turbulence.

Ce sont en général les mesures effectuées au cours d'un tour de cellule qui sont soumises à cette analyse. Elles ont en effet l'avantage d'obéir en première approximation, au critère de stationnarité (Fig. 8).

Précisons que l'appareillage qui permet d'effectuer l'analyse spectrale a été conçu et réalisé au laboratoire de Physique de l'Atmosphère de Toulouse (Picca et Elefterion).

Les observations faites sur les divers spectres obtenus peuvent faire l'objet d'études quantitatives et qualitatives, par exemple une classification élémentaire d'un certain nombre d'événements.

Dans le cadre des observations quantitatives, les périodicités observées ($t_i$) dans l'analyse spectrale, associées aux vitesses de déplacement de l'avion nous permettent d'évaluer les dimensions : $v_i \times t_i$, des "particules élémentaires" qui constituent l'environnement immédiat du nuage.

Il est important de remarquer que l'analyse spectrale n'est pas en mesure de localiser les différentes zones de turbulence. Cependant, cette indétermination peut être levée par les informations contenues dans les commentaires associés aux divers tours analysés.

**c** *Traitement des paramètres d'ascendances*

On procède tout d'abord à la préparation des données en vue de leur traitement sur l'ordinateur du Centre d'Etude et de Recherche de Toulouse.

Pour cela, les signaux à fréquence variable, correspondant aux cinq

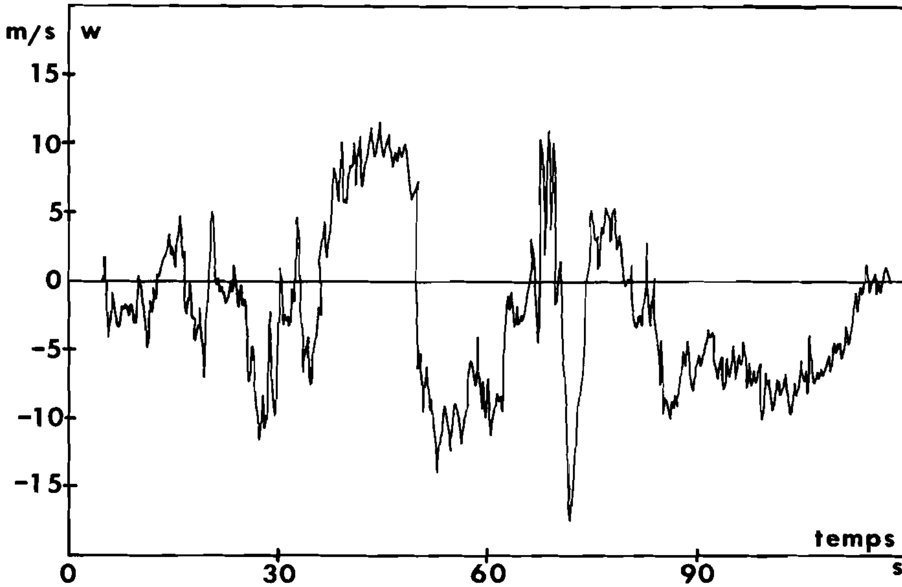


Fig. 9 Un exemple de variation de la composante verticale des courants aériens au cours de la traversée d'un cumulus-congestus.

paramètres sont transformés en signaux analogiques et transcrits sur une seule bande magnétique, les niveaux extrêmes étant −5 volts et +5 volts.

Ces nouveaux signaux sont ensuite échantillonnés avec une période de $^1/_{10}$ de seconde (celle-ci est choisie en accord avec le temps de réponse des divers capteurs). A ce niveau, l'échelle analogique −5v, +5v est convertie en une échelle numérique −32 000, + 32 000, qui permet alors l'introduction des données dans le calculateur.

Le programme de calcul de la composante verticale de la vitesse des courants ne présente aucune particularité notable. Précisons toutefois que les fonctions d'étalonnage correspondant à chaque paramètre sont introduites dans ce programme, ce qui permet d'obtenir un résultat final exprimé en unités physiques (en m/s).

De plus, ce résultat est obtenu d'une part sur table traçante, d'autre part sur cartes perforées pour être éventuellement exploité par ordinateur. La Fig. 9 donne un exemple de mesure de la composante verticale de la vitesse de l'air au cours d'une traversée.

## 5 Conclusion

L'équipement de conception classique qui vient d'être décrit est le résultat de mises au point successives faites au cours de plusieurs campagnes d'essais. Il nous a permis d'observer certains phénomènes relatifs aux échanges entre les nuages et le milieu extérieur. Nous espérons ainsi pouvoir apporter des précisions sur la génération et les mécanismes d'entretien des mouvements de compensation autour des cumulus congestus ou des cumulus mediocris, comme cela a été fait sur les cumulo-nimbus (Shmeter, Ackerman). Une étude plus approfondie de ces problèmes doit conduire à une meilleure compréhension des phénomiènes convectifs.

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# The 15–25 $\mu$m Barrier to Drop Growth in Warm Rain

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ABSTRACT

The condensation-accretion (warm rain) and Wegener-Bergeron (ice crystal) theories of rain formation are briefly outlined. In a number of situations the former process must be applicable. However, conventional models of the warm rain process lead to a 15–25 $\mu$m radius barrier to drop growth in observed cloud lifetimes. Modifications to the process to circumvent this difficulty have been only partially successful. These modifications will be critically reviewed. Giant hygroscopic and precipitable nuclei only explain a limited number of situations. Although uncertainties remain, especially in the treatment at small interaction distances, it is unlikely that changes in collision efficiencies alone can explain the observations, unless the modifications are due to turbulence effects. The stochastic nature of the accretion process also appears unsuccessful in explaining the observations, although there is some uncertainty concerning the applicability of the commonly used stochastic collection equation. It is important to model condensation and accretion simultaneously, but this consideration alone does not completely account for the short observed precipitation formation times. A recent model incorporating mixing processes has been successful in explaining the main characteristics of maritime cumuli clouds, but formation times for continental cumuli remain excessive. Although the considerations mentioned above reduce the discrepancy between theory and observation, it is not yet completely resolved. The importance of turbulence, the acceptability of the stochastic collection equation, and the suitability of present cumulus cloud models would appear to be the three most important unresolved questions.

## 1 Introduction

The formation of precipitation may conveniently be regarded as a two stage process. First, water vapour condenses on suitable nuclei, forming, after approximately 10 minutes, cloud droplets of radii 10 $\mu$m in typical cloud concentrations of 100 cm$^{-3}$. Secondly, large numbers (approximately $10^4$ to $10^7$) of these 10 $\mu$m droplets are brought together to form precipitation drops. This second process is the topic of the present review.

Two main mechanisms have been suggested to accomplish this growth from small (10 $\mu$m) droplets to precipitation size:

(a) Accretion: Growth by the condensation process will naturally lead to some dispersion in droplet radii at any one time. One may study the relative movement of different sized drops by equating the gravitational force to the Stokes drag force, obtaining, for spherical particles:

$$v = \frac{2\rho_w}{9\eta} g r^2 \tag{1}$$

where $v$ is the downward velocity of the droplet relative to the air, $\rho_w$ is the density of the water droplet, $\eta$ is the dynamic viscosity, $g$ is the acceleration due to gravity, and $r$ is the droplet radius. Thus we see that larger droplets will fall faster, thereby tending to sweep up smaller droplets as they fall. In convective clouds the time available for accretion is increased since the updraft velocity is greater than the downward velocity of the smaller droplets given by equation (1), and a drop will be carried upward until it grows large enough for its downward velocity to exceed the air updraft velocity. In order to quantify our model it is necessary to specify the efficiency of the accretion process. Two factors need be considered. First, one must determine the number of droplet collisions which occur, and secondly the probability that two drops, once they have collided, will coalesce must be specified. In a time $dt$ a large drop sweeps out a cylinder of length $(U\text{-}u)dt$ and radius $R + r$, where $U,R$ are the velocity and radius of the large drop and $u,r$ are the corresponding quantities for the small droplets (assumed identical). However, the large drop does not collide with all droplets in this volume since air motion around the large drop will tend to deflect smaller droplets from it. The ratio of the number of drops actually colliding to the number in the geometric volume mentioned above is termed the collision efficiency. The probability that a collision of a droplet with the drop will result in accretion is termed the coalescence efficiency, and the product of the coalescence and collision efficiencies is termed the collection efficiency.

Values of the coalescence efficiency are uncertain, but are believed to be close to unity for drops smaller than about 50$\mu$m radius, and perhaps even for considerably larger (60–200 $\mu$m) drops (Jayaratne and Mason, 1964). However, later experiments (e.g. Whelpdale and List, 1971; Levin *et al.*, 1973) suggest that in at least some types of collisions involving larger drops the value of the coalescence efficiency may be significantly less than unity. Theoretical calculations of the collision efficiency, based on non-turbulent flow about the large drop and several other simplifying hydro-dynamic assumptions, were carried out by Hocking (1959). He found that while collision efficiencies were near unity for collisions between large (greater than about 25 $\mu$m radius) drops and smaller ones of radii between about 0.4 and 0.9 that of the larger drop, collisions between drops of radii less than about 18 $\mu$m and smaller drops were forbidden. His results, coupled with the small relative velocities of such drops and the smaller cloud droplets, meant that growth of drops smaller than about 25 $\mu$m radius would not proceed by accretion, although accretion was the dominant process for growth of drops larger than this size. A problem then arose in explaining the growth of precipitation sized drops in some types of clouds in times of the order of 30 minutes and less. This is because although initial growth on common nuclei by condensation is quite rapid, growth to 25 $\mu$m by condensation under typical cloud conditions takes about $1^1/_2$ hours and the condensation-accretion mechanism appears to present a 15 $\mu$m to 25 $\mu$m barrier to rapid drop growth.

(b) Wegener-Bergeron Process: Partly because the above problem with the accretion process was suspected even before Hocking's formal calculations were carried out, an alternate theory of rapid precipitation formation was advanced by Wegener (1911) and Bergeron (1935) (the same process is sometimes referred to as the Bergeron-Findeisen process because of the work of Findeisen 1938, 1939). Because growth of ice crystals by sublimation-diffusion is much more rapid than the growth of water drops by condensation-diffusion, Wegener, Bergeron and Findeisen suggested that precipitation was often the result of rapid initial growth of ice crystals in the upper portions of clouds, followed by melting in the lower cloud and below the cloud base. Since rapid precipitation formation is often associated with convective clouds whose tops have temperatures below the freezing point, such a theory appears plausible.

However, although the Wegener-Bergeron process may be responsible for much of the precipitation in mid and high latitudes, there is widespread evidence of rapid formation of precipitation, especially in the tropics but also in mid-latitudes, in clouds whose tops do not even approach the freezing point (e.g. see pp. 288–297 of Mason, 1971). While it has been suggested (Braham and Spyers-Duran, 1967) that ice crystals falling from undetected higher (e.g. cirrus) clouds may explain such precipitation, there are clearly many cases when this process cannot be active. Furthermore, radar results of Braham (1964) suggest that even in ice crystal clouds, droplet growth by accretion often precedes freezing.

Thus we are faced with the problem of rapid formation of precipitation in non-freezing clouds (Mason and Ghosh, 1957; Mason, 1969). The classical condensation-accretion theory appears to present a 15–25 $\mu$m barrier to such rapid formation, while the Wegener-Bergeron process is clearly not active in such clouds. The remainder of this review will outline progress that has been made in modifying aspects of the condensation-accretion model to make its predictions consistent with observation. Such modifications fall into five classes; those dealing with:

(a) the presence of types of nuclei which either allow the accretion process to begin immediately on a nucleus of 25 $\mu$m or larger, or allow condensation to proceed to larger drop sizes rapidly;
(b) changes in the collection efficiency values arising from more realistic calculations and experiments;
(c) the discrete (versus continuous) nature of the accretion process, and possible splitting of drops;
(d) the importance of considering condensation and accretion simultaneously;

and
(e) the importance of mixing processes.

## 2 Presence of precipitable and/or giant hygroscopic nuclei

A solution to the problem is found if it can be shown that sufficient numbers of nuclei greater than about 25 $\mu$m radius (termed precipitable nuclei) exist prior to precipitation, since such particles are large enough to commence growth immediately by the accretion mechanism. A recent review of the role of aerosols in precipitation is given by Rosinski (1974). Ignoring, for the present, splitting of drops (see section 4), one requires at least one suitable nucleus for each raindrop formed. While under certain special conditions (e.g. severe convective storms when these large nuclei may be injected by updrafts; see Rosinski and Kerrigan, 1969) such concentrations may be present, in general this is certainly not true.

It should also be mentioned that the presence of larger than normal soluble condensation nuclei (termed giant hygroscopic nuclei) in suitable concentrations may also explain the rapid growth of precipitation drops, since with such nuclei the condensation process can rapidly produce drops of 25 $\mu$m radius. Mason (1969) points out that a $10^{-9}$ g hygroscopic nucleus will grow to 25 $\mu$m by condensation in less than 5 minutes under typical cloud conditions. Rosinski (1974) states that a 2.5 $\mu$m radius sea salt nucleus will grow to 50 $\mu$m radius during the lifetime of a typical maritime cloud, and precipitation may be accounted for if concentrations of the order of $10^3$ m$^{-3}$ of such nuclei are present. Some rapidly forming maritime clouds may be explained by the presence of giant hygroscopic nuclei.

## 3 Collection efficiency estimates

As pointed out in section 1, growth by accretion of droplets smaller than about 25$\mu$m is inhibited by the small velocity of such drops relative to typical (10 $\mu$m) cloud droplets, and forbidden by the zero collision efficiencies derived by Hocking (1959). Since unpublished calculations by Bartlett (related by Woods et al., 1972) have shown that the small relative velocities are not prohibitive in explaining observed precipitation growth rates if collision efficiencies near unity are assumed, it is reasonable to examine the collision efficiency calculations.

Collision efficiency calculations have been undertaken in three main ways. The simplest method assumes that the motion of each drop may be calculated from the flow pattern of air about the other, without taking into account the interactions between the two flows. Such calculations have been carried out by Langmuir (1948), Pearcey and Hill (1957) and Shafrir and Neiburger (1963). While calculations of this type are adequate for large separations, and give approximate collision efficiencies for relatively large (greater than about 50 $\mu$m) drops and typical cloud droplets, the treatment probably leads to erroneous values for the critical case of 10–30 $\mu$m drops.

A more refined approach, which considers the interactions between the two flows but assumes the Reynolds' number is low enough to ignore non-linear terms in the equation of motion, was undertaken by Hocking (1959). The Stokes equations result from such an assumption. While the results of Hocking differed considerably from the non-interaction treatments, both sets of results lead to such low collection efficiencies for 20 $\mu$m and smaller drops that observed precipitation growth rates could not be explained.

Davis and Sartor (1967) followed essentially the same approach as Hocking (1959), but carried out more accurate numerical calculations, and also included rotation effects (which they showed to be important), and Hocking and Jonas (1970) made similar more accurate calculations, still employing the Stokes equations.

One problem which arises from the Stokes equations is the failure of the equations at small interaction distances. This is because these equations lead to an infinite force as the interaction gap goes to zero, and thus collisions would never occur. This is clearly physically impossible since the air cannot provide an effective shield at distances less than one mean free path. It appears difficult to properly model the process at such distances, and the conventional procedure is to assume that the force goes to zero when the droplet approaches within a distance $eR$ of the larger drop (of radius $R$), where $\epsilon$ is an arbitrary constant which is usually assumed to be $10^{-3}$ to $10^{-4}$. Hocking and Jonas (1970) show that the specific value assumed for $\epsilon$ is not important for larger drops, but becomes critically important for the small drops with which we are most concerned. Michael and Norey (1969) point out that the generation of polarization image charges and surface irregularities, as well as the breakdown of the continuum air model, may be important at small separation distances. If the values of $\epsilon$ assumed are reasonable, Bartlett (1970) has shown that the revised collision efficiencies of Davis and Sartor (1967) and Hocking and Jonas (1970) (which lead to finite collision efficiencies for all radii but values of only a few hundredths and less for drops smaller than 20 $\mu$m) lead to more rapid growth of 25 $\mu$m drops but slower growth of 30 $\mu$m drops, and overall the results are not much different from those obtained by using the earlier values of the collision efficiency.

Davis (1972) has used slip flow theory to allow for gas kinetic effects and has thereby removed the arbitrary constant $\epsilon$ from the Stokes treatment. He obtained values of the collision efficiency somewhat higher than those resulting from earlier Stokes treatments, especially for small drops (e.g. 10 $\mu$m with 6 $\mu$m collision efficiency is increased by about 5 times whereas the 20 $\mu$m and 12 $\mu$m collision efficiency is only increased by a factor of about 1.5). He concludes that the increases are not great enough to account for the observed precipitation times.

Recently Klett and Davis (1973) have shown that even for the small (less than 30 $\mu$m) drops considered (Reynolds' number of about 0.2), the non-linear terms which were ignored by Davis and Sartor (1967) and Hocking and Jonas (1970) may indeed be important, and thus the Stokes equations may not be valid for the problem. Physically these non-linear terms represent the inertia of the air. Revised calculations taking into account such terms lead to collision efficiencies several times larger in the critical 15–25 $\mu$m range. Ryan (1974) has calculated the effect of these values on required growth times.

Even more recently Beard and Grover (1974) have recalculated collision efficiencies, but their results are most applicable to Reynolds' numbers corresponding to drops of 40 $\mu$m and greater, and thus are not useful for the problem of the 15 $\mu$m to 25 $\mu$m condensation-accretion barrier.

Manton (1974a) has used dimensional arguments to show that the collection kernel, $K$ (defined as the product of the collision efficiency and the geometric volume swept out by the drop as discussed in section 1), must, for small drops, have the functional form:

$$K \propto rR \ln(R/r) \tag{2}$$

where $r$, $R$ are the radii of the two drops. This is an important result since it allows one to extrapolate experimental or theoretical results to other values of $r$, $R$. Agreement of the functional form given by equation (2) with the calculated values of Davis (1972) for $R = 10$ $\mu$m and $r$ between 3 $\mu$m and 9 $\mu$m is good.

Considering the complexity and uncertainty in the theoretical collision efficiencies, several workers have attempted experimental determinations of the collection and/or collision efficiencies. Sartor (1954) modeled the droplet interaction process in air by that of mineral water in water, keeping Reynolds' numbers equivalent, and found collision efficiencies subtantially higher than those theoretically predicted (e.g. 0.5 for collision between 8 and 14 $\mu$m drops). However, Woods and Mason (1964) have pointed out that the difficulty with such experiments is the inability to simultaneously model both the Reynolds' number and the ratio of the density of the drop to that of the surrounding fluid.

An important experimental study of Kinzer and Cobb (1958) finds high values of the collision efficiency for drops less than about 12 $\mu$m with typical cloud droplets, as well as the theoretically predicted high values for drops with radii greater than 25 $\mu$m. Their experiments suggest that this increase at low drop radii is due to the turbulent wake behind the drop tending to catch droplets that would otherwise escape collision. The combined effect of the wakes of many drops is an induced turbulent flow field which leads to a random motion of the smaller drops similar to Brownian motion (although several orders of magnitude greater) and ultimately to higher collision efficiencies. They develop a simplified theory of such motion which agrees moderately well with their experimental results. The main difficulty in applying these results to natural clouds is that the process is dependent on the drop concentration (which was 1500 to 2900 $cm^{-3}$ in their laboratory studies). They predict the process would still be important for drop concentrations of about one-tenth this value, which are more typical of warm cumulus clouds. However, Woods and Mason (1964) point out several difficulties in the experimental arrangement used by Kinzer and Cobb which may explain the results they obtained.

The importance of the wake (which is not adequately modeled by the theoretical approximations) was earlier pointed out in a study of larger (about 75 $\mu$m radii) drops by Telford *et al.* (1955). They found collision efficiencies significantly greater than unity (up to about 12) and attributed the increase to the effect of the turbulent wake. However, a later study by Woods and Mason (1965) failed to confirm these findings in a study of the collision efficiency of droplets in the 30–110 $\mu$m range.

Finally, an experimental study of drops in the size range 30–55 $\mu$m (droplets of 1–12$\mu$m) by Woods and Mason (1964) approximately confirmed the theoretically determined collision efficiencies, although it is really the values for the smaller drops that we are more interested in.

Recently there has been interest in the possibility that natural cloud turbulence (independent of any turbulence produced by the wakes of falling drops) may be important in increasing collision efficiencies. A study by Bartlett and Jonas (1972) showed that turbulence had little or no effect on the growth, or size distribution of drops by the condensation process. Woods *et al.* (1972) showed that droplet collisions occur on length and time scales where the turbulence induced shears are much less than their peak values, which occur on the much larger dimensions of the order of the Kolmogoroff scale length. Thus, droplet collisions in a turbulent cloud can be modeled by a constant uniform shear. Experiments they conducted on 9.5 $\mu$m droplets collecting on larger drops in 17.5 $s^{-1}$ shears showed larger collection efficiencies than theoretically predicted for non-sheared flow. An increase of about 3 times was noted for collisions with 20 $\mu$m drops and about 20 times for 15 $\mu$m drops. These results were so encouraging that a more detailed study was carried out by Jonas and Goldsmith (1972). They largely confirmed the earlier results, and showed that the shear-induced increase in the collection efficiency could effectively be represented by a linear increase with shear after a certain cutoff value was reached. This shear cutoff value was lower for smaller drops, being about 2 $s^{-1}$ for 10 $\mu$m – 9 $\mu$m interactions and 9 $s^{-1}$ for 30 $\mu$m – 9 $\mu$m interactions. Tennekes and Woods (1973) looked more carefully at the question applying these results to actual cumulus clouds. Using the data of Jonas and Goldsmith (1972) and the best available shear spectrum data, they showed that turbulence in a typical warm cumulus cloud would be expected to increase the 20 $\mu$m – 9 $\mu$m collection efficiency by a factor of about 7. Recently Manton (1974a) has proposed a theoretical basis for the increased collection efficiencies in sheared flows. Qualitatively, he assumes that there is an interaction region associated with the wake of a drop such that any other drop swept into this region eventually collides with the initial drop. The effect of a vertical shear is to sweep drops horizontally into the interaction regions of other drops, and thus increase the coalescence rate. This is supported by the results of Jonas and Goldsmith (1972) which showed that vertical shears were more effective than shears inclined at 45° in increasing the collection efficiency. Manton (1974a) uses dimensional arguments to obtain the form of the collection kernel for sheared flow.

A number of workers have investigated whether electric fields and/or charged droplets could significantly increase the collection efficiency. An originally charged drop would be expected to polarize approaching drops and thereby increase the collection efficiency. An assembly of charged drops of mixed sign might be expected to have an increased collection efficiency due to electrostatic attraction. An experimental study by Woods (1965) and theoretical calculations by Semonin and Plumlee (1966) confirm that charged droplets enhance collision efficiencies linearly with increasing charge. However, the effect only commences after a certain threshold charge of about $10^{-16}$ C (about 800 elementary charges) per drop is reached. Since the precipitation we wish to explain falls from only moderately convective clouds, typical drop charges are 10 elementary charges and thus charged drops will play no role in increasing the collection efficiency.

Electric fields may enhance the accretion process either by inducing polarization charges which increase the efficiency of accretion, or by causing slightly charged drops to travel in straight lines due to the electric force, and thus not be deflected by aerodynamic forces from an impending collision with another drop. The effects of electric fields on accretion have been studied theoretically by Lindblad and Semonin (1963), Plumlee and Semonin (1966) and Hocking and Jonas (1970), and experimentally by Woods (1965) and Latham (1969). The results of the various workers do not vary significantly. For electric fields greater than about $5 \times 10^4$ V m$^{-1}$ the collection efficiency increases rapidly, and thus electric fields following lightning discharge in thunderstorms are of considerable importance in enhancing accretion (Moore *et al.*, 1962, 1964). However, in the less convective type of warm cumulus cloud with which this paper is concerned, it is unlikely that electric fields play a significant role in the condensation-accretion process.

## 4 The discrete nature of the accretion process

The early models of growth by the accretion process pictured the air as being uniformly filled with liquid water which the larger drops swept up continuously. According to such a model, all large drops (of the same initial size) grow at the same rate. However, as Telford (1955) first pointed out, such a treatment only calculates the average rate of growth of all drops. Because the actual process is discrete, and once a drop collects one smaller droplet its chances of gaining still another are enhanced over those of the other drops, some drops will grow to a large size relatively quickly. Such a discrete process is termed "stochastic", and an equation (popularly termed the kinetic, collection or coalescence equation) to describe the process may be derived (see e.g. Mason, 1969).

The stochastic equation was employed by Bartlett (1966, 1970), Berry (1967), Twomey (1966) and Warshaw (1968b) to study what differences from the predictions of the continuous model resulted. In general, although the growth of a few large drops is rapid, observed precipitation formation times cannot be explained unless higher liquid water contents or a significant (few $\mu$m) lowering of the radius at which collision efficiencies become negligible are assumed. Twomey (1966) points out that the observations can perhaps be reconciled if small regions of higher (1–3 g $m^{-3}$) liquid water content exist, even for short time periods (1 minute). Unfortunately, there are few observations of liquid water content in different small regions of a cloud (Mason, 1971).

The stochastic collection equation may be expressed as:

$$\frac{dN_i(t)}{dt} = \frac{1}{2}\sum_{j=1}^{i-1} N_j(t)N_{i-j}(t)C_{j,i-j} - \sum_{j=1}^{\infty} N_i(t)N_j(t)C_{i,j} \tag{3}$$

where $N_i(t)$ is the average concentration of cloud droplets of size $i$ at time $t$, and $C_{ij}$ is the collection probability for droplets of sizes $i$ and $j$. The first term in the equation represents growth of $i$ sized drops from smaller drops while the second term represents loss of $i$ sized drops by accretion with other drops. Several other equivalent forms of the equation are also commonly used. There has been widespread controversy over whether equation (3) properly models the accretion process. In particular, there has been disagreement over the statistical deviations to be expected from the average behaviour predicted (see e.g. Scott, 1967, 1968b, 1968c, 1972; Warshaw, 1967, 1968a; Long, 1971, 1972 and Slinn and Gibbs, 1971). Besides the question of its applicability, several workers (e.g. Scott, 1968a and Chin and Neiburger, 1972) have been concerned with the effects of different initial droplet size distributions. Gillespie (1972) has given a complete and illuminating discussion of the foundations of the stochastic coalescence equation. He shows that equation (3) may be derived subject to the assumptions that:
(a) the probability of having a certain number of droplets of one size present is independent of the numbers of droplets of other sizes present; and
(b) the probability of coalescence of equal sized drops is zero.
Also following from these two assumptions is a measure of the statistical fluctuations about the average number of drops a certain size to be expected after a time $t$. If $t$ is greater than the half life of drops of that size, the number probability distribution will be Poisson (independent of the original droplet size spectrum). However, the statistical fluctuations in the number of drops of a certain size present in any particular volume of a well mixed cloud will be somewhat greater than these stochastic fluctuations in the number of that size present in the entire cloud.

Assumption (b) is probably reasonable on the physical grounds that two drops of exactly the same size will have the same fall velocity and will not approach one another. Even if one argues that turbulent aerodynamic forces or electrical forces can bring such drops together, in any reasonably large spectrum of drop sizes collisions between drops of the same size will be mathematically unimportant. However, the grounds for assumption (a) are neither so obvious nor so convincing. One could envisage several circumstances in which the presence of droplets of one size would be correlated with the presence or absence of droplets of another size. For example, a large drop would be effective in sweeping out droplets of a certain size, and thus the region through which it had fallen would show a correlation amongst drops of different sizes which would affect future accretion in that region. Alternately, the distribution of nuclei and the condensation process might lead to some correlation of drops. Gillespie (1972) suggested that additional study of assumption (a) was required. Bayewitz *et al.* (1974) have recently considered this problem of correlation. In order to make the problem mathematically tractable they considered the special case of a constant collection coefficient, and thus their results are not immediately applicable to the case of warm rain formation. They found that in poorly mixed or small populations the true stochastic means may be significantly different from those predicted by equation (3). Furthermore, the tail of the size distribution is most affected, and thus the number of large raindrops produced in a coalescing system may be significantly different from that predicted by equation (3). Also, the authors suspect that these discrepancies will be even greater if more realistic (size dependent) collection coefficients are used. Until further work is undertaken, no definitive conclusions can be drawn at this time.

Monte Carlo simulation techniques (e.g. Chin and Neiburger, 1972 and Robertson, 1974) have recently been employed to study drop accretion. Such techniques offer promise in obtaining solutions for a variety of initial conditions, and also for modeling non-uniform cloud systems, although computer requirements may become prohibitive for truly realistic cloud calculations.

It occurred to List et al. (1970) that collisions between larger drops, which usually result in the creation of several smaller "satellite" drops which are still large enough to allow rapid growth by accretion, may, when combined with the stochastic nature of the process which permits rapid growth of a few large drops, explain the observed precipitation growth times. However, a later study by Brazier-Smith et al. (1973) showed that the effects on precipitation formation times are slight, although the process is important in altering the spectrum of drop sizes produced.

## 5 Simultaneous condensation-accretion

For the sake of mathematical expediency the early calculations assumed condensation to take place up until a certain time, producing an initial spectrum of drop sizes. Following that time no further condensation was considered, with accretion being the sole process for the growth of drops. In a real cloud situation, of course, condensation will continue after the coalescence process begins, and it may well have an important indirect effect on the resulting growth by coalescence through controlling the supply of small droplets. Kovetz and Olund (1969) considered condensation and coalescence simultaneously and confirmed that the rate of production of large drops was increased by considering continuing condensation. More recent models (e.g.

Leighton and Rogers, 1974; Ryan, 1974), some of them incorporating other processes as well (e.g. turbulence-induced coalescence – Manton, 1974b; breakup of drops – Young, 1975), have confirmed the importance of considering condensation and coalescence simultaneously and suggest that earlier treatments separating the two phenomena may be of limited value.

## 6 Mixing Processes

Any realistic treatment of the processes occurring within a cloud must surely take into account effects due to mixing with the surrounding environment. In general, such mixing will result in a loss of drops and a gain of fresh nuclei and water vapour. The first attempt to accurately include mixing processes in warm cumuli clouds was carried out by Mason and Chien (1962). The simplified model they used resulted in a broadening of the droplet size spectrum, but explained neither the rapid growth of drops nor the observed bimodal drop size distribution. A study by Bartlett and Jonas (1972) suggested mixing as the most probable agent for producing the observed drop size spectrum, and so recently several more detailed studies of mixing processes have been made. A study by Warner (1973) showed that simple mixing can not explain the observed drop size spectrum.

However, Mason and Jonas (1974) and Jonas and Mason (1974) considered the rise of successive thermals in cumulus clouds. They allowed the condensation and accretion processes to occur simultaneously for the reasons pointed out in section 5. Their model is successful in predicting the observed size spectrum and a rate of formation of precipitation sized drops in a typical maritime cloud, but the greater number of small drops initially present in a typical continental cumulus cloud leads to times that are still much longer than observed. The mixing model of Mason and Jonas has recently (Warner, 1975) been criticized. The main criticism of the model is that it apparently results in liquid water contents near the base of warm cumulus clouds which are significantly greater than measured values. Mason (1975) argues that the measured values refer to averages across a cloud cross-section composed of residues of a number of thermals, and not to the water content of an active rising thermal. In any case, the model of Mason and Jonas appears the most successful to date in approximating the bimodal drop size spectrum, and certainly more effort should be exerted in properly modeling cloud-environment mixing and the resulting drop growth rates. Finally, it should be noted that the problem of theoretical modeling of cumulus clouds has been thoroughly reviewed very recently by Cotton (1975).

## 7 Conclusions

A review of recent attempts to modify the condensation-accretion model of drop growth to overcome the apparent 15–25 $\mu$m radius barrier has been made.

(a) The presence of giant hygroscopic or precipitable nuclei can only explain certain of the observations. Giant hygroscopic nuclei probably play an important part in some warm maritime clouds.
(b) Recent, more accurate, calculation of collision efficiences slightly reduce the time required to produce precipitation sized drops, but not significantly. Problems in describing the movement of the drops at small interaction distances and difficulties in modeling the turbulent wakes of falling drops plague most such theoretical calculations. Experimental determinations of collision efficiencies meet with their own difficulties. Agreement between different experimental results is poor, and few definite conclusions can be drawn from the data available at this time. It can be said with some certainty that charged drops and/or electric fields play little role in the accretion of mildly convective clouds. Turbulence, either that introduced by the wakes of falling drops or that inherent in the cloud, may be of great importance in increasing collision efficiencies in the critical 15 to 25 $\mu$m range.
(c) The discrete nature of droplet accretion, if properly modeled by the conventional stochastic collection equation, does not, by itself, lead to reasonable precipitation formation times. However, discrepancies are less severe than for the simpler continuous model. The assumptions inherent in the stochastic collection equation have recently been examined. The most questionable assumption is that of no correlation between the numbers of drops of different sizes present in a certain volume of the cloud. This restriction may make the stochastic collection equation unacceptable in poorly mixed systems. The importance of the effect in real cloud situations is still uncertain.
(d) The times required to produce precipitation-sized drops are reduced somewhat if condensation and accretion are considered to occur simultaneously. This factor alone is insufficient to overcome the barrier, however.
(e) One model of mixing processes leads to realistic drop size spectra and growth times for maritime cumulus clouds, but excessively long times for continental cumuli persist. More realistic mixing models are, however, warranted.

In summary, it would appear that although the gap between observation and theory has been lessened in recent years, problems, at least with respect to warm continental cumuli, persist. Three aspects of the theory appear to warrant particular attention in the near future. Additional study, both theoretically and in the laboratory, of the importance of turbulence-related wind shears in enhancing collection should be undertaken. Also, it is important in this regard to obtain a better understanding of the turbulent structure of cumulus clouds. Secondly, the problem of droplet correlations in applying the stochastic collection equation to actual clouds must be resolved. Thirdly, more realistic models of complete cloud systems, incorporating both microphysical processes and cloud dynamics and thermodynamics, should be developed.

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## CALL FOR NOMINATIONS – 1975 AWARDS

Nominations are requested from members and Centres for the 1975 CMS Awards to be presented at the 1976 Annual Meeting. Four awards are open for competition: 1) the President's Prize for an outstanding contribution in the field of meteorology by a member of the Society; 2) the Prize in Applied Meteorology, for an outstanding contribution by a member in that field; 3) the Graduate Student Prize, for a contribution of special merit, and 4) the Rube Hornstein Prize in Operational Meteorology, for outstanding service in providing operational meteorological service. The awards will be made on the basis of contributions during the 1975 calendar year, except 4) which may be awarded also for work performed over a period of years.

Nominations are also requested for the award of citations to individuals or groups in Canada who have made some outstanding contribution in helping to alleviate pollution problems or in developing environmental ethics.

All nominations should reach the corresponding secretary not later than *March 1, 1976*.

# SNOW DEVILS – EARLY OBSERVATIONS

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ABSTRACT

Early observations of snow devils are reviewed and compared to those by Storr (1972). Possible mechanisms involved in their formation are suggested, based on studies of dust devils.

In the atmosphere there exists a hierarchy of small-scale vortices; those with roughly vertical axes range from whirlwinds or dust devils to tornadoes. In the formative process, vertical instablility due to intense surface heating or release of latent heat is a major factor; the role of mechanically induced eddies however is less well understood. Consequently, the observations by Storr (1972) of "snow devils" are valuable additions to the descriptive literature.

An earlier occurrence of snow devils was noted by Wegener (1914) during the "Danmark" Expedition (1906–08) to Greenland. Storr's observations and conclusions are similar to those of Wegener. Because of the snow cover, both authors surmised that surface heating played a minimal role, and that the snow devils were driven by purely mechanical forces. (Although emphasizing the difficulties in determining the day-time lapse rate near a snow-covered surface, Geiger (1966) affirms that an inversion may exist even within the lowest layers.) Storr offers a sketch to illustrate a possible relation between flow over a nearby ridge and the generation of snow devils in the lee. Wegener attributed the vortex formation to flow through the narrow Mörkefjord. Significantly, while Storr's snow devils were visible to only 15 m (40–50 ft.), those observed by Wegener extended to the height of the adjacent plateau – 800 m.

At Adelie Land during the Australasian Antarctic Expedition (1911–1914), Madigan (1929) notes that during periods of comparative calm and variable winds "small whirlwinds raising snow, like miniature 'willy-willys' with their dust columns in Australia, and also low fracto-cumulus cloud forming over the coast line, swirling round, drifting north and quickly evaporating" were characteristic phenomena. The "whirlies" (Mawson, 1915) were of up to a few hundred meters diameter. Mawson further describes how they "tracked

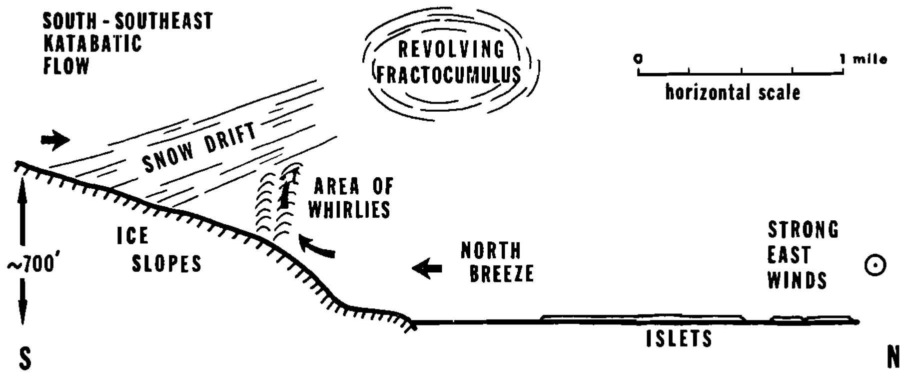


Fig. 1 Schematic diagram of conditions attending the appearance of whirlies at Cape Denison (after Madigan).

about in a most irregular manner''. The high velocity of the wind in the rotating column was evident when one snow devil moved a 300–350 pound lid fifty yards, then later another picked it up again and returned it to near the original vicinity. Over the sea, ''columns of brash-ice, frozen spray and water-vapour were frequently seen lifted to heights of from two hundred to four hundred feet, simulating waterspouts.''

Since the ''whirlies'' arose during periods of relative calm, Humphreys (1929) speculated that the snow devils resulted from strong currents aloft (rather than concentration of low-level vorticity). Indeed, horizontal vorticity due to vertical wind shear of lee-side eddies may be the source of angular momentum. Relief, however, from the usually fierce southerly winds may have come about due to various factors.

Inspection of the data reveals that, on occasion, the katabatic wind was probably temporarily counteracted by a northerly wind, during which time snow devils were observed. Detailed accounts note that the usual roar of wind was still present but apparently aloft. Madigan (1929) incorporated the observations into a single concept (Fig. 1). It would seem that the much colder katabatic wind should undercut the maritime air from the north; perhaps, however, due to the terrain, separation occurred allowing warmer air to undercut the very cold flow. Superadiabatic lapse rates could then develop setting the stage for the snow devil development.

If buoyancy is *not* a factor in the formation of snow devils, then these vortices may represent a simplification of the (possibly) rotor-induced dust devils observed by Hallett (1969, 1971). (Certainly the sketches by Storr and Madigan bear a resemblance to Hallett's photograph on the cover of *Weather*.) In any case, however, a mechanism must yet be found for transforming the ambient vorticity into the axial vorticity manifest in the snow devils. In this regard, Maxworthy (1973) contends that a tilted vortex may ''scoop up'' and subsist on boundary layer vorticity and, moreover, by self-induction, maintain the motion relative to the ambient wind field required to persist.

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## APPEL AUX CANDIDATURES POUR LES PRIX ET CITATIONS 1975

On demande aux membres et aux centres locaux de soumettre leurs nominations aux candidatures pour les prix de la Société pour l'année 1975. Il y a quatre prix: 1) le prix du président pour un travail exceptionnel en météorologie par un membre de la Société, 2) le prix de météorologie appliquée pour un travail exceptionnel dans ce domaine par un membre, 3) le prix aux étudiants gradués et 4) le prix de météorologie opérationnelle Rube Hornstein pour un travail exceptionnel dans l'exploitation des services météorologiques.

Tous les prix seront attribués pour un travail qui a été effectué durant l'année 1975, à l'exception de 4) qui peut aussi être attribué pour un travail effectué durant une période couvrant plusieurs années.

On demande aussi des nominations des candidats susceptibles de se voir décerner une citation par la Société. Une citation peut être décernée à un individu ou à un groupe qui a apporté une contribution exceptionnelle à la solution des problèmes de la pollution, à l'amélioration de l'environnement ou au développement d'une éthique écologique.

Toutes les nominations recueillies par le secrétaire-correspondant avant *le 1$^{er}$ mars 1976* seront remises aux comités des récompenses et des citations, selon le cas.

## CALL FOR PAPERS – TENTH ANNUAL CONGRESS

The Tenth Annual Congress and Annual General Meeting of the Canadian Meteorological Society will be held at Laval University, Quebec City, Quebec at the end of May 1976, the exact date to be announced in due course. The theme of the congress will be *Observational Networks*. In addition to sessions on the theme, there will be the usual sessions on other topics in meteorology according to the papers submitted.

The new Oceanography Division of the CMS will be organizing sessions on oceanography at this congress, as at the last, and we cordially invite contributions.

Titles and definitive abstracts (less than 300 words) should reach the Program Committee by *1 February 1976*. Papers on meteorology should be sent to Dr. Gaston Paulin, Service de la Météorologie du Québec, Ministère des Richesses Naturelles, 194 rue St-Sacrement, Québec, Québec. Papers on oceanography should be sent to Dr. C. Mann, Bedford Institute of Oceanography, P.O. Box 1006, Dartmouth, Nova Scotia.

## CONTRIBUTIONS SCIENTIFIQUES AU DIXIEME CONGRES ANNUEL

La fin de mai 1976 sera marquée par la tenue du dixième congrès annuel et de la réunion générale annuelle de la Société météorologique du Canada dans la ville de Québec. Les dates exactes du congrès seront publiées plus tard. Les réseaux d'observation seront le thème central du prochain congrès. Toutefois, suivant la coutume, plusieurs autres sujets en météorologie pourront être présentés après soumission.

Pour la deuxième année consécutive, la division nouvelle d'Océanographie à la S.M.C. organisera des sessions traitant de cette science. Nous espérons recevoir de nombreuses contributions scientifiques dans ce domaine.

Les titres ainsi que les sommaires définitifs (300 mots ou moins) devront parvenir au comité du programme d'ici le *1er février 1976*. Les textes scientifiques en météorologie devront être envoyé au Dr. Gaston Paulin, Service de la Météorologie du Québec, Ministère des Richesses naturelles, 194, Ave St-Sacrement, Québec, tandis que ceux en océanographie devront être envoyés au Dr. C. Mann, Bedford Institute of Oceanography, P.O. Box 1006, Dartmouth, Nova Scotia.

## INFORMATION FOR AUTHORS

**Editorial policy.** *Atmosphere* is a medium for the publication of the results of original research, survey articles, essays and book reviews in all fields of atmospheric science. It is published quarterly by the CMS with the aid of a grant from the Canadian Government. Articles may be in either English or French. Contributors need not be members of the CMS nor need they be Canadian; foreign contributions are welcomed. All contributions will be subject to a critical review before acceptance. Because of space limitations articles should not exceed 16 printed pages and preferably should be shorter.

**Manuscripts** should be submitted to: the Editor, *Atmosphere*, West Isle Office Tower, 5th Floor, 2121 Trans-Canada Highway, Dorval, Quebec H9P 1J3. Three copies should be submitted, typewritten with double spacing and wide margins. Heading and sub-headings should be clearly designated. A concise, relevant and substantial abstract is required.

**Tables** should be prepared on separate sheets, each with concise headings.

**Figures** should be provided in the form of three copies of an original which should be retained by the author for later revision if required. A list of legends should be typed separately. Labelling should be made in generous size so that characters after reduction are easy to read. Line drawings should be drafted with India ink at least twice the final size on white paper or tracing cloth. Photographs (halftones) should be glossy prints at least twice the final size.

**Units.** The International System (SI) of metric units is preferred. Units should be abbreviated only if accompanied by numerals, e.g., '10 m', but 'several metres.'

**Footnotes** to the text should be avoided.

**Literature citations** should be indicated in the text by author and date. The list of references should be arranged alphabetically by author, and chronologically for each author, if necessary.

## RENSEIGNEMENTS POUR LES AUTEURS

**Politique éditoriale.** *Atmosphère* est un organe de publication de résultats de recherche originale d'articles sommaires, d'essais et de critiques dans n'importe lequel domaine des sciences de l'atmosphere. Il est publié par la SMC à l'aide d'une subvention accordée par le gouvernement canadien. Les articles peuvent être en anglais ou en français. Il n'est pas nécessaire que les auteurs soient membre de la SMC; les contributions étrangères sont bien-venues. A cause des limitations d'espace les articles ne doivent pas dépasser 16 pages dans le format final. Tout article sera soumis à un critique indépendant avant d'être accepté.

**Les manuscrits** doivent être envoyés à : le Rédacteur, *Atmosphère*, West Isle Office Tower, 5e étage, 2121 route Trans-canadienne, Dorval, Québec H9P 1J3. Ils doivent être soumis en trois exemplaires dactylographiés à double interlignes avec de larges marges. Les titres et sous-titres doivent être clairement indiqués. Chaque article doit comporter un résumé qui soit concis, pertinent et substantiel.

**Les tableaux** doivent être préparés et présentés séparément accompagnés d'un titre et d'un numéro explicatifs concis.

**Les graphiques** doivent être présentés en trois copies dont les originaux devraient être conservés par l'auteur au cas où ils seraient nécessaire de les reviser. Une liste des légendes des graphiques doit être dactylographiée séparément. L'étiquettage doit être de grand format de façon à ce qu'il soit facilement lisible après réduction du format. Le traçage des lignes doit s'effectuer au moyen d'encre de chine en doublant, au moins, le format final, le tout sur papier blanc ou sur papier à calquer et identifié adéquatement. Les photographies (demi-teintes) devraient être présentées sur épreuves glacées au double du format final.

**Les unités.** Le Système International (SI) d'unités métriques est préférable. Les unités devraient être abrégées seulement lorsqu'elles sont accompagnées de nombres, ex: "10m", mais "plusieurs mètres".

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## The Canadian Meteorological Society / La Société Météorologique du Canada

The Canadian Meteorological Society came into being on January 1, 1967, replacing the Canadian Branch of the Royal Meteorological Society, which had been established in 1940. The Society exists for the advancement of Meteorology, and membership is open to persons and organizations having an interest in Meteorology. At nine local centres of the Society, meetings are held on subjects of meteorological interest. *Atmosphere* as the scientific journal of the CMS is distributed free to all members. Each spring an annual congress is convened to serve as the National Meteorological Congress.

Correspondence regarding Society affairs should be directed to the Corresponding Secretary, Canadian Meteorological Society, c/o Dept. of Meteorology, McGill University, P.O. Box 6070, Montreal, P.Q. H3C 3G1

There are three types of membership – Member, Student Member and Sustaining Member. For 1975 the dues are $20.00, $5.00 and $60.00 (min.), respectively. The annual Institutional subscription rate for *Atmosphere* is $15.00.

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Toute correspondance concernant les activités de la Société devrait être adressée au Secrétaire-correspondant, Société météorologique du Canada, Département de Météorologie, l'Université McGill, C.P. 6070, Montréal, P.Q. H3C 3G1

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