ZENITH MEASUREMENTS OF CLOUD EMISSIVITY IN THE 8 - 13 MICRON WAVEBAND

by


ABSTRACT

To determine the emissivities of clouds, measurements of the radiative intensity in zenith in the atmospheric window were made for nearly one year. The height of the cloud base was measured by stereophotography for high and medium cloud and by ceilometer for low cloud. The radiosonde sounding was used to find the cloud base temperature and, together with the known absorption coefficients of water vapour and carbon dioxide, to compute the effects of the intervening atmosphere.

The emissivity of low cloud was found to be $100 \pm 2.5\%$. The emissivity of alto-cumulus was found to range from $30\%$ to $100\%$, with a mean of $80 \pm 6\%$. The emissivity of cirrus was found to range from zero to $100\%$ with a mean of $35 \pm 5\%$. This latter value is less than the value of $50\%$ which has previously been used, and this affects calculations of the longwave radiation field in the atmosphere.

Department of Meteorology, McGill University. Ph.D. Thesis

March, 1969.
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A thesis submitted to the Faculty of Graduate Studies and Research in partial fulfilment of the requirements for the degree of Doctor of Philosophy.

Department of Meteorology,
McGill University,
Montreal.

March 1969

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ABSTRACT

To determine the emissivities of clouds, measurements of the radiative intensity in zenith in the atmospheric window were made for nearly one year. The height of the cloud base was measured by stereophotography for high and medium cloud and by ceilometer for low cloud. The radiosonde sounding was used to find the cloud base temperature and, together with the known absorption coefficients of water vapour and carbon dioxide, to compute the effects of the intervening atmosphere.

The emissivity of low cloud was found to be 100 ± 2.5%. The emissivity of alto-cumulus was found to range from 30% to 100%, with a mean of 80 ± 6%. The emissivity of cirrus was found to range from zero to 100% with a mean of 35 ± 5%. This latter value is less than the value of 50% which has previously been used, and this affects calculations of the longwave radiation field in the atmosphere.
INTRODUCTION

In recent years there has been considerable interest in calculations of radiative heating in the atmosphere. One motivation for such calculations has been the determination of the radiative components in heat budget studies such as that of Julius London (1957). Another has been the necessity of incorporating heating rates into dynamic models of the atmosphere. The advent of high-speed electronic computers has made it possible to perform such calculations, at least in an approximate way, in a reasonable time. The studies of Manabe and Strickler (1964), for example, are a first step in this direction. But perhaps the greatest stimulus to determine the radiative properties of the atmosphere is that such information is essential in order to interpret the copious radiometric data which is now available from artificial satellites.

For solar (shortwave) radiation, the radiative properties of both clear and cloudy atmospheres are reasonably well known. With terrestrial (longwave) radiation the experimental techniques are more difficult and the properties of the radiationally active gases (principally water vapour, carbon dioxide, and ozone) have been determined by a mixture of experiment in the laboratory, experiment in the atmosphere and quantum-mechanical calculation. As far as cloud is concerned, few systematic measurements have been reported and only a small number of approximate calculations have been attempted. Most workers have taken the emissivity of low and medium cloud as 100%, and the emissivity of cirrus as 50%, for want of any more precise values.

The simplest possible assumption for calculating the radiative properties of a cloud is to suppose that it will absorb as a slab of
liquid water of the same mass. This assumption was used by MacDonald (1960) to show, for example, that a cloud with a liquid water content of 1 gm m\(^{-3}\) would absorb 50% of the incident longwave radiation within a depth of only 5 metres. A more precise approach is to take into account the multiple scattering taking place within the cloud. The problem is complicated by the fact that the size of the cloud particles is of the same order of magnitude as the wavelength of the radiation. Computations have been made by Havard (1960) for clouds consisting of spherical water droplets of uniform size and for monochromatic radiation. He shows, for example, that, for a cloud of droplets of radius 6 microns and for radiation of wavelength 10 microns, the reflectivity may be as high as 10% and that significant amounts of radiation may penetrate through 250 metres of cloud.

A more recent study by Yamamoto, Tanaka and Kamitani (1966) considers a cloud drop-size distribution given by \(n(r) = r^6 e^{-1.5r}\), where \(n(r)\) is the volume concentration at radius \(r\), with monochromatic radiation of wavelength 10 microns. They also find significant transmission through 100 metres of cloud (though not through 1000 metres), but the maximum value of the reflectivity is 3%. This is discussed further in section 4(e).

Few results have so far been published of calculations for clouds consisting of ice crystals.

The general conclusion of these limited calculations is that the emissivity of thick clouds of water droplets lies somewhere between 85 and 100%, while that of thin clouds is smaller.

In the present work an attempt has been made to deduce the emissivities of clouds from measurements of the radiative intensity at the ground. A radiometer measures continuously the flux received in a narrow vertical
beam in the 8 - 13 micron waveband (the atmospheric window). Details concerning the construction and operation of the radiometer are given in chapter 2. The 8 - 13 micron waveband is chosen partly because it contains the peak and approximately 30% of the total energy of the terrestrial spectrum, and partly because the atmosphere is particularly transparent in this region and the effects of absorption and emission in the atmosphere between the radiometer and the cloud are therefore minimised. Theoretical work is not yet sufficiently advanced to indicate whether results in the 8 - 13 micron waveband are representative of the whole terrestrial spectrum. However, it will be in this waveband that the effect of clouds will be most important. In the carbon dioxide and water vapour bands, the atmosphere is fairly opaque anyway.

Although the effects of the intervening atmosphere are small in this waveband, they are not negligible. In an earlier work (Allen, 1965) a computer program was used to compute these effects. Since then, a number of important refinements have been made in this program, the details of which are discussed in chapter 4.

In order to calculate the emissivity of the cloud, it is necessary to know not only the intensity of the emitted radiation but also the true temperature. This is found by measuring the height of the cloud and by using a radiosonde sounding to find the temperature at this height. A discussion of the difficulties involved in using a radiosonde sounding in this way, and of the methods used to measure the height of the cloud base are discussed in chapter 3.

The results obtained are given in chapter 5.
2. THE RADIOMETERS

At the start of the project, a Huggins model 5038 radiometer was used. This instrument was designed specifically for observing clouds and has a number of desirable features which are discussed in section 2(a). In the course of two year's operation numerous difficulties arose with this instrument and a second radiometer, a Barnes model I.T.3, was obtained. The latter is not specifically designed for cloud observation and it was necessary to modify it for this purpose. The sensitivity, the field of view and the filters are less suitable, but it was found to have excellent stability as far as calibration was concerned. The calibration of the Huggins instrument on the other hand was found to drift quite rapidly. For this reason, most of the results were obtained with the Barnes radiometer.

Some time after the Huggins radiometer was acquired, the Huggins business was taken over by the Barnes Manufacturing Company. It is to be hoped that this may result in the production of a radiometer which combines the convenience and sensitivity of the old Huggins instrument with the stability of the Barnes I.T.3. If, furthermore, a filter could be found with a sharp cut-off at 12.5μ, many of the difficulties and uncertainties of this work would be overcome.
(a) The Huggins model 5038

The Huggins model 5038 Radiometer measures infra-red radiation in the 8 - 13 micron waveband. It receives radiation in a narrow beam (nominal 2.16° x 2.16°) which is normally pointed vertically upwards. The focusing is achieved by a coated Irtran II lens which also serves as a filter to cut out radiation of wavelength greater than 14 microns. The lens is kept free from dust and moisture by means of a warm air purge which passes over the lens and out through the aperture above the lens. Radiation of wavelength less than 8 microns is cut out by a germanium substrate filter. The combined filter transmission is shown in figure 2.5. The temperature sensitive element is a thermistor flake 1mm x 1mm. Between the lens and the filter is an aluminum chopper which interrupts the incident radiation at a frequency of 30 cps. The surface of the chopper blade is reflecting so that the thermistor flake senses alternately the sky and the detector block, which is maintained at a constant temperature (50 C). A block diagram of the circuit is shown in figure 2.1.

A 30-cycle signal is thus generated which is a function of the temperature difference between the sky and the detector block. Since the reference temperature is 50 C, the signal increases as the temperature decreases and is non-linear with temperature. An electronic network converts this signal into one which is linear with temperature over the range -80 C to +20 C. The output is shown on a meter with a linear temperature scale and is connected to a chart recorder for a permanent record. The recorder has a chart speed of 1 in/hr for continuous operation and the 100 divisions of the chart correspond conveniently to the 100 centigrade degree range from +20 C to -80 C.
Figure 2.1 Block diagram of the Huggins radiometer.
The calibration of the radiometer is achieved by placing a black-body source at known temperature over the aperture. With the source at or near -80 C, the zero control is adjusted until the indicated temperature is the same as the source temperature. Then with the source at or near +20 C, the calibration control is adjusted until the indicated temperature is the same as the source temperature. The adjustments are not entirely independent; it is necessary to repeat the procedure a number of times in order to obtain the optimum setting. The source is then maintained at a series of temperatures between -80 C and +20 C, and the indicated temperature noted in each case so that a complete calibration curve can be plotted. The low temperature source consists of a brass cone which is blackened on the inside and cooled with a mixture of ethanol and dry ice, the temperature being measured with a toluene-in-glass thermometer.

A number of difficulties arose in connection with this calibration procedure. The manufacturer suggested that the warm air blower be switched off, the cover removed, and the black-body source placed close to the lens. It was found however that switching off the warm air blower could change the indicated temperature by as much as 15 C, (presumably because of the change of temperature of the lens) so that it was necessary to perform the calibration with the radiometer in its normal operating state. This means that a jet of warm air is issuing from the aperture and the cold source has to be placed well above this. A second difficulty concerned the size of the black-body; with a 2° beam and a 1 inch diameter lens, the radiometer should sense an area of diameter 1.4 inches at the height of 12 inches above the lens. It was found however that this
Figure 2.2: Field of view of the Huggins 5038 Radiometer
(a series of circular aperture stops with radii from 0 to 2 inches were placed 12 inches above the radiometer which was pointed at a constant cold source (the sky); knowing the flux when it was completely uncovered and the flux for this series of intermediate cases, the percentage of the total flux contained within any given radius was found.)
area contained only 70% of the beam. By using a series of aperture stops, the percentage of the beam within any given radius was found. The results are plotted in figure 2.2, and it will be seen that, in order to include at least 99% of the beam, a source at least 4 inches in diameter is needed. In practice, the source was made 5 inches in diameter to allow for small errors in alignment and edge effects.

The final and most serious difficulty was the formation of frost on the black-body. Initially the indicated temperature falls slightly, presumably because the 'blackness' of the surface is reduced. As the layer of ice thickens it acts as an insulator and the surface temperature rises, giving a higher reading. Various methods were attempted to prevent the formation of this frost, the most successful of which was to coat the black surface with a thin film of ethanol which dissolves the frost as it forms. The ethanol makes the surface somewhat reflecting as far as visible light is concerned, but for the infra-red region it should still radiate as a black body, especially as the source is in the shape of a cone. For a 60° cone, incident light parallel to the axis must suffer 3 reflections before escaping, so that even if the reflectivity of a single surface is as high as 10%, the reflectivity of the cone will be only 0.1%.

The formation of frost was further reduced by inverting the apparatus so that the radiometer was looking down into the cone, and the cold air tended to remain in the cone (see figure 2.3). With this arrangement, calibration could be taken down to -40 C without any formation of frost, and down to -75 C if the frost was removed periodically with ethanol.
Figure 2.3  Calibration apparatus.
(b) The Barnes Model I.T.3. Radiometer

The Barnes Model I.T.3. radiometer is similar to the Huggins in its general mode of operation; the essential components being a thermistor detector and a chopper which allows the detector to sense alternately the sky and a fixed reference temperature. The standard reference temperature of the I.T.3. is 50 C but this was reduced to approximately 40 C in order to extend the range of the instrument to lower temperatures.

The output of the radiometer appears on a meter with a non-linear scale of temperature ranging from +30 C to -80 C and is also fed through an amplifier to a Rustrak recorder. Since the full range of the instrument was not needed, the amplifier was adjusted so that the recorder operated only in the range of +20 C to -80 C. In this way the maximum use was made of the recorder chart. The chart speed was one inch per hour, and the calibration technique was the same as that used with the Huggins radiometer.

The main difference between the two radiometers lies in the sequence of optical components. In the Huggins, the first component is the Irtran II lens (which acts both as a lens and as a filter to cut off radiation beyond 14 microns) and this is followed by the chopper blade. In the Barnes radiometer the chopper is the first component and this is followed by a Germanium lens and filters of Irtran II and indium antimonide. The advantage of putting the lens first is that it can be sealed into its mounting and the system is then reasonably weatherproof. The disadvantage is that the readings are quite sensitive to changes in the temperature of the lens. In the Huggins instrument, such changes are kept small by the thermostatically controlled warm air purge.
The advantage of putting the chopper first, as in the Barnes radiometer, is that the changes in the temperature of the lens or filters do not affect the reading. On the other hand, the instrument cannot be pointed upwards at the sky in all weathers. Also, it is difficult to find any material to cover the instrument which will not absorb and emit to an appreciable extent in this waveband. In order to overcome this difficulty, the radiometer was placed in a weatherproof box and viewed the sky in a mirror (see figure 2.4). Since glass is strongly absorbing in this waveband, a front-silvered mirror had to be used. Ordinary front-silvered mirrors were found to be too fragile for outdoor use so a mirror was made by plating a brass plate with rhodium. The resulting mirror was not optically perfect, but this is unimportant, and the mirror has withstood frequent cleaning without any sign of deterioration.

Figure 2.4 - Arrangement of the Barnes I.T.3 Radiometer.
In order to reduce the amount of dust collecting on the mirror, air was drawn in through a filter and blown out through the hole in the top of the box. An ordinary hair-drier was used for this purpose so that in winter the warm setting could be used and the radiometer mirror kept at a reasonable temperature.

It was found that the presence of the mirror in the optical train caused a small increase in the temperature recorded by the radiometer. This occurs because the mirror is not a perfect reflector in the 8 - 13 micron waveband. The calibration of the instrument was therefore performed with the mirror in place. Of course, there will still be an error if the temperature of the mirror changes from its value at the time of calibration. However, it was thought that the error would be quite small provided the temperature of the mirror was kept within reasonable limits by the use of the air purge in winter. Ideally the temperature of the mirror should be thermostatically controlled. The effects of changes in the mirror temperature are discussed further in chapter 5.

The other main differences between the two radiometers are that the field of view of the Barnes is approximately 3° as opposed to 2° for the Huggins; that the Barnes does not have any shaping circuitry so that the signal is approximately linear in intensity rather than in temperature; and that the filter cut-off is less sharp in the Barnes instrument. The absence of the shaping network is in some respects an advantage; when the recorder reading is plotted against the intensity in calibration, a straight line graph is obtained. This makes it possible to eliminate small errors in calibration and to extrapolate the calibration curve down to lower temperatures where the calibration technique becomes particularly difficult.
The transmission functions of the two radiometers are shown in fig 2.5.
It will be seen that the transmission of the Huggins system is better in the centre of the band; this is because the components are coated to give maximum transmission at 10 microns. At the shortwave end, the germanium substrate filter of the Huggins has a sharper cut-off than the indium antimonide of the Barnes. At the longwave end, both instruments use Irtran II, but the Huggins uses a thicker piece and hence has a better cut-off. The net result is that the Barnes instrument sees more of the water vapour and carbon dioxide bands and hence is affected more by the temperature and humidity of the lower layers of the atmosphere. Some typical results are shown in table 2.1.

The net result of these differences is that the Huggins instrument is more sensitive and should give more accurate results for the emissivity of the cloud. However, the calibration of this instrument was found to be rather unstable. It was necessary to recalibrate it at least once a week. The cause of this instability is not known. The Barnes instrument, on the other hand, gave consistent results over a period of one year. For this reason, the majority of the readings were obtained with the Barnes radiometer.
Figure 2.5  Transmission functions of the Huggins and Barnes radiometers.
Table 2.1 - Typical computed intensities for different filter functions. The soundings are mean values for each season taken from London (1957). Note that the Barnes instrument always records the higher temperature as it sees more of the carbon dioxide and water vapour bands. The difference increases as the cloud height increases.

<table>
<thead>
<tr>
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<th>Huggins filter function</th>
<th>Barnes filter function</th>
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<td></td>
<td>Base km</td>
<td>Temp. C</td>
<td>Intensity Watt m^-2sr^-1</td>
</tr>
<tr>
<td>Summer</td>
<td>none</td>
<td>7.02</td>
<td>-42.8</td>
</tr>
<tr>
<td>Fall</td>
<td>none</td>
<td>5.10</td>
<td>-55.0</td>
</tr>
<tr>
<td>Winter</td>
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<td>3.53</td>
<td>-67.7</td>
</tr>
<tr>
<td>Fall</td>
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<td>20.69</td>
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<td>-</td>
<td>1.13</td>
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<tr>
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<td>3.3</td>
<td>-6.4</td>
<td>16.31</td>
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<tr>
<td>-</td>
<td>5.5</td>
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<td>13.32</td>
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<tr>
<td>-</td>
<td>8.3</td>
<td>-42.2</td>
<td>10.00</td>
</tr>
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3. MEASUREMENT OF CLOUD BASE TEMPERATURE

(a) Height measurement

In order to determine the emissivity of a cloud it is necessary to know its temperature to a reasonable degree of accuracy. Since the radiometer can be read to the nearest degree (though the absolute accuracy is less than this), it is reasonable to attempt to measure the cloud base temperature also to the nearest degree. In the absence of any direct method, it has been necessary to measure the height of the cloud base and to rely on the radiosonde soundings for finding the temperature at the measured height. Some of the difficulties involved in the use of radiosonde soundings are discussed in section 3(b). A further difficulty in this case is that the cloud may itself modify the sounding, but we have no means of knowing how great this effect may be. It is also pertinent to ask what exactly is meant by the term 'cloud base', where radiation in the 8 - 13 micron waveband is concerned. It is certain that such radiation will penetrate further into the cloud than will visible radiation and that the radiometer therefore 'sees' further into the cloud than does the eye (or camera). However, rough calculations suggest that most of the radiation is absorbed (or emitted) in the first 300 feet, and as the temperature will not drop by more than $1^\circ$C in this distance, it should be sufficient to use the visible cloud base.

The problem therefore is to determine the height of the cloud base to the nearest 300 feet. Except in the case of severe inversion, this will give the temperature to the nearest degree centigrade, provided the sounding is correct. Two methods have been used: the ceilometer, and stereographic photography.
(a) (i) The ceilometer.

The ceilometer is a standard instrument used at airfields to determine the height of the cloud base. It employs a pulsed light beam which points vertically upwards and a slowly rotating receiver tuned to the same frequency (figure 3.1). A stylus moves over heat-sensitive paper in phase with the rotation of the receiver and marks the paper whenever a signal is received. The position of the mark thus gives the zenith angle of the receiver when the signal is returned. Knowing the separation of the source and the receiver, we can convert this scale of angles into a scale of height. The scale of height will, of course, be non-linear and will give the greatest accuracy for heights which are approximately equal to the separation of source and receiver (see figure 3.2). The nearest available ceilometer is at Montreal International Airport, Dorval, and the separation in this case is 600 feet. This gives maximum accuracy for cloud bases of the order of 600 feet, which is the region of greatest interest to the commercial airlines. The accuracy decreases with increasing height and above 5000 feet the results are virtually useless. Dorval Airport is approximately 9 miles west of the McGill campus where the radiometer is situated, so that it is not possible to match the ceilometer record exactly with the radiometer record. The ceilometer results are used whenever there is a uniform deck of low cloud persisting for more than one hour.
Figure 3.1 - The ceilometer.

Figure 3.2 - Sample of ceilometer record.
(a) (ii) Stereographic Photography.

Two 16-mm movie cameras have been set up approximately 3000 ft apart and pointing vertically upward. Each is triggered by a time-lapse device to take one frame every minute; this device consists of a synchronous motor geared to drive a wheel at $1/6$ rpm. The wheel contains a number of set-screws (in this case, six) which operate a micro-switch when they pass a certain point. By varying the number of screws, the frequency of the exposures can be altered. The micro-switch does two things: it energises the electric motor which replaces the usual clockwork motor for driving the camera; and it starts another electric motor which drives a cam to trigger the camera (see figure 3.3). Since both time-lapse devices are driven by synchronous motors using the 60 cycle mains supply, once the two cameras have been synchronised, they will continue to be triggered simultaneously until there is a break in one power supply.

There is a time switch with each camera which interrupts the power from the micro-switch during the hours of darkness. The synchronous motors continue to run while the cameras are switched off, so that the two systems are still in phase when they are switched on again in the morning. A small mirror is placed at the edge of the field of view of the camera to reflect the face of a digital clock. In this way, the time of each frame is recorded. A lamp connected to the triggering circuit lights up just before each exposure to illuminate the clock face.

The whole apparatus (with the exception of the lamp and the mirror) is contained in a weather-proof box with windows for the camera and clock face. In order to prevent condensation on the inside of these windows, a blower provides a constant stream of air over them. In summer the blower
Figure 3.3 - Schematic diagram of camera and time-lapse

- **M₁** Synchronous motor
- **M₂** Triggering motor
- **M₃** Camera motor
- **S₁** Micro-switch
- **S₂** Time-switch
- **S₃** Switch to stop camera without destroying synchronisation
- **S₄** Switch to adjust synchronisation
draws fresh air from outside the box, but in winter the air is re-
circulated inside the box and is heated.

The cameras are synchronised to within 1/5 second by means of
a stop-watch. Small variations in the time of triggering occur, depending
on which of the six screws is operating the micro-switch, but these do
not exceed ± 0.2 sec. The maximum error in timing is therefore ± 0.4 sec.
A high cloud moving at 50 mph will travel a distance of 30 ft in 0.4
sec and this will cause an error of approximately 300 ft in the measure-
ment of height. However, only the component of the wind velocity which
is parallel to the line through the two cameras is important and, by
making this line run north and south (i.e. perpendicular to the direction
of the prevailing wind), the chances that this component would exceed
50 mph are fairly small.

Approximately 4½ ft above each camera are cross wires with one
wire parallel and one wire perpendicular to the line joining the cameras.
The centre of the cross-wires is adjusted to be vertically above the
centre of the lens by means of a plumb-line. This adjustment is fairly
critical; if the error in height measurement is to be less than 300 ft,
the centre of the cross-wires must be within 0.05 inch of the vertical
line through the centre of the lens. With sufficient care, this degree
of accuracy can be achieved. It is necessary to protect the plumb line
from the wind and this can conveniently be done by enclosing it in a
cardboard tube with a plastic window at the bottom through which to
view the plumb-bob. It is particularly important that this adjustment
be correct as any error here will produce a consistent error in all
subsequent height measurements. When the film is projected, all
measurements are made from the image of the cross-wires, so that the exact angle of the cameras and the projectors and the exact position of the film in the camera and projectors is not critical.

It was not possible to obtain an automatic aperture control for a 16-mm camera lens so a fixed aperture had to be used. For this reason, black and white film was used in preference to colour film as the latter is rather more sensitive to changes in light conditions. A red filter was used to increase the contrast between white clouds and blue sky. It was found to be quite satisfactory to use the negative rather than a positive print, as the viewer quickly grows accustomed to seeing black clouds on a white background instead of white clouds on a black background. The single aperture setting (f/16), with a 2x red filter and black and white negative film (Kodak Plus-X, ASA 80), was found to give good results from about an hour after sunrise to an hour before sunset for all types of cloud and all seasons.

The camera lenses had a focal length of 15 mm, giving a field of view of approximately 36° x 30°. The longer axis was aligned parallel to the line through the two cameras. The distance between the cameras was 3070 ft (this distance being necessary to give the required accuracy at a height of 30,000 ft), so that the two fields of view intersected at a height of 4600 ft (see figure 3.4). Clouds below this level could not be measured. This minimum level could have been reduced by tilting the cameras towards each other but it was felt that this gain in range would have been more than offset by the greater errors incurred by the difficulty of precise alignment. Also, high clouds, which require the greatest accuracy in measurement, would have appeared at the edge of the
Fig 3.4. Arrangement of cameras \((C_1, C_2)\) and projectors \((P_1, P_2)\) (assuming equal focal lengths).

Fig B. Effect of different focal lengths in camera and projector lenses.
field of view where the lens aberrations are greatest.

At a height of 30,000 ft, the field of view of each camera covers approximately 20,000 ft x 16,000 ft. A cloud moving at 200 ft/sec would therefore take a minimum of 80 seconds to cross the field of view. The time interval of 60 seconds between frames was chosen to ensure that each cloud would appear on at least one frame. The beam widths of the radiometers (approximately $2^\circ$ for the Huggins and $3^\circ$ for the Barnes) are only a small fraction of the field of view of the cameras.

The developed films are projected onto a screen by two Dunning Animatic projectors. These projectors are designed for showing single frames of 16-mm film and the film is advanced one frame at a time by a solenoid. Both projectors were connected to a single switch and, every time this switch is depressed, the two films advance by one frame. The original concept of this projection system was derived from a similar system designed by Reznick (1966). Assuming that the focal lengths of the camera and projector lenses are the same, the positions of the cloud image on the screen and the two projectors will be an exact scale model of the positions of the cloud and the two cameras. Referring to figure 3.4 again we have

\[
\frac{h}{3070} = \frac{d}{x}
\]

where $h$ is the height of the cloud in feet, $d$ is the distance of the screen from the first principal plane of the projector lens, and $x$ is the separation of the two projectors.

In practice, projector lenses of focal length 15 mm were not available and the nearest approximations were of 5/8 inch (15.88 mm) focal length. It is necessary therefore to make a small correction
for this change in focal length. In figure 3.5 we have imagined the two cameras and the two projectors moved together. The cloud positions relative to the two cameras are now 3070 ft apart and the separation of the two images on the film (s) is given by

\[ \frac{h}{3070} = \frac{f_c}{s} \]

where \( f_c \) is the focal length of the camera lens.

Similarly for the projectors we have

\[ \frac{d}{x} = \frac{f_p}{s} \]

where \( f_p \) is the focal length of the projector lens. Eliminating \( s \) between these two equations gives

\[ h = 3070 \frac{d f_c}{x f_p} \]

Substituting \( f_c = 15 \) mm, \( f_p = 15.88 \) mm and \( d = 40 \) in, gives

\[ h = 116,520 / x \]

where \( x \) is the distance between the two images in inches.

In practice, it is much easier to adjust the projectors so that the two cloud images coincide and to measure the distance between the two cross-wires than it is to adjust the projectors so that the cross-wires coincide and measure the distance between the two cloud images. This is because there is seldom any clearly defined point of the cloud to which one can measure. In adjusting the cloud images to coincide, the whole field of view can be used rather than just a single point and the cross-wires have sharp images whose separation can be easily measured. In practice, one projector is kept stationary and the other is moved along a track to make the adjustment. On the screen there is a scale
which gives the height of the cloud directly in kilofeet. By making all measurements on the screen, the exact alignment of the projector and the exact position of the film in the projector are not critical. The projectors are on leveling screws to adjust the height of the image and, if necessary, to rotate the image. This latter adjustment is required if the axis of the camera is not exactly parallel to the line through the two cameras.

So that the two projectors do not get in each other's way, one of the projectors is placed behind the screen (which is translucent) and the film is loaded back to front. It is necessary that the screen should reflect as much light as it transmits if the two images are to be equally bright. Ground glass was found to transmit far too much, but ordinary tracing paper, fixed to a sheet of glass to keep it flat, was found to be quite satisfactory. A rheostat was placed in series with one of the projector lamps to make small adjustments in brightness. The most critical part of the measurement is the part that involves judging when the two cloud images exactly coincide. It was found that when both projectors were shone on the screen simultaneously, the appearance was that of a single large cloud and the two separate images could not be distinguished. To overcome this difficulty, rotating chopper blades were placed in front of each projector so that the images are flashed on the screen alternately with a frequency of about four flashes per second. Then, when the alignment is not quite correct, the image appears to oscillate to and fro, and when the images coincide exactly, this oscillation disappears. With this technique, the adjustment becomes quite sensitive. Provided there
is some recognisable feature or variation in the cloud base, the two images can consistently be aligned to within 0.05 inch. This corresponds to an error of less than 300 ft at a height of 30,000 ft. In order to establish the height of a cloud deck in the first place, some definitely recognisable feature is necessary but, once the approximate height is known, only a very small variation in appearance anywhere in the field of view is sufficient to make the final adjustment.

In many cases, the height of the cloud base will be seen to vary throughout the field of view. Now the film in the fixed projector is the one which comes from the camera which is immediately adjacent to the radiometer. The field of view of the radiometer therefore can be, and is, marked on the screen. Where different heights are observed, a value can be taken for a feature which is within, or as near as possible to, the field of view of the radiometer. It had been hoped originally that the cloud decks would have a fairly uniform base so that many stereoscopic pairs could be viewed without any adjustment being necessary. In practice, this was found not to be the case. There is usually a variation in height of several hundred feet between one frame and the next. Of course, it must be borne in mind that only those clouds with some visible features can be measured. Uniform decks of altostratus or cirrostratus can not be measured with this apparatus and these may well be more uniform in height.

We can now summarize the errors inherent in this method of cloud height measurement:
A. Consistent errors -

Alignment of cross-wires; for each camera ±300 ft

total ±600 ft

B. Random errors -

Errors in synchronisation of cameras ±300 ft
Errors in superposition of cloud images ±300 ft

total ±600 ft

We see that the maximum consistent error is ±600 ft and the maximum random error is ±600 ft, making a total possible error of ±1200 ft in any one reading, or of ±600 ft in the average of many readings. These values correspond to errors in temperature of ±3°C and ±1.5°C respectively. These errors are calculated for high cloud (30,000 ft). For medium cloud the angle subtended by the cameras is greater and the error is less although, due to the fact that the cloud is viewed from two different angles, it is not always possible to superimpose the two cloud images exactly.

There is also the effect of parallax in these measurements. This is illustrated in figure 3.6 where the edge of the cloud is the feature whose height we are attempting to measure. Of course, the error has been much exaggerated in this diagram. If it were as great as this it would be quite apparent when the films are projected. The fact that consistent results are obtained over periods of an hour or more suggests that the error can not be too great. This type of error will be most significant with cumulus and alto-cumulus clouds. High clouds are usually comparatively thin and the error should be small in this case.
Finally, in table 3.1, the range, limitations and accuracy of the two methods of cloud height measurement are compared.

Table 3.1 Summary of cloud height measuring methods

<table>
<thead>
<tr>
<th></th>
<th>i) Ceilometer</th>
<th>ii) Stereophotography</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Range</strong></td>
<td>0 - 5000 ft</td>
<td>5000 - 40,000 ft</td>
</tr>
<tr>
<td><strong>Limitations</strong></td>
<td>Uniform cloud deck (St, Ns, Sc, etc.)</td>
<td>Daylight</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Cloud with some structure</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(i.e. not As or Os)</td>
</tr>
<tr>
<td><strong>Accuracy</strong></td>
<td>±300 ft at 1000 ft</td>
<td>±1200 ft at 30,000 ft</td>
</tr>
<tr>
<td></td>
<td>±2000 ft at 5000 ft</td>
<td>±800 ft at 15,000 ft</td>
</tr>
</tbody>
</table>

Figure 3.6 - Effect of parallax on measurement of cloud base.
It will be seen that the only method that is applicable to high clouds is stereophotography and that this is limited to hours of daylight and to clouds with some structure. In order to overcome these limitations it would be necessary to use either some form of ceilometer with a longer base-line or a vertically pointing 'lidar' unit (i.e. a unit employing the principles of radar with a visible light beam).

(b) Radiosonde soundings

In order to compute the intensity at the surface, it is necessary to know the temperature and water vapour content of each layer of the atmosphere. The temperature and dewpoint are measured as functions of pressure by radiosondes which are sent up at 00 Z and 12 Z each day. The nearest radiosonde station to Montreal is at Maniwaki, approximately 130 miles to the north-west. In order to be able to make computations on a routine basis, a sub-program was written which accepts the sounding as nearly as possible in the form in which it is transmitted,¹ and carries out the following procedures:

i) Decodes the first decimal place of the temperature and dew-point,
ii) Decodes the negative values,
iii) Locates the highest pressure and takes that to be the surface pressure,

¹ The only changes to be made when punching the transmitted data are the omission of the height and wind data, the insertion of 99 when the dew-point is missing or less than -50 C, and the insertion of a 1 before the temperature when it is less than -50 C (the figures 04, for example, may mean +4 C or -54 C; the decision as to which is meant involves a degree of judgement which is difficult to incorporate in a computer program. On Jan 1st 1968, the code for transmitting radiosonde data was changed. With the new code, this ambiguity is removed and no changes are needed in punching the data.
iv) Calculates the values of the 18 pressure levels,
v) Sorts the values in order of pressure,
vi) Inserts the values of temperature and dew-point for the 18 computing levels and for the cloud levels by linear interpolation,
vii) Calculates the mass of water vapour and the mean temperature of each layer.

One of the main problems in computing intensities is the fact that the sounding is at Maniwaki, while the radiometer is at Montreal, and that the soundings are only taken every twelve hours. In order to obtain an estimate of the sounding over Montreal, a rough calculation was first made of the time taken for the air mass over Maniwaki to reach Montreal. This was done using that component of the wind velocity which was in the direction of Montreal. The velocity component perpendicular to this direction was ignored, but in practice the upper winds were found to be nearly always westerly so that this latter component was small. Of course, the wind velocity varies with height; the value chosen was that for the cloud height since it is at this height that it is most important to have the correct temperature. To estimate the sounding at intermediate times (between the twelve-hourly values), a linear interpolation was made.

The temperature and humidity in the lower layers varies considerably with local conditions, so the surface data for Maniwaki were replaced by the surface data from the McGill Observatory. These are measured continuously so that actual values could be used without any need for interpolation. A constant lapse rate was assumed between the surface and the first upper air values, unless this lapse rate was super-adiabatic, in which case an adiabatic lapse rate was assumed until the upper air sounding was reached.
Temperatures (but not humidities) are also measured at the top and bottom of the television transmitting tower on Mount Royal (heights about 900 ft and 700 ft respectively above the site of the radiometers). Computations made for a period of two weeks, both with and without the tower temperatures, showed that in most cases the addition of the tower data did not alter the computed temperature by more than 0.2 C. The only cases for which the tower data made any significant difference were those for which the cloud base was near the level of the tower, but even in these cases, the difference did not amount to more than 0.5 C. It was therefore decided to discontinue the use of the tower data on a routine basis and only use it when special circumstances made this desirable.

On some occasions, soundings taken at other stations - namely Sault Sainte Marie, Buffalo and Caribou - were taken into account but, since the nearest of these is about 300 miles away, it was felt that the gain in accuracy was very slight. The most critical part of the sounding is the temperature at the height of the cloud. It was found that the temperature aloft seldom changed by more than 3 C over a twelve hour period, so that our estimated values should certainly be accurate to within ±1 C. This means that the error in the sounding is no greater than the possible errors in the measurement of cloud height and in the radiometer itself.
4. THEORY AND CALCULATIONS

(a) The computer program.

The experimental program used to explore the feasibility of emissivity measurements at the surface was described in an earlier work (Allen, 1965). Since then a number of improvements have been made.

Since ozone variations turned out to be unimportant, the transmissions due to ozone have been computed for each season and are now fed in as data, instead of being computed afresh each time.

A sub-routine has been written to enable the radiosonde soundings to be fed in very nearly in the form in which they are transmitted (see section 3(b)). In the original program, the atmosphere was divided into 18 layers for the purpose of numerical computations, and cloud was inserted effectively in between the layers. This meant that there were only 17 possible levels at which cloud could be placed. This did not matter so long as purely hypothetical situations were being studied, but as soon as accurately measured cloud heights became available, it was necessary to modify the program by the addition of reference levels at the level of the cloud base. Since a varying number of additional levels were being added, it was decided to include also all the levels transmitted in the radiosonde report, rather than reduce the sounding to 18 fixed levels. The levels now used are therefore

i) 18 fixed levels - the inclusion of these ensures that no layer will be too thick,

ii) 10 to 24 reported levels, and
iii) up to 15 cloud levels, which can be placed at any height.

(b) Filter transmissions

In order to test the program, it was applied to a series of measurements made by Bolle (1960). These measurements were made with a spectrometer so that each part of the atmospheric window could be evaluated individually. It was found that, although there was broad agreement in the centre of the window, there were considerable differences at the edges. The effect of these differences was reduced by the fact that the transmission of the filters is small in these regions, but there was still a significant error. In calculating the effects at the edges of the waveband, the precise transmission characteristics of the filters becomes rather important. This is because the spectrum of the clear sky emission is quite different from the spectrum of the calibrating black body. The black body radiation has a maximum somewhere in the 8 - 13 micron range (the exact position depending on its temperature) and falls off slowly on either side (curve 1, figure 4.1). The clear sky emission, on the other hand, is quite small in the centre of the band and rises rapidly at the edges (curve 2, figure 4.1). It becomes most important therefore to know the precise transmission of the filters at the edges. In the case of the Huggins radiometer, the filters were tested in an infra-red spectrometer and the transmission functions were multiplied together (graphically, on semi-log-coordinate paper) to obtain the transmission function of the entire optical system. It was thought that the fact that one of the filters is a lens might affect the performance of the spectrometer. A plane specimen of the same material (Irtran II),
Figure 4.1 - Relevant spectra:

Curve 1: Intensity spectrum of a black body at 270K.

Curve 2: Intensity spectrum of the clear sky near zenith, as measured by Bolle (1960) at 2020 Z on 10 December 1957 at Hamburg.

Curve 3: Huggins radiometer filter function in arbitrary units.

Curve 4: Intensity spectrum computed for the conditions of curve 2 without band corrections.

Curve 5: Intensity spectrum computed for the conditions of curve 2 with band corrections.
coated in the same way, was therefore obtained and tested in the spectrometer, but its transmission function in fact did not differ significantly from that of the lens. In the case of the Barnes radiometer the transmission function was supplied by the manufacturer. The two curves were shown in figure 2.5.

(c) Absorption coefficients

In view of the differences between the measured and the computed clear sky spectra (curves 2 and 4, figure 4.1), an analysis was made of the 6.3 micron and the rotation bands of water vapour and of the 15. micron, 10.4 micron and 9.4 micron bands of carbon dioxide. An intensive study was also made of all available data on the absorption coefficients of water vapour in the 'atmospheric window'. Details of this work are given in appendix 1. The results of these studies were that the absorption coefficients were left unchanged but that corrections were made for the effects of the water vapour and carbon dioxide bands, based mainly on empirical formulas derived from the measurements of Howard, Burch and Williams (1955). The modified program was applied to the ideal soundings used for the previous study, (Allen, 1965) and the results are shown in table 4.1. The changes are significant, particularly in the case of a very dry atmosphere (winter sounding) where the effect of carbon dioxide becomes dominant. The conclusions of the thesis, that the intensity at the surface is strongly dependent on the transmissivity of the cloud, are not altered, though the sensitivity of the surface intensity to this parameter is somewhat reduced. This is to be expected, as we have now faced the fact that the cut-off of the filters is not as sharp as we would like, and that some radiation is transmitted from
outside the 'window'.

The modified program was applied to the results obtained by Bolle (1960). It was found that the corrections made for the bands were in all cases too large (see curve 5, figure 4.1). The error in the filtered intensity is however, quite small. For the continuum, the computed intensities were found to agree well with the measured intensities for low humidities, but tended to be too small when the humidity is high. This suggested that a self-broadening coefficient should be included.

<table>
<thead>
<tr>
<th>Sounding</th>
<th>Cloud</th>
<th>Without band corrections</th>
<th>With band corrections</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Base km</td>
<td>Emissivity</td>
<td>Temp. °C</td>
</tr>
<tr>
<td>Summer</td>
<td>none</td>
<td>2.61</td>
<td>-57.8</td>
</tr>
<tr>
<td>Fall</td>
<td>none</td>
<td>1.63</td>
<td>-72.5</td>
</tr>
<tr>
<td>Winter</td>
<td>none</td>
<td>0.93</td>
<td>-87.6</td>
</tr>
<tr>
<td>Fall</td>
<td>0.07</td>
<td>1.0</td>
<td>+10.2</td>
</tr>
<tr>
<td>-</td>
<td>4.13</td>
<td>+6.0</td>
<td>+10.2</td>
</tr>
<tr>
<td>-</td>
<td>3.3</td>
<td>-6.4</td>
<td>9.44</td>
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<td>7.29</td>
</tr>
<tr>
<td>-</td>
<td>8.3</td>
<td>-42.2</td>
<td>4.92</td>
</tr>
<tr>
<td>Fall</td>
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<td>-42.2</td>
</tr>
<tr>
<td>-</td>
<td>0.75</td>
<td>4.11</td>
<td>-41.7</td>
</tr>
<tr>
<td>-</td>
<td>0.50</td>
<td>3.27</td>
<td>-50.0</td>
</tr>
<tr>
<td>-</td>
<td>0.25</td>
<td>2.45</td>
<td>-59.9</td>
</tr>
<tr>
<td>-</td>
<td>0.0</td>
<td>1.63</td>
<td>-72.5</td>
</tr>
</tbody>
</table>

Table 4.1 - Computed intensities with and without band corrections

(N.B. These results can not be compared with table 2.1, as a different filter function was used for the above computations.)
(d) Self-broadening effect.

As a further test of the program, the computed intensity was compared with the measured intensity for a number of cases when no cloud was present in July 1966 and July 1967. It was again found that the agreement was reasonable in the case of moderate temperatures and humidities but that when the temperature and humidity (and hence the intensity) was high, the computed values were too small. The laboratory experiments of Bignell (1965) to measure the absorption coefficients of water vapour have also produced very large results at high water vapour pressures, and he has suggested that this may be due to self-broadening and that the absorption coefficients should be given by

\[ k_v = k_1 p + k_2 e \]

where \( k_1 \) and \( k_2 \) are constants, \( p \) is the total pressure and \( e \) is the water vapour pressure. Taking the wavenumber of 901 cm\(^{-1}\) as an example, measurements in the atmosphere at water vapour pressures in the 0 - 10 mb range give \( k_{901} = 0.09\) gm\(^{-1}\) cm\(^2\), whereas Bignell's measurements at a water vapour pressure of 30 mb give \( k_{901} = 0.4\) gm\(^{-1}\) cm\(^2\).

Taking the linear relationship we obtain curve A of figure 4.2. Unfortunately, since the absorption of water vapour is so small in this waveband, it is not possible to make measurements with lower water vapour pressures.

Although the curve fits the data at \( e = 0 \) and at \( e = 30\) mb, it clearly gives far too large values at \( e = 10\) mb. In order to remove this discrepancy, it was thought than an equation of the form

\[ k_v = k_1 p + k_2 e^2 \]

might be used (curve B, figure 4.2). When this was suggested to Bignell,
Figure 4.2 - Absorption coefficient of water vapour at wavenumber 901 cm$^{-1}$ and pressure one atmosphere, as a function of the water vapour pressure.

Curve A: \[ k_{901} = k_1 p + k_2 e \]

Curve B: \[ k_{901} = k_1 p + k_2 e^2 \]

Typical atmospheric measurements: $\times \times \times \times$

Bignell's laboratory measurement: $\circ$
he agreed that this would be more consistent with the measurements but stated that the accuracy of his experiments was insufficient to distinguish between the two formulae. He pointed out that a term of the form $k_2e^2$ could be explained theoretically by the existence of polymers of water such as $(H_2O)_2$. He further suggested that suitable values of $k_2$ would be 400, 330, 430, and 570 $gm^{-1}cm^{-2}atm^{-1}$ at wavenumbers of 1118.8, 901, 832, and 790 $cm^{-1}$ respectively.

Rearranging the equation in the form

$$k_v = k_1 (p + k_2e^2),$$

we obtain values of $k_2/k_1$ equal to 2500, 3700, 3600 and 3600 $atm^{-1}$ respectively. Since this coefficient was fairly constant, except at the edge of the waveband it was decided that, for simplicity, the value 3600 $atm^{-1}$ would be used throughout the entire waveband.

In figure 4.3, the computed and measured clear sky intensities are compared both with the self-broadening term (crosses) and without the self-broadening term (circles). It will be seen that the change is quite small for small values of the intensity, but for large values the effect of adding the self-broadening term is to bring the computed values close to the measured values. A comparison for the month of July 1966 showed very similar results.

It seems clear therefore that equations of the form

$$k_v = k_1p + k_2e^2$$

adequately describe the absorption coefficients of atmospheric water vapour in this waveband. Whether or not this is a genuine self-broadening effect is another question. The effect could equally be caused
Figure 4.3 - Clear sky intensities for July 1967.

Computed with 'self-broadening term': +
Computed without 'self-broadening term': ◇
by hygroscopic nuclei in the atmosphere which become water droplets with diameters of the order of ten micron when the relative humidity exceeds about 70% (see Mason, 1957, p.30). The presence of such droplets is quite obvious in the visible spectrum as haze. In the 8 - 13 micron waveband we might expect the effect to be even more marked as water vapour is particularly transparent here and liquid water is nearly opaque. If hygroscopic nuclei are the main cause of the high absorption, we would expect the effect to be related to the relative humidity. If self-broadening is the cause, on the other hand, it will be related to the absolute humidity. It would be interesting to compare the results of a winter day with high relative and low absolute humidity with those of a summer day with low relative and comparatively high absolute humidity. Unfortunately, other sources of error make such a comparison difficult with present data. It is most probable, in any case, that both mechanisms are operating.

(b) Calculations of Emissivity.

The computer program was designed to take any given atmospheric sounding together with cloud of any given emissivity and reflectivity, and compute the radiative intensity at the surface, as seen through a given filter system. In practice, the intensity at the surface is measured by the radiometer and we wish to use this measurement to calculate the emissivity of the cloud. The components of the radiation reaching the ground at any given wavelength are as follows (see figure 4.4) i) the radiation emitted by the cloud is \( E B(T_c) \), where \( E \) is the emissivity and \( B(T_c) \) is the black body function for the temperature of the cloud \( T_c \).
ii) the radiation transmitted by the cloud is $T I_1$, where $T$ is the transmissivity of the cloud and $I_1$ is the radiative intensity incident on the top surface of the cloud;

iii) the radiation reflected from the base of the cloud is $R I_2$, where $R$ is the reflectivity of the cloud base and $I_2$ is the upward intensity at the cloud base.

The above terms give the downward intensity immediately below the cloud. To obtain the intensity at the ground we must multiply them by the transmission of the intervening atmosphere, $T$, and add -

iv) the radiation emitted by the atmosphere below the cloud, $I_3$.

The intensity at the ground in a wavenumber interval from $\nu$ to $\nu + d\nu$ is therefore given by -

$$I_\nu = T \{E B(T_c) + T I_1 + R I_2\} + I_3$$

(Note that all the terms in this equation are functions of $\nu$).

Then, if the transmission of the filter system at wavenumber $\nu$ is $X_\nu$, the total intensity received is given by -
\[ I = \int I_v I_v' dv, \]
the integration being performed between appropriate limits.

In practice, the computer program calculates the fluxes in a slightly different way, but it is entirely equivalent to the above equations.

Now, if \( A \) is the absorptivity of the cloud, we know that \( R + A + T = 1 \) and, by Kirchhoff's law, that \( A = E \). Hence \( T = 1 - E - R \) and

\[ I_v = T \left( E B(T_c) + (1 - E - R) I_1 + R I_2 \right) + I_3. \]

This assumes, of course, that the transmissivity of the cloud is the same for downward radiation as it is for upward radiation, which is probably reasonable.

\( I \) is measured and \( T, B(T_c), I_1, I_2, I_3 \) can be computed once we know the temperature and humidity profile of the atmosphere and the height of the cloud. This leaves the values of \( E \) and \( R \) as unknowns. It had been hoped originally that, for the case of very thick clouds for which \( T = 0 \), we could write \( E = 1 - R \) and hence determine the mean value of \( R \) from the equation

\[ I = \int \left[ T \left( E B(T_c) - R (B(T_c) - I_2) \right) \right] I_v dv. \]

In practice this turned out to be impossible since, for thick clouds, the cloud base temperature is not very different from the ground temperature. This means that the term \( (B(T_c) - I_2) \) is very small compared to \( B(T_c) \). Thus the effect of errors in \( T_c \) is far greater than the effect of the term containing \( R \). As an alternative, the possibility of using a theoretical value for \( R \) was considered.

Initially, the value of \( R = 0.1 \) was used, as computed by Havard (1960) for a cloud of water droplets of radius 6 microns with incident radiation of wavelength 10 microns. Subsequently, the results of other computations became available; Feigel'son (1964), for example,
quotes values of \( R \) from 0.15 to 0.19, depending on the mean drop size in the cloud. Like Havard, he considered monochromatic radiation and droplets of uniform radius. A more recent paper by Yamamoto, Tanaka and Kamitani (1966) however, using a realistic drop size distribution, gives a value of \( R = 0.03 \). It seems probable that the reflectivity of water clouds in this waveband is small and, in view of the uncertainty in its precise value, it seemed best to set \( R = 0 \) as a first approximation, while retaining the facility to include values of \( R \) if and when these become available.

Neglecting \( R \) then, we have

\[
I = \int \left[ E B(T_c) + (1 - E) I_1 \right] X_\nu \, d\nu
\]

A problem now arises concerning the integration over wavelength, since \( E \) is a function of wavelength. Since we have measured only the integrated intensity \( I \), we can only obtain a mean value of \( E \) for the entire waveband. Rearranging the above equation we have

\[
I = \int E X_\nu T(B(T_c) - I_1) \, d\nu + \int (T I_1 + I_3) X_\nu \, d\nu
\]

or

\[
I = E' \int X_\nu T(B(T_c) - I_1) \, d\nu + \int (T I_1 + I_3) X_\nu \, d\nu
\]

where

\[
E' = \frac{\int E X_\nu T(B(T_c) - I_1) \, d\nu}{\int X_\nu T(B(T_c) - I_1) \, d\nu}
\]

In other words, \( E' \) is a weighted mean of \( E \), weighted by the function \( X_\nu T(B(T_c) - I_1) \). Typical values of the functions \( X_\nu, B(T_c) \), and \( I_1 \) are shown by curves 3, 1 and 2 respectively of figure 4.1. It will be seen that within the 'atmospheric window', \( I_1 \) is always small compared to \( B(T_c) \). It appears therefore that, if \( E \) varies at all rapidly with wavelength, \( E \) will be a function of the cloud temperature. The calculations of Havard (1960) and of Yamamoto, Tanaka and Kamitani (1966) suggest that the variation of \( E \) with temperature is quite small.
This is partly because the peak of the black body curve lies between 10 and 14 microns for all observed cloud temperatures, so that the shape of the $B(T_c)$ curve does not vary very greatly. Karshunova (1959) has claimed that the emissivity of stratus clouds in the Arctic may fall from 1.00 at $-4$ C to 0.85 at $-26$ C. However, the effect is more likely to be caused by changes in the phase, liquid water content or thickness of the clouds than by the temperature change itself.

From the above equations we see that $I$ is a linear function of $E^*$. This is very convenient as it means that it is not necessary to carry out a computation for each reading of the radiometer. All that is necessary is to compute the intensity for an emissivity of zero (i.e. clear sky), $I_0$, and the intensity for an emissivity of 100% (i.e. black body), $I_{100}$. $I_0$ is a function only of the sounding and $I_{100}$ is a function of the sounding and the cloud height and is computed for all the cloud heights observed on a given day. Then, if $I_m$ is the measured intensity at the surface, the emissivity is given by

$$E = \frac{I_m - I_0}{I_{100} - I_0} \times 100\%$$

Values of $I_m$ are read every minute, to coincide with the measurements of cloud heights which are also made every minute for high and medium cloud. Values of $E$ are then calculated for each reading.

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1 For the remainder of this thesis, we shall drop the * and use $E$ to denote the mean emissivity over the 8 - 13 micron waveband.
The computed value of $I_0$ can be compared with the measured value whenever there is no cloud and frequently a serious difference is found. The reasons for this lie partly in errors in the radiometers and partly in errors in the sounding and computation. This is discussed more fully in section 5(a). For the moment we merely note that, whatever the cause of the error, it occurs mainly below the levels at which cloud is observed. This means that the error will be approximately the same in the calculation of the cloudy sky intensity $I_{100}$ as it is in the calculation of the clear sky intensity $I_0$ and the difference $(I_{100} - I_0)$, which appears in the denominator of the expression for $E$ should be correct. In the numerator of the expression for $E$ we have $(I_m - I_0)$, where $I_m$ is the measured intensity. In order to obtain a correct difference, $I_0$ should also be a measured value. In the case of low cloud which may persist for many hours or even days, this could not always be found. But in the case of the medium and high clouds which were observed, periods of cloudiness were nearly always preceded and followed by periods without cloud so that it was possible to obtain a measured value of $I_0$. The equation for $E$ thus becomes

$$E = \frac{I_m - I_0}{I_{100} - I_0} \times 100\%$$

where $I_m$ is the actual measured intensity with a particular cloud, $I_{100}$ is the measured intensity with clear skies, $I_{100}$ is the computed intensity for cloud of 100% emissivity at the appropriate height, and $I_0$ is the computed intensity for clear skies.
5. RESULTS

(a) Clear Sky Results

As a check on the method of calculation, the measured radiation was first of all compared with the computed radiation for clear skies. The soundings are taken at 00 Z and 12 Z each day so the comparison was made for these times. Where there was cloud present at these times it was frequently possible to obtain an approximate value of the 'measured' clear sky intensity by interpolation. Some of the results of this work were shown in section 4(a) where it was shown that better agreement between computed and measured intensities was obtained if the water vapour and carbon dioxide bands and the self-broadening effect were taken into account in the computations.

In figure 5.1 the computed and measured intensities for the period July 1967 to May 1968 are compared. It will be seen that the points are spread fairly evenly about the line of equality and that the majority lie within the ±10% lines. Some spread of values is to be expected as the soundings are taken at Maniwaki, 140 miles away from Montreal. In this particular series of computations, the soundings were not modified except that the Maniwaki surface data were replaced by the surface data measured at the McGill Observatory.

When the results are broken down by season of the year, as in figures 5.1 (b), (c), (d) and (e), an interesting pattern emerges. Whereas the summer and winter results - figures 5.1 (b) and 5.1 (d) - show very good agreement, during the autumn the computed intensity tends to be too small, particularly at low values, and during the spring the computed intensity tends to be too high, particularly at high values.
Figure 5.1 - Computed intensity against measured intensity for clear skies.

(a) July 1967 to May 1968.
Figure 5.1 - Computed intensity against measured intensity for clear skies.
One possible explanation of this behaviour is as follows: the Barnes radiometer sees the sky in a mirror whose reflectivity, though high, is not 100%. It is calibrated indoors so that the mirror temperature is approximately 20°C. During the summer, the outdoor temperature will be approximately the same, so that no error will arise. In the autumn the outdoor temperature will fall, the mirror temperature will fall, and the measured intensity will be too low. This effect will be most serious on the colder days and these are the days when the intensity too tends to be lower. We would expect the lower values of the measured intensity to be too small and this is exactly what is found in figure 5.1(c). In winter, the temperature of the mirror is maintained by the heater which heats the air flowing over it, so that good results are obtained again. In spring, with the heater still on, there may be some days when the warmer weather and the sun shining on the box (these are all clear sky results) combine to raise the mirror temperature above its value at the time of calibration. We will expect that, for the larger values of intensity (which correspond to higher temperatures), the measured intensity will be too high. That this happens, indeed, is seen in figure 5.1(e).

So far the discussion of the effect of changing the mirror temperature has been entirely qualitative. In order to determine the orders of magnitude involved, the radiometer was calibrated with the mirror at temperatures 17 degrees above and 17 degrees below the usual calibration temperature of 20°C. It was found that a change in the mirror temperature of 17°C produced a change in the measured intensity of approximately 0.6

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1 This work was carried out by K.S. Han-Shun-Cheong.
watt m^{-2}sr^{-1}. (From these measurements we calculate the absorptivity of the mirror to be approximately 9\% and the reflectivity 91\%. ) A change in the measured intensity of this magnitude is shown in figures 5.1 (c) and (e) by a dashed line. It will be seen that the temperature changes in the mirror could account for much of the consistent error. Of course, there remain random errors such as the inaccuracy of the humidity and temperature sounding, variations in the ozone content of the atmosphere and the effect of pollution.

Fortunately, errors arising from changes in the mirror temperature do not affect the measurement of emissivity. This is because we are only interested in the difference between the clear sky intensity and the cloudy sky intensity, and a change in mirror temperature will alter these two intensities by equal amounts. The random errors will of course affect the measurement of emissivity. From figure 5.1 we see that, once the consistent errors are removed, approximately 90\% of the computed values will lie within ±0.5 watt m^{-2}sr^{-1} of the measured value. The effect of these random errors on the emissivity is discussed in section 5(d). One of the possible causes of the difference between the measured and computed intensities is the atmospheric pollution. The latter is measured continuously at the McGill Observatory with a filter-type pollution meter. In figure 5.2, the difference between the measured and the computed clear sky intensities is plotted against the measured pollution for the month of February 1968. There does not appear to be any significant correlation between the two quantities. An examination of the data for the month of August 1965 showed a similar lack of correlation. Of course, the radiometer is pointing vertically so that we really require the total pollution in a vertical column rather than the pollution density
Figure 5.2 - Error in computed intensity against measured pollution for the month of February 1968.
at the surface. For a few days in March 1968, information was available on the depth of the mixing layer so that it was possible to estimate the total pollution in a vertical column. However, for the few occasions where this was done, there still did not appear to be any correlation with the radiation measurements. The main effect of allowing for the depth of the mixing layer was to reduce the range of the pollution values; this is hardly surprising as high pollution counts always occur when the mixing layer is shallow and low values when it is deep. It appears therefore, that the contribution of dry pollution to the radiative intensity at the surface is quite small.\(^1\) It is still possible that hygroscopic nuclei may be important at high humidities (when they become small droplets), but our pollution meter does not measure this type of pollution.

(b) Qualitative Results

Qualitatively the radiometer produces an excellent record of the times when cloud was present over the instrument. The temperature recorded also gives a very approximate idea of the cloud temperature. If better filters were available, the difference between the indicated temperature and the cloud temperature could be considerably reduced.

\(^1\) This conclusion has been confirmed by Han-Shun-Cheong (1969) who measured the radiative intensity both at the McGill Campus in the centre of Montreal and at Ste. Anne de Bellevue, thirty miles west of Montreal, and was unable to detect any difference.
In principle, a very thin cloud at low level might give the same reading as a dense cloud at high level; in practice, it is usually the low clouds that are dense and the high clouds that are more diffuse, so that no ambiguity arises, and the approximate height of the cloud can be estimated from the radiometer trace.

The changes in humidity and temperature which are associated with the passage of fronts are shown by corresponding changes in the intensity received. Since the radiometer responds to conditions throughout the whole depth of the atmosphere (or at least that layer of the atmosphere which is below cloud), the changes take place over several hours and may be masked by the presence of various layers of cloud. In the case of warm fronts, the intensity usually starts to increase some hours before the appearance of the cloud. Then as the cloud appears and the cloud base lowers, so the intensity increases further. An example of this is shown in figure 5.3 (a).

The tendency for the intensity to start increasing before the appearance of cloud (and conversely to continue decreasing after its disappearance) is typical of layer type clouds. The effect is presumably caused by a layer of high humidity which exists before the appearance and persists after the disappearance of the visible cloud. In other words, clouds are bigger when viewed at 10 microns than when viewed at 0.5 micron. A typical case is shown in figure 5.3 (b). This phenomenon creates difficulties when we try to determine the measured clear sky intensity as it is difficult to decide exactly when the sky is 'clear'. It should be emphasized that this effect is not always observed. There were even some cases where alternate periods of cloud and clear sky were recorded on the
Figure 5.3 (a) - Radiometer trace for July 11th, 1967, showing the passage of a warm front.

-20
Radiometer recording in degrees C
-40
-60
onset of high cloud
-80
Time in hours - E.S.T.
10 12 14 16 18 20

Figure 5.3 (b) - Radiometer trace for July 27th 1966.

The shading indicates when visible cloud was present. The cloud bank at about 1830 could not be photographed as it was getting dark. The dashed line is the value of \( I_{m0} \); it slopes upwards as a warm front is approaching. The cloud type was cirro-cumulus.
film without any measurable change being recorded by the radiometer.

With convective clouds, the reverse effect was sometimes noted. That is to say, the region immediately adjacent to the cloud gave less intensity than did the undisturbed atmosphere well away from the clouds. This could be explained by downdrafts outside the convective cloud producing regions of low humidity.
(c) Barnes - Huggins comparison.

During July and August 1967 both the Barnes and the Huggins instruments were operating and it was possible to compare results. The readings of the two instruments are plotted in figure 5.4 (a) for a number of sample occasions. Some difference is to be expected in view of the difference in the filter characteristics. The Barnes instrument has a less sharp cut-off at 13 microns and hence responds more to the carbon dioxide and water vapour bands. It will therefore be affected more by temperature and humidity in the lowest layers than will the Huggins instrument. It is for this reason that the temperature measured by the Barnes radiometer is always greater than that measured by the Huggins.

In figure 5.4 (b) a comparison is shown of the computed intensities for the two instruments. In this case, identical data are put into the program except for the filter functions so that the latter are the only possible cause of differences. This figure therefore gives an idea of the difference in values to be expected from the different filters alone. The additional spread in figure 5.4 (a) must be due to random errors. The large spread of values around 10°C is almost certainly caused by rainwater entering the Huggins radiometer and collecting on the lens. Ignoring this, we find that the random spread ranges from about ±2°C at 0°C to ±4°C at -50°C.

During the period when both radiometers were operating, the emissivities of the clouds were calculated separately for each instrument and the difference between the two emissivities was found in each case. For high level clouds there were 270 instances and the mean value of this difference was 0.1 percent points with a standard deviation of 6.5 percent points. The fact that the difference between the results for
Figure 5.4 - Comparison of Huggins and Barnes readings.
the two radiometers was so small gives us confidence in the results. In particular, we can use the results of the more convenient Barnes radiometer and be confident that the mean results are very close to those that would have been obtained by the more sensitive Huggins instrument.

(d) Analysis of errors

Before we examine the actual results obtained for the emissivities of clouds, it will be useful to estimate the errors which are inherent in the method. In doing so, we can make use of the results of the preceding three sections.

(i) Errors in the measurement of intensity

The noise level of the Huggins radiometer runs from ±0.5 C at 20 C to ±2 C at -80 C, corresponding to a variation of ±0.12 watt m⁻²sr⁻¹ in intensity over the whole range. The Barnes radiometer can be read to the nearest division on the chart, which corresponds to ±0.2 watt m⁻²sr⁻¹ in intensity. These random errors will apply to individual readings of \( I_m \), the measured cloudy sky intensity.

Both instruments are subject to some calibration error; in the Huggins, the calibration is liable to drift by several degrees in a week; in the Barnes the instrument itself is stable but variations will occur due to changes in the mirror temperature as is discussed in section 5(a). These errors will apply both to the cloudy sky intensity \( I_m \) and to the clear sky intensity \( I_{m_0} \). When the difference \( (I_m - I_{m_0}) \) is found, the calibration errors, being the same in each term, will cancel out.

In determining \( I_{m_0} \), the instrument noise is unimportant as we can take an average over a period of time and eliminate fluctuations.
On the other hand, there is the problem of interpolation in between the periods of clear sky, to find the value of $I_{m_0}$ when cloud is present. Some of the difficulties encountered in finding $I_{m_0}$ were discussed in section 5(b). Another difficulty in interpolation arises in the case of frontal clouds since there is a large change in $I_{m_0}$ between the first appearance of the cloud in one air mass and its final disappearance in the other. Fortunately, in the case of high clouds, there are usually sufficient breaks in the cloud for the changes in $I_{m_0}$ to be followed. The error in $I_{m_0}$ is therefore of the same order of magnitude as that in $I_m$ and the total error in $(I_m - I_{m_0})$ is thus of the order $\pm 0.24$ watt m$^{-2}$sr$^{-1}$ for the Huggins and $\pm 0.40$ watt m$^{-2}$sr$^{-1}$ for the Barnes.

The situation with alto-cumulus will be similar. In the case of low cloud, the cover is often continuous for many hours and there may be no alternative but to use the computed clear sky intensity $I_o$. As shown in section 5(a), this should not differ from the measured value by more than 0.5 watt m$^{-2}$sr$^{-1}$.

Typical values of $(I_{100} - I_o)$ for high cloud are 3 to 4 watt m$^{-2}$sr$^{-1}$ in the summer and 4 to 5 watt m$^{-2}$sr$^{-1}$ in winter when the colder drier atmosphere gives a lower value of $I_o$. Taking 4 watt m$^{-2}$sr$^{-1}$ as an average we get an error in the emissivity $E$ due to errors in measurement of 6 percentage points for the Huggins and 10 percentage points for the Barnes. Corresponding figures for medium and low cloud are given in table 5.1. The maximum difference between the Barnes and Huggins emissivities for high clouds should therefore be of the order of 16 percentage points. In practice, the standard deviation was found to be 6.5 percentage points (see section 5(c)) which is well within this range.

\[p.65\]
(ii) Errors in the computed intensity

The values of $I_{100}$ and $I_0$ are computed from the measured cloud height and the measured temperature and humidity profile of the atmosphere. The errors in the measurement of cloud height were discussed in section 3(a), where it was shown that the maximum consistent error was of the order of ±600 ft for high cloud and the maximum random error was also of the order of ±600 ft. Assuming a lapse rate of 2 C per 1000 ft, this gives a temperature error of ±1.2 C in each case, which corresponds to an error in intensity of the order of ±0.25 watt m$^{-2}$ sr$^{-1}$. Taking 4 watt m$^{-2}$ sr$^{-1}$ as a typical value of $(I_{100} - I_0)$, we obtain both consistent and random errors of the order of ±6% for high cloud. This is the only consistent error, but there are other random errors. In particular, the temperature and humidity sounding is not known exactly. This sounding is used in two ways: firstly to find the cloud base temperature from its height and secondly to find the transmission of the atmosphere below the cloud. Fortunately, provided we are above the surface mixing layer, the temperature at the cloud base level does not usually vary by more than 3 C over a twelve hour period. By interpolation both in space and time (as explained in section 3(b)), a sounding can be obtained which should give the cloud base temperature to within ±1 C and the emissivity therefore within ±5%. The error in the transmission of the atmosphere is more difficult to determine. From table 2.1 we see that a high cloud at a temperature of -42.2 C will only record -23 C on the Barnes radiometer. This means that the transmission of the intervening atmosphere is of the order of 65% while the absorptivity is 35%. It was found that this absorptivity rarely changed by more than 20% over a twelve-hour
period. Our process of interpolation therefore should give us this absorptivity to within ±6%, the intensity received to within ±0.25 watt m⁻²sr⁻¹ and the emissivity within ±6%.

The value of \( I_0 \) is also computed from the temperature and humidity sounding and is subject to the same errors. However, by far the greatest contribution to \( I_0 \) comes from the lower layers of the atmosphere and these will be contributing equally to the clear sky intensity \( I_0 \) and to the cloudy sky intensity \( I_{100} \). In the difference \( (I_{100} - I_0) \), these errors will cancel out.

Adding together the random errors in the computed intensities, we have ±6% due to errors in height measurements, ±5% due to errors in the temperature sounding and ±6% due to errors in the humidity sounding, making a total possible error of ±17%. This, of course, is for high cloud; for medium cloud the error in height measurement is less but this is partly compensated by greater parallax error in height measurement and the greater variation in temperature at the lower level. We should also consider the fact that a cumulus cloud will modify the temperature and humidity profiles in its vicinity. The final error in \( I_{100} \) due to height and temperature errors is therefore probably much the same as for high cloud but, since \( (I_{100} - I_0) \) is larger, the percentage error is less.

For low cloud the effect of the intervening atmosphere is quite small and, although the temperature varies more rapidly, its fluctuations can be followed more easily as we have continuous temperature measurement at the surface. The errors in height measurement with the ceilometer vary from about ±300 ft at 1000 ft to ±2000 ft at 5000 ft. Most data were collected for cloud below 2000 ft so we can take ±600 ft as a typical
error. The error in emissivity is thus of the order of $\pm 6\%$ due to height errors and $\pm 4\%$ due to temperature errors. These errors are summarized in table 5.1. In order to find the total possible error in any one measurement it is necessary to add items 4, 5 and 6 in this table. Thus for a cirrus cloud of emissivity 35%, for example, the maximum consistent error would be $\pm 2$ percentage points ($6\%$ of 35) and the maximum random error would be $\pm 12$ percentage points ($6 + 17\%$ of 35). For alto-cumulus cloud of emissivity 90%, the consistent error would be $\pm 3$ percentage points ($3\%$ of 90) and the random error $\pm 14$ percentage points ($3 + 12\%$ of 90).

These values are for the Huggins radiometer.

Table 5.1 - Errors in emissivity for the Barnes radiometer (Huggins errors in brackets, where different)

<table>
<thead>
<tr>
<th>Cloud level</th>
<th>high</th>
<th>medium</th>
<th>low</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Error in $I_m$ (Watt $m^{-2}sr^{-1}$)</td>
<td>0.20 (0.12)</td>
<td>0.20 (0.12)</td>
<td>0.20 (0.12)</td>
</tr>
<tr>
<td>2. Error in $I_{mo}$ (&quot; )</td>
<td>0.20 (0.12)</td>
<td>0.20 (0.12)</td>
<td>0.5</td>
</tr>
<tr>
<td>3. Typical value of $(I_{100} - I_o)$ &quot;</td>
<td>4</td>
<td>8</td>
<td>11</td>
</tr>
<tr>
<td>4. Error in $E$ due to 1. and 2. (percentage points)</td>
<td>10 (6)</td>
<td>5 (3)</td>
<td>6</td>
</tr>
<tr>
<td>5. Errors in $I_{100}$ (%):</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(a) Alignment of cross-wires</td>
<td>6</td>
<td>3</td>
<td>0</td>
</tr>
<tr>
<td>(b) Height measurement</td>
<td>6</td>
<td>3</td>
<td>6</td>
</tr>
<tr>
<td>(c) Temperature measurement</td>
<td>5</td>
<td>4</td>
<td>4</td>
</tr>
<tr>
<td>(d) Humidity measurement</td>
<td>6</td>
<td>5</td>
<td>0</td>
</tr>
<tr>
<td>6. Total of 5(b), (c) and (d)</td>
<td>17</td>
<td>12</td>
<td>10</td>
</tr>
</tbody>
</table>
These errors apply to a single measurement of the emissivity. When we take the mean of a number of statistically independent measurements, the consistent error is unaltered but the probable random error is reduced by a factor equal to the square root of the number of measurements.

(iii) Error in reflectivity

There remains the error due to the omission of the reflectivity term in the equations. Estimates of $R$ range from 0.03 (Yamamoto, 1966) to 0.19 (Feigel'son, 1964) via 0.10 (Havard, 1960). There is some reason for preferring the Yamamoto computations as he has used a fairly realistic spectrum of droplet sizes in his model cloud, whereas the other workers have only considered droplets of a single size. If we take the Yamamoto value then the equation for $E$ becomes

$$E = \frac{I_m - I_{m0}}{I_{97} - I_0} \times 97\%$$

Where $I_{97}$ represents the intensity for an emissivity of 97% and a reflectivity of 3%. The use of this equation implies a linear relationship between $E$ and $I_m$, so that if $R \neq 0$, $R$ must vary linearly with $E$. The calculations of Havard and of Yamamoto show that this is not quite true, but that it is a good approximation. For low cloud, $I_{97} = I_{100}$, since the temperature of the cloud is not very different from the temperature of the ground. The values of $E$ have therefore simply to be reduced by 3%. For high cloud, the upward intensity at the cloud base will be greater than the black body intensity for the cloud temperature, so that the effect of the reflectivity will be to increase the intensity received at the surface.
This increase was calculated in each case for \( R = 0.03 \) and it was found to be typically of the order of 0.3 watt \( m^{-2}sr^{-1} \) for high clouds. This means that the value of \( E \) should be reduced by approximately 10\% for high clouds. The reason why this percentage is so high is that the upward intensity at the base of a high cloud is much greater than the downward intensity.

The errors of the previous paragraph apply if we are interested in \( E \) for its own sake. If we are interested in \( E \) in order to calculate intensities, the situation is not so serious. \( E \) has here been calculated from the intensity, assuming \( R \) to be zero; if we reverse this process we shall obtain the same answer, provided the surface temperature is the same. If the surface temperature is not the same, then the error will be \( 3\% \) (following Yamamoto) of the difference between the upward intensity at the time of measurement and the upward intensity at the time of calculation.

In the case of low cloud, this difference will always be small since the temperature at the surface is closely correlated with the cloud temperature. In the case of high cloud, the correlation with surface temperature will be small but the surface temperature is unlikely to vary by more than \( \pm 25^\circ C \) from a mean value. This would give an error of only 2\% in intensity at the surface.

Finally, it should be noted that all calculations of \( R \) which have been quoted have been made for water droplets. Few results have been published for the much more difficult case of ice crystals.
(e) Low Cloud

Cloud heights below 5000 ft could not be measured photographically but the Dorval ceilometer records were used, as was explained in section 3 (a). Since the ceilometer is approximately nine miles west of the McGill campus, it was not possible to match the ceilometer record exactly with the radiometer record and the method was used only when a uniform deck of low cloud persisted for at least an hour. Mean values of the cloud height and of the radiation received were taken for each hour. This procedure contrasts with that for medium and high cloud where spot values were taken at one minute intervals. The mean emissivity was calculated for each hour as explained in section 4 (e). A total of 248 values were obtained, from 34 days, spread throughout the period July 1967 to February 1968. The mean of these values was 99.7% and the standard deviation was 6.8 percentage points.

From table 5.1 we obtain an estimated maximum error of ±16% for a low cloud of $E = 99.7\%$ and the observed deviations are well within this range. In estimating the probable error of the mean, we clearly cannot take all 248 values as statistically independent since many of them are taken on consecutive hours and will be strongly correlated. On the other hand, to suppose that we have only 34 independent values is unduly conservative, as this is equivalent to taking only one reading each day instead of an average of 7. The correct number must be somewhere between 34 and 248 but, since we do not know what it is, we will take the more conservative figure of 34 as the number of independent observations. The probable maximum error in the mean is therefore \( 1/\sqrt{34} \) of the maximum error in one value, or approximately 2.5%. Our value for the emissivity
of low cloud is therefore $99.7 \pm 2.5\%$.

All authorities agree that the transmission of low thick clouds in the infra-red is zero. The reflectivity of such clouds will have very little effect on the intensity at the surface as the cloud temperature is close to the surface temperature. Ignoring reflectivity as we have, the emissivity of such clouds should be 100%. The fact that the mean measured value is $99.7\%$ is therefore excellent confirmation of the accuracy of the instruments and the validity of the method. Of course, the effect of the intervening atmosphere is smallest for low clouds and the calibration of the radiometers is most reliable at the higher temperatures, so that we cannot assume that the same accuracy will apply to high clouds. Nevertheless, at least at one end of the range, the method produces excellent results.

(f) Medium Cloud

A total of 1188 observations of medium cloud were made on 19 days during the period July 1967 to May 1968. All of these observations were of alto-cumulus, since alto-stratus could not be measured by the stereophotographic technique. The emissivity was calculated for each observation and the distribution of these emissivities is shown by the dashed lines in figure 5.5. It will be seen that values range down to zero. The reason for this is that in some cases the field of view of the radiometer was

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1 The term 'maximum error' has been used rather loosely in this work. This is inevitable, as the errors have only been estimated approximately. In this case the 'maximum error' is approximately twice the standard deviation so that, assuming a normal distribution, the probability of an error greater than the 'maximum error' is less than 0.05. Such a definition of the term 'maximum error' is consistent with its use elsewhere in this thesis.
Figure 5.5 - Distribution of emissivity for alto-cumulus for the period July 1967 to April 1968.

Dashed line: all observations (1188 values).

Solid line: cases for which the cloud completely covered the field of view of the radiometer (865 values, mean 78.5%).
not completely covered by cloud. If we exclude all such cases, we are left with 865 values with the distribution shown by the solid lines in figure 5.5. It will be seen that, although the lowest values have been excluded, there are still some values as low as 25 to 30%. We must conclude therefore that although most values of the emissivity lie in the 75 to 100% range, there are some parts of the cloud which have a lower emissivity than this. Alto-cumulus is cellular in structure and it is at the edges of these cells that the low values occur. Of course, the method used to measure the cloud height could only be used when the cloud had some visible structure. It is entirely possible that this may have biased our selection of data in favour of the edges of cloud cells so that, the mean emissivity of 79% which we found is lower than the mean for all alto-cumulus. It is certainly not applicable to all medium cloud.

From table 5.1 we obtain an estimated consistent error of ±3% and an estimated random error of ±15 percentage points (5+12% of 79) for a medium cloud with an emissivity of 79%. Taking, as before, the more conservative value for the number of independent data (i.e. 19 rather than 865) the probable maximum random error in the mean is a little over 3 percentage points. Combining the consistent and the random errors, the mean value of the emissivity of alto-cumulus was found to be 79 ± 6%.

When the emissivity for each day is plotted against time over the whole year (see figure 5.6) it appears that lower values tend to occur in the winter months. This is in keeping with the findings of Marshunova (1959) who found a reduction in the emissivity of Arctic stratus clouds from 100% in the summer to 85% in the winter. Unfortunately, we do not have sufficient winter observations to be able to confirm or reject this hypothesis with any great degree of confidence.
Figure 5.6 - Variation of mean emissivity of alto-cumulus with time of year.
(g) High Cloud

A total of 1845 values for the emissivity of high cloud were obtained on 26 days during the periods July - August 1966 and July 1967 - May 1968. The mean of these values is 34.7% and the distribution is shown in figure 5.7. It should be borne in mind that these 1845 values are not statistically independent. Observations were made at one minute intervals whenever high cloud was observed and a particular cloud type usually persists for many minutes. In view of this, an average emissivity was found for each of the 26 days on which high cloud was observed. These values are plotted against the time of year in figure 5.8 and in this case there does not appear to be any particular seasonal trend. This is quite reasonable as the temperature of such clouds remains fairly constant throughout the year and only the altitude changes with the season. If we take the average of the 26 daily mean values we obtain an emissivity of 29.2%. The difference between this and the mean of all values, 34.7%, reflects the fact that cloud banks which persist for several hours are likely to be denser than those which last only a few minutes; giving equal weight to each cloud bank therefore favours the low values, while giving equal weight to each minute of observation favours the high values.

The total of 1845 values is made up of 1384 observations of cirrus, 316 observations of cirro-stratus and 145 observations of cirro-cumulus. The mean emissivities were 34.4% for cirrus, 33.4% for cirro-stratus and 40.0% for cirro-cumulus. The higher value for cirro-cumulus is certainly reasonable in view of the denser appearance of this type of cloud. It was felt however, that the emissivities were sufficiently close for all types of high cloud to be placed in a single category for the
Figure 5.7 - Distribution of emissivity for high cloud for the periods July-August 1966 and July 1967 - May 1968.

Mean emissivity 34.7%
Figure 5.8 - Variation of mean emissivity of high cloud with time of year.
purpose of radiation calculations.

For a high cloud of emissivity 35%, table 5.1 gives an estimated consistent error of ±2 percentage points (6% of 35) and a random error of ±16 percentage points (10 + 17% of 35). Taking, as before, the more conservative value for the number of statistically independent observations (26, rather than 1845), the probable maximum random error in the mean is ±3 percentage points (16 /√26). Combining the consistent and the random errors, the mean value of the emissivity of high clouds was found to be 34.7 ± 5%.

The mean value obtained (approximately 35%) is somewhat lower than the value of 50% which has commonly been used in the past. One of the physical reasons for the emissivity being so low is undoubtedly the fact that cirrus generally consists of strands of cloud interspersed with bands of blue sky. Our radiometer has a field of view of 3° (1500 ft at a height of 30,000 ft) and hence gives us a mean value for both the patches of cloud and the intervening blue sky. It would be possible to estimate the percentage of the field of view which was covered by actual areas of cloud and come up with an emissivity for the cloud alone. Such a process would be difficult since the cloud fades away very gradually at the edges without any sharp demarcation. Furthermore, it is not the general practice to record cloud cover in this way. When an observer records 4/10 of cirrus, this implies that 4/10 of the sky contains some cirrus cloud, while 6/10 is completely clear; he does not stop to add up the areas of each tiny strand of cloud and each tiny band of blue sky. It was felt that this study would be most useful if the same procedure were followed since it is now possible to use observers records of cloud cover (which are readily
available) to calculate the longwave radiation field. The fact that
the cloud emissivity is low should not surprise us in view of the
procedure adopted.

Figure 5.9 - An example of cirrus cloud as seen by the 16-mm camera
(1040 E.S.T. on 31 January, 1968)

At the left is the mirror showing the reflection of the digital
clock; the numbers are a little hard to read, as the mirror was not
quite properly aligned on this occasion, and they are laterally inverted
by the mirror (the film was normally projected from the other side
so that the numbers appeared the right way round on the screen).
The black lines in the photograph are the cross-wires which were
approximately 1.5 m above the camera. The dashed lines drawn on
the photograph indicate the field of view of the Barnes radiometer (30);
that of the Huggins radiometer would be a little smaller. The difficulty
of distinguishing between cloud and clear sky can be readily appreciated.
Virtually no point in this photograph is completely free of cloud
although in some parts it is clearly extremely tenuous. It is not
hard to see why the emissivity of such a cloud bank ranges from a few
percent to nearly 100%.
6. DISCUSSION OF RESULTS

The aim of this work was to measure the emissivity of clouds in the 8-13 micron waveband. This was successfully done, the mean values being approximately 100% for low stratus, 80% for alto-cumulus and 35% for cirro-stratus and cirrus. The accuracy of these mean values was approximately ±2.5, ±6 and ±5 percentage points respectively. Of course, particularly for high clouds, the mean is not necessarily the most significant parameter, as the quantity being measured has such a wide range. For heat budget studies, the mean value will be useful, but in many other cases it will be necessary to know the complete distribution of emissivities (as in figure 5.7).

The accuracy of individual emissivity measurements could be improved in the following ways: In the first place, it would be an advantage to use a longwave cut-off filter with a sharp cut-off at 12.5 microns. (The shortwave cut-off is reasonably good, at least in the Ruggins radiometer). This would eliminate much of the error caused by uncertainty in calculating the effects of the carbon dioxide and water vapour bands. Since the clear sky intensity $I_0$ would be much less, the difference $(I_{100} - I_0)$ would be greater and the errors in $I_m$ would have less effect on the emissivity. Furthermore, errors in measuring the humidity of the atmosphere would be less important. As a second step, some improvement in the temperature and humidity soundings could be obtained by making measurements at the same place that the radiosonde is released. This would obviate the necessity for making interpolations in space as well as time. Finally, the method of measuring cloud height could be improved by using a vertically pointing lidar, or optical radar, unit. Such a system has been described by Barrett and Ben-Dov (1967) and has been used successfully for pollution measurements. The advantages of such a
system would be that it could measure the cloud base at any height and the accuracy would be independent of the height; it would not be limited to hours of daylight, and it would measure uniform cirro-stratus and alto-stratus just as well as clouds with visible structure. This would eliminate any possible bias in the selection of clouds. It might even be possible, using a carbon dioxide laser as a source, to make the height measurements in the 8-13 micron waveband. Using these improvements in technique, the errors in individual measurements of emissivity could be halved.

It is clear from our results, and even from casual observation, that clouds in general, and high clouds in particular, vary enormously from one moment to another. In order to interpret the results of emissivity measurements, it is necessary either to relate the emissivity to the physical properties of the cloud (size distribution and concentration of particles, thickness, etc.) or to find the distribution of emissivity over a long period of time. In the present work, the latter course has been attempted and the essence of the results is the histogram of figure 5.7. Presumably, if more values were available, a smoother curve would be obtained but it is unlikely that the general shape or the mean value would be greatly altered. Not many results have been published by other workers in this field but those that have been are entirely consistent with our findings. Brewer and Houghton (1956) measured both upward and downward intensities from an aircraft at various heights and, for the two cirrus clouds that they observed, obtained emissivities of 5% and 80%. More recently, Valovcin (1968) has measured the upward intensity only from a U-2 aircraft flying
above, through and below jet-stream cirrus. With the data from 23 flights, each of several hours, he has deduced emissivities varying from 5% to 100% but he does not give the distribution of values. A more accurate method is to measure the downward intensity from an aircraft since the cloud is then contrasted against the cold background of the sky rather than the warm background of the earth. This procedure has been adopted by Kuhn and Weickmann (1969) who, in the course of twelve flights, obtained values ranging from 5% to 50% for the emissivity of cirrus clouds. They have found quite good correlation between the emissivity of the cloud and its thickness, with values ranging from about 15% for clouds one kilometer thick to 45% for clouds five kilometers thick. They also claim that cirrus clouds with a thickness of only 100 m occur very frequently in the neighbourhood of the tropopause. Such clouds are visible from an aircraft flying immediately below them but are generally not visible from the ground. The existence of such 'invisible' cirrus would also explain the anomalous decrease in the upflux near the tropopause which has been found by Riehl (1962) and others using radiometersondes. This theory has been carefully studied by Zdunkowski et al. (1965) who has made calculations for spherical ice particles of radius 120 microns\(^1\) and finds that a particle concentration of approximately \(1.5 \times 10^{-3}\text{cm}^{-3}\) would be invisible but would decrease the upflux by as much as 5%.

\(^1\) This radius seems rather large but it is based on a claim by Schaefer that ice crystals of less than 100 microns radius could not persist for sufficient length of time.
It seems very likely that it is such a cloud that causes the increase in intensity which we have observed just before the arrival of visible cirrus. Of course, we have only calculated the emissivity when visible cirrus was present, so that we have probably missed numerous values in the $0-5\%$ range. If the number of values in this range were increased by a factor of three, all the points in figure 5.7 would lie close to a smooth curve and the mean would drop to $31\%$. Very few results have been published of computations of the emissivity of cirrus apart from those of Zdunkowski et al. (1965) mentioned above. One of the problems about computing the emissivity of cirrus is that, quite apart from the computational difficulties, very little is known about the size, shape or concentration of the ice particles. All of the results discussed here are for the 8-13 micron waveband.

Since the emissivity of cirrus appears to range from zero to $100\%$, it is natural to wonder whether we are really looking at the same cloud type all the time, or whether we have included several different cloud types under one heading. Some light is shed on this problem by looking at the distribution of emissivities for high cloud on a single day, so that we are considering only one type of cirrus. Unfortunately, the number of values obtained on any one day is not sufficient to give a smooth distribution curve; it would be interesting and useful to measure the radiation on a few days with a higher chart speed and take the photographs every six seconds, say, instead of every sixty seconds, so that a much larger number of values could be obtained. This would also enable us to

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1 The author has recently been informed of unpublished works by J.H. Joseph of Tel-Aviv University and by F.F. Hall of the McDonnel Douglas Corporation.
compare the infra-red emissivity with the visible opacity. However, even with the number of values which we have obtained (up to 260 on one day) it is possible to get some idea of the distribution, and the answer is that, even on a single day, the range is very large. Values usually ranged down to zero, three-quarters of the days having values below 5\% and all days having values below 15\%. The maximum value recorded on any one day ranged from 25\% to 100\%. A simple model of the distribution of cirrus emissivities would be as follows: for any one cloud bank, values are uniformly distributed between zero and some maximum value $E_m$; for different cloud banks, the values of $E_m$ are uniformly distributed between zero and 100\%. With this model, the distribution curve for all values would be a straight line with a maximum at $E = 0$, falling to zero at $E = 100\%$, and the mean emissivity would be 33.3\%. The distribution of the daily means would be uniform between zero and 50\%. An examination of figures 5.7 and 5.8 shows that these are quite close to our measured distributions. We appear to have rather more values in the 5-30\% range, but this is off-set by the lack of values in the 0-5\% range. But, as we discussed earlier, there is strong evidence that the number of values in the 0-5\% range has been underestimated. Although our model is not quantitatively perfect, it does give an infinitely better picture of cirrus emissivities than the simple statement that the mean is 35\%.

It appears from our results that widely ranging emissivities are characteristic of cirrus clouds and are not caused by grouping different cloud types under one heading. Attempts to categorize high clouds confirm this; the mean emissivity for cirro-stratus is almost the same as that for cirrus, while that of cirro-cumulus is only slightly larger.

The distribution of emissivities must, to some extent, be a function
Suppose the sky were divided into equal bands of 'black' cloud and clear sky each with a width of 500 metres: a radiometer with a field of view of 5000 metres would record a constant emissivity of 50%; one with a field of view of 500 metres would record nearly half the values as zero and half as 100%; while one with a field of view of 5000 metres would record a uniform distribution between zero and 100%. To find the effect of increasing the field of view we can simply average a number of our readings; but we cannot tell what the effect might be of decreasing the field of view. We can only record our impression that this would not change the distribution by very much. On the other hand it seems unlikely that the width of the cloud bands was never greater than the field of view; and yet a bimodal distribution was never observed. Also, an examination of cirrus clouds, or of the photograph in figure 5.9, shows that the opacity varies over a wide range.

In view of the fact that we obtain a value of 35% for the emissivity of cirrus, whereas a value of 50% has generally been used in the past for radiation calculations, it is interesting to estimate the contribution of high cloud to the radiation budget. If we take the 35% rather than the 50% figure, the radiative intensity reaching the ground when high cloud is present will be reduced by about 15% in the 8-13 micron waveband. This waveband accounts for only about 12% of the total longwave radiation reaching the ground however, since most of the flux is in the carbon dioxide and water vapour bands. The reduction in the total longwave intensity is therefore less than 2%. The radiation going out into space will be increased, as more of the upflux now comes from the ground and less from the cloud. The increase amounts to about 10% for the atmospheric window and, since this waveband accounts for roughly half of the total
intensity, the latter is increased by 5%. The mean cloudiness and the
frequency of cirrus occurrence have been given by Vowinckel (1962) for
latitudes greater than 65°N and by McDonald (1938) for the oceans.
By multiplying the percentage cloudiness by the fractional frequency of
cirrus it is possible to obtain an approximate figure for the percentage
of the earth's surface covered by cirrus. The mean for the northern
hemisphere is 12%, with values ranging from 16% in the tropics to 9%
in temperate latitudes.¹ The change in the net longwave radiation loss
is therefore only 0.6% (12% of 5%) for the whole northern hemisphere
and 0.8% (16% of 5%) for the tropics.

We should also note the importance of this work to satellite
measurements in the 8-13 micron waveband. It has been the practice to
use the radiation in this band to determine the height of the cloud tops,
assuming them to radiate as a black body. The results obtained here show
that this assumption is frequently false. The same radiation might be
received from a dense medium cloud as from a tenuous high cloud and
measurements in a single waveband will not detect the difference. Finally
we note the importance of precise values for cloud emissivity in calculations
of radiative heating rates in the atmosphere. The main result of reducing
the cloud emissivity will be to reduce the heating rate in the vicinity
of the cloud but the radiation field is complex and there will presumably
be secondary effects.

In conclusion, we claim that the distribution of the emissivities
of high clouds has been measured here for the first time, and that the
values obtained will be useful for radiation calculations in the atmosphere.

¹ This information was supplied by S.Woronko.
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APPENDIX I - ABSORPTION COEFFICIENTS OF ATMOSPHERIC ABSORBERS

(a) The water vapour 8 - 13 micron continuum.

One of the problems in calculating the precise effect of water vapour is to obtain reliable values for the generalised absorption coefficients. Most workers have concerned themselves with the absorption in the main water bands and the 'window' region has received little attention. Furthermore, measurements in the window region pose special problems: since the absorption is small, long path lengths are needed and these are difficult to obtain in the laboratory. Bignell (1965) has attempted such work but has been forced to use vapour pressures which are higher than those commonly encountered in the atmosphere. In the atmosphere itself, measurements can be made in two ways; either by measuring the attenuation of the solar beam as has been done by Roach and Goody (1958) and by Saiedy (1960) or by measuring the clear sky emission as has been done by Bolle (1965). The results of some of these measurements are presented in figure A1.1. It will be seen that there are considerable discrepancies between the results, extreme values differing by as much as a factor of four. Particularly high values have been obtained where measurements are made with large water vapour concentrations and Bignell has produced evidence that this may be due to a large self-broadening effect. However, the measurements are not sufficiently precise for the self-broadening coefficient to be determined with any degree of accuracy.¹

¹ Absorption lines are broadened by molecular collisions. This effect is known as 'Lorentz' or 'pressure' broadening. It appears that collisions with other water molecules are more effective than are collisions with inert molecules such as nitrogen. This is
High absorption coefficients were also obtained by Roach and Goody (1958) in London and by Bolle (1965) at Beer Sheva in Israel. In both cases there was probably an important contribution by the atmospheric aerosol - smoke in London and sand at Beer Sheva. So far, it has not been possible to determine the effect of atmospheric aerosols in any precise quantitative way. It may be that the high values of the absorption coefficients which are obtained at high humidities are also partly due to the atmospheric aerosol. Many of the particles are hygroscopic and can exist as droplets of solution even at quite low relative humidities (see Mason, *Physics of Clouds*, p.30). This effect is observed in the visible spectrum where visibility is found to depend strongly on relative humidity. In the atmospheric window, the effect will be even more marked, since liquid water absorbs very strongly in this region and water vapour only very weakly.

Disregarding then the results of Bignell (1965) which were obtained in the laboratory with high water vapour pressures, those of Roach and Goody measured in London and those of Bolle measured in Beer Sheva, we are left with two groups of data: the data deduced from solar measurements by Roach and Goody in Ascot, by Saiedy and by Bignell, and the data deduced from the clear sky emission at St.Agata by Bolle. The values of

1 presumably because the water molecules may cling together for some time on collision and the effect is known as 'self-broadening'. Instead of writing \( k_v = k_u \frac{p}{P_0} \), where \( k_v \) is the absorption coefficient at a total pressure of \( p \) and \( k_u \) is the absorption coefficient at standard pressure \( P_0 \), we must now write \( k_v = k_u \frac{(p + (B - 1)e)}{P_0} \), where \( e \) is the partial pressure of water vapour and \( B \) is called the self-broadening coefficient. Estimates of \( B \) for the 8 - 13 micron waveband vary from 5 to 30.
Figure A1.1 - Absorption coefficients of water vapour in the atmospheric window after various authors.

<table>
<thead>
<tr>
<th>Author</th>
<th>Method</th>
</tr>
</thead>
<tbody>
<tr>
<td>Saiedy (1960)</td>
<td>solar</td>
</tr>
<tr>
<td>Bignell (1963)</td>
<td>solar</td>
</tr>
<tr>
<td>Roach and Goody (1958), Ascot</td>
<td>solar</td>
</tr>
<tr>
<td>Bolle (1963), St.Agata</td>
<td>emission</td>
</tr>
<tr>
<td>Vigroux (1959)</td>
<td>emission</td>
</tr>
<tr>
<td>Bolle (1965), Beer Sheva</td>
<td>emission</td>
</tr>
<tr>
<td>Roach and Goody (1958), London</td>
<td>solar</td>
</tr>
<tr>
<td>Bignell (1965)</td>
<td>laboratory</td>
</tr>
</tbody>
</table>
$k_v$ deduced from emission measurements tend to be 20 - 40\% higher than those from solar measurements. It is not easy to see why this should be: the theory has, of course, ignored the possibility of scattering by the atmospheric aerosol. However, if this were significant, then the transmission of the solar beam would be reduced and the values of $k_v$ deduced from solar measurements would be greater, not less. The variation of $k_v$ with frequency is fortunately very similar in both sets of measurements so that if the values of, say, Saiedy are in error, they will be so by a constant factor. By comparing the calculated with the measured intensity at the surface, it should be possible to determine the value of this factor. Unfortunately, the spread of values caused by inaccurate soundings is so large that this factor cannot be determined with any degree of accuracy. Values which have been obtained suggest that the Saiedy coefficients may be low by a few percent but certainly not by as much as 20\%. The values of the absorption coefficients given by Saiedy have therefore been used in the computer program.

Some of the early results produced further evidence for the existence of a self-broadening effect and this was eventually included in the computer program also. Details of this work were given in section 4(a). Finally, no attempt has been made to allow for the variation of $k_v$ with temperature. The results of Bignell (1965) obtained in the laboratory at various temperatures between 22 and 44 C suggest that $k_v$ may decrease as the temperature rises. However, the evidence at this stage is far from conclusive and, in any case, the variation is quite small.

It was found that the computed intensity for clear skies has an approximately linear variation with $k_v$, with a 10\% increase in $k_v$ causing an increase of 4 to 5\% in the intensity.
(b) The 6.3 micron water vapour band

At the shorter wavelength end of our waveband (8 micron) the germanium substrate filter permits some radiation to pass from the extreme wing of the 6.3 micron water vapour band. This band was therefore considered in the computation.

The 6.3 micron band of water vapour has been measured in some detail by Howard, Burch and Williams (1955). They have obtained empirical formulae relating the total absorption ($\int A \, dv$, where $A$ is the absorptivity at wavenumber $v$ and the integration extends over the whole band) to the amount of water vapour ($u \text{ gm.cm}^{-2}$) and to the pressure. In the present study, these formulae were not applicable, as only the extreme edge of this band is passed by the filters, and it was necessary to go back to the original absorption curves. However, since most of the water vapour in the atmosphere lies below the 500-mb level, only the curves for high pressures were used (in practice, those for 742 and 736 mm of Hg). The area under the absorption curves for each of the frequency intervals used ($A_j$) was measured separately and plotted as a function of $u$ (see figure A1.2).

Now the intensity received from a given layer is proportional both to the absorptivity and to the filter transmission ($X$) so that we require to find not $\int A \, dv$ but $\int X \, A \, dv$; and only if $X$ remains reasonably constant over the wavenumber interval, may we replace this by $X_j \int A \, dv$ where $X_j$ is the mean filter transmission for interval $j$. For interval 12 (1204 to 1250 cm$^{-1}$), $X$ varies rapidly with wavenumber (see curve 3, figure 4.1) and this interval was
Figure A1.2 - Absorption in the 6.3 micron band of water vapour, based on the data of Howard, Burch and Williams (1955).

Band 9    - 1066 to 1112 cm\(^{-1}\)
Band 10   - 1112 to 1158 cm\(^{-1}\)
Band 11   - 1158 to 1204 cm\(^{-1}\)
Band 12.1 - 1204 to 1227 cm\(^{-1}\)
Band 12.2 - 1227 to 1250 cm\(^{-1}\)
therefore divided into two parts, 12.1 and 12.2. The mean absorption for interval 12 was then obtained from

\[ A_{12} = \frac{A_{12.1} X_{12.1} + A_{12.2} X_{12.2}}{2 X_{12}}. \]

It was found that \( X_{12.2} \ll X_{12.1} \) so that \( A_{12} \approx A_{12.1} \) and the curve for \( j = 12.1 \) was adopted for the whole of interval 12. A similar procedure applied to interval 11 led to a slight lowering of this curve. For intervals 10 and 9, no change was necessary as \( X \) remains fairly constant in these intervals.

It was found that the absorption curves could be adequately represented by an empirical formula,

\[ A_j = C_j + D_j \log u. \]

This formula is similar to that used by Howard, Burch and Williams for the total absorption of the whole band except that the pressure dependence has been omitted. The values of \( C_j \) and \( D_j \) are listed in table A1.1. Some error is caused by neglecting the pressure dependence

<table>
<thead>
<tr>
<th>( j )</th>
<th>&lt;9</th>
<th>9</th>
<th>10</th>
<th>11</th>
<th>12</th>
</tr>
</thead>
<tbody>
<tr>
<td>( C_j )</td>
<td>0</td>
<td>2</td>
<td>9</td>
<td>21</td>
<td>38</td>
</tr>
<tr>
<td>( D_j )</td>
<td>0</td>
<td>5</td>
<td>9</td>
<td>13</td>
<td>22</td>
</tr>
</tbody>
</table>

Table A1.1 Values of constants for the 6.3 micron band
but since the range of effective pressures is quite small (from 760 mm to about 500 mm), this error should not be too great. In any case, the filters are such that the 6.3 micron vibration-rotation band makes only a small contribution to the total intensity received. The empirical formula does not automatically limit \( A_j \) to values between 0 and 100 \%; this limitation is imposed subsequently.

(c) The 15 micron carbon dioxide band.

The absorption measurements of Howard, Burch and Williams (1955) were used for the 15 micron \( \text{CO}_2 \) band, and band areas (\( \int A \, dv \)) were measured for each relevant wavenumber interval. The mean pressure of the \( \text{CO}_2 \) layers will vary from about 760 mm of Hg for the lowest layer to about 380 mm of Hg for the whole atmosphere. Since we are only interested here in calculating the radiation at the surface, it is not necessary to consider pressures lower than 380 mm of Hg. Two sets of results were plotted, one for pressures of 750 and 745 mm and one for pressures of 350 and 375 mm. As noted previously, the radiation is limited by a filter system of transmission \( X \), and the expression

\[
\int A X \, dv
\]

can only be replaced by \( A_j X_j \) (where \( A_j = \int A \, dv \) and \( X_j = \int X \, dv \) over the appropriate limits) if either \( A \) or \( X \) remains fairly constant. In the present case, both \( A \) and \( X \) are varying rapidly. However, if the curves cannot be replaced by constant values, they can, to a fair degree of accuracy, be replaced by straight lines. That is, the absorption can be considered to vary linearly from \( A_j - \frac{1}{2} \Delta A_j \) at one limit of the waveband to \( A_j + \frac{1}{2} \Delta A_j \) at the other. Similarly,
the filter transmission varies linearly from \( X_j - \frac{1}{2} \Delta X_j \) to \( X_j + \frac{1}{2} \Delta X_j \).

In this case it may be shown that (Allen, 1965, p.18)

\[
\int A \, \text{d}v = A_jX_j + \frac{1}{12} \Delta A_j \Delta X_j
\]

Thus the mean absorption \( \bar{A}_j = \frac{1}{X_j} \int A \, \text{d}v = A_j + \frac{\Delta A_j \Delta X_j}{12X_j} \).

This quantity was calculated in each case and plotted against the CO2 amount (\( w \) atmo-cm) in figure A1.3. It will be seen that the absorption is generally less at the lower pressure. However, since the difference is not great, and since the low pressures will always be associated with high CO2 amounts and the high pressures with low CO2 amounts, it was decided to draw a single line for each wavenumber interval, giving more weight to the low pressure values when \( w \) is large. These lines are shown on figure A1.3. The absorption is thus given by the empirical formula

\[
\bar{A}_j = C_j + D_j \log w
\]

where \( C_j \) and \( D_j \) are constants given in table A1.2. If the formula gives a negative value, \( \bar{A}_j \) is set equal to zero and if it gives a value greater than 1.0, then it is set equal to 1.0.

<table>
<thead>
<tr>
<th>( j )</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>&gt;3</th>
</tr>
</thead>
<tbody>
<tr>
<td>( C_j )</td>
<td>38</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>( D_j )</td>
<td>25</td>
<td>20</td>
<td>3</td>
<td>0</td>
</tr>
</tbody>
</table>

Table A1.2 Values of constants for the 15 micron CO2 band.
Figure Al.3 - Absorption in the 15 micron band of carbon dioxide
from the data of Howard, Burch and Williams (1955).

Band 1 - 698 to 744 cm$^{-1}$
Band 2 - 744 to 790 cm$^{-1}$
Band 3 - 790 to 836 cm$^{-1}$
(d) The water vapour rotation band.

For the water vapour rotation band, few measurements are available. It was decided therefore to use the results of Elsasser (1960). These are based on theoretical values computed from quantum mechanics by Yamamoto and confirmed by the measurements of Dr. C. Harvey Palmer at John Hopkins University and of Professor E. E. Bell at Ohio State University. A series of typical absorptions were calculated from the data supplied by Elsasser and the mean absorption for each wavenumber interval (weighted according to the filter transmission function as described for the 15 micron CO₂ band) was plotted as a function of the amount of water vapour (figure A1.4). It was found that these points could be adequately represented by the empirical formula

\[ A_j = C_j + D_j \sqrt{\lambda} \]

where \( C_j \) and \( D_j \) are constants given in table A1.3. This empirical formula was therefore used in place of the Elsasser transmission function for ease of computation. As in previous cases, whenever the formula gave a negative value, \( A_j \) was set equal to zero.

<table>
<thead>
<tr>
<th>( j )</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>&gt;3</th>
</tr>
</thead>
<tbody>
<tr>
<td>( C_j )</td>
<td>-6</td>
<td>-5.3</td>
<td>-5.1</td>
<td>0</td>
</tr>
<tr>
<td>( D_j )</td>
<td>19</td>
<td>11.7</td>
<td>6.9</td>
<td>0</td>
</tr>
</tbody>
</table>

Table A1.3 Values of constants for the water vapour rotation band
Figure A1.4 - Absorption in the water vapour rotation band based on Elsasser (1960).

Band 1 - 698 to 744 cm$^{-1}$
Band 2 - 744 to 790 cm$^{-1}$
Band 3 - 790 to 836 cm$^{-1}$
There remain the weak bands of CO$_2$ which are centered at 1064 and 961 cm$^{-1}$ (9.4 and 10.4 micron) and are therefore in the middle of the window. Burch et al. (1962) deduce the following empirical formulae for the band areas at 26°C:

For the 1064 cm$^{-1}$ band:
$$\int A\,dv = 0.023\, (w\,P_e^{0.3})^{0.75},$$

For the 961 cm$^{-1}$ band:
$$\int A\,dv = 0.016\, (w\,P_e^{0.25})^{0.78},$$

where $\int A\,dv$ is the band area, $w$ is the CO$_2$ content in atmo-cm and $P_e$ is the mean effective pressure in mm of Hg. Taking values of $w = 240$ atmo-cm and $P_e = 380$ mm of Hg as typical for the whole atmosphere, we obtain band areas of 5.34 cm$^{-1}$ and 3.66 cm$^{-1}$ for these two bands. These band areas are small compared to typical values of the 9.6 micron ozone band of the order of 50 cm$^{-1}$ but, since these bands lie near the middle of the window where the filter transmission is greatest, and since most of the CO$_2$ is at a higher temperature than the ozone, it was deemed advisable to include these bands in the computation.

(f) Corrections for the Barnes radiometer

All of the band corrections which have been discussed in this appendix were worked out on the basis of the Huggins filter function. The Barnes filters have a less sharp cut-off and extend slightly beyond the waveband for which the computer program was written. To avoid writing a new program, the transmission which was beyond these extremes was added on to the last of the 12 wavenumber intervals which are used for computing. That is to say, the transmission above 1250 cm$^{-1}$ was
added to interval number 12 (1204 - 1250 cm\(^{-1}\)) and the transmission below 698 cm\(^{-1}\) was added to interval number 1 (698 - 744 cm\(^{-1}\)). The justification for this procedure is that the absorption is already very strong in intervals 1 and 12 (see curve 5 of figure 4.1). The further increase in the absorption coefficients which occurs as we move further from the atmospheric window can have very little effect therefore. In any case, only a very small fraction of the transmitted waveband is involved. Rough calculations showed that the error produced by this approximation was small compared to other errors.