EVAPORATION FROM AN AGRICULTURAL CATCHMENT

A field and theoretical study of evaporation

Thesis submitted for the degree of

Doctor of Philosophy

University of Melbourne

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Department of Civil Engineering
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ADDENDUM

1. In Table 2.4, \( \beta \) was calculated for the period of the day when \( R_n > 0 \). In Tables 2.5 to 2.11 and Figs 2.22b and 2.24b, \( \beta \) was calculated for the period when \( R_g > 0 \).

2. On page 239, line 4, insert "here" after "made".

3. On page 259, lines 22/23, D.B. means the moisture content of the soil calculated on the dry basis, i.e. mass of water / mass of dry soil.

20th September, 1983.
SUMMARY

This is a field and theoretical study of the evaporation component of the hydrology of an agricultural catchment. The aim of the study was to improve the evaporation section of the Australian Representative Basins Model, ARBM. ARBM is a physically based computer model of catchment hydrology. It was developed for use in conjunction with the Representative Basins Project. The catchment selected for this study was the Warrambine Creek Representative Basin, No. 2.15.

A major part of the study was the development of equipment to directly measure evaporation in the field. Two sets of equipment were developed. One system was based on the energy balance method of evaporation measurement. The other system was based on the eddy correlation method.

In its final form, the energy balance equipment proved most satisfactory. At the study site over 100 days of complete data was collected. However this data was discontinuous due to instrument and data recording faults. The data collected was used in the testing of various components of the evaporation model. A detailed analysis was undertaken to estimate the likely errors in the computer values of evaporation. This analysis showed that little error could be expected when evaporation was measured over a wet surface. For a dry surface, the analysis predicted that erratic results could occur. This prediction was confirmed with experimental evidence.

The development of the eddy correlation system was less successful. This occurred because of the late start to this section of the work and because of the innovative nature of the equipment which was developed. A microprocessor system was designed to replace the eddy correlation analogue computers. This system offers many advantages over conventional equipment. Unfortunately it was only developed to the laboratory testing stage. Also an infrared hygrometer was designed. This too was only tested in the laboratory. Although no field measurements of evapo-
ration were made with this equipment, additional theory and insight into this method was obtained.

The performance of any hydrological model depends on the reliability and representativeness of its input data. This data is supplied by a Bureau of Meteorology climate station situated only 3 km from the study site. The location of the climate station appears to be non-representative of the catchment in general. Thus climatic data from the two sites was compared. Little difference was noted in dry bulb temperature and rainfall data. Wet bulb depressions were slightly different. A poor correlation of pan evaporation was found. It was also found that the global radiation data collected at the climate station was worthless due to an instrument fault.

The evaporation model of ARBM was examined in detail. In particular, the parameter known as potential evaporation, PE, was examined from first principles. It was found that PE is poorly defined. PE was re-defined and prediction formulae were derived. These formulae were used in a new PE model for ARBM. The model which predicts the diurnal variation of evaporation was examined and found to be satisfactory.

Particular improvements to the equipment and topics of further study were recommended.
ACKNOWLEDGEMENTS

This study would not have been possible without the assistance, advice and support given by various people. All these people are sincerely thanked for giving so generously of their time; knowledge and experience.

There are several people who were particularly helpful. Mr. P. Hyson and other staff of the CSIRO, Division of Atmospheric Physics kindly lent various pieces of equipment and offered advice and guidance. Mr. R.J. Williamson provided assistance in understanding the computer model, ARBM, and provided detailed information about the catchment area.

The technical staff of the Department of Civil Engineering, in particular Mr. R.G. McIlroy and Mr. P. Jacobs assisted greatly in the design, construction and installation of field equipment. Much credit is due to the electronics staff, viz. Mr. D. Tuck, Mr. R. Nankevill and Mr. M. Cook. The ability and patience of these people is to be admired and is greatly appreciated.

I thank my supervisor, Dr. D.E. Angus, for introducing me to the subject of evaporation, conceiving this project and providing a friendly and encouraging environment in which to work. The staff and post-graduate students of the Agricultural Engineering Section all, at one time or another, provided physical or moral assistance. Their contributions are gratefully appreciated.

Miss Jacqui Wise and Mrs. Emma James are heartily thanked for their patience and devotion in typing this thesis. Most of all I thank my wife, Yvonne, for her sustained support throughout the whole study.
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CHAPTER ONE

1.0 INTRODUCTION

Water is Australia's most valuable resource. Without it, agriculture would decline and the economy falter. Urban and rural communities have an ongoing need for the design of dependable water supply systems which function soundly both in times of drought and flood. Engineers are thus being called upon continually to provide designs relating to the management of Australia's water resources.

However, sound engineering design is based on the premise that one can confidently predict the outcome of particular events and, as water resources are influenced by the erratic patterns of Australian weather, this is particularly difficult. The limited amount of reliable and complete data coupled with its stochastic nature continues to limit understanding of hydrological processes.

1.1 A.W.R.C. REPRESENTATIVE BASINS PROJECT

To help remedy this problem, the Australian Water Resources Council (A.W.R.C.) established in 1963 a network of "representative basins". These are well defined catchment areas which are hydrologically representative of surrounding areas and include nearly all the significant hydrological regimes in Australia. It is hoped that, if the hydrology of each of these small catchments can be understood, confident prediction of hydrological phenomena can be extended across the continent.

A representative basin is defined as "a catchment which contains within its boundaries a complex of landforms, geology, land use and vegetation which can be recognised in many other catchments of similar size throughout a particular region." (A.W.R.C. (1975)). It should be noted that this definition refers only to the physical and biological characteristics of the catchments. However, as the hydrology of an area depends on its climate, land form and geology,
and as these characteristics predetermine the biological elements, the above definition can be used successfully as a basis for basin selection.

As a further development of the representative basins project, a computer model has been developed to simulate the rainfall-runoff process. The Australian Representative Basin Model (ARBM) is not simply a statistical correlation between rainfall and runoff, but rather it is a deterministic or process model, i.e. it attempts to model physically each facet of the hydrological system. A.W.R.C. (1969) recommended the use of a deterministic mathematical model of the catchment processes to predict runoff because it is hoped that:

(i) it will lead to a better understanding of the various processes involved and their relative importance in the whole hydrological cycle;

(ii) by measuring the physical characteristics of a catchment, the model can be applied firstly to similar basins in the same representative region and secondly be extrapolated to basins with different combinations of geology, land form, soils and vegetation;

(iii) it will be possible to forecast possible hydrologic effects of changes in land use (and management) in any catchment.

To provide data to test this model, many of the representative basins have been instrumented. In general, this involves the establishment of stream gauging sites and meteorological stations. The amount of data collected so far is summarised below (Anon. (1980)).

- Rainfall at 6-minute intervals: 831.8 station years for 151 stations in 55 basins.
- Daily rainfall: 941.5 station years for 62 basins.
- Stream discharge: 206.2 station years for 32 basins.
- Evaporation: 370.6 station years for 54 stations in 42 basins.
However, the evaporation data listed above is pan evaporation, the usefulness of which will be discussed in detail later.

In fact, of the 93 basins only one or two have had any intensive evaporation studies conducted within them. This at first seems odd when one realises that up to 90% of the rain which falls in Australia will be re-evaporated and thus not contribute to runoff. (A.W.R.C. (1976)) However, the complexity of the equipment required to measure accurately actual evaporation rates has restricted the amount of long term research in this field.

1.2 THEESIS OBJECTIVES

It is then the aim of this thesis to measure evaporation from a typical area within a representative basin and to propose and test physically based prediction models. However, because this thesis contains aspects of both the "pure" science of micrometeorology and the "applied" science of hydrological engineering, it is necessary from the outset to state clearly the direction and emphasis of the thesis.

Primarily this thesis is about hydrological modelling. That is, the aim of the study is not necessarily to gain further insight into the evaporation process, but to apply the present theories concerning evaporation and, the equipment needed to measure it accurately, to a specific situation and in a realistic manner. This implies that, because the data available for use by the model, either as climatological input or fixed catchment parameters, is often limited in both quantity and quality, some of the subtleties and refinements of the present theories may not be relevant since their inclusion in the model may not be justified by any appreciable improvement in model performance. Model performance is measured not only by the "accuracy" of prediction, but also by the computing time (and thus cost) involved in achieving that result.
As a further note, it is apparent that some hydrologists have a poor understanding of the evaporation process. This is often limited to a belief in evaporation pans as the only reliable and accurate instrument for measuring evaporation and a vague notion of Penman's formula as a method of estimating evaporation. To help remedy this, a short section has been included which describes the various methods used to measure evaporation, listing their advantages and disadvantages. Also, deliberately, many references are cited. Some of these are particularly relevant; some are less relevant, but often quoted; whilst others are included for historical interest only. This collection is intended to act as a sound basis for any hydrologist interested in this field.

1.3 WARRAMBINE CREEK REPRESENTATIVE BASIN

1.3.1 Catchment Description

The area chosen for the study was the Warrambine Creek representative basin. (See Plate 1.1) It is a volcanic plainlands basin of approximately 55 km² situated approximately 40 km south of Ballarat, Victoria (≈ 37°50'S, 144°E), and has an elevation range of 180-400 m above sea level. The agriculture of the area is predominantly sheep farming with some cattle and occasional grain crops. The vegetation is described as temperate tussock grassland, converted largely to improved pasture (A.W.R.C. (1975)).

The Warrambine Creek basin is presently monitored by a climate station operated by the Bureau of Meteorology and by stream gauging stations and groundwater observation bores, all operated by the State Rivers and Water Supply Commission of Victoria (See Fig. 1.1). Williamson (1979) reported on a network of access tubes which has been established in the catchment so that soil moisture changes can be monitored using a neutron moisture meter. Thus, with the rainfall, runoff, groundwater, soil moisture and infiltration characteristics of the catchment investigated, the only significant parameter in the water balance of the catchment
Fig 1.1 CATCHMENT AND INSTRUMENTATION
yet to be studied is the evaporation loss. For a more
detailed description of the catchment, see Williamson and

PLATE 1.1 General view of the catchment (looking south
from Mt. Lawauk). The experimental site is
3 km away and is just visible to the left of
the long grove of trees in the middle distance.

1.3.2 Instrument Site Description

The location of the actual instrument site (See Fig. 1.1)
within the catchment needed some consideration. A.W.R.C.
(1970 b) noted that "evaporation from natural surfaces
shows considerable spatial variability, both on a local,
and on a regional basis (and) it is therefore of primary
importance that the location of sites for the determination
of evaporation, and the surface treatment of these sites,
should be representative of the areas for which the evapor-
ation estimates are to be applied". The site chosen satis-
fies this and several other selection criteria.
Firstly, it is representative of the vegetation, agriculture and geology of the catchment. It is open grassland consisting mainly of perennial rye grass (Lolium perene), barley grass (Hordeum hystric) and Yorkshire fog (Holcus lanatus). Some clover is present, but only in small quantities and no fertiliser has been applied for many years. The area is presently used, though not intensively, as pasture for sheep. These sheep were present throughout the whole study and, in order to maintain typical pasture conditions, no instruments were fenced off. This of course meant that all the equipment had to be "sheep-proofed". Unfortunately much valuable data was lost in the initial stages due to the inquisitive nature of the sheep (as well as mice, frogs, crows, galahs and insects). The second selection criteria satisfied by this site is ready access by road and the availability of mains (240V AC) power. Thirdly, the site is topologically representative of the catchment, i.e. open, rolling plains. Finally, it satisfies the fetch requirements of the evaporation measurement techniques chosen. This aspect will be discussed more fully later. A contour plan of the immediate area is shown in Fig. 1.2, and views of the surrounding area are given in Plates 1.2 to 1.5.

The soil at the site has been classed as a yellow sodic duplex soil of coarse structure with low permeability (A.W.R.C. (1975)). The soil is approximately 1.5 m deep with 200-500 mm of sandy clay loam overlaying a heavy clay subsoil. Three access tubes for the neutron moisture meter were installed in this paddock by Williamson and were still available for use.

1.3.3 Climate

Since climatic conditions directly influence the amount of evaporation from the catchment, this section has been included to illustrate in general terms, the area's climate and dominant weather systems. This information is useful when analysing and interpreting results obtained in the field and when proposing changes to the model.
Fig 1.2 Experimental Area
PLATE 1.2 The experimental site showing the instrument shed. This view was taken looking south from the farm access road.

PLATE 1.3 Mt. Lawauk. This view was taken looking north from the instrument shed. Note the same electricity pole as in Plate 1.2.
PLATE 1.4  Experimental site looking east from the instrument shed. Note the pluviometer and bulk rain gauge in the foreground and the presence of sheep.

PLATE 1.5  Experimental site looking west. This grove of trees surrounds the farm house and outbuildings and is the nearest windbreak (see Fig. 1.2 and Plate 1.1).
The area is described by A.W.R.C. (1975) as a medium uniform rainfall zone with a median annual rainfall of 560 mm and an average annual pan evaporation of 1090 mm. The rainfall is spread relatively evenly throughout the year with a maximum occurring in the winter-spring period. In any one year the rainfall can be variable, perhaps with drought periods or heavy storms during summer. A notable feature of the catchment is the decrease in the amount of rainfall from north to south; the northern (upper) end of the catchment has a typical yearly rainfall of 700 mm, (BM 89094, 89021) while at the stream gauging station (BM 89084) it is about 500 mm (Williamson & Turner (1980)).

Since no long term records of climate (apart from rainfall) are available from within the catchment, data for two nearby stations has been obtained to illustrate monthly climate parameters (See Table 1.1). Although approximately 45 km west of the catchment the climate of Lismore is likely to be similar to the lower elevation, lower rainfall southern section of the catchment. Durridwarrah (~30 km to the east) is probably typical of the northern end.

In order to illustrate the dominant weather patterns of the region, typical summer and winter synoptic charts are shown in Fig. 1.3. In winter, the weather is characterised by the continual passage of cold fronts across southern Australia. (See Fig. 1.3a) As a consequence, the winds predominantly come from the NW-SW sector, and are generally quite strong and cold. The passage of a cold front is typified by a drop in air temperature and often by brief showers of rain. These storms typically have an intensity of less than 4 mm/hr. In summer, large stable high pressure systems direct the cold fronts south of the continent. Under these conditions, winds are generally light and are not restricted to the NW-SW sector. Storm events are less frequent, but more intense.

A situation occurs periodically in spring and summer which is of particular interest to evaporation modelling. A high pressure system can direct hot, dry, gusty NW winds from
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<td>Min. Daily Temp °C</td>
<td>11.3</td>
<td>12.6</td>
<td>10.4</td>
<td>8.8</td>
<td>6.2</td>
<td>4.1</td>
<td>3.6</td>
<td>4.1</td>
<td>5.2</td>
<td>6.9</td>
<td>8.1</td>
<td>10.0</td>
<td>7.6</td>
</tr>
<tr>
<td>Rainfall mm</td>
<td>42</td>
<td>50</td>
<td>54</td>
<td>59</td>
<td>59</td>
<td>57</td>
<td>54</td>
<td>60</td>
<td>69</td>
<td>67</td>
<td>60</td>
<td>53</td>
<td>684</td>
</tr>
<tr>
<td>Rain Days No.</td>
<td>8</td>
<td>8</td>
<td>10</td>
<td>13</td>
<td>16</td>
<td>17</td>
<td>18</td>
<td>18</td>
<td>17</td>
<td>16</td>
<td>13</td>
<td>11</td>
<td>165</td>
</tr>
<tr>
<td>Dry Bulb 9am °C</td>
<td>17.7</td>
<td>18.6</td>
<td>16.0</td>
<td>13.4</td>
<td>9.7</td>
<td>7.0</td>
<td>6.6</td>
<td>7.8</td>
<td>9.9</td>
<td>12.7</td>
<td>13.9</td>
<td>16.3</td>
<td>12.5</td>
</tr>
<tr>
<td>Wet Bulb 9am °C</td>
<td>14.2</td>
<td>15.2</td>
<td>13.4</td>
<td>11.4</td>
<td>8.7</td>
<td>6.3</td>
<td>5.8</td>
<td>6.7</td>
<td>8.0</td>
<td>10.1</td>
<td>11.1</td>
<td>12.6</td>
<td>10.3</td>
</tr>
<tr>
<td>Humidity 9am %</td>
<td>66</td>
<td>68</td>
<td>73</td>
<td>78</td>
<td>87</td>
<td>90</td>
<td>88</td>
<td>85</td>
<td>76</td>
<td>70</td>
<td>69</td>
<td>63</td>
<td>76</td>
</tr>
</tbody>
</table>

Lismore | No. 089018 | Lat. 37°58'S | Long. 143°21'E | Elevation 159.7m
Duridwarrah | No. 087021 | Lat. 37°49'S | Long. 144°13'E | Elevation 365.8m
Warrambine Basin No.3 | No. 089094 | Lat. 37°50'S | Long. 143°53'E | Elevation 311.0m

Table 1.1 Regional Climatic Data (Mean & Total Data)
Fig 1.3a  Typical Winter Weather System

Fig 1.3b  Typical Summer Weather System

Fig 1.3  SYNOPTIC CHARTS FOR AUSTRALIA

* Catchment - see Fig 1.1
central Australia over SE Australia for some days at a time. (See Fig. 1.3b) This means that sensible heat is regionally advected into the area, thus raising daily maximum temperatures by up to \(10^\circ\text{C}\) and greatly increasing evaporation rates. This situation is illustrated with data presented in Chapter 2.

1.4 PROJECT PLANNING

An important decision to be made in the initial planning of the project is exactly what parameter should be measured and how to measure it. Over the last 50 years, many different methods of "evaporation" measurement have been developed and these methods vary greatly in the accuracy of the result obtained (and, consequently, the complexity of the equipment involved); the time scale resolution possible; and the dimensions of the area investigated.

Before deciding on which method to employ, it is necessary to discuss briefly how the A.W.R.C. model works and thus what data should be collected. The ARBM estimates, as part of its water budgeting routine, the amount of water lost by evaporation from various "water stores" (e.g. interception store, depression store, upper soil store). This is performed either as a daily total on days when no rain falls, or as the loss per part day (e.g. per hour) when either rain is falling or when water is present above ground.

To do this, the model first estimates the maximum amount of evaporation possible for that day. (This is called Potential Evaporation and will be discussed in detail in Chapter 5.) Any water present in the interception and/or depression store is evaporated at this rate. If the evaporative demand is not satisfied after these stores have been emptied, evaporation from the soil profile then occurs. This evaporation rate may be restricted by the current soil moisture content. Potential evaporation was estimated by Williamson (1979) from pan evaporation data collected at the climate station.
Thus the evaporation section of ARBM can be tested by comparing evaporation measurements made over intervals as short as one hour and to an accuracy of $\sim \pm 0.1$ mm with estimates of evaporation made by the model. Since the ARBM is a process model, a physically based model using data collected in the catchment should be used.

The following sections briefly describe the evaporation process and the various techniques used to measure evaporation from a natural surface, noting the parameters measured, the time scale resolution involved and the relative advantages and disadvantages in this particular application.

1.5 THE PROCESS OF EVAPORATION

Evaporation is the process of transferring liquid water from the catchment to vapour in the atmosphere. Water can be directly evaporated from the catchment surface (e.g. from grass recently moistened by rain or from water bodies) but more usually the water must move, in the liquid phase, from the soil profile to the plant's stomata or the soil surface before being evaporated. Thus, for evaporation to take place, three conditions must be met and it is on one or more of these conditions that each evaporation measurement technique is based.

(i) There must be a source of water for evaporation. This water may be stored above the ground as interception, on the ground as depression storage, or in the soil profile, and it can be evaporated directly or transpired by the vegetation.

(ii) Since evaporation involves a change of state, there must be a source of energy to provide latent heat of vaporization. The source of this energy may be radiation from the sun and sky and/or sensible heat transferred from the adjacent air and/or soil.

(iii) The atmosphere must have the ability to transfer water vapour away from the surface. This depends on the air
having suitable humidity and temperature gradients and
turbulent mixing characteristics to allow water vapour to
diffuse into it.

1.6 EVAPORATION MEASUREMENT TECHNIQUES

1.6.1 Water Balance Methods

The first factor mentioned above leads to the water balance
methods of evaporation measurement. These are all based on
a conservation of mass (water) type formula, of which a
simple form is:

\[ E = P - (Q_R + Q_D + \Delta Q_S)/A \]  \hspace{1cm} (1.1)

where
- \( E \) = evaporation (units of depth, e.g. m)
- \( P \) = precipitation (m)
- \( Q_R \) = net runoff out of area (m³)
- \( Q_D \) = net seepage/drainage out of area (m³)
- \( \Delta Q_S \) = change in moisture content of the soil (m³)
- \( A \) = defined investigation area (m²).

Evaporation is then determined over a suitable time inter-
val as the residual of many quantities which are either
measured or estimated.

For the Warrambine Creek study this technique could have
been applied to the whole catchment area as \( P, Q_R \) and \( A \)
have been carefully measured; \( \Delta Q_S \) has been investigated and
could be estimated with some confidence; and \( Q_D \) is, from all
indications, fairly minimal (<10 mm yr⁻¹, Williamson,(1979)).
However, this method does not give evaporation rates accur-
ately on a short time scale and could only be statistically
related to the data collected at the climate station.

1.6.2 Soil Moisture Depletion

In this method a much smaller area than above is defined in
which \( \Delta Q_S \) is carefully measured and \( Q_R \) is either measured or
restricted to zero. Usually \( Q_D \) is assumed to be zero and not
measured (thus possibly leading to substantial errors).
Soil moisture is often measured with a neutron probe and, in the present situation, reasonably good calibration data is available for the access holes installed. However, A.W.R.C. (1976) noted that this method has "an error of the order of 5 mm" and that evaporation rates "based on such data constitute average values, and their application in a short period deterministic model where the catchment response for individual events is sought, is questionable". Therefore this method was rejected but it may be useful in verifying the predictions made by the model. An example of the use of this method is the work of Calder (1976).

1.6.3 Lysimetry

A lysimeter is a block of soil with appropriate vegetation placed in a water-tight container and set into the soil at ground level. There are two main types of lysimeter: the drainage lysimeter and the weighing lysimeter. In both cases $Q_R$ is usually restricted to zero, $Q_D$ is collected at the base of the container and then measured. $P$ and $A$ are also measured. The difference between the two types lies in the determination of $\Delta Q_S$. In a drainage lysimeter, $\Delta Q_S$ is determined by tensiometers, gypsum blocks or a neutron meter. In the second case, the whole soil block is weighed extremely accurately, and evaporation is then recorded as a change in mass (and thus soil moisture content) of the block. The weighed lysimeter is more accurate, but much more expensive.

There are also several strict requirements for the design of lysimeters which must be met to ensure similarity between the lysimeter and its surroundings. These include thermal, roughness and drainage characteristics. Notwithstanding these, this system has been used very successfully (e.g. Pruitt & Angus (1960), Angus (1963), McIlroy & Angus (1964)), even to the extent of placing a 28 meter fir tree in a weighing lysimeter (Fritschet al (1973)). For a review of the use of lysimetry, see Harrold (1966) or Tanner (1967).
While this type of device can produce accurate measurements of evaporation on a daily or hourly time scale, it was not chosen for this study because of:

(i) the high initial cost of construction and installation (assuming the use of a weighing lysimeter)

(ii) the total lack of mobility for later work investigating different areas within the catchment

(iii) the problems involved in developing and maintaining a representative regime in the lysimeter

(iv) the problems of having stock and machinery on or near the lysimeter

(v) the possible difficulties of using a cracking soil in the lysimeter

(vi) the shallow bedrock (1.0 m to 1.5 m) which would have made installation extremely difficult.

1.6.4 Evaporimeters

Classed under the general name of evaporimeters are a number of devices designed by meteorological organisations. In all these instruments, the amount of water loss per day by the device is considered to be a measure of evaporation and, hence, these devices have been included here under the general heading of water balance techniques.

The most common device used is the evaporation pan or tank. In the past, various organisations designed their own versions (e.g. U.S. Class "A", Australian (Sunken) tank, "CGI 3000" Russian evaporimeter, etc. (Hounam (1966))); each varying in diameter, colour, construction material, depth and position with respect to the ground level. Consequently, it was found that evaporation from these devices was influenced not only by meteorological conditions, but also by pan characteristics. More recently, the U.S. Class "A" pan has been accepted as the standard, but all that can really be
said of this device is that it gives a measure of the free water evaporation from a 4’ diameter, 10" deep galvanized iron tank, the maintenance and siting of which varies from location to location. To illustrate this point, comparative data for two pans will be presented in Chapter 4.

Notwithstanding the artificiality of the device, significant correlation has been found in certain situations between pan evaporation and evaporation rates as determined by other methods (Campbell & Phene (1976), Dilley & Shepherd (1972), Mukammal & Neuman (1977)). As a consequence of this, evaporation pan measurements are often considered as representing a "lumped sum" of all the meteorological factors which influence evaporation (e.g. wind speed, humidity, solar radiation) and pan evaporation data are therefore commonly used in hydrology and agriculture.

Another approach has been to design a device which requires water to diffuse through some porous medium before it evaporates. Examples of these are the Piche, Bellani, Wilde and Livingston atmometers. Again, these have been used to measure or estimate actual evaporation (Stanhill (1961), Shannon (1968), Read (1968)). For a review of all the types of evaporimeters, see Hounam (1966) or Mukammal (1961).

In spite of their limitations, both a Class "A" pan and atmometers were installed at the Warrambine Creek site. The pan had been installed by Williamson three years earlier and observations were continued during this study while the atmometers (based on the work of Dilley & Helmond (1973)) were part of an associated, but separate, project.

1.6.5 Aerodynamic Formulae

The third factor mentioned in Section 1.5 has lead to the development of the "aerodynamic" type formulae. These consist of a hydrodynamic treatment of the process of turbulent diffusion; evaporation being represented as a turbulent transfer of matter by a particular form of the general flow law. Thus, evaporation (conveniently expressed as an energy
flux) can be expressed as:

$$LE = L \rho K_w \frac{\partial q}{\partial z}$$  \hspace{1cm} (1.2)

where \( \rho \) = density of air
\( L \) = latent heat of vaporisation of water
\( K_w \) = eddy diffusivity of water vapour
\( \frac{\partial q}{\partial z} \) = gradient of specific humidity.

Although \( \frac{\partial q}{\partial z} \) can be fairly readily measured, \( K_w \) which is a function of the turbulent mixing characteristics of the lower atmosphere, is not directly measureable. Hence it is necessary to either introduce an empirical formula of \( K_w \) or to derive it from other data.

Dalton, in the eighteenth century, was implicitly aware of this process when he proposed a formula for evaporation estimation. His formula, now called the "bulk aerodynamic" formula, is of the form:

$$LE = L \rho f(u) (q_2 - q_1)$$  \hspace{1cm} (1.3)

where \( f(u) \) = some wind speed function
\( q_2, q_1 \) = specific humidity at heights 1 and 2.

It has been found that this formula works very well for large bodies of water where the lower measurement height is taken as the water surface. Water temperature is measured and \( q_1 \) is then taken as the saturation value at the water temperature. Empirical wind functions have been determined for different measuring heights. For further information on this technique, see Webb (1960) and A.W.R.C. (1970a). This type of approach is not used over land surfaces, since each surface roughness requires the determination of a new wind function and the measurement of \( q_1 \) at the surface is extremely difficult.

Another procedure has been to assume \( K_w = f(K_M) \), where \( K_M \) is the eddy diffusivity of momentum or eddy viscosity.
$K_M$ can be determined from wind profile measurements (hence the name "profile analysis") and is a function of wind speed, surface roughness and atmospheric stability. A review of these types of formulae is given by Pierson & Jackman (1975), King (1966) and Webb (1965). These equations require accurate measurements of wind speed, humidity and temperature to be made at two heights in the atmosphere and are applicable only over extensive uniform surfaces.

This method, although not chosen as a measurement technique, is of interest as it is, in part, incorporated in one of the prediction formulae discussed in Chapter 5.

1.6.6 Energy Balance Techniques

The first factor mentioned in Section 1.5 has lead to the development of the "energy balance" technique of evaporation measurement. As evaporation represents an energy loss from the surface in the form of latent heat, evaporation can then be determined as the residual of all the other energy fluxes across that surface.

This technique was chosen for this study because:

(i) the method has been used extensively elsewhere and is a proven and reliable method of determining actual evaporation losses;

(ii) the sensors and recorders needed for this system were either readily available or easily constructed;

(iii) the system can be designed to operate unattended for periods of up to two weeks;

(iv) the equipment is readily transportable, hence enabling investigations to be carried out in other areas within the catchment, if required;

(v) many of the parameters measured in the energy balance method are also measured at the climate station, viz. solar radiation, temperature, humidity, and were thus useful
in verification of the model's performance and the reliability of the climate station's data.

The selection of this method as the primary evaporation measurement technique agrees with the findings of an A.W.R.C. Committee on the topic. (A.W.R.C. 1970b) However, they stated that "a level of accuracy of better than ± 0.01" (± 0.25 mm) can be obtained." Analysis presented in Section 2.3 demonstrates that this is not necessarily the case, particularly in arid environments.

Details of the theory behind this method, and the equipment used, are given in Chapter 2, along with details of the system's performance.

1.6.7 Eddy Correlation Techniques

The only evaporation measuring technique yet to be described is the eddy correlation method. In this system, the vertical vapour transport is measured directly in the atmosphere by sensing the turbulent fluctuations of both humidity and vertical wind and then calculating the mean of the instantaneous product of these parameters. This approach is theoretically the most fundamental method of evaporation measurement. However, it is also technically the most complex and its application to the routine measurement of evaporation has always been restricted by the lack of suitable sensors and eddy covariance computers. With the rapid advances occurring in the electronics industry today, new devices are constantly becoming available and the application of these to the eddy correlation technique can be investigated.

From the outset it was decided that the energy balance system would be the source of the long term evaporation measurements required to test the model. However, it was further decided that, once the energy balance system was operating satisfactorily, an eddy correlation system incorporating some new innovations should be designed and constructed and that this system would be used occasionally
to act as a comparison with the energy balance system. Details of the eddy correlation technique and the equipment used are given in Chapter 3.
Chapter 2
THE ENERGY BALANCE METHOD

2.0 INTRODUCTION

2.1 THEORY

2.1.1 Energy Balance of an Idealized Surface
2.1.2 Energy Balance of a Crop Volume
2.1.3 The Bowen Ratio
2.1.4 The $K_H = K_W$ Assumption

2.2 INSTRUMENTATION

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2.2.2 Net Radiation
2.2.3 Global Radiation
2.2.4 Soil Heat Flux
2.2.5 Temperature Measurements
2.2.6 Temperature Gradient Measurements
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2.3 ERROR ANALYSIS

2.3.1 Introduction
2.3.2 Definitions
2.3.3 Relative Error in LE
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2.3.5 Relative Error in $\beta$

2.4 SELECTED RESULTS

2.5 SUMMARY
CHAPTER TWO
THE ENERGY BALANCE EVAPORATION MEASURING SYSTEM

2.0 INTRODUCTION
This chapter includes a detailed discussion of the theory involved in the energy balance method of measuring evaporation. Details of the equipment used are presented and data collected on particular days are used to illustrate various aspects of the equipment's performance and aspects of the micrometeorology of the site. Data was collected during the spring-summer periods (i.e. October + February) of 1979-80 and 1980-81.

2.1 THEORY
2.1.1 Energy Balance of an Idealized Surface

![Energy Balance Diagram](image)

FIG. 2.1 Energy Balance of a Natural Surface

Consider the ideal case of a vanishingly thin layer right at a natural surface. Then, the energy fluxes across that surface (See Fig. 2.1) can be expressed as:

\[(R_n - G) - (H + LE) = 0\]

where:

(i) \( R_n \) is the net radiation flux. This is the net result of all incoming and outgoing radiative energy fluxes
and may be positive or negative. The various components of $R_n$ will be discussed in Chapter 5.

(ii) $G$ is the soil heat flux. This is the rate at which energy is transferred, usually by conduction, away from the surface into the soil profile. In general this is positive by day and negative by night.

(iii) $H$ is the sensible heat flux. This is the rate at which the surface loses energy due to convective cooling.

(iv) $LE$ is evaporation or the latent heat flux. This is the rate at which the surface loses energy in the form of latent heat via a net water vapour transfer.

2.1.2 Energy Balance of a Crop Volume

This ideal situation is never actually realised because a natural surface consists of a finite volume of soil and vegetation, thus making the precise definition of the surface nearly impossible and also introducing heat storage effects. The situation is further complicated because the sensors used cannot, for various reasons, be placed exactly at this surface. A more thorough examination of the energy balance of an evaporating crop is given by Fritschen (1966) as taken from Tanner (1963) or Tanner (1960). This is shown in Fig. 2.2.

![Energy Fluxes in a Crop](image)

FIG. 2.2 Energy Fluxes in a Crop
Consider all the energy fluxes into and out of a volume with unit dimensions in the horizontal plane and defined in the vertical plane by the instrument heights, \( z_1 \) and \( z_2 \), above and below the evaporating surface.

Then, the complete energy conservation equation for the defined volume is:

\[
R_n - G - H - LE - P + C_p \left[ \int_{0}^{z_1} \nabla (\rho_a u T) \, dz + \int_{0}^{z_1} \rho_a u q \, dz + \int_{0}^{z_1} C_c \rho_c \frac{3T_c}{3t} \, dz \right] + \int_{0}^{z_1} C_s \rho_s \frac{3T_s}{3t} \, dz + \int_{0}^{z_1} L \rho_a \frac{3q}{3t} \, dz = 0 \quad 2.2
\]

where  
\( P \) = photosynthesis  
\( u \) = horizontal wind speed  
\( T \) = temperature  
\( q \) = specific humidity  
\( C_p \) = specific heat of air (\( \equiv C_a \))  
\( C \) = heat capacity  
\( \rho \) = density  
\( L \) = latent heat of vapourisation  
\( z_1, z_2 \) = upper and lower instrument heights, note \( z_2 \) is negative  
\( \nabla \) = two dimensional partial derivative \( \left( \frac{\partial}{\partial x} + \frac{\partial}{\partial y} \right) \) of the horizontal wind, \( u \)

and \( R_n, G, H \) and \( LE \) are as defined in Equation 2.1

The subscripts denote  
\( c \) = crop  
\( s \) = soil  
\( a \) = air

**N.B.** For the dimensions of these and all other variables used, see Appendix A.
Hence, if evaporation is to be determined by an energy conservation approach, all the terms in Equation 2.2 (except LE) must be measured. Alternatively, the experiment should be designed so that certain of the terms in Equation 2.2 can be neglected. The following is a discussion of the relative importance of several of these terms.

(i) Photosynthesis, $P$. This term represents the energy used by the photosynthetic action of the crop. Estimates of this vary from $<5\%$ of $R_n$ (Slatyer & McIlroy (1961)) to $<2\%$ of $R_n$ (A.W.R.C. (1970(b))) for most crops. $P$ becomes zero at night and is not relevant in the case of bare soil or dead vegetation. Therefore it is generally ignored.

(ii) Energy storage effects. The last four terms in Equation 2.2 represent energy stored in the defined volume due to temperature changes in the air, crop and soil and due to humidity changes. All these terms are small, as will be seen below.

(a) For example, consider that in a one hour period, the temperature changes from $18^{\circ}C$ to $26^{\circ}C$ and that the relative humidity decreases from $80\%$ to $60\%$. (This is a larger change than commonly observed and can be regarded as the worst case). Then this represents an energy storage rate, $\Delta E$, in a 2 m deep defined volume of air of:

$$\Delta E = \int_0^z C_p \rho a \frac{\partial T}{\partial t} \, dz + \int_0^z L \rho a \frac{\partial g}{\partial z} \, dz$$

$$= C_p \rho a \frac{\Delta T}{\Delta t} z + L \rho a \frac{\Delta g}{\Delta t} z$$

$$= \frac{1.01 \times 1.18 \times 8 \times 2}{60 \times 60} + \frac{2450 \times 1.18 \times 0.002}{60 \times 60} \text{ KW m}^2$$

$$= 5.3 + 3.2 \text{ W/m}^2$$

$$= 8.5 \text{ W/m}^2$$
This rapid change in temperature could occur in mid-morning on a clear summer day when $R_n$ would be ~450 W/m$^2$. Thus this term represents ~2% of $R_n$ in the worst instance. Furthermore, this rate of change would only be sustained for a short period during the day.

(b) The heat stored in the soil between the surface and the level of the soil heat flux plates can be significant if these devices are placed deep in the soil, i.e. $z_2$ is large (Fuchs & Tanner (1968)). In this investigation, these plates were placed at ~15 mm (See Section 2.2.4), thus reducing this term as much as possible.

(c) The heat stored in the crop itself can be shown to be insignificant for short pasture (Angus (1963)); however, Dilley (1974) indicates that this term can reach 5% of net radiation in certain instances. (When dealing with forests, this term can be quite considerable and must be measured).

(iii) Advection. This is represented in Equation 2.2 by the terms

$$C_p \int_{0}^{z_1} v (\rho_a u T) \, dz + L \int_{0}^{z_1} v (\rho_a u q) \, dz$$

(The symbol, $v$, represents the two dimensional partial derivative $\left[ \frac{\partial}{\partial x} + \frac{\partial}{\partial y} \right]$ of the horizontal wind, $u$).

These terms represent the net horizontal transport of sensible and latent heat respectively into or out of the defined volume. This is known as advection and is defined as "the process of transport of an atmospheric property solely by the mass motion of the atmosphere" (Blad & Rosenberg (1974)).

Advection can occur because of two quite separate sets of circumstances. On the large scale, sensible heat can be advected into an area due to the large scale movement of hot, dry weather systems. This is called regional advection.
and, as has been already discussed in Section 1.3.3, this phenomenon occurs periodically in S.E. Australia. This has been observed during other evaporation studies in this region (McIlroy & Angus (1964)). However, this type of advection is not represented by Equation 2.3, since the net horizontal divergence of $C_p \rho u T$ and $L \rho u q$ is negligible over the experimental site. The occurrence of regional advection is revealed by the term $H$ in Equation 2.2 becoming negative, and evidence of this will be presented in Section 2.4.

On the small scale, advection can occur when wind blows across a surface which is discontinuous in temperature, humidity or roughness. The classic example of this is a limited area of irrigated (i.e. wet) land located in an arid region. This is often known as an "oasis". Sensible heat can then be advected from the surrounding hot, dry area to the cooler, wetter oasis. The effect of this local advection on evaporation measurements can be minimised by allowing the air to come into "equilibrium" with the evaporating surface before any measurements are made (See Fig. 2.3). That is, sufficient fetch must be provided upwind of any discontinuity so as to allow the air temperature and humidity to adjust to its surroundings before reaching the actual instrument site.

![Diagram](image-url)

**FIG. 2.3 Simplified Representation of Local Advection**
Argument exists as to the minimum length of this fetch. Some workers simply recommend an instrument height to fetch ratio, e.g. 1:100 (Slatyer & McIlroy (1961), Blad & Rosenberg (1974)), while others include the effect of windspeed as well. Fritschen (1966), using the results of Dyer (1963), suggests that, for an instrument height of 2 m, a fetch of 420 m is required. This is somewhat longer than the 200 m as suggested by the 1:100 ratio. Penman, Angus & van Bavel (1967) perhaps best summed up the situation by stating that "local rules based on local research may be the only solution".

In any event, local advection is not likely to occur at the Warrambine Creek site. As can be seen from Plates 1.2, 1.3 and 1.4, the experimental site is a relatively homogeneous area of pasture for several hundred metres in every direction and no irrigation is used in the area. The nearest major discontinuity is the farm buildings area with its surrounding trees (See Fig. 1.2 and Plate 1.5) which is over 450 m away from the instruments. A dirt road runs along the northern boundary of the property, but this is over 200 m away and should only constitute a minor discontinuity since it is only 5 m wide and constructed of dirt and gravel, not bitumen.

Thus, by ensuring that sufficient fetch is provided downwind of a discontinuity and by locating the various sensors as close as practicably possible to the evaporating surface (i.e. reducing the magnitude of $z_1$ and $z_2$), it can be shown that Equation 2.2 can be reduced without significant error to Equation 2.1, and this equation can then be used to determine evaporation.

2.1.3 The Bowen Ratio

In practice, $R_n$ and $G$ can be measured directly and this will be discussed in Section 2.2. However, $H$ is almost as difficult to measure directly as LE since both represent the turbulent diffusion of an entity (i.e. heat or water vapour) into the atmosphere.
Bowen (1926), when criticising Cummings (1925) for completely neglecting the sensible heat loss component, proposed what is now known as the Bowen Ratio, $\beta$. This is the ratio of the sensible heat flux to the latent heat flux, as defined by

\[ \beta = \frac{H}{LE} \tag{2.4} \]

and therefore Equation 2.1 becomes

\[ LE = \frac{R_n - G}{1 + \beta} \tag{2.5} \]

Bowen considered two extreme cases of laminar flow over a water surface and arrived at an expression for $\beta$ as

\[ \beta = 6 \times 10^{-4} \rho \frac{(T_o - T_a)}{(e^*_o - e_a)} \tag{2.6} \]

where $\rho$ = atmospheric pressure (mb)

$T_o$ = surface temperature ($^\circ$C)

$T_a$ = air temperature ($^\circ$C)

$e^*_o$ = saturation water vapour pressure at $T_o$ (mb)

$e_a$ = actual water vapour pressure of the air mass (mb)

However, $\beta$ can also be derived by considering the energy fluxes as turbulent diffusion processes, as in Section 1.6.5. Thus:

\[ H = C_p \rho K_H \frac{\partial T}{\partial z} \tag{2.7} \]

\[ LE = L \rho K_W \frac{\partial q}{\partial z} \tag{2.8} \]

where $K_H$ and $K_W$ are the eddy diffusivities of heat and water vapour respectively.

**N.B.** Strictly speaking, the potential temperature gradient should be used, i.e.

\[ \frac{\partial \theta}{\partial z} = \frac{\partial T}{\partial z} + \Gamma \]

where $\theta$ = potential temperature

$\Gamma$ = dry adiabatic lapse rate

$= 0.01^\circ$C/m
This small correction can generally be neglected for small height intervals. Hence,
\[
\beta = \frac{H}{LE} \frac{\partial T}{\partial z} = \frac{\rho C_p K_H \partial T/\partial z}{\rho L K_W \partial q/\partial z}
\]
and assuming that the profiles of \(q\) and \(T\) are of similar form, then
\[
\beta = \frac{C_p}{L} \frac{K_H}{K_W} \frac{\Delta T}{\Delta q}
\]
and by further assuming that \(K_H/K_W = 1\), this reduces to
\[
\beta = \frac{C_p}{L} \frac{\Delta T}{\Delta q}
\]
Furthermore, on introducing the expression relating \(q\) and \(e\), the vapour pressure (Harrison (1965(a))
\[
q = \frac{b e}{(p - (1-b) e)}
\]
where \(b = 0.62198\)
then \(q = \frac{b e}{p}\)
and, by substitution
\[
\beta = \frac{C_p}{b L} \frac{\Delta T}{\Delta e} = \gamma \frac{\Delta T}{\Delta e}
\]
If the vapour pressure gradient is to be determined using psychrometry, then using the psychrometric equation
\[
e = (e^*)_{Tw} - A p (T - Tw)
\]
where \((e^*)_{Tw}\) = saturated vapour pressure at the temperature, \(Tw\)
and \(A = \) accurately known psychrometric constant which very nearly equals \(\frac{C_p}{b L}\) (Slatyer & McIlroy (1961)) (See Appendix B)
Then:
\[
\Delta e = e_2 - e_1
\]
\[
= (e^*)_{T_{w_2}} - (e^*)_{T_{w_1}} - \frac{C_p b}{L} \left( (T_2 - T_{w_2}) - (T_1 - T_{w_1}) \right)
\]
\[
= s(T_{w_2} - T_{w_1}) - \frac{C_p b}{L} \left( (T_2 - T_1) - (T_{w_2} - T_{w_1}) \right)
\]
\[
= (s + \gamma) \Delta T_{w} - \gamma \Delta T
\]
\[
= \gamma \left( \frac{s + \gamma}{\Delta T_{w} - \Delta T} \right)
\]
\[
\beta = \frac{C_p b}{L} \Delta T \left( \gamma \left( \frac{s + \gamma}{\Delta T_{w} - \Delta T} \right) \right)^{-1}
\]
\[
\beta = \left( \frac{s + \gamma}{\Delta T_{w} - \Delta T} \right)^{-1}
\] 2.15

where:

(i) \( \gamma = \frac{C_p b}{L} \) 2.16

This "constant", gamma, is in fact a slowly varying function of pressure, temperature and humidity (Storr & den Hartog (1975)). Typical values for the parameters involved are given in Appendix B.

(ii) \( s = \frac{(e^*)_{T_{w_2}} - (e^*)_{T_{w_1}}}{T_{w_2} - T_{w_1}} \)

which can be very closely approximated to

\[
s = \left( \frac{3}{3 T_{w_{a}}} \right) (e^*)_{T_{w_{a}}}
\] 2.17

where:

\[
T_{w_{a}} = \left( \frac{T_{w_1} + T_{w_2}}{2} \right)
\]

Therefore, to a good approximation, s is the slope of the saturated vapour pressure vs. wet bulb temperature curve. This is a slowly varying function and can be determined
using empirical relationships, e.g. Dilley (1968), Lowe (1977). For details of these, see Appendix B.

Thus, evaporation, LE, can be determined by the simultaneous measurement of $R_n$, G, $\Delta T_w$, $\Delta T$ and Tw.

2.1.4 The $K_H = K_W$ Assumption

Before discussing the equipment used to measure evaporation, it is necessary to discuss the validity of the assumption made in the derivation of the Bowen ratio that $K_W = K_H$.

This assumption was implicit in Bowen's original derivation and has since been the subject of debate by many researchers. Apparently conflicting results have been obtained between different studies and this is due to three factors.

(i) Many studies were conducted over wet surfaces when $\beta$ is likely to be small. As will be shown in Section 2.3.1, if $\beta$ is much less than 1.0, then errors of 20% in the estimated value of $\beta$ give only small errors in the computed evaporation rate. Thus, the fact that LE as determined by a Bowen ratio-energy balance approach is nearly equal to LE as determined by some other method (e.g. lysimetry), does not necessarily verify the $K_H = K_W$ assumption.

(ii) Many experiments were conducted over a limited range of atmospheric stabilities and, as some evidence indicates that $K_H/K_W$ is a function of atmospheric stability (Campbell (1972)), then again erroneous conclusions (and extrapolations) have been made.

(iii) The reference measure of evaporation to which the Bowen ratio results are compared may itself be in error, or be non-representative. This has been suggested by several workers of the results of Pasquill (1949). Denmead & McIlroy (1970) went to some length to point out that there may have been shortcomings with their lysimeters, and Brost (1979) uses the same argument when criticising Verma et al (1978). For a summary of some of the published results on this topic, see Table 2.1. From Table 2.1, a clearer picture of the relationship between $K_H$ and $K_W$ is beginning to emerge. This is summarised below.
<table>
<thead>
<tr>
<th>AUTHOR</th>
<th>TECHNIQUE</th>
<th>SITE CONDITIONS</th>
<th>CONCLUSIONS</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pasquill (1949)</td>
<td>Bowen ratio: lysimeter</td>
<td>Pasture (turf)</td>
<td>$K_H = K_W$ stable conditions</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>$K_H &gt; K_W$ unstable conditions</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>$K_H = 2K_W$ at $R_l = -0.1$</td>
</tr>
<tr>
<td>Suomi &amp; Tanner (1958)</td>
<td>Bowen ratio: lysimeter LE$_L$</td>
<td>Alfalfa ($-3.2 &lt; R_l &lt; 1.8$)</td>
<td>$K_H = K_W$ Suomi &amp; Tanner's conclusion</td>
</tr>
<tr>
<td></td>
<td>Bowen ratio: lysimeter LE$_B$</td>
<td>Stubble ($-1.2 &lt; R_l &lt; 0.02$)</td>
<td>$LE_L/LE_B = 1.0$ for $R_l &gt; 0.01$</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>$LE_L/LE_B = 0.65-0.87$ for $R_l &lt; -1.1$ (Campbell's interpretation)</td>
</tr>
<tr>
<td>Pruitt &amp; Aston (1963)</td>
<td>Bowen ratio: lysimeter</td>
<td>n.a.</td>
<td>$K_H = K_W$ unstable conditions</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>$K_H/K_W = 1.2-1.3$ at $R_l = 0$</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>$K_H/K_W = 2-3$ very stable conditions</td>
</tr>
<tr>
<td>Dyer (1967) and Swinbank &amp; Dyer (1967)</td>
<td>Profile analysis</td>
<td>Freely evaporating pasture</td>
<td>$K_H = K_W$ for $-0.005 &lt; R_l &lt; -0.8$ (unstable) (limited range of $R_l$)</td>
</tr>
<tr>
<td>Webb (1970)</td>
<td>Analysis of profile data</td>
<td>Short grass</td>
<td>$K_H = K_W$ for $0 &lt; R_l &lt; 0.2$ (stable)</td>
</tr>
<tr>
<td>Oke (1970)</td>
<td>Profile analysis</td>
<td>Smooth, flat bare soil</td>
<td>$K_H = K_W$ for very stable conditions</td>
</tr>
<tr>
<td>Denmead &amp; McIlroy (1970)</td>
<td>Bowen ratio: lysimeter</td>
<td>Wheat, non-potential conditions</td>
<td>$K_H = K_W$ for $-0.001 &lt; R_l &lt; -0.026$ (unstable) (limited range of $R_l$)</td>
</tr>
<tr>
<td>Campbell (1972)</td>
<td>Synthesis of empirical relationships</td>
<td>n.a.</td>
<td>$K_H/K_W = 1$ for $0 &lt; R_l &lt; 0.5$</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>$K_H = K_W$ at $R_l = 0.5$</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>$K_H = K_W$ for $-2.5 &lt; R_l &lt; 0.025$</td>
</tr>
<tr>
<td>Blad &amp; Rosenberg (1974)</td>
<td>Bowen ratio: lysimeter</td>
<td>Soybeans, potential conditions</td>
<td>$K_H = K_W$ for non-advective conditions</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>$K_H/K_W = 1.2-1.5$ for sensible heat advection</td>
</tr>
<tr>
<td>Warhaft (1976)</td>
<td>Theoretical modelling of heat &amp; moisture transport</td>
<td>n.a.</td>
<td>$K_H = K_W$ when buoyancy effects are negligible</td>
</tr>
</tbody>
</table>
|                        |                   |                          | $K_H 
eq K_W$ when temperature gradient is opposite sign to humidity gradient  |
| Verma et al (1978)      | Bowen ratio: lysimeter | Alfalfa, soybeans, potential conditions | $K_H > K_W$ for sensible heat advection                                    |
| Motha et al (1979)      | Eddy correlation & profile analysis | Alfalfa | $K_H > K_W$ for regional sensible heat advection                           |
(i) When the stability is "within the range \(-2.5 < R_i < 0.025\), the assumption of equal diffusivities for heat and water vapour (i.e. \(K_H = K_W\)) leads to an error of less than 10% in estimating the Bowen ratio." (Campbell (1972)) This result has been confirmed by many studies, e.g. Dyer (1967), Denmead & McIlroy (1970), Blad & Rosenberg (1974). This stability range includes normal daytime lapse (unstable) conditions and the neutral stability conditions which occur during heavy overcast days. Both these stability ranges frequently occur at the Warrambine Creek site.

(ii) For inversion (stable) conditions, apparently conflicting results have been obtained. For example, Campbell (1972) states that \(K_H\) will be greater than \(K_W\) such that when \(R_i = 0.5\), errors of more than 10% in the estimation of \(\beta\) can occur." Pruitt & Aston (1963) results agree with this, but contrary findings (i.e. \(K_H = K_W\)) are reported by Webb (1970) and Oke (1970). The reason for this appears to be that, at night, all the energy fluxes are small and winds light, and thus large errors are likely to occur in the measured energy fluxes or profiles of wind, temperature and humidity. Fortunately, these conditions occur at night when the evaporation rate is low and hence any error in \(\beta\) causes very small errors in the daily totals of evaporation.

(iii) Perhaps more importantly for this study, "the use of the energy balance method in arid regions over very dry surfaces, where \(\beta\) will be large and positive and \(R_i < 2.5\), must be approached with caution." (Campbell (1972)) This conclusion has been drawn by extrapolating empirical relationships of \(K_H/K_W\) into stability ranges where experimental results do not presently exist. Campbell admits Dyer & Hicks (1970) caution against doing this, but, as for most of the summer period, the Warrambine Creek site is very dry (i.e. \(\beta\) is large and positive) and, presumably, strongly unstable conditions can occur, then the data collected at this time should be viewed with caution.

(iv) Under conditions of regional sensible heat advection, it appears that \(K_H/K_W = 1.2 - 1.5\) (Blad & Rosenberg
(1974)), Verma et al (1978), Motha et al (1979)). This result has since been disputed by Brost (1979) by using the theoretical analysis of Warhaft (1976). These conditions occur periodically at the Warrambine Creek site and the consequence of the above finding will be discussed in the example cited in Section 2.4.

2.2 INSTRUMENTATION

2.2.1 Summary

The following sections describe in detail the evaporation measuring system, based on the Bowen ratio concept, which was designed and constructed for this study. The main features of this system are summarised below.

(i) An eight channel data logger continuously integrates the analogue voltage outputs from the various sensors and, at 15 minute intervals, dumps these data on paper-tape.

(ii) The data recorded are:
  (a) Net radiation, $R_n$
  (b) Global radiation, $R_g$
  (c) Soil heat flux (2 sites), $G_1$, $G_2$
  (d) Dry bulb temperature, $T$
  (e) Wet bulb temperature, $T_w$
  (f) Dry bulb temperature difference, $\Delta T$
  (g) Wet bulb temperature difference, $\Delta T_w$.

(iii) A thermometer interchange system is used to greatly reduce errors in the determination of the temperature gradients.

(iv) The system can be left operating unattended for periods of up to two weeks.

2.2.2 Net Radiation

Net radiation represents the net radiative energy flux being emitted or absorbed by the vegetative surface and has both incoming and outgoing, and short-wave and long-wave compon-
ents. These various components will be discussed in detail in Chapter 5. This energy flux is measured with a device known as a net radiometer or net pyrradiometer. Basically this device consists of two aluminium plates, spray painted with an optical matt-black lacquer. One plate faces upwards and the other downwards. The radiation falling upon the surface of each plate is absorbed, causing each surface to be heated to a particular temperature. Since the sensor elements are very close to being radiation black bodies, the incident radiation flux, \( R \), is related to the temperature of the surface, \( T \), by the Stefan-Boltzmann law

\[
R = \sigma T^4
\]

where \( T \) is the temperature of the sensor face and \( \sigma \) is the Stefan-Boltzmann constant. The temperature difference between the two plates is measured by a set of thermopiles and the net radiation flux (i.e. \( R_{\text{down}} - R_{\text{up}} \)) is indicated by an analogue voltage which is proportional to the temperature difference of the two plates. For this study, a Swissteco (Type S-1) net radiometer, mounted 1.5 m above the evaporating surface, was used. For further details of this instrument, see Funk (1959) and Funk (1962).

Since it is necessary to protect each black surface from the convective cooling effects of the wind, each side must be covered by a membrane which is transparent to all the wavelengths of radiation involved. It has been found that a thin polythene hemisphere is suitable for this and these hemispheres were installed and inflated according to the manufacturer's recommendations. A dry nitrogen cylinder supplied the required pressure and low flow rate necessary to keep the domes adequately inflated. The hemispheres had to be replaced periodically due to the ravages of the weather and for a second and unexpected reason. An unknown species of bird attacked and destroyed the hemispheres, often scratching the blackened surfaces in the process. To overcome this, a bird guard was constructed. This bird guard was designed so as to cause the minimum disruption possible
to the radiative flux (See Plate 2.1). After its introduction, no further damage occurred. A small fan was also used to continuously blow air over the hemispheres so as to prevent the build up of dust or the condensation of dew.

While the mast (Plate 2.2) is quite substantial in construction (necessary due to the strong winds in the area and the continual presence of the sheep), its effect on modifying the radiative flux was minimised by:

(i) pointing the mast in a northerly direction so shadows from the mast would not be cast across the "sensed" area, and

(ii) painting the mast a matt dull green colour such that its reflectivity and long-wave emissivity would approximate (at least better than a glossy white mast) its natural surroundings.

An estimate of the fraction of the radiative flux obscured by the mast can be made using the formula cited by Angus (1966).

If there is a rectangular obstruction of width "b", projecting vertically for a distance "c" beyond the plane of the radiometer, and at a distance "a" from it (See Fig. 2.4), then the fraction of radiation, ΔR, which is obscured is:

\[ ΔR = \frac{6.5}{2π(110)} \left( \frac{150^2}{110^2 + 150^2} \right) \]

2.18

\[ ΔR = 0.6\% \]

Thus, the fraction of the radiative flux obscured by the mast is minimal and when considering that the mast itself will compensate by re-emitting radiation, it can be seen that the effect of the mast is negligible.
PLATE 2.1  Detail of net radiometer showing the inflated polythene hemispheres and bird guard.

PLATE 2.2  Radiation Mast. Note the net radiometer, solarimeter and dry nitrogen supply at the base of the mast. One set of soil heat flux plates was buried approximately 10 m from this mast, near the steel peg.
FIG. 2.4 Schematic Representation of an Exposed Radiometer

2.2.3 Global Radiation

Global radiation is the incoming short-wave component of net radiation. It was decided to measure this because

(i) the Bureau of Meteorology climate station, which is only 3 km away, also measures global radiation and a comparison with the data there would be useful (See Section 4.4.2), and

(ii) global radiation is the basic variable necessary for the estimation of net radiation (See Chapter 5) and the simultaneous measurement of $R_n$ and $R_g$ would therefore aid the proposal and testing of various models.

The instrument used is called a solarimeter or pyranometer. This device has one upward facing blackened surface and, in a similar way to the net radiometer, the incident radiative flux causes this surface to reach a black-body temperature as predicted by the Stefan-Boltzman law. This temperature is measured with a set of thermopiles and the incident
radiation is then indicated by an analogue voltage. As
glass is opaque to long-wave radiation but transparent to
short-wave, the blackened surface is covered by two con-
centric glass domes (See Plate 2.3). The inner dome is
used to reduce fluctuations in the amount of long-wave rad-
iation reaching the black plate due to temperature variations
of the outer dome. The instrument used was a Swissteco
(Type SS-5) pyranometer mounted 1.5 m above the surface on
the same mast as the net radiometer. For calibration data
of both the net radiometer and the pyranometer, see Table
2.2.

TABLE 2.2

Calibration Data for Radiation and Soil Heat Flux Instruments

<table>
<thead>
<tr>
<th>Instrument</th>
<th>Maker</th>
<th>Instr. No.</th>
<th>Sensitivity (mV/W/m²)</th>
<th>Accuracy</th>
<th>Date of Calibration</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pyranometer</td>
<td>Swissteco</td>
<td>7215</td>
<td>0.0268</td>
<td>±2%</td>
<td>17/1/77</td>
</tr>
<tr>
<td>Type SS-5</td>
<td></td>
<td>0.0255</td>
<td></td>
<td>&quot;</td>
<td>22/8/80</td>
</tr>
<tr>
<td>Net Radiometer</td>
<td>Swissteco</td>
<td>7226</td>
<td>0.0514 S-W*</td>
<td>±2½%</td>
<td>21/2/77</td>
</tr>
<tr>
<td>Type S-1</td>
<td></td>
<td>0.0517 L-W</td>
<td></td>
<td>&quot;</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.0497 S-W**</td>
<td></td>
<td>&quot;</td>
<td>15/4/79</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.0501 L-W</td>
<td></td>
<td>&quot;</td>
<td></td>
</tr>
<tr>
<td>Soil Heat</td>
<td>Middleton</td>
<td>419</td>
<td>0.0157 In***</td>
<td>5%</td>
<td>11/77</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>0.0157 Out</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>506</td>
<td>0.0180</td>
<td>&quot;</td>
<td>11/77</td>
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<td></td>
<td></td>
<td>0.0176</td>
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</tr>
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<td></td>
<td>511</td>
<td>0.0163</td>
<td>&quot;</td>
<td>11/77</td>
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<td></td>
<td></td>
<td>0.0165</td>
<td></td>
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</tr>
<tr>
<td></td>
<td></td>
<td>F99</td>
<td>0.0209</td>
<td>&quot;</td>
<td>11/77</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>0.0203</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

* Long-wave and short-wave calibration coefficients
** Instrument repainted and recalibrated after damage by birds
*** Calibration coefficients into and out of the numbered face.

N.B. All calibrations were done by the CSIRO Division of
Atmospheric Physics, Aspendale. This laboratory is
registered by the National Association of Testing
Authorities, Australia.
PLATE 2.3 Global radiometer. Note the results of birds perching on the mast.

2.2.4 Soil Heat Flux

Soil heat flux is the flow of heat, usually by conduction, into or out of the soil profile. This can be measured by two methods. Firstly, temperature gradients can be measured in the soil profile and the soil heat flux, $G$, can then be calculated from a knowledge of the soil heat capacity, $C_s'$, or the thermal conductivity of the soil, $k_s$. However, $C_s'$ and $k_s$ are both difficult to determine accurately and also change with soil moisture content.

The second method of soil heat flux measurement is the use of heat flux plates (Deacon (1950)). These are thin plates of approximately the same thermal conductivity as that of the soil which are buried horizontally in the soil. The heat flux through the plate sets up a measurable temperature difference between the upper and lower surfaces equal to the flux divided by the plate's thermal conductivity. This temperature difference is measured by a set of thermopiles in a similar manner to the operation of the net radiometer.
This method was used in this study and the sensors were kindly supplied and calibrated by the CSIRO Division of Atmospheric Physics, Aspendale, Victoria. Details of the calibration data are given in Table 2.2.

This is an attractively simple method of flux measurement, but is not without its problems. Some of these are discussed below.

(i) Care must be taken in the design of these plates to ensure that the thermal conductivity of the plates, \( k_p \), is similar to \( k_s \). If this is not the case, heat flux divergence or convergence can occur near the plate. (Philip (1961)).

(ii) The plates must be located as close as possible to the soil surface so as to minimize the effects of heat storage in the soil above the plates. However, when these plates are placed in the top few centimetres of soil, root growth can easily twist the plates out of a horizontal plane. In this case the plates measure \( G \cos \theta \) where \( \theta \) is the angle of tilt. The plates in this study were placed at \( \sim 15 \) mm below the soil-grass surface and were held in steel frames in order to minimize the effects of root growth (See Plates 2.4 and 2.5).

(iii) The presence of these impermeable obstructions in the soil profile can locally affect drainage and soil moisture content, thus locally modifying soil heat flux. Also, the plates cannot measure heat flux in the form of water or vapour movement.

(iv) Good thermal contact between the plates and the surrounding soil cannot always be ensured, especially when shrinking clays are being investigated.

In order to get some idea of the spatial variability of this parameter, \( G \) was measured at two sites and at each site two plates were used. Comparative results from the two sites are shown in Fig. 2.5a to Fig. 2.5f.
Fig 2.5a  SOIL HEAT FLUX DATA  20/10/1979
Diurnal Variation

\[ G_1 = -0.908 + 0.763 \times G_2 \quad r^2 = 0.975 \]

Half Hourly Averages

Fig 2.5b  SOIL HEAT FLUX DATA  1/1/1980
Diurnal Variation

\[ G_1 = -0.881 + 0.812xG_2 \quad r^2 = 0.972 \]

Half-Hourly Averages

Fig 2.5c  SOIL HEAT FLUX DATA  6/1/1980
Diurnal Variation

\[ G_1 = -4.957 + 0.734 \times G_2 \quad r^2 = 0.971 \]

Half Hourly Averages

Fig 2.5d  SOIL HEAT FLUX DATA  30/1/1980
Diurnal Variation

G1 = -0.696 + 0.572xG2  \( r^2 = 0.985 \)

Half Hourly Averages

Fig 2.5e  SOIL HEAT FLUX DATA  12/10/1980
Diurnal Variation

\[ G1 = -3.282 + 0.518xG2 \quad r^2 = 0.980 \]

Figure 2.5f  SOIL HEAT FLUX DATA   18/10/1980
In the 1979-80 season, the plates were installed in early spring when the grass was less than 50 mm tall. The two sites were approximately 20 m apart. Initially the data from the two sites correlated very well (See Fig. 2.5a), with G1 generally < 10% lower than G2. However, as the season progressed, this difference became greater, eventually reaching ~ 20%-30% (See Fig. 2.5b). Also, on some days, a marked hysteresis occurred, where G2 was considerably larger than G1 in the morning, but by the afternoon G2 equalled or was less than G1 (See Figs. 2.5c and 2.5d). Since there is very nearly a 1:1 correlation between G1 and G2 at night (negative fluxes), it seems unlikely that any of these differences can be attributed to calibration or instrument errors.

One possible explanation for the daytime difference could be that the vegetative cover over, and adjacent to, the plates grew to different heights and/or densities, thus effectively causing different net radiation loads on the soil surface immediately above the plates. This also could explain the hysteresis. If, for instance, there was a large clump of grass to the north-east of one set of plates, this would cause a shadow to fall across the sensed area in the morning which would be absent in the afternoon. This would in turn mean, that this sensor received proportionately less radiation in the morning than the afternoon when compared with the other "unshaded" sensor. This effect would only occur on cloudless days and the absence of the hysteresis on other days is consistent with overcast skies. An alternate explanation is that the grass over one set of plates grew more densely immediately above the plates. This would include more root growth. Thus the amount of heat storage immediately above the plates could change with respect to the other site. However, this theory does not explain the good night-time agreement, or the hysteresis.
Installation of soil heat flux plates (September 1980). A section of turf approximately 15 mm thick was cut and rolled back (Plate 2.4). The plates in their supporting frames were then installed and the grass was rolled back (Plate 2.5). The growth of grass over the plates was not noticeably different during the season.
In any case, it is impossible to tell whether one set of readings is "right" or "wrong", or whether these differences are real indicators of field variability. When the data were used to calculate evaporation rates during the 1979-80 season, the results from the lower reading set of plates (i.e. G1) were used. This was done only because, for a significant period of time, an intermittent fault was present in the cable connecting the second set of plates to the data logger and consequently much data was lost. Use of an averaged value of soil heat flux on those days when both sets of plates were working would add a small bias to that evaporation data.

In order to see if the differences between G1 and G2 could be reduced, the two sets of plates were placed immediately adjacent to each other for the 1980-81 season (See Plate 2.4). Surprisingly, the difference between the two was not reduced, but rather increased, such that G1 was only half of G2 during the day (See Figs. 2.5e and 2.5f). The reason for this is completely unknown, since the same pairs of sensors and system calibrations, as for the previous summer, were used. These results reflect the inability of a 40 mm by 60 mm plate to adequately sense the "average" soil heat flux in a field and therefore indicate that one sensor, alone, is inadequate and that for any future studies, the average flux from several plates should be measured. The relationship between soil heat flux and net radiation will be discussed in Chapter 5.

2.2.5 Temperature Measurements

All the temperature measurements (i.e. T, Tw, ΔT, ΔTw) were made on the same mast and using similar sensors. The following discussion on the design of the temperature circuits and the sensor housings applies equally well to the temperature difference circuits as to the absolute temperature circuits. However, certain aspects of the temperature difference system will be discussed in more detail in the next section.
All temperatures were measured using platinum resistance thermometers (nominal resistance 100Ω) placed in out-of-balance bridge circuits (See Fig. 2.6). The four complete bridge circuits (including separate reference voltage supplies) were placed in the one container which was mounted on the temperature gradient mast. An alternative method to this would have been to place the main components of the bridge circuits in the instrument shed, thus having only the sensors (connected via 50 m cables) mounted on the mast. This method was not used because of the possibility of temperature effects in the long arms of the bridge (i.e. the 50 m cables) introducing errors, particularly in the ΔT and ΔTw circuits. All the electronic components were chosen for their high temperature stability and the reference voltage was supplied by 1.35V mercury cells. The circuits were designed so that the current through each thermometer was low enough (< 2 mA) to ensure that there were no self-heating effects. In the Tw and T circuits, the 10K variable resistor (See Fig. 2.6) was used to set the temperature at which the bridge was balanced, thus effectively determining the range of measurement. In the ΔT and ΔTw circuits, this variable resistor was used to balance the bridge for a zero temperature difference.

Errors in the measurement of air temperature can be caused by the extraneous flow of heat by conduction or radiation to the sensing element. The conduction of heat to the sensors was reduced by mounting each sensor in a small plastic bracket. This plastic was chosen because of its low thermal conductivity. Extraneous radiation exchange between the thermometers and their surroundings was reduced by the use of two concentric radiation shields. The large outer shield (200 mm dia.) was made of aluminium and painted glossy white. Firstly, this prevented any direct sunlight (short-wave radiation) from reaching the inner shield. Secondly, because of its high reflectivity, it absorbed little of the radiation incident upon it and thus re-radiated only a small amount of long-wave radiation towards the inner shield. The inner shield (25 mm dia.) was chrome plated and thus reflected
Fig 2.6  BRIDGE CIRCUITS FOR TEMPERATURE MEASUREMENTS.
most of this long-wave radiation. Also, as the thermometers were force ventilated, the inner shields were further cooled convectively to the ambient air temperature, thus reducing the net radiative exchange between the sensor and its surroundings to nearly zero.

In the case of the wet bulb thermometer, further extraneous heat flow can occur when water is supplied to the wet bulb at a temperature different to the actual wet bulb temperature. To prevent this, a system similar to that described by Collins (1965) and Swinbank & Dyer (1967) was used. The water supply to that part of the wick which covers the sensing element is via a stainless steel tube which passes under an extension of the wick itself. Hence, the water is pre-cooled to the ambient wet bulb temperature before it reaches the active part of the wick.

Water was supplied to the wicks from a large water container (~2 weeks supply) via two smaller constant head devices. The essential components of these devices are the needle valve and float components of a normal motor vehicle carburettor. It has been found elsewhere that this type of device works unsatisfactorily because the floats and/or valves tend to jam, causing a flood or drought as the case may be. However, in this system the whole arrangement was shaken vigorously once every 15 minutes during the sensor interchange and prolonged jamming was prevented. Consequently the system worked extremely well. The supply rate to the wicks could be controlled by changing the head difference between the wick and the supply chamber by using the slotted mounting bracket provided. Through a process of trial and error, a head difference was found which ensured an adequate, but not excessive, flow of water under all conditions. The constant head devices could pivot on an axis in line with the thermometer elements and, with the help of pendulum-like weights, the whole arrangement would remain vertical during and after the sensor interchange, even though the sensor housing had been inverted. This is shown in Fig. 2.7 and Plate 2.6. Distilled water was used throughout and commonly
Fig 2.7  CONSTANT HEAD WATER SUPPLY - see Plate 2.6 also.
PLATE 2.6 Detail of the thermometer housing, radiation shield and constant head water supply

PLATE 2.7 Sensor interchange system
For detail of components, See Fig. 2.13
available facial tissue (Kleenex) was used as the wick material. Because of contamination by dust, these wicks were replaced on every visit to the site.

As is well known, the wet bulb temperature attained is a function of the wind speed past the wicks. Collins (1965) recommends that a wind speed of 4 m/s should be used to ensure that the full wet bulb depression is reached. This aspiration rate was achieved using a 250 W electric fan which sucked air, firstly past the dry bulb element, then the wet bulb element, and finally through the whole mast structure. A hot-wire anemometer was used to measure the aspiration rate and this was found to exceed the required 4 m/s. Although no attempt was made to accurately balance the flow rate in each arm of the mast, no error should be introduced in $\Delta T_w$, since in both cases the maximum wet bulb depression would be reached.

Unlike some other Bowen ratio systems, e.g. the Energy Partition Evaporation Recorder (EPER) (McIlroy (1971)), the mast designed for this study was not free to rotate like a wind vane and, hence, some consideration of which direction the mast should face was necessary. Since the predominant winds come from the west, pointing the mast in this direction would aid aspiration of the wet bulbs. However, this would also allow direct solar radiation to reach the inner radiation shield during late afternoon. This can be eliminated by facing the mast either north or south, but this greatly increases wind loads on the mast, particularly during sensor interchange. As a compromise, the mast was initially set facing the north-west, but, as will be shown in the next section, this still allowed the late afternoon sun to affect temperature measurements, particularly $\Delta T$ and $\Delta T_w$. Hence, the mast was left facing due north for the rest of the study.

The temperature bridges were calibrated by placing the sensors in a stirred water bath (two in the case of the temperature difference circuits). In both cases the water
temperature was measured with mercury-in-glass thermometers. For the absolute temperatures, a thermometer graduated to 0.1°C was used, while for the temperature differences a Beckmann type differential thermometer graduated to 0.01°C was used. The output of each bridge was measured with a digital voltmeter capable of resolving to ± 1 μV.

In order to test these calibrations, the system was periodically operated in the field with the water supply disconnected. Under these circumstances, Tw = T and ΔTw = ΔT. This was done to see if any errors were present (due, for example, to one sensor being affected by an extraneous radiation exchange). A plot of consecutive half hourly averages of T and Tw for a four day period is shown in Fig. 2.8. To better illustrate the magnitude and occurrence of the differences between T and Tw, Fig. 2.9 was plotted. From this, it can be seen that for 82% of cases, |T-Tw| < 0.1°C. This, however, does not preclude the possibility of a systematic error since both T and Tw could be equally affected. To test this, a third and independent measure of T is required. The only available data for this were the temperature data collected at the Bureau of Meteorology climate station some 3 km away. When the 0900 and 1500 dry bulb temperatures for both sites were plotted (See Fig. 4.3 and Section 4.3.1), no systematic differences were detected. Thus, from the above evidence, it can be reasonably assumed that the likely error in T is approximately ± 0.1°C and that the error is unlikely to exceed 0.5°C. The term, "likely" error described above is only an estimate of the error in the temperature measurement based on experience and Fig. 2.9. It does not purport to have any statistical validity since this can only be determined from measurements against some "temperature standard". The statistically derived error, if it was possible to obtain, is the "probable" error. When considering Tw, this error may be increased due to shortcomings in the aspiration, water supply or wicking, but the magnitude of this is not measurable. A discussion of the correlation between ΔT and ΔTw will follow the next section.
Fig 2.8  PLOT OF HALF HOUR AVERAGES OF DRY BULB AND WET BULB
( WICKS REMOVED )

166 Consecutive Half Hour Averages.
Max. Dry Bulb - 32.3°C
Min. Dry Bulb - 3.2°C

Tw = -0.10 + 1.007 T
r² = 1.000
Fig 2.9 FREQUENCY PLOT OF (T - Tw) for Data shown in Fig 2.8

\[ \bar{x} = 0.010 \]
\[ \sigma = 0.146 \]
2.2.6 **Temperature Gradients**

In the theoretical development of the Bowen ratio, no mention is made of any specific heights between which the temperature and humidity gradients should be measured. In fact, the only requirement is that $\Delta T$ and $\Delta Tw$ should be measured between the same two heights. However, because of possible advection effects, these gradients should be measured within the boundary layer developed over the evaporating surface. As has been shown in Section 2.1.2, a maximum instrument height of 2.0 m is acceptable at the Warrambine Creek site and, because of the presence of sheep, a minimum height of 1.0 m is required. If the temperature differences are to be measured between 1.0 m and 2.0 m, it becomes necessary to measure these differences ($< 1.0^\circ C$) very accurately. In fact, under certain circumstances even the slightest error in $\Delta T$ or $\Delta Tw$ can cause very large errors in the computed value of LE. This will be discussed in Section 2.3.3.

In order to achieve greater accuracy, the concept of a sensor interchange system has been used by various researchers (e.g. Angus (1963), McNeil & Shuttleworth (1975), Black & McNaughton (1971), Sargent & Tanner (1967)). In this system, the sensors are interchanged regularly between the two desired measuring heights and the average temperature difference over two time periods is calculated.

It is instructive to further examine this concept with respect to this particular study. Errors in the final computed values of $\Delta T$ and $\Delta Tw$ can occur due to two different effects. Firstly, the temperature of the sensing element may be different from the actual air temperature because for example, an extraneous radiation exchange exists between the sensor and its surroundings. Hence consider the situation in Fig. 2.10.
Consider that the actual air temperatures at 1.0 m and 2.0 m are $T_1$ and $T_2$ respectively and these remain constant over the 30 minute period. Furthermore, assume that element A is sensing a temperature, $T_A$, which is in error by $\delta T_A$ and similarly for element B. Then, for the first 15 minute period, the apparent temperature difference is

$$\Delta T' = T_A - T_B$$

$$= (T_2 + \delta T_A) - (T_1 + \delta T_B)$$

and, for the second 15 minute period after the sensors have interchanged,

$$\Delta T'' = T_B - T_A$$

$$= (T_2 + \delta T_B) - (T_1 + \delta T_A)$$

Averaging these apparent temperature differences over 30 minute yields
\[ \Delta T = \left( \Delta T' + \Delta T'' \right)/2 \]

\[ = \left( T_2 + \delta T_A - T_1 - \delta T_B + T_2 + \delta T_B - T_1 - \delta T_A \right)/2 \]

\[ \Delta T = T_2 - T_1 \]

Thus the apparent temperature difference equals the actual temperature difference and therefore systematic errors are eliminated by this procedure. In practice, because the actual temperature difference is fluctuating and because extraneous errors are unlikely to be exactly the same after the sensors have interchanged, some errors will persist in the final result. This is demonstrated in Fig. 2.11.

The second source of error is the possibility of incorrect calibration factors. The system requires two calibrations to be determined. Firstly, the output voltage, \( V \), of the bridges must be calibrated against the temperature difference. The form of this calibration is

\[ \Delta T = a + b \, V \]

As was noted in Section 2.2.5, these bridges can be balanced such that for \( \Delta T = 0 \), \( V = 0 \). That is, in theory, the calibration coefficient, "\( a \)" should be zero.

The second calibration involves the data logger. This device will be discussed in detail in the next section, but it is sufficient to say that the calibration is of the form

\[ V = c + d \, C \]

where \( V \) = average voltage input over 15 minute period

\( C \) = digital count difference as recorded on paper tape

\( c, d \) = calibration coefficients

Thus, the final calibration curve used for data reduction is

\[ \Delta T = a + b \left( c + d \, C \right) \]

\[ = \left( a + b \, c \right) + \left( b \, d \right) \, C \]

or

\[ C = \frac{-(a + b \, c)}{(b \, d)} + \frac{\Delta T}{(b \, d)} \]
Now, consider that the sensors measure the actual temperature difference with no error and that this difference, $\Delta T$, is constant over 30 minutes. Then, for the first 15 minutes, the output, $C'$, is given by

$$C' = \frac{-(a + b c)}{b d} + \frac{\Delta T}{b d}$$

and for the second time period, after the interchange, the sensors measure the same difference, but this is now negative. Thus

$$C'' = \frac{-(a + b c)}{b d} + \frac{(-\Delta T)}{b d}$$

Then, over 30 minutes, the average output is determined by

$$\bar{C} = (C' - C'')/2$$

$$= \frac{\Delta T}{b d}$$

or

$$\Delta T = (b d) \times \bar{C}$$

$$= \frac{b d}{2} (C' - C'')$$

Thus, the interchange system eliminates errors which can occur due to errors in the calibration coefficients "a" and "c". This result is important since the bridge may drift out of balance and thus "a" becomes non-zero and yet no error is introduced.

As mentioned above, the system was occasionally tested by operating the sensors with the water supply disconnected. The results for 240 consecutive half hour periods (i.e. five days) are plotted in Fig. 2.11. One half hour period represents the average $\Delta T$ before and after a sensor interchange. To more clearly show the differences between $\Delta T$ and $\Delta Tw$, these data were again plotted as a frequency diagram. These differences were plotted for the whole sample (Fig. 2.12b) and for the daylight period of each day only (defined by $R_n > 0$) since evaporation essentially only occurs during this period and it is during this period that extraneous radiation errors are likely to occur. From
Fig 2.11  PLOT OF HALF HOUR AVERAGES OF $\Delta T$ AND $\Delta T_w$

( WICKS REMOVED )
Fig 2.12a

Daytime Data

\[ R_n > 0 \]

\[ \bar{x} = 0.022 \]

\[ \sigma = 0.030 \]

Fig 2.12b

All Data

\[ \bar{x} = 0.016 \]

\[ \sigma = 0.026 \]

Fig 2.12  FREQUENCY PLOT OF \((\Delta T - \Delta T_w)\)

for Data shown in Fig 2.11
Fig. 2.12b, it can be seen, that 80% of the sample lies within the range $-0.01^\circ C \leq \Delta T - \Delta T_w \leq 0.02^\circ C$. The cause of the few times when $(\Delta T - \Delta T_w) > 0.05^\circ C$ needs explanation. A detailed examination of the data revealed that this large difference only occurred in late afternoon. Knowing that the mast was pointing to the NW during this test, it seems probable that the late afternoon sun affected the $\Delta T$ sensors. When the mast was later pointed due north, this error was largely eliminated.

Since there is no independent standard to which these measurements can be compared, it is impossible, as in the $T$ and $T_w$ measurements, to statistically define the probable error involved in these measurements. However, the following comments can be made. Firstly, it seems that the sensor interchange system largely eliminates systematic errors, as predicted. Secondly, knowing how well the $T$ and $T_w$ circuits performed, and knowing that all the circuits, sensors and housings are essentially the same, there is no reason to suspect that any systematic error should exist in the $\Delta T$ circuits when no similar error exists in the $T$ circuits. Hence, it is assumed that the likely error in the measurement of $\Delta T$ is approximately $\pm 0.02^\circ C$. When considering the error in the measurement of $\Delta T_w$, again aspiration and wicking problems are likely, so that the error here can only be guessed at.

The essential components of the sensor interchange system are shown in Fig. 2.13 and Plate 2.7. The control circuitry is shown in Fig. 2.14. The electric motor used was a commonly available automobile windscreen wiper motor. This can rotate in both directions, but contains a worm gear assembly which prevents any movement of the motor when the power is off. This motor was attached to the rotating arms via a bicycle chain and gears. On the rotating arms axle were mounted two cam-activated micro-switches. At the end of every 15 minute period, the data logger sent a pulse to the control circuitry which, via the logic circuit, closed a
Fig 2.13  SCHEMATIC DIAGRAM OF SENSOR INTERCHANGE SYSTEM
relay which started the motor. Once the arms had rotated through 180°, the cam closed one of the switches which consequently stopped the motor and reversed the polarity of the motor's voltage supply. Then, when the next time pulse occurred, the motor started, but rotated in the opposite direction until the second switch was closed; the motor was stopped and the polarity reversed again. Complete rotation takes ~ 3.5 seconds. This system worked very well over long periods of time and required little maintenance.

2.2.7 Data Logger

The data logger which was used to integrate and record the analogue signals from the various sensors was designed by Mr. R. H. Hill of the CSIRO Division of Atmospheric Physics and was constructed within the University. Essentially, it is a similar system to that described by Francey (1975), except that up-to-date components were used and the data are dumped onto punched papertape rather than magnetic tape. The logger outputs data in a binary format for each of eight input channels and a time counter once every 15 minutes.

The essential feature of this device which distinguishes it from many other types of data loggers is that rather than sampling and averaging "spot readings" of the data, this system continuously integrates each input over the whole 15 minute period. The operation of each data channel (See Fig. 2.15) is as follows. Firstly, the input signal is amplified to a suitable level by A. Then a constant bias voltage is added to this at B so as to compensate for negative inputs. The signal (plus bias) is then fed into C, which causes the 0.1 μF capacitor to begin charging. As the capacitor charges, the output of C increases until a critical level which causes the Schmitt trigger, D, to fire. This causes the capacitor to discharge and, simultaneously, a pulse is sent to the binary counters, E. The charge rate of the capacitor (and thus the count rate) is a linear
Fig 2.15  TYPICAL DATA LOGGER INTEGRATOR CHANNEL
function of the input voltage, i.e. \( V = c + d \cdot C \). On the initiation of the 15 minute pulse from the timer circuit, the clock counter is incremented by one and then the current values of the eight channel counters and the time counter are dumped to papertape. These counters are not reset after each output.

For the time counter and the first six channels, 11 bit counters are used, while for the last two channels (used for G1 and G2), 7 bit counters are used. When the calibration data for each channel and sensor are combined, it is possible to obtain the range and resolution (per count) of each channel. Typical values are given below.

<table>
<thead>
<tr>
<th>Channel</th>
<th>Resolution</th>
<th>Range</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \Delta T ) ( ^\circ C )</td>
<td>0.003</td>
<td>6.2</td>
</tr>
<tr>
<td>( \Delta T ) &quot;</td>
<td>0.003</td>
<td>6.2</td>
</tr>
<tr>
<td>Tw &quot;</td>
<td>0.02</td>
<td>42</td>
</tr>
<tr>
<td>T &quot;</td>
<td>0.02</td>
<td>42</td>
</tr>
<tr>
<td>( R_n ) W/m(^2)</td>
<td>0.8</td>
<td>1580</td>
</tr>
<tr>
<td>( R_g ) &quot;</td>
<td>1.6</td>
<td>3380</td>
</tr>
<tr>
<td>G1 &quot;</td>
<td>1.7</td>
<td>210</td>
</tr>
<tr>
<td>G2 &quot;</td>
<td>1.7</td>
<td>210</td>
</tr>
</tbody>
</table>

**FIGURE 2.16** Idealised output of \( \Delta T \) bridge over 30 minute period
Special mention must be made of the recording of the \( \Delta T \) and \( \Delta Tw \) signals. Consider that the temperature gradient, \( \Delta T \), remains constant over a 30 minute period and, at any time, the temperature gradient can be related to the output of the bridge circuit by

\[
\Delta T = a + b \, V
\]

and the count, \( C \), recorded on papertape can be related to the average voltage over the time period by

\[
\bar{V} = c + d \, C
\]

Now, after each interchange, the \( \Delta T \) sensor must come into equilibrium with their surroundings. Chart recordings taken in the field indicate that this response time is approximately 1.5 minutes. Thus, the average voltage, \( \bar{V} \), for the time period is 0.9V (See Fig. 2.16). Thus, for the first time period,

\[
C' = \frac{-c}{d} + \frac{\bar{V}}{d}
\]

\[
= \frac{-c}{d} + \frac{0.9V}{d}
\]

and, for the second 15 minutes,

\[
C'' = \frac{-c}{d} + \frac{0.9}{d} (-V)
\]

Therefore, the average count over 30 minutes is

\[
\bar{C} = \frac{C' - C''}{2}
\]

\[
= \frac{0.9V}{d}
\]

and, from the formula developed in Section 2.26, it can then be shown that

\[
\Delta T = 0.9 \, b \, V
\]

Thus, this system under-estimates \( \Delta T \) and \( \Delta Tw \) by approximately 10%. However, to calculate \( b \), the ratio of \( \Delta Tw \) to \( \Delta T \) is required and this is calculated by
\[
\frac{\Delta Tw}{\Delta T} = \frac{0.9 b'V'}{0.9 b V} = \frac{b'V'}{b V}
\]

where \( V' \) is the voltage from the \( \Delta Tw \) bridge and \( V \) is the voltage from the \( \Delta T \) bridge.

Thus, when the ratio (and hence \( \beta \)) is calculated, this error is eliminated. Of course, in the field it is unlikely that the response time of the two sensors will be exactly the same and the values of \( \Delta T \) and \( \Delta Tw \) will be constantly fluctuating. Hence, some error will remain.

To overcome this, the logger could be modified to halt the integration of \( \Delta T \) and \( \Delta Tw \) for approximately 2 minutes after each sensor interchange. Time constraints prohibited doing this for this study, but this modification should be considered for any future work.

2.3 ERROR ANALYSIS OF BOWEN RATIO METHOD

2.3.1 Introduction

Having described all the individual components of the energy balance system, it is now necessary to assess the system as a whole and to try to estimate the likely errors involved in the computed values of evaporation. As will be shown below, it is impossible to estimate the accuracy of \( LE \) from a knowledge of the equipment alone. This is because the likely errors are strongly dependent on the nature of the evaporating surface itself.

Previous attempts to assess the accuracy of the Bowen ratio method have generally centred on comparing the computed value of \( LE \) to an independent measure of evaporation. This was usually done using lysimeters (e.g. Fritschen (1965)). This was often done in situations where \( \beta \) was small (i.e. a freely evaporating surface) and, as will be shown below, good agreement between Bowen ratio evaporation rates and
lysimeter readings is expected even when large errors in \( \beta \) are present. Fritschen's conclusions that "the relative errors (in LE) were less than 5\%" is not based on any error analysis procedure and is not qualified by stating over what range of \( \beta \) this applies.

One of the few papers to discuss the errors produced in LE due to errors in the measured parameters is Fuchs & Tanner (1970). They used the method discussed below. It should be pointed out that, strictly speaking, it is not possible to calculate the errors in LE because it is not possible to measure or calculate the probable errors in the measured parameters (particularly \( \Delta T \) and \( \Delta T_w \)). Thus this section might be more correctly described as a sensitivity analysis of the Bowen ratio method following essentially the techniques discussed by McCuen (1973). He defines sensitivity as "the rate of change of one factor with respect to change in another factor", while the analysis below discusses the error produced in one parameter due to errors present in other parameters. In any case, the aim of both methods is the same. Both try to predict the error (change) in LE due to errors (changes) in the measured parameters.

2.3.2 Definitions
The following definitions and derivations are taken from Scarborough (1962).

The quantity, \( \delta X \), is defined as the absolute value of the error in the measurement of the parameter, \( X \), i.e. the difference between the true value of \( X \) and the measured value of \( X \). The relative error in \( X \) is then defined as \( \frac{\delta X}{X} \) and the percentage error in \( X \) is \( \delta X.100\% / X \). To estimate the absolute error in the calculated value of \( X \) when it is derived from other measured parameters, consider that \( X \) is given by

\[ X = f (U_1, U_2, \ldots, U_n) \] 2.19
then it follows that

$$X + \delta X = f (U_1 + \delta U_1, U_2 + \delta U_2, \ldots, U_n + \delta U_n)$$  \hspace{1cm} (2.20)

To find the absolute error in $X$, the right hand side is expanded by Taylor's theorem and, after some rearrangement and with second order terms neglected, it can be shown that

$$\delta X = \frac{\partial X}{\partial U_1} \cdot \delta U_1 + \frac{\partial X}{\partial U_2} \cdot \delta U_2 + \ldots + \frac{\partial X}{\partial U_n} \cdot \delta U_n$$  \hspace{1cm} (2.21)

This is essentially the same as McCuen's Eqn. 4, which he calls the linearized sensitivity equation.

By definition, it follows that the relative error in $X$ is

$$\frac{\delta X}{X} = \frac{\partial X}{\partial U_1} \cdot \frac{\delta U_1}{X} + \frac{\partial X}{\partial U_2} \cdot \frac{\delta U_2}{X} + \ldots + \frac{\partial X}{\partial U_n} \cdot \frac{\delta U_n}{X}$$  \hspace{1cm} (2.22)

2.3.3 The Relative Error in LE

From eqn. 2.5, the general energy balance equation

$$LE = \frac{R_n - G}{1 + \beta}$$  \hspace{1cm} (2.5)

it follows that

$$\frac{\partial LE}{\partial R_n} = \frac{1}{1 + \beta}$$  \hspace{1cm} (2.23)

$$\frac{\partial LE}{\partial G} = \frac{-1}{1 + \beta}$$  \hspace{1cm} (2.24)

$$\frac{\partial LE}{\partial \beta} = \frac{-(R_n - G)}{(1 + \beta)^2}$$  \hspace{1cm} (2.25)

and, by substitution in Eqn. 2.22,

$$\frac{\delta LE}{LE} = \frac{\delta R_n}{(1 + \beta) \cdot LE} + \frac{-\delta G}{(1 + \beta) \cdot LE} + \frac{-\delta \beta \cdot (R_n - G)}{(1 + \beta)^2 \cdot LE}$$

$$= \frac{\delta R_n - \delta G}{R_n - G} + \frac{-\delta \beta}{1 + \beta}$$
Since the errors in \( R_n \), \( G \) and \( \beta \) are just as likely to be negative as well as positive, then all the error terms must be expressed in the positive sign in order to be sure that the maximum error of the function is obtained.

Thus

\[
\frac{\delta \text{LE}}{\text{LE}} = \frac{\delta R_n}{R_n - G} + \frac{\delta G}{1 + \beta}
\]

2.26

The immediate consequence of this is that the net radiation and soil heat flux terms can be lumped into one term which is independent of the Bowen ratio. An estimate of this can readily be made. From Table 2.2

\[
\left| \frac{\delta R_n}{R_n} \right| = 0.025
\]

and

\[
\left| \frac{\delta G}{G} \right| = 0.05
\]

From the data presented in Section 2.2.4, it appears likely that the error in \( G \) will be much larger than the instrument error because of sampling problems and spatial variability. Thus, as a guess, it will be assumed that \( \left| \frac{\delta G}{G} \right| = 0.20 \).

Furthermore, as indicated by field data, it can be assumed that for daytime conditions

\[ G = 0.05 \times R_n. \]

Thus, by substituting these results into the first term of Eqn. 2.26

\[
\frac{\delta R_n + \delta G}{R_n - G} = \frac{0.025 \ R_n + 0.2 \ G}{R_n - G}
\]

\[ = \frac{0.025 \ R_n + 0.2 \times 0.05 \ R_n}{R_n - 0.05 \ R_n} \]

\[ = 0.037 \]

Thus, for this study, the contribution to the relative error in LE due to both the net radiation and soil heat
flux terms is approximately ±4%. This term can be regarded
as constant and is independent of the effects of the Bowen
ratio.

In order to illustrate the effects of different relative
errors in β on LE, consider the following.

\[
LE = \frac{R_n - G}{1 + \beta}
\]

2.5

and, by considering only errors due to β, then by defin-
ition

\[
LE + \delta LE = \frac{R_n - G}{1 + (\beta + \delta \beta)}
\]

Thus

\[
\delta LE = \frac{R_n - G}{1 + \beta + \delta \beta} - \frac{R_n - G}{1 + \beta}
\]

and

\[
\frac{\delta LE}{LE} = \frac{\frac{R_n - G}{1 + \beta + \delta \beta} - \frac{R_n - G}{1 + \beta}}{\frac{R_n - G}{1 + \beta}}
\]

\[
= \frac{1 + \beta}{1 + \beta + \delta \beta} - 1
\]

\[
= -\frac{\delta \beta}{1 + \beta + \delta \beta}
\]

Thus

\[
\frac{\delta LE}{LE} = \frac{\delta \beta}{1 + (\beta + \delta \beta)}
\]

2.27

This is a slightly different result to that obtained in
Eqn. 2.26. This is due to the omission of second order
terms in the derivation of Eqn. 2.22. As would be expected,
this omission produces very small differences between
Eqn. 2.26 and Eqn. 2.27 for small values of β and δβ.
However, when β, and more particularly δβ/β, become large,
Eqn. 2.26 underestimates the relative error in LE by
several percent. In order to illustrate Eqn. 2.27 more
clearly, Fig. 2.17 was plotted.
Fig 2.17  PLOT OF RELATIVE ERROR IN LE FOR DIFFERENT RELATIVE ERRORS IN $\beta$

Using Eqn. 2.27 and assuming no error in ($R_n - G$)
Several important conclusions can be drawn from Fig. 2.17.

(i) For freely evaporating surfaces, i.e. $-0.2 < \beta < 0.2$, it can be seen that errors of up to 30% in $\beta$ will produce errors of less than 5% in $\text{LE}$ (9% when the $R_n$ and $G$ terms are included). Hence when evaporation rates are high, so too is the relative accuracy of the computed value of $\text{LE}$, even if $\beta$ is poorly measured.

(ii) As $\beta$ tends towards $-1$, $\Delta\text{LE}/\text{LE}$ becomes infinite. This is to be expected since when $\beta = -1$, the energy balance formula (Eqn. 2.5) is indeterminate except when $R_n - G = 0$. This situation often occurs at sunrise and sunset when $\Delta T_w$ approaches zero and $R_n - G$ is small. However, this only occurs for a brief period each day and the error introduced in the daily total of $\text{LE}$ is quite small.

It is worthwhile to note that an apparent anomaly exists here between the relative error in $\text{LE}$ as estimated by Eqn. 2.26 or Eqn. 2.27 and the equation presented by Fuchs & Tanner (1970). Their equation is

$$\frac{\delta \text{LE}}{\text{LE}} = \frac{\delta R_n + \delta G}{|R_n - G|} + \frac{\delta \beta}{1 + |\beta|}$$

Why $|\beta|$ is used is unclear, but Eqn. 2.28 implies that the errors in $\text{LE}$ are the same when $\beta = -\beta$. Fig. 2.17 clearly shows that this is not the case. Furthermore, Eqn. 2.28 implies that when $\beta = -1$, a finite error in $\text{LE}$ can be calculated (assuming $(R_n - G) \neq 0$).

(iii) As the surface dries and $\beta$ increases, the relative error in $\text{LE}$ due to errors in $\beta$ becomes larger. If, for example, the accuracy of the measurement of $\text{LE}$ is required to be $\pm 10\%$, then, subtracting the error due to $R_n$ and $G$ (i.e. $\pm 4\%$), this requires the contribution due to $\beta$ to be approximately $\pm 6\%$. This then requires $\beta$ itself to be measured to $\pm 5\%$ (approximately). As will be shown in the following sections, this accuracy is extremely difficult to achieve under dry conditions when $\beta$ is to be determined by wet and dry bulb psychrometry.
2.3.4 The Absolute Error in LE

The above discussion has centred on the estimation of the relative error in LE, i.e. under certain circumstances and using certain equipment, evaporation can be estimated to ±x%. From a modelling point of view, it is usually more important to discuss the absolute error in LE, i.e. ±x mm/day. Consider again Eqn. 2.26 (incorporating the improved β component of Eqn. 2.27).

\[
\frac{\delta LE}{LE} = \frac{\delta R_n + \delta G}{R_n - G} + \frac{\delta \beta}{1 + (\beta + \delta \beta)} \quad 2.29
\]

Then, rearranging this and substituting in the error due to \(R_n\) and \(G\) (i.e. ±4%), it can be shown that using Eqn. 2.5

\[
\delta LE = \frac{0.04 (1 + \beta) + 1.04 \delta \beta}{(1 + \beta + \delta \beta)(1 + \beta)} \cdot R_n - G \quad 2.30
\]

Assuming that, for a particular day, the total \(R_n - G\) was equivalent to 6 mm of evaporation (i.e. 14.7 MJ/m²/day), then the absolute error in LE can be determined for different surface conditions (i.e. varying \(\beta\)) and this is shown in Fig. 2.18. Again, different relative errors in \(\beta\) are plotted. This graph has opposite implications to Fig. 2.17, viz. when \(\beta\) is small, the error in LE is also small. Rather, Fig. 2.18 implies that the greatest absolute errors in LE are likely to occur when \(\beta\) is small (and hence LE is large).

Summarising the above discussion, it can be seen that, for a constant relative error in \(\beta\), the largest absolute error in LE is likely to occur when \(\beta\) is small (even though in this situation \(\delta LE/LE\) is small). Similarly, the absolute error in LE is likely to be small when \(\beta\) is large simply because LE is itself small. In any event, the accuracy of the computed value of LE is strongly dependent on the accuracy of \(\beta\) and it is now necessary to estimate the likely errors in \(\beta\).
Fig 2.18 PLOT OF ABSOLUTE ERROR IN LE FOR DIFFERENT RELATIVE ERRORS IN $\beta$

$(R_n - G) = 6 \text{mm/day.}$

Using Eqn. 2.30
2.3.5 The Relative Error in $\beta$

In order to estimate the relative error in $\beta$ due to errors in the measured parameters which are used to calculate $\beta$, the same procedure as in Section 2.3.3 is used. As pointed out by Fuchs & Tanner (1970), errors due to a $\pm 0.1^\circ C$ error in $T_w$ cause a resulting error in $(s + \gamma)/\gamma$ which is an order of magnitude smaller than the errors due to the temperature gradient measurements. As can be seen from Table 2.3, even errors of $\pm 0.5^\circ C$ cause errors in $(s + \gamma)/\gamma$ of less than 2%. Thus errors in this term are neglected in the following discussion.

<table>
<thead>
<tr>
<th>$T_w$ $^\circ C$</th>
<th>$\delta T_w$ $^\circ C$</th>
<th>$\frac{\delta s}{s} %$</th>
<th>$\frac{\delta K}{K} %$</th>
</tr>
</thead>
<tbody>
<tr>
<td>10</td>
<td>0.1</td>
<td>0.6</td>
<td>0.3</td>
</tr>
<tr>
<td></td>
<td>0.2</td>
<td>1.2</td>
<td>0.7</td>
</tr>
<tr>
<td></td>
<td>0.5</td>
<td>3.0</td>
<td>1.7</td>
</tr>
<tr>
<td></td>
<td>1.0</td>
<td>6.0</td>
<td>3.4</td>
</tr>
<tr>
<td>15</td>
<td>0.1</td>
<td>0.6</td>
<td>0.4</td>
</tr>
<tr>
<td></td>
<td>0.2</td>
<td>1.1</td>
<td>0.7</td>
</tr>
<tr>
<td></td>
<td>0.5</td>
<td>2.9</td>
<td>1.8</td>
</tr>
<tr>
<td></td>
<td>1.0</td>
<td>6.0</td>
<td>3.6</td>
</tr>
<tr>
<td>20</td>
<td>0.1</td>
<td>0.5</td>
<td>0.4</td>
</tr>
<tr>
<td></td>
<td>0.2</td>
<td>1.1</td>
<td>0.7</td>
</tr>
<tr>
<td></td>
<td>0.5</td>
<td>2.7</td>
<td>1.9</td>
</tr>
<tr>
<td></td>
<td>1.0</td>
<td>5.5</td>
<td>3.8</td>
</tr>
<tr>
<td>25</td>
<td>0.1</td>
<td>0.5</td>
<td>0.4</td>
</tr>
<tr>
<td></td>
<td>0.2</td>
<td>1.0</td>
<td>0.8</td>
</tr>
<tr>
<td></td>
<td>0.5</td>
<td>2.6</td>
<td>1.9</td>
</tr>
<tr>
<td></td>
<td>1.0</td>
<td>5.3</td>
<td>3.9</td>
</tr>
</tbody>
</table>

**TABLE 2.3**

Errors in $\frac{s + \gamma}{\gamma}$ due to errors in $T_w$

(i) $s$ is calculated by the method of Dilley (1968); See Appendix B

(ii) $\frac{\delta s}{s} = \frac{s(T_w + \delta T_w) - s(T_w)}{s(T_w)} \times 100\%$

(iii) $K = \frac{s + \gamma}{\gamma}$; $\frac{\delta K}{K}$ as in (ii)

(iv) $\gamma = 0.66 \text{ mb } ^\circ C^{-1}$
From Section 2.1.3

\[ \beta = \left( \frac{S + \gamma}{\gamma} \Delta Tw \frac{\Delta T}{\Delta T} - 1 \right)^{-1} \]

Thus

\[ \frac{\partial \beta}{\gamma \Delta Tw} = -\left( \frac{S + \gamma}{\gamma} \Delta Tw \frac{\Delta T}{\Delta T} - 1 \right)^{-2} \frac{S + \gamma}{\gamma} \frac{1}{\Delta T} \]

\[ \frac{\partial \beta}{\Delta T} = \left( \frac{S + \gamma}{\gamma} \Delta Tw \frac{\Delta T}{\Delta T} - 1 \right)^{-2} \frac{S + \gamma}{\gamma} \frac{\Delta Tw}{\Delta T^2} \]

and by substitution into Eqn. 2.22

\[ \frac{\partial \beta}{\beta} = \left( \frac{S + \gamma}{\gamma} \Delta Tw \frac{\Delta T}{\Delta T} - 1 \right)^{-2} \frac{S + \gamma}{\gamma} \frac{\delta \Delta Tw}{\Delta T^2} \]

\[ + \left( \frac{S + \gamma}{\gamma} \Delta Tw \frac{\Delta T}{\Delta T} - 1 \right)^{-2} \frac{S + \gamma}{\gamma} \frac{\Delta Tw}{\Delta T^2} \frac{\delta \Delta T}{\Delta T^2} \beta \]

\[ = \frac{S + \gamma}{\gamma} \beta \left( \frac{\delta \Delta Tw}{\Delta T} + \frac{\Delta Tw \delta \Delta T}{\Delta T^2} \right) \]

\[ = (1 + \beta) \left( \frac{\delta \Delta Tw}{\Delta T \Delta Tw} + \frac{\delta \Delta T}{\Delta T} \right) \]

This is essentially the same result as Fuchs & Tanner (1970), except they again use \(|\beta|\) instead of \(\beta\).

In order to visualise this result, Fig. 2.19 has been plotted. As there are an infinite number of combinations of \(\Delta T, \Delta Tw, \delta \Delta T, \delta \Delta Tw\) and \(Tw\), this situation has been simplified by assuming \(Tw = 10^\circ C, \Delta T = 0.5^\circ C\) (common daytime value at Warrambine Creek) and \(\delta \Delta T = 0.02^\circ C\) (from Section 2.2.6). By doing this, the \(\Delta Tw\) axis also defines the absolute value of \(\beta\). While realising that this is only one of many possible scenarios, some general observations can be made.

(i) When \(\Delta Tw\) is large with respect to \(\Delta T\) and, thus, \(\beta\) is small, large errors in the temperature gradients (+0.05°C) cause only small errors in \(\beta\) (≈ ±10%) and, from Fig. 2.17, the relative error in LE due to \(\Delta T\) and \(\Delta Tw\) is thus less than 2%.
Fig 2.19  PLOT OF RELATIVE ERROR IN $\beta$ FOR DIFFERENT ERRORS IN $\Delta Tw$
(ii) The error in $\beta$ becomes infinite at a certain value of $\Delta Tw$. This occurs when $\frac{S + \gamma}{\gamma} \frac{\Delta Tw}{\Delta T} = 1$ and thus $\beta$ itself is infinite. This implies that, as the surface dries and $\beta$ increases, this method of determining $\beta$ becomes unstable and even the smallest errors in $\Delta T$ or $\Delta Tw$ cause gross errors in $\beta$. For example, if the error in $\Delta Tw$ was $\pm 0.02^\circ C$, then when $\beta = 1$, $\frac{\Delta \beta}{\beta} = \pm 17\%$ and when $\beta = 2$, $\frac{\Delta \beta}{\beta} = \pm 30\%$.

Consider the situation when $\beta = 10$ (i.e. dry conditions). How accurately must $\Delta T$ and $\Delta Tw$ be measured to ensure that the error in $\beta$ is $\pm 10\%$ (assume that $\Delta T = \delta \Delta Tw$)?

By substituting into Eqn. 2.31 and assuming $\Delta T = 0.5^\circ C$ and $\Delta Tw = 0.2^\circ C$ (taken from typical field data), then

$$0.1 = (1 + 10) \left( \frac{\Delta T}{0.5} + \frac{\Delta Tw}{0.2} \right)$$

$$\delta \Delta T = \delta \Delta Tw = \pm 0.0013^\circ C$$

This accuracy is an order of magnitude greater than could possibly be expected from the equipment at the Warrambine Creek site. In fact, it is doubtful that any equipment could achieve this degree of accuracy and the only way of improving the situation would be to substantially increase $\Delta T$ and $\Delta Tw$. This could be done by sampling over a larger height interval, but the fetch requirements previously mentioned may prohibit this (See Section 2.1.2). It is therefore evident that the Bowen ratio method using wet and dry bulb psychrometry cannot accurately determine evaporation rates under very dry conditions.

The Bowen ratio method therefore apparently fails to meet the requirements of an evaporation measuring system for this project. However, this method only becomes highly inaccurate when $\beta$ is large (i.e. $>10$). That is, $LE < 0.1 (R_n - G)$ and if $(R_n - G) = 17 \text{ MJ/m}^2/\text{day}$, then $LE$ is 0.7 mm/day. Thus, under these conditions $LE$ can even be guessed with sufficient absolute accuracy. An example of this is given in the next section.
To summarise this whole section on errors, Fig. 2.20 and Fig. 2.21 have been plotted. These graphs are intended to illustrate the likely maximum errors in the measurements of LE under dry and wet conditions. The hypothetical situations chosen are again only a few of the infinite number of combinations of parameters, but are typical of the conditions in the Warrambine Creek basin. The equations used to plot these are Eqns. 2.5, 2.15, 2.29 and 2.31 and the parameters chosen were

<table>
<thead>
<tr>
<th>Wet (Spring)</th>
<th>Dry (Summer)</th>
</tr>
</thead>
<tbody>
<tr>
<td>(Rₙ - G) mm/day</td>
<td>5</td>
</tr>
<tr>
<td>MJ/m/day</td>
<td>12.3</td>
</tr>
<tr>
<td>δRₙ + δG</td>
<td>0.04</td>
</tr>
<tr>
<td>Rₙ - G</td>
<td></td>
</tr>
<tr>
<td>Tw °C</td>
<td>15</td>
</tr>
<tr>
<td>ΔT °C</td>
<td>0.2, 0.5</td>
</tr>
<tr>
<td>δΔT °C</td>
<td>0.02</td>
</tr>
<tr>
<td>δΔTw °C</td>
<td>0.05</td>
</tr>
<tr>
<td>ΔTw °C</td>
<td>Defined by β</td>
</tr>
</tbody>
</table>

From Fig. 2.20 and Fig. 2.21, it can be seen that the maximum error in the computed values of LE will, in most cases, not exceed ±0.6 mm/day. This error seems large but it must be remembered that these figures are intended to illustrate the maximum errors which may occur. There is no reason to suspect that the errors in the variables (particularly ΔT and ΔTw) are not random. Consequently, the errors in LE are similarly random. It is then reasonable to assume that over a period of days or weeks, the error in the cumulative total of LE will be small. From this, it can be concluded that the data collected by the equipment described in this chapter will be adequate for the testing of the A.W.R.C. digital process model (ARBM).

If sufficient data were available it would be possible to quantify the random nature of the errors in LE. Scarborough (1962) defines the probable error, R, in X as
Fig 2.20  ABSOLUTE ERROR IN LE (Summer)

Dry Summer Conditions
\( \beta > 1 \)
\( R_n - G = 7 \text{ mm/day} \)
\( \Delta T_w = 0.05^\circ \text{C} \)

Fig 2.21  ABSOLUTE ERROR IN LE (Spring)

Wet Spring Conditions
\( \beta < 1 \)
\( R_n - G = 5 \text{ mm/day} \)
\( \Delta T_w = 0.05^\circ \text{C} \)
\[ R = \left[ \left( \frac{\partial X}{\partial U_1} \right)^2 r_1^2 + \left( \frac{\partial X}{\partial U_2} \right)^2 r_2^2 + \ldots + \left( \frac{\partial X}{\partial U_n} \right)^2 r_n^2 \right]^{\frac{1}{2}} \]  \quad 2.32

where \( r_i \) = the probable error in \( U_i \)

= 0.6745 \sigma \quad 2.33

where \( \sigma \) = standard deviation of the measured errors in \( U_i \) i.e. \( \delta U_i \)

Scarborough (1962) then defines the probable relative error in \( X \) as

\[ \frac{R}{X} = \left[ \left( \frac{\partial X}{\partial U_1} \right)^2 \frac{r_1^2}{X^2} + \left( \frac{\partial X}{\partial U_2} \right)^2 \frac{r_2^2}{X^2} + \ldots + \left( \frac{\partial X}{\partial U_n} \right)^2 \frac{r_n^2}{X^2} \right]^{\frac{1}{2}} \]  \quad 2.34

This represents a root-sum-square error and it expresses the law of propagation of errors. Rather predictably, this formula produces errors substantially smaller than those calculated by Eqn. 2.22. This method was used by Blad & Rosenberg (1974) and Sinclair et al (1975) to calculate the errors in their Bowen ratio data. This approach was not used for two reasons.

(i) The analysis in this thesis is not intended to accurately predict the error in the computed value of LE (since this is impossible), but rather it is used to illustrate the maximum errors which may occur. That is, this is essentially a type of sensitivity analysis. It is used to illustrate which parameters need to be measured most carefully and which circumstances are likely to cause the greatest errors.

(ii) It is impossible to measure the error (or the probable error) in \( \Delta T \) and \( \Delta Tw \). The data presented in Fig. 2.12 clearly shows that even if the \( \Delta T \) and \( \Delta Tw \) sensors are calibrated in the laboratory to \( \pm 0.01^\circ C \), it does not necessarily follow that this accuracy can be reproduced in the field. Collins (1965) agrees with the conclusion. Sinclair et al (1975) make the statement that "for our
instrumentation, the value of $\delta T$... for differential measurements was taken as $\pm 0.01^\circ C$ but this is not substantiated by any data. The use of this value together with Eqn. 2.34 produces estimates of probable errors in LE. These are considerably smaller than those found in the above sections. It is a debatable point whether Sinclair's analysis underestimates the errors which are present or whether the analysis in this thesis overstates the importance of the errors.

One final point needs to be made. This concerns the use of psychrometry (i.e. wet and dry bulb thermometers) to determine the Bowen ratio. In the analysis above, it has been assumed that no significant errors exist in the determination of $s$ and $\gamma$. This is not always the case. Revfeim & Jordan (1976) point out that failure to take account of the pressure and temperature dependence of the psychrometric constant can cause large errors if the wet bulb depression is large. A detailed analysis of this is beyond the scope of this thesis.

2.4 SELECTED RESULTS

This section gives details of the data collected by the Bowen ratio system on individual days during the 1979-80 season. These days were selected because they illustrate certain aspects of the micrometeorology of the site, the performance of the equipment and the errors which may occur. For each day, diurnal plots of the various energy fluxes, the Bowen ratio and air temperature and humidity are included. Tabulated results are given for each day and a summary of the Bureau of Meteorology climate station data with the corresponding University of Melbourne data is given in Table 2.4. (For details of the climate station, see Chapter 4.) The daily totals of net radiation and evaporation given in Table 2.4 have been calculated for the period when net radiation is positive. The reason for this will be given below.

To calculate the estimated errors in evaporation, the equations developed in Section 2.3 have been used. It has
### Bureau of Meteorology

<table>
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<th></th>
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<th>4.11.79</th>
<th>8.11.79</th>
<th>10.12.79</th>
<th>11.12.79</th>
<th>1.1.80</th>
<th>10.1.80</th>
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<tr>
<td>Temperature (°C) Max.</td>
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<td>26.0</td>
<td>19.0</td>
<td>23.5</td>
<td>19.5</td>
<td>21.5</td>
<td>27.5</td>
</tr>
<tr>
<td>Min.</td>
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<td>3.5</td>
<td>6.5</td>
<td>9.5</td>
<td>5.0</td>
<td>8.0</td>
<td>14.5</td>
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<tr>
<td>Dry 0900*</td>
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<td>22.0</td>
<td>11.5</td>
<td>15.0</td>
<td>13.0</td>
<td>14.0</td>
<td>15.5</td>
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<td>25.5</td>
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<td>20.0</td>
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<tr>
<td>Wet 1500</td>
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<td>18.0</td>
<td>15.0</td>
<td>14.0</td>
<td>13.0</td>
<td>14.5</td>
<td>18.5</td>
</tr>
<tr>
<td>Rainfall (mm)</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.6</td>
<td>0.2</td>
<td>0.8</td>
</tr>
<tr>
<td>Wind run (km)</td>
<td>32</td>
<td>68</td>
<td>101</td>
<td>37</td>
<td>121</td>
<td>39</td>
<td>184</td>
</tr>
<tr>
<td>Pan evaporation (mm)</td>
<td>3.2</td>
<td>4.4</td>
<td>2.6</td>
<td>3.4</td>
<td>4.6</td>
<td>2.2</td>
<td>2.6</td>
</tr>
</tbody>
</table>

### University of Melbourne

|               | 23.6     | 27.0    | 18.1    | 21.8     | 19.2     | 20.3   | 26.8    |
| Min.          | 8.3      | 15.7    | 6.2     | 10.4     | 5.1      | 6.5    | 13.9    |
| Dry 0900      | 12.7     | 19.6    | 10.0    | 13.4     | 9.9      | 11.8   | 15.1    |
| Wet 0900      | 10.0     | 13.4    | 8.1     | 10.9     | 8.7      | 8.5    | 14.4    |
| Dry 1500      | 22.6     | 24.4    | 17.0    | 21.2     | 18.7     | 19.0   | 25.9    |
| Wet 1500      | 16.1     | 16.4    | 13.2    | 13.8     | 10.9     | 11.9   | 17.6    |
| Rainfall (mm) | -        | -       | -       | 0.6      | -        | -     | 0.4     |
| Global radiation (mJ/day) | 28.8     | 21.1    | 28.1    | 21.0     | 31.3     | 31.3   | 29.4    |
| Net radiation (mJ/day) | 14.9     | 11.0    | 15.4    | 10.1     | 16.3     | 16.2   | 15.5    |
| Evaporation LE (mJ/day) | 13.0     | 15.4    | 8.4     | 4.5      | 4.9      | 0.8    | 2.8     |
| (mm/day)      | 5.3      | 5.7     | 3.4     | 2.0      | 2.0      | 0.3    | 1.2     |
| Average Bowen ratio | 0.13    | -0.21   | 0.73    | 1.1      | 2.2      | > 20   | 4.1     |
| Estimated error, δE (mm/day) | ±0.3    | ±0.6    | ±0.4    | ±0.5     | ±0.7     | -      | ±0.6    |
| δLE x 100%    | ±6%      | ±10%    | ±11%    | ±23%     | ±35%     | -      | ±47%    |

* 0900 E.S.T. = 1000 Daylight Saving Time.

**TABLE 2.4 SUMMARY OF DATE COLLECTED FOR SELECTED DAYS**
been assumed that
\[ \frac{\delta R_n + \delta G}{R_n - G} = 0.04 \]

\[ \delta \Delta T = \pm 0.02^\circ C \]

\[ \delta \Delta Tw = \pm 0.05^\circ C \]

Because the Bowen ratio fluctuates during the day, the errors have been weighted to reflect that period of the day when evaporation is highest. It must be emphasised that these are only estimates of the maximum error in LE based on assumptions concerning the errors in the measured parameters.

Example 1

These results were collected on 31/10/1979 and are shown in Fig. 2.22 a,b,c and Table 2.5.

During early spring, the annual pastures of the Warrambine Creek catchment have begun to grow vigorously. The spring rains saturate the soil profile and water frequently ponds on the soil surface. In the jargon of ARBM, the depression store is full (See Chapter 5). Under these conditions, potential evaporation occurs.

The average Bowen ratio for this day (i.e. $\frac{\Sigma H}{\Sigma LE}$ for the period when $R_n > 0$) is 0.13 and $\beta$ ranges from 0.20 in the morning to zero throughout the afternoon. In Table 2.5, note the values of $H$ and $LE$ at 0730. These spuriously large values were calculated because $\beta = -1$. This often occurs near sunrise and sunset when $\Delta Tw = 0$. To prevent these results from unduly affecting the daily totals of LE, the computer used to analyze the field data was programmed to calculate $\Sigma LE$ only for the period when $R_n > 0$. This procedure effectively excluded all odd data. On this particular occasion, this procedure may slightly underestimate the 24 hour total of evaporation since some evaporation appears to occur at night. However, examination of Fig. 2.24a, 2.25a and 2.26a shows that this is generally not the case.
Example 2

These results were collected on 4/11/1979 and are shown in Fig. 2.23 a,b,c and Table 2.6.

This day was chosen because it is an example of regional sensible heat advection. Hot, dry air blows across the catchment (See Section 1.3.3) and sensible heat is advected into the area. This causes $H$, and consequently $\beta$, to remain negative throughout the day as the whole surface of the catchment absorbs sensible heat. Under these circumstances, the dominating influence of solar radiation on the local micrometeorology is greatly reduced and the characteristics of the weather system itself dominate. As an illustration of this, note the lack of a marked diurnal pattern in the results (particularly temperature and humidity) when compared with the other days data.

As a consequence of this, a significant amount of evaporation occurs at night and thus, the total of 5.7 mm of evaporation is an underestimate. Furthermore, if the findings of Section 2.1.4 are correct and $K_H/K_W$ is in the range of 1.2 to 1.5 then both $\beta$ and LE are underestimated. The total evaporation for this day could exceed 8 mm.

Examples 3, 4, 5

These results were collected on 8/11/1979, 11/12/1979 and 10/1/1980. The results are shown on Fig. 2.24 a,b,c; Fig. 2.25 a,b,c and Fig. 2.26 a,b,c and are tabulated in Table 2.7, 2.8 and 2.9 respectively.

These three days illustrate how the average Bowen ratio progressively increased throughout the summer as the soil dried. Every year the catchment changes from an area of green pasture and saturated soils in early spring to an area of dry, dead vegetation and parched, dry, cracked soils. These changes are shown in Plates 1.1 to 1.5.

It is interesting to note that the absolute error in LE remains moderately constant throughout the period even though the relative error becomes progressively larger (See
Table 2.4). Also, no evaporation occurs at night. For Example 5, (10/1/1980), 1.8 mm of rain fell on the preceding day and 0.4 mm fell just before dawn. This rain was quickly re-evaporated at a potential rate (β = 0) in the early morning, but for the rest of the day little evaporation occurred.

Example 6

These results were collected on 10/12/1979 and are shown in Fig. 2.27 a,b,c and Table 2.10.

This day was selected to show the effect of brief showers during the day. The rain fell in two showers (0.4 mm at 1200 and 0.2 mm at 1500). On both occasions, the rain was rapidly evaporated at a potential rate (i.e. β = 0) and β returned to pre-storm values within an hour of the finish of each storm. (N.B. A short power black-out appears to have occurred during the 1200 storm and this accounts for the odd data recorded at this time).

It is interesting to note that the Bureau of Meteorology shows this rain as occurring on the next day. The Bureau's data are taken for the 0900 to 0900 period each day and thus, this rain would not have been recorded until the next morning. This rain event would normally be back-dated but this was apparently not done this time.

Example 7

These results were collected on 1/1/1980 and are shown on Fig. 2.28 a,b,c and Table 2.11.

This day was chosen to illustrate how under very dry conditions, the Bowen ratio method can fail because the errors in the computed value of LE become excessive. Firstly, examine the values of β in Table 2.11. This erratic data, ranging from +37 to -14, immediately suggests malfunction. However, when the diurnal patterns of ΔT and ΔTw are compared with the same parameters on another dry day (See Fig. 2.28b), it is evident that the data is sensible and that the equipment is functioning satisfactorily. To under-
stand the erratic values of \( \beta \), refer again to Fig. 2.19. Under certain circumstances, \( \beta \) becomes infinite. When this is close to occurring, the slightest error in \( \Delta T_w \) causes gross errors in \( \beta \). This infinite error occurs when

\[
\left( \frac{S + Y}{Y} \right) \frac{\Delta T_w}{\Delta T} - 1 = 0
\]

It is then possible to define the value of \( \Delta T_w \) which causes \( \beta \) to become infinite as

\[
\Delta T_w' = \frac{\Delta T}{S + Y} \tag{2.35}
\]

When \( \Delta T_w' \) is also plotted on Fig. 2.28b, it is evident that on both days, the measured value of \( \Delta T_w \) was within \( 0.05^\circ C \) of \( \Delta T_w' \). On 10/1/1980, \( \Delta T_w > \Delta T_w' \) and consequently "sensible" Bowen ratios were calculated. On 1/1/1980, \( \Delta T_w \) was often less than \( \Delta T_w' \) and consequently negative Bowen ratios were calculated. Under these conditions, the daily total of \( LE \) can only be guessed. It would not be unreasonable to assume that for most of the day, \( \beta \) was greater than 20. Thus \( LE = 0.05(R_n - G) \) which yields 0.3 mm of evaporation. The error in this estimate is completely unknown.

The above discussion has assumed that the \( \Delta T_w \) data was in error. However, there is no reason why this should be so, particularly as the \( \Delta T_w \) data on later days appears to be correct. Thus, there is a possibility that a Bowen ratio of -37 actually occurred.

Normally, negative Bowen ratios indicate sensible heat advective (See Example 2), but in this case, the negative \( \beta \) may indicate a negative \( LE \), not negative \( H \). This implies that the surface was so dry that it absorbed water from the more humid atmosphere above it. This is an unlikely event and, to the author's knowledge, such an occurrence has never been reported in the literature. However, this possible explanation of the data collected on this day still exists.

2.5 SUMMARY

Evaporation, \( LE \), can be determined as the residual of all the energy fluxes across a natural surface. With careful
planning, the experiment can be designed so that only four energy fluxes are significant (i.e. $R_n$, G, H and LE). Of these, $R_n$ and G can be measured directly and H can be determined by use of the Bowen ratio. This involves the measurement of $\Delta T$, $\Delta Tw$ and $Tw$. An error analysis is presented which indicates that for these five parameters, the most precision is required in the measurement of $R_n$, $\Delta T$ and $\Delta Tw$. Errors in the measurement of G and Tw cause far smaller errors in the computed value of LE. The analysis also demonstrates that under dry surface conditions, large errors in LE could be expected and this was confirmed with experimental data. The data collected by the equipment described in this chapter can be used to propose and test the evaporation sequences in ARBM.
Fig 2.22a
EXAMPLE 1 - ENERGY FLUX DATA 31/10/1979

- $R_g$
- $R_n$
- LE
- G

Energy Flux (W/m²)

Time (Hr.)

0 2 4 6 8 10 12 14 16 18 20 22

-100 0 100 200

Fig 2.23a  EXAMPLE 2 - ENERGY FLUX DATA  4/11/1979
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Minimum -87 11.3 15.6 0 -12 -1.90 -290 -494 -41
Mean 16.3 24.3

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| -87.1   | 0.72 | Rg  | t² = 0.985
| -3.4   | 0.032 | Rn  | t² = 0.620

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Fig 2.24b   EXAMPLE 3 - BOWEN RATIO DATA   8/11/1979
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Rn = -55.7 + 0.64 * Rg  \( r^2 = 0.975 \)
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| Maximum | 65   | 11.8| 19.2| 1075| 52  | 9.00 | 172 | 434 | 342 |
| Minimum | -80  | 3.9 | 4.5 | 0  | -17 | -6.73 | -57 | -117 | -40 |
| Mean    | 10.4 | 16.5|     |     |     | 2.14 |
| Total   | 16.67| 10.83| 0.96| 4.87 | 10.83| 8.85 |

Rn = 113.9 + 0.71 × Rr \quad \quad r^2 = 0.982

G = 11.5 + 0.099 × G \quad \quad r^2 = 0.976

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| Minimum | -77 | 12.7| 13.9| 0 | -16 | -12.71| -15| -64 | -41 |
| Mean    | 16.7| 22.8| 6  | 4.59 |
| Total   | 15.58| 29.07| 1.06| 2.83| 11.70| 9.47 |
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Fig 2.27b  EXAMPLE 6 - BOWEN RATIO DATA
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Total  10.28  20.78  0.22  4.88  5.18  6.01
Mean  12.7  17.2

Rn = 2.0 + 0.49 * Rg  r^2 = 0.0621
G  = 11.1 + 0.029 * Rn  r^2 = 0.071

Table 2.10 Example 6 - Data for 10/12/1979
Fig 2.28a  EXAMPLE 7 - ENERGY FLUX DATA 1/1/1980

Note: No LE data calculated see text.
Fig 2.28b  EXAMPLE 7
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**Maximum** 611 12.2 20.3 1095 42 37.23 7021 1205 340

**Minimum** -63 5.7 6.5 0 -22 -14.68 -990 -7042 -23

**Mean** 10.8 16.7

**Total** 16.19 30.92 0.79 -4.42 19.82 8.8

\[ \text{Rn} = -53.3 + 0.60 \times Rg \quad r^2 = 0.970 \]

\[ \text{GL} = -12.83 + 0.084 \times Rn \quad r^2 = 0.972 \]

**Table 2.11 Example 7** Data for 1/1/1980
Chapter 3

THE

EDDY CORRELATION

SYSTEM

3.0 INTRODUCTION

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CHAPTER 3

THE EDDY CORRELATION EVAPORATION MEASURING SYSTEM

3.0 INTRODUCTION

In 1951, Swinbank proposed a method of evaporation measurement based on what is now known as the Eddy Correlation Technique (Swinbank (1951)). This approach is the most theoretically fundamental method of measuring evaporation, since it actually senses the vertical flux of water vapour above an evaporating surface. Because no simplifying assumptions are made in the derivation of the theory, this appears to be an attractive technique, but its widespread implementation has been hindered by the lack of suitable sensors and computing systems.

As was noted in Section 1.6.7, the development of the eddy correlation equipment for this project did not commence in earnest until the energy balance equipment was working satisfactorily. This late start, coupled with problems encountered in attempting to use innovative technology, has meant that the instrumentation has only been developed to the laboratory testing stage. Very little field data was collected and therefore questions concerning the system's performance under adverse conditions have yet to be answered. An A.W.R.C. grant has been obtained to continue research into evaporation from the Warrambine Creek catchment and, as part of that programme, field testing of the equipment described below (or developments of it) should commence soon. It is in the light of the future programme that many comments and suggestions have been made.

For this project, it was initially intended to use an existing "Fluxatron" system with a single Gill propeller. An infrared hygrometer, similar to that described by Hyson & Hicks (1975), would be constructed. However, a detailed review of the performance of the Gill propeller revealed several serious sensor deficiencies which cannot be corr-
ected with a simple "Fluxatron". It was then decided to replace the "Fluxatron" with an "on-line" microprocessor. It was intended to use this not only to perform conventional eddy correlation calculations, but also to perform additional operations, notably (i) corrections for sensor deficiencies and non-linearity, and (ii) provision of additional meteorological data, e.g. $\bar{u}$, $\bar{q}$, $\bar{T}$. Unfortunately time constraints prevented the satisfactory completion of all these objectives.

3.1 THEORY

Consider a point in the atmosphere above an evaporating surface. Then the instantaneous vertical flux of water, expressed as an energy flux, is given by

$$LE = L \rho \bar{w} q$$  \hspace{1cm} 3.1

where

LE = latent heat flux (W/m$^2$)
L = latent heat of vaporization of water (J/kg)
$\rho$ = density of air (kg/m$^3$)
$\bar{w}$ = vertical wind velocity (m/s)
$q$ = specific humidity (kg/kg)

and thus the average flux over some time period can be defined as

$$\overline{LE} = L \overline{\rho \bar{w} q}$$  \hspace{1cm} 3.2

Now, both $q$ and $w$ are rapidly fluctuating atmospheric variables which can conveniently be expressed as

$$w = \bar{w} + w'$$  \hspace{1cm} 3.3
$$q = \bar{q} + q'$$  \hspace{1cm} 3.4

where the bar and prime denote the mean and fluctuating components respectively. This is shown in Fig. 3.1.

Thus, the instantaneous energy flux can be represented as

$$LE = L \rho (\bar{w} + w')(\bar{q} + q')$$  \hspace{1cm} 3.5

and the average flux can be represented as

$$\overline{LE} = L \overline{\rho (\bar{w} + w')(\bar{q} + q')}$$

$$= L \overline{\rho (\bar{w}\bar{q} + \bar{w}q' + w'\bar{q} + w'q')}$$  \hspace{1cm} 3.6
By definition, \( \overline{w'q'} \) and \( \overline{w''q} \) equal zero. Also, it is generally assumed that \( \overline{w} = 0 \) since any non-zero value would imply a net upward or downward draft near the surface. Then Eqn. 3.6 reduces to

\[
\overline{LE} = \frac{L \rho \overline{w'q'}}{P} \tag{3.7}
\]

This is the fundamental equation upon which all eddy correlation systems are based. A similar procedure can be used to determine the fluxes of other atmospheric properties, e.g. sensible heat, \( H \), and momentum, \( \tau \).

\[
H = \frac{\rho C_w \overline{w'T'}}{P} \tag{3.8}
\]
\[
\tau = \rho \overline{u'w'} \tag{3.9}
\]

Before reviewing the methods used to apply this theory, it is instructive to extend the analysis. Consider the case where both \( q \) and \( w \) could be expressed as a Fourier series, as below.

\[
w = a_0 + a_1 \sin \omega_1 t + \ldots + a_n \sin \omega_n t \tag{3.10}
\]
\[ q = b_0 + b_1 \sin \omega_1 t + \ldots + b_n \sin \omega_n t \quad 3.11 \]

The average flux over a period of time, \( T \), can be calculated as

\[
\overline{LE} = \frac{1}{T} \int_0^T L \rho w q \, dt \quad 3.12
\]

and after integration over a large enough time interval, this yields

\[
\overline{LE} = a_0 b_0 + \sum_{i=1}^{n} \frac{a_i b_i}{2} \quad 3.13
\]

This important result shows that only fluctuations of the same frequency correlate and contribute to the energy flux. In an ideal case where \( w = a \sin \omega_1 t \) and \( q = b \sin \omega_2 t \), where \( \omega_1 \neq \omega_2 \), the net energy flux is zero.

This result is used to determine the required frequency response of the sensors and the averaging periods used. If there exists in one entity a fluctuation of a frequency which is not present in the other (e.g. the diurnal fluctuations of \( T \) and \( q \) which are not present in \( w \)), then correlations at that frequency do not exist and can be ignored. Several workers have examined the cospectra of \( wq \) and \( wT \) and have determined the frequency range relevant to eddy correlation work. Fig. 3.2 shows the normalized cospectra for vertical velocity (\( w \)) and an atmospheric property (\( s \)) as taken from Garratt (1975). Normalized frequency is used to allow for the effect of wind speed and measurement height on the variation of cospectral density with natural frequency.

3.2 COVARIANCE COMPUTING SYSTEMS

3.2.1 Analogue Computing Systems

Historically, only one method has been used to compute the various fluxes in the field. The technique has been to convert the atmospheric properties (e.g. \( w, T, q \)) into analogue voltages which were then manipulated electronically.
This cospectra is taken from Garratt (1975).
Based on previous experimental results, cospectra will probably apply for momentum, heat, water vapour and carbon dioxide in near-neutral stability.
Normalized frequency, $f = nz/u$ where $n = $ natural frequency, $z = $ height, $u = $ horiz. windspeed.

Fig 3.2 NORMALIZED COSPECTRA FOR VERTICAL VELOCITY ($w$) AND PROPERTY ($s$).
to produce another analogue voltage which was proportional to the instantaneous energy fluxes. These signals were then integrated, using conventional d.c. integrators.

One of the first of these devices was designed by the CSIRO Division of Atmospheric Physics and was called the "Evapotron" (Dyer & Maher (1965a), Dyer & Maher (1965b)). This system was later modified, improved and subsequently renamed the "Fluxation" (Dyer et al (1967), Hicks (1970)). This device worked successfully in several investigations (e.g. Linacre et al (1969), Hicks (1973), Hicks et al (1975)) and is still in use today (Raupach (1977)).

The system is shown diagrammatically in Fig. 3.3. A fundamental component is the high-pass RC filter. This filter effectively subtracts from the input (i.e. w, q, T) a running mean (i.e. \( \bar{w}, \bar{q}, \bar{T} \)) whose averaging period is determined by the time constant of the filter. The resultant analogue voltage is then directly proportional to the fluctuating component of the input (i.e. \( q', w', T' \)). The time constant chosen must be sufficiently long so that all the fluctuation frequencies contributing to the flux are included. Early systems used time constants which were too short (e.g. 40s, Dyer et al (1967); 160s, Hicks (1970)) and consequently a proportion of the flux was lost due to low-frequency filtering. A better understanding of the contributing frequencies now exists and longer time constants are used (e.g. 600s, Raupach (1977)).

The above system, while being a proven research tool, does have some disadvantages. Firstly, in most cases, the average values of w, q and T are not calculated and thus, if for any reason \( \bar{w} \neq 0 \), this will be undetected and errors in the computed flux will result. Secondly, unless extremely long time constants are used, some low frequency filtering loss will occur. Thirdly, the analogue multipliers and integrators can introduce errors. This was noted by Hicks (1969). Fourthly, it is not readily possible to correct for sensor deficiencies or to use sensors with a non-linear output.
Fig 3.3a  BLOCK DIAGRAM OF TYPICAL "FLUXATRON" SYSTEM

Fig 3.3b  CIRCUIT DIAGRAM OF "FLUXATRON" (w channel only)

Fig 3.3  ANALOGUE COVARIANCE COMPUTER
3.2.2 Digital Computing Systems

The idea of digitizing the analogue sensor signals and then using a computer to calculate the various fluxes is not new. Several researchers have recorded field data on magnetic tape and later analysed this in a laboratory. This approach is attractive since not only can the fluxes be calculated accurately, but also the frequency spectrum can be analysed. However, this method has only been used in "pure" research applications. Because of the inconvenience and cost of storing and analysing massive amounts of data, it does not lend itself to the routine measurement of evaporation.

Recently an offshoot of the computer industry has emerged which requires a re-evaluation of this position. The microprocessor is a miniature computer; a device which can be programmed to perform all the usual operations of large computers and yet is small, compact and cheap. The concept of using a microprocessor in the field to calculate evaporation fluxes is extremely inviting, since not only can additional information (e.g. \( \overline{q} \), \( \overline{T} \), \( \overline{w} \)) be obtained at no extra cost, but also non-linear sensors can be used and sensor deficiencies can be corrected for instantaneously. Also, the system offers great flexibility since no hardware changes are necessary to modify the system's performance.

This type of system was discussed by Moore et al (1976), but, to the author's knowledge, there were at the time of writing no published results of the use of a microprocessor for "on-line" eddy correlation calculations.

3.2.3 Hardware Details

The system is shown diagrammatically in Fig. 3.4. The microprocessor is a Signetics 2650 with 4K of RAM. Its input ports are connected to an 8 bit, 16 channel A-D (analogue to digital) converter, of which only four channels are presently used. An external clock is also attached. Output is to a FACIT paper-tape punch. The paper-tape is normally analysed in the laboratory, but if a paper-tape
Fig 3.4  BLOCK DIAGRAM OF MICROPROCESSOR SYSTEM
reader/printer is available in the field, the ASCII formatted data can be viewed immediately.

3.2.4  Programming

Before discussing the microprocessor's programming, it is necessary to review the eddy correlation theory in the light of the microprocessor's capabilities. Generally, microprocessors have limited memory space. Initially it appears that this characteristic will cause problems when dealing with the massive amounts of data required in eddy correlation analysis. The most perplexing problem is: how can the mean component of the atmospheric entity be removed from the total to leave the fluctuating component only, when this mean is not known previously? One method might be to continuously calculate a running mean and thus, in a digital sense, simulate the high-pass filter of the analogue system. This, however, is a relatively complicated task requiring large amounts of memory space. Clearly the problem needs to be re-examined from first principles.

Swinbank's initial equation for the instantaneous flux is

\[ \text{LE} = \text{L} \rho \text{w} \text{q} \]  \hspace{1cm} 3.1

Thus, the average flux, \( \overline{\text{LE}} \), over some specified time period, is

\[ \overline{\text{LE}} = \overline{\text{L} \rho \text{w} \text{q}} \]  \hspace{1cm} 3.2

That is, the average value of the instantaneous product of \( \text{w} \) and \( \text{q} \) should give the correct energy flux and the requirement of separation of the mean and fluctuating components is not necessary. The apparent anomaly that Eqn. 3.2 produces an equivalent result to Eqn. 3.7 is a direct result of the assumption that \( \overline{\text{w}} = 0 \). As long as this condition holds (a vital aspect which will be discussed later), then these equations are equivalent.

Thus, the primary aim of the microprocessor system described below is to calculate evaporation as defined by Eqn. 3.2. Also, to confirm energy balance data collected by the equip-
ment described in Chapter 2, it is necessary to calculate \( \bar{H} \), the average sensible heat flux. Furthermore, for reasons which will become apparent later, it is necessary that the average value of each of the four inputs, i.e. \( \bar{w}, \bar{q}, \bar{T} \) and \( \bar{u} \) should also be determined.

To see how this data can be obtained, the operation of the system (See Fig. 3.4) must be examined step by step. The various sensors, which will be discussed in later sections, each produce an analogue voltage which can be expressed as

\[
s = a + b V_1
\]

where \( s \) = atmospheric entity
\( V_1 \) = analogue voltage out of sensor (Point 1 in Fig. 3.4)
\( a, b \) = calibration coefficients

(N.B. Some of the sensors (e.g. \( q, T \)) have slightly non-linear outputs and others (e.g. \( w \)) have response deficiencies. It is anticipated that in future developments, the microprocessor will be used to compensate for these deficiencies as each signal is measured. Some of these corrections are outlined in Section 3.3.)

The signal, \( V_1 \), is fed into a pre-amplifier which amplifies and/or biases it so that it will suit the input range of the A-D converter (i.e. 0 - 10V).

Thus,

\[
V_1 = c + d V_2
\]

where \( V_2 \) = analogue voltage into A-D converter
\( c, d \) = gain, bias parameters

When the A-D converter is told by the microprocessor to sample a channel, it converts the 0-10V analogue input to a 0-2^8 binary output. This number is then used by the microprocessor. Thus,

\[
V_2 = e + f \, B
\]
where $B =$ binary input to the microprocessor
$e, f =$ conversion coefficients

Thus, the atmospheric entity, $s$, can be instantaneously related to the binary number entering the microprocessor by

$$s = g + hB$$  \hspace{1cm} 3.17

where $g = a + b + c + b + d + e$
$h = b + d + f$

Now, one of the required outputs is $\bar{s}$. That is

$$\bar{s} = \frac{1}{n} \sum_{i=1}^{n} B_i$$

where

$$\overline{\bar{B}} = \frac{1}{n} \sum_{i=1}^{n} \bar{B}_i$$

$\bar{B}_i =$ the $i^{th}$ sample of the input, $B$
$n =$ total number of samples in the period

The second output required is $\overline{ws}$. That is

$$\overline{ws} = \frac{1}{n} \sum_{i=1}^{n} (g + hB)(g'' + h''B'')$$

where the double prime denotes the $w$ channel. Thus

$$\overline{ws} = \frac{1}{n} \sum_{i=1}^{n} (gg'' + hg''B + gh''B'' + hh''BB'')$$

$$= gg'' + hg''\overline{B} + gh''\overline{B}'' + hh''\overline{F}$$  \hspace{1cm} 3.19

where $\overline{F} = \frac{1}{n} \sum_{i=1}^{n} \bar{B}_i B''$

and all the other terms are as previously defined.

Eqn. 3.19 indicates that the mean values of each atmospheric entity (i.e. $\overline{B}$) are not only interesting by-products of this type of system, but are essential components in the determination of the various energy fluxes.
Thus, in its most basic form, the microprocessor's programme must calculate $\overline{E}$ for each input and $\overline{F}$ for the WT and WQ channels. This is done using the flow chart shown in Fig. 3.5.

There are a number of interesting features of this system. Firstly, it is not the external clock (See Fig. 3.4) which determines when to finish an averaging period. Rather, this is determined by the time the system takes to take $n$ samples. The value of $n$ chosen was $256 \times 256 \times 16$. This means that each channel is sampled 1.04 million times in each averaging period. The duration of this period is determined by the sampling rate of the A-D converter and by the computational time required by the microprocessor. This turned out to be 44 minutes, a time which adequately measures the lowest eddy frequencies encountered. Conversely, this produced a system which sampled each channel at approximately 400Hz. This is more than adequate high-frequency sampling for eddy correlation measurements. A second feature of the system is the method of calculating $\overline{E}$ and $\overline{F}$. This "averages of averages" method reduces the size of the numbers handled and thus reduces the memory space required. This system appears to be a most satisfactory method for calculating eddy fluxes when using an "on-line" microprocessor.

3.2.5 Laboratory Testing

In order to confirm the performance of the microprocessor system, laboratory tests were performed. Initially, constant voltages, $V_2$, were fed into each channel. The results of this test are shown in Fig. 3.6. Then, a sine wave was added to this constant input voltage, such that

$$V_2 = \overline{V} + V' \sin \omega t$$  \hspace{1cm} 3.20

The system operated as expected. The output, $\overline{E}$, for each channel corresponded to the value predicted for $\overline{V}$ (since the average of the sine wave is zero). For $\overline{F}$, the value was in accordance with that predicted by Eqn. 3.13 when the added sine waves were of the same frequency. However, when different frequencies were used, no eddy correlation
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Start

Input Time, Date

Initialize Parameters

Input T,q,w,u from A-D converter

T(0) = T(1) + T
T(2) = T(2) + q
T(3) = T(3) + w
T(4) = T(4) + u
T(5) = T(5) + w.T
T(6) = T(6) + w.q

N = N + 1

N < 256

Yes

N = 0

For J = 1 to 6
R(J) = R(J) + T(J)/256

M = M + 1

M < 256

Yes

M = 0

For J = 1 to 6
Q(J) = Q(J) + R(J)/256

K = K + 1

K < 16

Yes

K = 0

B1 = Q(1)/16
B2 = Q(2)/16
B3 = Q(3)/16
B4 = Q(4)/16
P1 = Q(5)/16
P2 = Q(6)/16

Output Subroutine

Fig 3.5 FLOWCHART OF MICROPROCESSOR PROGRAMME
Fig 3.6  LABORATORY CALIBRATION OF MICROPROCESSOR PROGRAMME
occurred. Thus, as far as laboratory tests were concerned, the new covariance computer worked well.

3.2.6 Field Testing

Because of the late start to this section of the project and because of difficulties in getting the microprocessor system operational, very little field data was collected. However, experience was obtained in the use of a microprocessor system in the field and a number of operational problems were discovered.

Most of the problems occurred because the microprocessor is similar to a programmable calculator. Firstly, each time the calculator is switched on, the programme must be "keyed" in. In this case, the programme was stored on paper-tape and had to be loaded into the microprocessor each time the system was set up. While this procedure presented no difficulties in the laboratory, it was a different situation in the field. It proved difficult to control 30 metres of paper-tape in windy conditions when the paper-tape reader was mounted in the back of a utility.

The second problem encountered was more serious. Like a pocket calculator, the microprocessor loses all programme information whenever the power is switched off. Furthermore, for the microprocessor, a complete power failure is not necessary. On several occasions, the microprocessor, which was left operating correctly after a visit to the site, mysteriously stopped after only a few hours. It then remained inoperative until the next site visit. A mains-powered clock in the instrument shed indicated that no power failure had occurred. After considerable investigations, it was discovered that when the farmer who owned this property switched on a pump, a voltage drop occurred which was sufficient to disrupt the microprocessor, yet did not disrupt any of the other instruments. The third problem was that the time and date had to be manually keyed into the programme each time the system was set up.
Solutions to all these difficulties were proposed, but time constraints prevented their implementation. However, as an aid to future developments, the necessary modifications will be outlined below.

Firstly, the system should have a full date and time clock with a separate battery power supply. In this case, the clock would continue operating even when the microprocessor is switched off. The time can be accurately set in the laboratory. The microprocessor can then be programmed to read the date and time as input data each time the system is switched on.

Secondly, when development of the microprocessor programme is complete, a PROM should be "blown". A PROM (Programmable Read Only Memory) is an electronic component in which the programme can be stored permanently. It remains there even when the microprocessor is switched off. This programme should include an automatic reload and restart sequence which would reactivate the microprocessor after any power failure.

Thirdly, a simple OCTAL keyboard should be included in the system. This would then allow minor modifications to the programme to be made in the field without the necessity of carrying a large teleprinter to the site.

While many problems were encountered, both in the laboratory development and field use of this system, the experience gained here has shown that the microprocessor is a viable alternative to the analogue computing systems. All the modifications to the eddy correlation theory necessary because of the microprocessor's own characteristics have been examined. Also, further uses of the microprocessor's capabilities have been proposed. Field experience with a first prototype has demonstrated the need for additional features in the system and when these are incorporated, it is hoped that a reliable, versatile alternative to the "Fluxatron" will emerge.
3.3 VERTICAL WIND SENSOR

3.3.1 Sensor Selection

A number of different sensors are presently available to measure the vertical component of the wind vector. These include hot-wire anemometers (e.g. Dyer & Maher (1965a)), pressure spheres (e.g. Goltz et al (1970)), sonic anemometers (e.g. Kaimal & Businger (1963)) and propeller anemometers. This diverse range of sensors indicates that a final solution to the measurement of $w$ has yet to be found.

For this study, a Gill propeller (Holmes et al (1964)) was used. This was chosen mainly because one was available at the commencement of the project, but the Gill propeller is probably the optimum choice at this stage. Moore et al (1976), when reviewing all the sensors presently available, concluded that "the sonic anemometer appears to be the most suitable instrument for measuring the atmospheric wind components". However, they noted that significant problems still exist in the design of the device. The same general comments apply to pressure spheres. Hot-wire anemometers, although accurate, are ruled out because of their lack of durability in the field. Of the propeller anemometers, Moore et al (1976) concluded that the Gill propeller "probably represents the best mechanical, linear windspeed transducer available".

The Gill propeller is available from R.M. Young & Co., Michigan, U.S.A. It consists of a four-bladed helicoid polystyrene propeller which drives a miniature d.c. generator. The analogue voltage output of this generator is a linear function, both in direction and magnitude of the windspeed along the propeller axis. The Gill propeller used is shown in Plate 3.1.

This device has several advantages over the other types of sensors. These are reliability, simplicity, ease of mounting, and light weight. It does have several deficiencies in its response to $w$. Fortunately, because this device has been widely used, these deficiencies have been extensively
PLATE 3.1 Gill Propellor and cup anemometer mounted on the eddy correlation mast.

PLATE 3.2 Eddy correlation mast with instruments mounted at 5m. Plate shows replacement of a thermistor being undertaken without lowering the mast.
studied. The nature of these deficiencies and possible corrections for them will be discussed below.

3.3.2 Cosine Response

The Gill propeller was designed to measure only that component of the wind parallel to its axis. That is, its output should be proportional to $U \cos \theta$, where $U$ is the total wind vector and $\theta$ is the angle between $U$ and the propeller's axis (See Fig. 3.7). Several studies (e.g. Camp et al (1970), Hicks (1972), Gill (1975)) have shown that

\[
\bar{U} = \text{Total wind vector} = \text{Vector sum of components} = \bar{U} + \vec{V} + \vec{W}
\]

![Diagram](attachment:image.png)

FIG. 3.7 Definition of the Direction of the Total Wind Vector, $U$, with Respect to the Gill Propeller

this device fails to achieve a perfect cosine response, particularly when $60^\circ < \theta < 120^\circ$. To evaluate the cosine response of the anemometer used in this study, it was calibrated in the laboratories of the CSIRO Division of Atmospheric Physics. This calibration data is shown in Table 3.1 and Fig. 3.8. To illustrate the cosine response, Fig 3.9
$w = 0.128 + 0.0170 \, E$

Fig 3.8a  AXIAL CALIBRATION OF GILL PROPELLOR

Windspeed along axis i.e. $U \cos(\theta)$
where $U = 4 \text{ m/s}$
and $\theta = 60^\circ - 120^\circ$

Fig 3.8b  CALIBRATION OF GILL PROPELLOR TILTED INTO THE WIND

Fig 3.8  CALIBRATION OF GILL PROPELLOR
Fig 3.9  PLOT OF $V/V_0 \cos \theta$ vs. $\theta$
was plotted. In this figure, $V_0 \cos \theta$ is the actual axial windspeed during the test and $V$ is the windspeed predicted using the axial ($\theta = 0^\circ$) calibration data. The perfect cosine response is indicated when $V/V_0 \cos \theta = 1$. From Fig. 3.9, it is clear that there is a significant deviation from the ideal cosine response in the region $60^\circ < \theta < 120^\circ$ and that this deviation is a function of windspeed.

<table>
<thead>
<tr>
<th>Windspeed (m/s)</th>
<th>Output (mV)</th>
<th>Angle of Attack</th>
<th>Output (mV)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.34</td>
<td>10</td>
<td>85</td>
<td>13</td>
</tr>
<tr>
<td>0.49</td>
<td>23</td>
<td>80</td>
<td>30</td>
</tr>
<tr>
<td>1.00</td>
<td>53</td>
<td>75</td>
<td>44</td>
</tr>
<tr>
<td>2.02</td>
<td>111</td>
<td>70</td>
<td>62</td>
</tr>
<tr>
<td>3.01</td>
<td>169</td>
<td>65</td>
<td>80</td>
</tr>
<tr>
<td>4.03</td>
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<td>94</td>
</tr>
<tr>
<td>4.99</td>
<td>288</td>
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<td>126</td>
</tr>
<tr>
<td>5.95</td>
<td>343</td>
<td>40</td>
<td>162</td>
</tr>
<tr>
<td>7.10</td>
<td>410</td>
<td>30</td>
<td>194</td>
</tr>
<tr>
<td>8.05</td>
<td>468</td>
<td>20</td>
<td>214</td>
</tr>
<tr>
<td>8.98</td>
<td>520</td>
<td>10</td>
<td>228</td>
</tr>
<tr>
<td>10.04</td>
<td>585</td>
<td>0</td>
<td>230</td>
</tr>
<tr>
<td>10.90</td>
<td>635</td>
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<td>-226</td>
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<td>11.95</td>
<td>695</td>
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<tr>
<td>13.00</td>
<td>760</td>
<td>-30</td>
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</tr>
<tr>
<td></td>
<td></td>
<td>-85</td>
<td>-18</td>
</tr>
</tbody>
</table>

**TABLE 3.1 Calibration data for Gill propeller**

Unfortunately, when the Gill propeller is used as a vertical wind sensor, this is the region in which most wind measurements are made. To compensate for this Hicks (1972) noted that in this region the response is nearly linear and proportional to $\cos \theta$. He therefore recommends that the calibration coefficient which should be used is not that obtained from axial tests (Fig. 3.8a), but rather that obtained for tests when $60^\circ < \theta < 120^\circ$ (Fig. 3.8b). This calibration coefficient is approximately 80% of the axial value.
This solution to the cosine response problem is only a first order approximation. It does not take into account varia-
tions in cosine response due to changes in $g$ or $V$. To im-
prove this situation, it was one of the objectives of the
microprocessor system that "on-line" corrections for cosine
response deficiencies would be made to the $w$ data. This can
be done by using the methods suggested by Moore et al (1976),
Gill (1975) or Horst (1973). Unfortunately, time constraints
prohibited any attempt at those calculations.

3.3.3 High Frequency Response

It has already been shown that an eddy correlation system
does not have to respond to all fluctuation frequencies of
the atmospheric entities. However, the obvious corollary
of this is that the sensors must fully respond to all fre-
quencies contributing to the energy flux. The inability of
the Gill propeller to respond to high frequency fluctuations
has always limited its use in eddy correlation studies.

Unlike many other sensors, the frequency response of the
Gill propeller is characterised not by its response time,
but by its response length. This is defined by Camp et al
(1970) to be "the length of a column of air that must pass
a windspeed sensor (after a sharp-edge gust or partial lull
has occurred) for the sensor to reach 63\% (1 - $1/e$) of the
new equilibrium value". Tests on Gill propellers have
shown that when $0 = 0^\circ$, a response length of 1 m is typical,
but that this increases rapidly as $0$ approaches $90^\circ$ (Carratt
(1975)). Thus, when used as a vertical wind sensor, the
Gill propeller has a relatively slow response time (e.g. 2 s;
Wesley & Hicks (1975)). McBean (1972) clearly showed this
with comparative $w$ measurements made by Gill propellers and
sonic anemometers.

Generally, eddies of higher frequencies occur closer to the
ground. Thus, because of its poor high frequency response,
it is unacceptable to use a Gill propeller close to any sur-
face. To illustrate this, Wesley & Hicks (1975), when crit-
icising Desjardins & Lemon (1974) for mounting a Gill pro-
peller only 1 m above a crop, calculated that in this situation the true energy flux was being underestimated by 48% due to high frequency losses. Garratt (1975) has thus concluded that if high frequency losses are to be kept to less than 10%, then a measurement height of 5 m or more over land should be used. Therefore, an instrument height of 5 m was chosen for this project (See Plate 3.2).

Even at this height, some high frequency loss will still occur and again it was intended to use the microprocessor to make "on-line" corrections for this. Garratt (1975) has shown that these corrections can be estimated from the horizontal windspeed and therefore a horizontal cup anemometer was included in the system. Again, time constraints prohibited the implementation of this concept.

3.3.4 Threshold Response

Because of its inertial and frictional characteristics, the Gill propeller has a finite threshold windspeed below which it ceases to rotate. This is typically 0.15-0.20 m/s. Furthermore, its calibration is non-linear below 1 m/s. When used as a vertical wind sensor, the propeller must frequently pass through this "dead" region as w approaches zero or changes sign. This occurs so frequently that McBean (1975) estimated that at a height of 10 m above a corn crop, the anemometer would be operating in its linear calibration zone for less than 5% of the time. Obviously, this situation must cause errors in the computed flux if the linear calibration is used.

The solution most frequently proposed to this (e.g. McBean (1975), Horst (1973)) is to use not one, but two or more anemometers. These are tilted into the wind, but are at an angle to each other. This then ensures that both anemometers are continuously operating in their linear calibration zone. This technique necessitates the instantaneous resolution of the vertical and horizontal components and the introduction of the microprocessor offers an ideal method of doing this. Once again, time constraints prevented any work in this
area, but, as will be shown below, the concept of using more than one anemometer has further advantages and must be considered in any future project.

3.3.5 **Tilt Errors**

This is an error caused not by deficiencies in the sensor itself, but by errors in sensor alignment. When a sensor is not correctly aligned, it does not measure $w$ alone, but rather some combination of $u$, $v$ and $w$. The computed energy flux then includes contributions due to $\bar{u} \bar{q}$, $\bar{u}' \bar{q}'$ and $\bar{v}' \bar{q}'$. These terms represent the horizontal transport of water vapour by mass movement of the air and by eddy turbulence. They do not represent the vertical transport of water from the surface to the atmosphere.

Previously, most eddy correlation studies have been conducted over extensive flat surfaces. Under these circumstances, an exact definition of the "vertical" direction of sensor alignment was not necessary. The mean wind vector, $\bar{U}$, would be parallel to the surface and thus horizontal (i.e. perpendicular to the direction of gravity). Sensors were then aligned using a plumb-bob method. However, the situation does not apply at the Warrambine Creek site (See Fig. 3.10). Three choices (which are conveniently concurrent for an extensive flat surface) exist for the definition of the vertical.

They are:

(a) Perpendicular to the earth's surface, this being defined by gravity.

(b) Perpendicular to the plane of the surrounding surface. For this site, this would be approximately $1.5^\circ$ from the line of gravity (See Fig. 1.2).

(c) Perpendicular to the direction of the mean wind vector, $\bar{U}$.

Definition (a) is the alignment most commonly recommended. Definition (b) however, seems correct since it implies that one is trying to measure the evaporation flux perpendicular
to the surface being studied. However, only definition (c) provides a situation which will ensure that $\bar{w} = 0$ and that no horizontal energy flux will be measured. This subtle definition of the direction of the "vertical", based only on the components of the wind vector, is rarely mentioned in the literature, though Hyson et al. (1977) have implicitly used the same definition in their work.

![Diagram showing definitions (a), (b), and (c) of the direction of the "vertical".]

**FIG. 3.10** The Definition of the Direction of the "Vertical"

This definition causes some problems in the field alignment of sensors. The first problem is that the direction of $\bar{U}$ is not initially known. Tests at Warrambine Creek confirmed this. The sensor was initially aligned according to direction (a), using a plumb-bob suspended inside the square steel section of the mast. Not surprisingly, a large apparent $\bar{w}$ was calculated by the microprocessor. Then an alignment parallel to direction (b) was attempted and again a large $\bar{w}$ was calculated. Then it was decided to adjust...
the mast by watching a chart recording of $\tilde{w}$ and trying to visually ensure that $\tilde{w} = 0$. While this method initially produced values of $\tilde{w}$ which were close to zero, the situation changed when the wind direction changed. This then illustrated the second problem with this definition. It appears that for an undulating surface, the direction of $\tilde{U}$ with respect to that surface will change with changes in wind direction. This is due to differences in surface geometry on the approach paths of the wind towards the instrument site. Thus, it appears that for the Warrambine Creek site, a single, rigidly mounted Gill propeller will retain some tilt error in all but one wind direction.

To correct for this, Hyson et al (1977) suggest a method using a cup anemometer in conjunction with the Gill propeller. This system may work satisfactorily, but only field testing will tell. The optimum solution would be to use a three-dimensional array of Gill propellers and then to use the microprocessor to resolve these three signals into another coordinate system defined by $\tilde{w} = 0$ and $\tilde{v} = 0$. A number of propeller arrays exist for this purpose, e.g. the U VW array (Gill (1975)), the 30° UVW array (Christensen (1971)) or the UVR array (Horst (1973)).

In summary, it can be seen that the use of a Gill propeller as the vertical wind sensor is desirable because of its simplicity and reliability. Furthermore, many of its response deficiencies can be corrected for with the introduction of "on-line" corrections by the microprocessor. However, the serious problem of tilt errors at this site can only be overcome completely by the use of a three-dimensional array of sensors. Again, the microprocessor can be used to resolve instantaneously these wind components.

3.4 TEMPERATURE SENSOR

Moore et al (1976) listed the requirements of an eddy correlation temperature sensor as the following. It should:
(a) Be small and have a good spatial resolution;
(b) Have a small thermal inertia and good high frequency response;
(c) Preferably have an output linear with temperature;
(d) Have negligible self-heating and not act as a thermo-anemometer;
(e) Not disturb the wind field;
(f) Not disturb the temperature field;
(g) Have a stable calibration;
(h) Not be affected by rain, wind or air contaminants;
(i) Not be affected by radiation heating or temperature gradients in the air: this requirement is sometimes equivalent to (a);
(j) Preferably be capable of removing the low frequency trend in temperature itself, leaving only the fluctuating part: this is not strictly essential, but may be useful in meeting requirement (c);
(k) Not be easily damaged, or at least be easily replaced.

Of the three commonly used types of sensors, viz. resistance elements, thermistors and thermocouples, Moore et al (1976) found that thermistors are the most commonly used and are probably the best method presently available.

The sensor proposed for this project is a STC P-23 microbead thermistor. In light of the above sensor requirements, the following should be noted.

The sensor itself is extremely small (See Plate 3.3), particularly when compared with the w and q sensors. Therefore, as long as its mounting system is also small, it will not significantly disturb the local wind field. Its small size also means that it has very little thermal inertia and
comparative measurements with other temperature sensors (Verma et al. 1979) indicate that it has an adequate high frequency response.

Even though its resistance varies with temperature according to

$$R = A \exp \left( \frac{B}{T} \right)$$

it can, when placed in a Wheatstone bridge designed by the method cited by Raupach (1977), produce a linear response over a large temperature range. As long as a low bridge voltage is used (Raupach (1977) used 0.09V), little self-heating will occur and thus the device will not modify the local temperature field, nor will its output be affected by windspeed.

PLATE 3.3 The Temperature Sensor - The STC P-23 Micro-Bead Thermistor

Since it is intended to use the thermistor to measure $\bar{T}$ as well as $T'$, it is important to ensure that it is not affected
by radiation heating. Raupach (1977) argues that because of its small size, any radiation heating effect would be small (\(< 0.3^\circ\) C) and the results of Verma et al (1979) seem to confirm this. Fourrier et al (1970), however, found significant radiation heating errors occurred when using this device. They concluded that heat was being conducted to the thermistor from its metal mounting arms. They then proposed an alternate mounting method to reduce this problem.

One shortcoming of this device is that it fails to satisfy requirements (h) and (k) cited above. Field tests at Warrambine Creek showed that the fine wire supports of the bead frequently broke during rain. While they can be easily replaced (See Plate 3.2), this then requires a large supply of calibrated thermistors. Furthermore, the changing of a thermistor requires the changing of calibration factors in the microprocessor programme. This problem also greatly reduces the period for which the system can run unattended. This is one problem which has yet to be resolved.

Again, time constraints prohibited adequate field testing of the device. The few results obtained seemed encouraging, but the sensor's durability seems to be a problem.

3.5 THE HUMIDITY SENSOR

3.5.1 Humidity Sensor Requirements

The routine application of the eddy correlation technique to the direct measurement of evaporation has been restricted due to the lack of a suitable instrument to measure the fluctuating humidity component. The requirements of such a sensor have been summarised by Raupach (1977) and are:

(i) fast response time (\(< 0.1\) s);

(ii) small size to approximate a point measurement;

(iii) aerodynamic shape so as not to interfere with the wind and temperature fields;
(iv) freedom of the humidity signal from cross-talk with other atmospheric variables, e.g. temperature;

(v) freedom of the sensor from contamination by airborne particles, e.g. dust, dew and rain;

(vi) simplicity, reliability, low maintenance and robust construction to facilitate continuous field use;

(vii) a stable, low noise output, preferably linearly related to humidity or humidity fluctuations. (When a microprocessor is used, a linear output is not so important).

To this list can be added the requirement of low power consumption in applications where mains power is not available.

3.5.2 Review of Humidity Sensors

The search for a suitable sensor has been in progress for the last three decades. The method with the longest history is the psychrometric approach, using "fast response" wet and dry bulb thermometers, e.g. Swinbank (1951), Dyer (1961), Dyer & Maher (1965a, 1965b). Such sensors have an unacceptably long response time and require frequent attention to the wicking and water supply. Many inventive and elaborate concepts have also been tried. These have included microwave refractometry, sensitized quartz crystals (Hicks & Goodman (1971)) and barium fluoride films (Goitz et al 1970)). All of these methods have in one way or another failed to satisfy the sensor requirements discussed above.

Recently, most attention has been placed on radiation absorption hygrometry. This method is based on the principle that water vapour strongly absorbs radiation of certain wavelengths. Devices have been constructed using absorption bands in both the ultra-violet (e.g. Wesley & Hicks (1978), Miyake & McBean (1970)) and the infrared (Bogomolova et al (1974), Raupach (1978)) spectra. The performance of these devices has been most promising. Moore et al (1976), when reviewing the presently available humidity sensors, con-
cluded that "radiation absorption devices are likely to be the most useful ... of all existing sensors". With this recommendation in mind, it was decided to build a radiation absorption hygrometer.

3.5.3 The Absorption Bands of Water Vapour

It has been known for many years that water vapour strongly absorbs radiation of specific wavelengths. To a first approximation, it is possible to regard the energy of a given water molecule as the sum of its rotational, vibrational and electronic energies. Transitions between two states give rise to the emission or absorption of radiation equal in energy to the energy difference between the two states. There are three fundamental energy levels due to the vibration of a water molecule and these give rise to emissions at 6.27 μm, 2.74 μm and 2.66 μm (Johns (1965)). As well as these absorption bands, there are a large number of overtones, including 1.87 μm, 1.37 μm, 1.12 μm and 0.94 μm (See Fig. 3.11). Theoretically, any of these absorption bands can be used for radiation absorption hygrometry. For the devices described below, the 2.7 μm band was chosen. This band has a relatively strong absorption (thus allowing a short path length to satisfy requirement (ii) listed above) and, more importantly, suitable infrared sources, filters and detectors are available at this wavelength. Also, glass optical components can be used. Glass is opaque at wavelengths of approximately 3 μm and greater. This precludes the use of glass in a device utilising the strong 6.3 μm absorption band, while making it acceptable at 2.7 μm. The only problem with the 2.7 μm is that in some circumstances the weak CO₂ absorption between 2.65 μm and 2.81 μm may need to be considered.

3.5.4 Theory of Band Absorption

(This section is an abbreviated version of Raupach (1977)).

A beam of perfectly monochromatic radiation will always obey the Beer-Bouguer-Lambert law when passing through an absorb-
Transmission characteristics of infrared filter.

Absorption bands of water vapour.

Fig. 3.11a
ing medium. Thus, the absorption at a single wavenumber, \( \nu \), may be written as

\[
A = 1 - \exp(-k_\nu \, c)
\]

where \( k_\nu \) is the absorption coefficient and \( c \) is the mass of absorbent per unit area along the path of the beam. In principle, it should be possible to calculate the absorption of a band by integrating Eqn. 3.25 over all the frequencies in that band. However, the required functional form of \( k_\nu \) is far too complicated to be accurately measurable. Hence it is necessary to develop some model to enable the absorption of a band to be calculated. For the 2.7 \( \mu \)m band, the random statistical model described by Goody (1964) is suitable. The resulting expression for the band transmission is

\[
\tau = \frac{-S_0 \, c}{d \, (1 + S_0 \, c) \frac{1}{\pi \, \alpha}}
\]

where

- \( \tau \) = band transmission
- \( S_0 \) = parametric line intensity
- \( d \) = mean spacing of the lines within the band
- \( \alpha \) = pressure and temperature dependent factor defined by

\[
\alpha = a_0 \left( \frac{P}{P_0} \right) \left( \frac{T}{T_0} \right)^{\frac{r}{4}}
\]

where

- \( P \) = pressure
- \( T \) = temperature

and the zero subscript denotes some reference state.

The many assumptions inherent in the random statistical model can be justified only by comparing the predicted band transmission with measured transmissions. Since the parameters \( S_0 \), \( d \) and \( a_0 \) in Eqn. 3.26 and Eqn. 3.27 are difficult to infer accurately from measured spectra, Eqn. 3.26 can be regarded as containing two arbitrary coefficients which must be determined by experiment. These are

\[
X = S_0 / d; \quad Y = S_0 / \pi a_0
\]
By collecting together all these equations, the predicted band transmission can be written in a suitable form for experimental testing as

\[ \tau = \exp \frac{-Xc}{1 + Yc \left( \frac{P}{P_o} \right) \left( \frac{T}{T_o} \right)^{\frac{1}{2}} \left( \frac{c}{c_o} \right)^{\frac{1}{2}}} \]  

This equation was tested experimentally by Howard et al (1956). For the 2.7 μm band, their measurements of \( \tau \) as a function of water vapour concentration agree very well with Eqn. 3.29 and lead to the following empirical values for the coefficients \( X \) and \( Y \):

\[ X = 1.97/c_o; \quad Y = 6.56/c_o \]  

where \( c_o \) is the water vapour concentration required to give a transmission of one half at the reference state (\( T = T_o, \ P = P_o \)). The behaviour of \( c_o \) about the 2.7 μm band is shown in Fig. 3.12.

To apply this theory to the infrared hygrometer, it is only necessary to convert the measure of water vapour concentration in the beam, \( c \), to absolute humidity, \( a \). The relationship is

\[ c = a \ell \]  

where \( \ell \) = path length of the beam

The hygrometer transmission function, with the weak temperature and pressure dependence terms omitted, can now be written as

\[ \tau = \exp \frac{-Xa}{(1 + Ya)^{\frac{1}{2}}} \]  

where \( X \) and \( Y \) have been redefined to include the path length, \( \ell \), i.e.

\[ X = S_o \ell /d = 1.97/A_o \]
\[ Y = S_o \ell /\pi a_o = 6.56/A_o \]  

N.B. Raupach (1978) has estimated that the correction due to the temperature and pressure terms on the
Fig 3.11b  ABSORPTION BANDS OF WATER VAPOUR
in the 2.7µm range, with the transmission spectrum of various filters superimposed.

Fig 3.12  RELATIONSHIP BETWEEN $c_0$ and WAVELENGTH.
( after Howard, Burch & Williams (1956) )
measured evaporation flux is "never more than 2% and is often negligible". Thus this dependence will be neglected.

Unfortunately, it is not possible to calculate $X$ and $Y$ from the data of Howard et al (1956), firstly because the $c_o$ data are only graphical and thus do not permit sufficient accuracy, and secondly, because the exact wavelength at which $c_o$ applies for each instrument is unknown. However, Eqn. 3.33 implies that, to within the accuracy of the empirical coefficients,

$$\frac{X}{Y} = \frac{1.97}{6.56} = 0.300$$

This eliminates one of the arbitrary coefficients, $X$ and $Y$, from the transmission function. Thus, the calibration depends only on one parameter, $X$.

It should be pointed out that the strength of the transmission function of each individual hygrometer nominally operating in the 2.7 $\mu$m band can vary considerably from one device to another. This is because each combination of infrared source, filter and detector, produces different and particular wavelength response functions over the 2.6 $\mu$m to 2.8 $\mu$m range. This effectively changes the specific wavelength at which $c_o$ is measured and, as can be seen from Fig. 3.12, $c_o$ can vary almost by an order of magnitude over this range. Results confirming this concept and the consequences of it will be discussed later.

3.5.5 Historical Review of Infrared Hygrometer

The concept of developing a device to use infrared absorption spectra to measure humidity has been applied many times previously, but the performance of each device has always been restricted by the lack of a suitably stable infrared detector and source. To overcome this problem, Foskett et al (1953), Wood (1958), Staats et al (1965), and Togulev et al (1977) all used similar techniques in which the radiation intensity in the absorption band was detected and continuously compared with the radiation intensity in a nearby
non-absorption band. A feedback system was then used to adjust the source intensity to maintain a constant output for the non-absorption band. This procedure compensates for both source and detector instability. This approach always leads to a system of complex double-beam optics and/or rotating multiple filters, which results in a bulky instrument with a large power consumption. This means that these devices are generally unsuitable for field use, although Yelagina et al (1970) and Bogomolova et al (1974) used double-beam hygrometers for field evaporation studies.

More recently, Hyson & Hicks (1975) introduced a single beam device specifically for eddy correlation studies. They recognised that only $q'$, the fluctuating component of the humidity signal, is required for eddy correlation and that the long term mean component can be ignored. Because of the exponential form of the transmission function, it can be shown that if the output of a device is proportional to $r'/r$, then it is also proportional to $q'$. That is, for their instrument,

$$V = K \frac{r'}{r}$$

$$= k q'$$

where $V =$ output voltage of the device
$r =$ transmission of the beam
$q =$ specific humidity
$K,k =$ calibration factors

and the prime denotes the fluctuating component (See Eqn. 3.3).

There is in fact a weak dependence on the mean humidity, but this is dependent on the specific transmission function. This dependence is negligible for weak transmission functions, i.e. when $X$ is small.

Because of the encouraging results obtained from field studies using this type of device (e.g. Hicks et al (1975), Raupach (1977)), it was decided to build an infrared hygrometer based on this principle.
3.5.6 The Infrared Hygrometer - Model #1

The device described below was built following discussions with P. Hyson of the CSIRO Division of Atmospheric Physics. Essentially, it was a duplication of their current infrared hygrometer, which itself was a development of the device described by Hyson & Hicks (1975). The only significant difference was that an Optoelectronics OE-25-5 detector was used instead of a Mullard 62SV. The physical layout of this device is shown in Fig. 3.13 and the circuitry is shown in Fig. 3.14. A beam of radiation from a 12V torch bulb was chopped at 330 Hz and passed through an 80 mm atmospheric path to the detector. Two planar-convex lenses were used; one to form the beam and the other to focus the beam onto the detector. A band-pass interference filter, whose characteristics are shown in Fig. 3.11, was placed immediately in front of the detector to ensure that only radiation of the appropriate wavelength was measured. The output of this device is proportional to $\tau'/\tau$, which is proportional to $q'$.

Calibration of this device is difficult. Raupach (1977) calibrated his device by placing the whole instrument in a chamber. He slowly changed the humidity inside this chamber while maintaining it at a constant (± 0.05°C) temperature. He then plotted humidity against the voltage measured at point G in Fig. 3.14 in order to determine $X$. While producing good results, this method requires specialised equipment and is time consuming. This method does highlight the extreme temperature dependence of these PbS detectors.

Since the equipment necessary for the above type of calibration was not available, it was decided to calibrate the output of the device directly against $q'$. This was done using the system shown in Fig. 3.15. Two high flow rate air streams, one wet and one dry, were mixed in reproducible proportions in a valve similar to that described by Vaisala (1965). This mixture was then passed via the sensing path of the hygrometer to a wet and dry bulb psychrometer. By repeatedly rotating the valve from one set position to
Fig 3.13  PHYSICAL CONFIGURATION
MODEL NO.1  HYGROMETER
Fig 3.14 DETECTOR CIRCUIT - MODEL NO.1 HYGROMETER
Fig. 3.15

CALIBRATION SYSTEM - MODEL NO.1 HYGROMETER
another and then back again after 30s, it was possible to change the humidity measured by the hygrometer rapidly from \( q_1 \) to \( q_2 \) and back again. This changeover took less than 1s. By using the psychrometer to measure \( q_1 \) and \( q_2 \) before and after the test, the repeated fluctuations in humidity, \( q' = q_1 - q_2 \), could be plotted against the change in output voltage, \( \Delta V \). A typical calibration run is shown in Fig. 3.16, where \( q_1 = 7.2 \text{ g/kg} \) and \( q_2 = 3.0 \text{ g/kg} \).

While this calibration system worked well, the results obtained were most unsatisfactory. This occurred because the high noise level of the hygrometer (See Fig. 3.16) made the exact determination of \( \Delta V \) difficult. Thus considerable scatter existed in the final calibration plot when the data from several runs was collated. The noise level was estimated to be approximately equivalent to a humidity fluctuation of 0.7 g/kg and this is much larger than the equivalent levels reported by Hyson & Hicks (1975) and Raupach (1977). It is theoretically true that this noise should not cause errors in the computed energy fluxes because it is random and should not correlate with  \( w \). However, it did mean that a satisfactory calibration curve could not be obtained. Furthermore, an essential design feature of these devices is that freedom from drift must be ensured below the lowest contributing eddy frequency (\(< 0.0005 \text{ Hz}\)) (Raupach (1978)). The main instrumental factor which could cause drift was the responsivity of the detector which is very strongly temperature dependent. In an attempt to overcome this, the detector was temperature lagged by mounting it in a block of brass. Tests revealed that the time constant for response to temperature changes was much less than the 35 minutes suggested above.

While it was something of a mystery as to why this hygrometer should perform so badly when compared with apparently similar devices, it was evident that the whole hygrometer should be completely redesigned and rebuilt. (The main reason for this poor performance was only discovered many months later and will be explained in the following sections.)
Note large background noise level and high-pass filter decay.

$q' = 4.3 \text{g/kg}$

Fig 3.16  TYPICAL CALIBRATION RUN - MODEL NO.1 HYGROMETER
This decision to redesign occurred concurrently with the decision to replace the analogue Fluxatron system with a microprocessor. Since the theory involved in using the microprocessor (See Section 3.2) eliminates the procedure of separating the mean and fluctuating components of the humidity signal, the concept of designing a device to measure $q$ and not $q'$ was born. As a further incentive, the use of a microprocessor also meant that the exponential (i.e. non-linear) relationship between output voltage and humidity was no longer a problem.

While recognizing that the design of a single beam hygrometer for field use which could measure absolute values of humidity had not been successfully achieved elsewhere to date, it was felt that a fresh attempt at the problem was warranted. This was because most of the instability in the device appeared to be due to the temperature dependence of the detector and that new detectors were available which apparently overcame this problem. In any event, it was always recognized that if the objective of long-term stability was not achieved, then the re-introduction of the $\tau'/\tau$ divider circuit (or a digital equivalent thereof) would then still produce a hygrometer suitable for eddy correlation studies.

The specific aims of the new design were to

(i) eliminate the temperature dependence of the detector;
(ii) increase the sensitivity of the device by increasing its path length;
(iii) improve the signal demodulation technique;
(iv) reduce the noise level of the bias voltage to the detector (since any noise here is seen as a humidity fluctuation and is subsequently greatly amplified);
(v) reduce random noise in the whole circuit by reducing supply voltage noise, 50 Hz noise, earth loops, etc.;
(vi) stabilise the output of the lamp and the speed of the chopper.
3.5.7 The Infrared Hygrometer - Model #2

(i) Physical description: Essentially, the physical configuration of this model (See Fig. 3.17) is similar to the earlier design. The path length has been extended to 200 mm. Kaimal (1968) has shown that this path length is still equivalent to a point measurement for all practical purposes. A larger and more robust chopper motor has been used and the beam is now chopped at 100 Hz. Several modifications were made to the chopper system to produce an off-on light beam whose output closely resembled a square-wave. This physical improvement in the design greatly reduced electronic noise at the chopping frequency after demodulation. At present, the squareness of the output wave from the detector is only limited by the response time of the detector.

A significant improvement to the stability of the device comes from the introduction of the beam diffuser; in this case, two layers of tracing paper. In the first model, the filament of the bulb was placed exactly at the focus of the lens. This was done by adjusting screws specially provided for this purpose. Similarly, at the detector end, the image of the filament was carefully focused onto the detecting surface. This arrangement produced a sharp, intense image of the filament (~1 mm x 3 mm) on the detecting surface. As this surface is 2 mm x 2 mm, the whole of the surface was not evenly irradiated. It was found experimentally (and later in the literature) that these detecting surfaces typically have a highly non-homogeneous spot-to-spot response to incident radiation. Consequently, even the slightest movement (<0.1 mm) of the source assembly relative to the detector would cause movement of the image across the detecting surface and this would cause gross changes in the output of the detector. This problem was overcome by illuminating the whole detecting surface and surrounding area with a spot of even intensity diffuse radiation. This technique has a second desirable outcome. It means that the bulb itself no longer needs to be placed at the focus of the lens (since this is where the diffuser is) and thus can
Fig 3.17  PHYSICAL CONFIGURATION - MODEL NO.2 HYGROMETER
be placed anywhere in the cavity behind the diffuser. This allows a more convenient configuration of lamp and motor and provides the opportunity of using larger or different types of bulbs.

(ii) **Electronic description:** The detector used in this model (Optoelectronics OTC-21-52T) is the same size and has the same detecting surface as the OE-25-5 detector. The manufacturers have recognised that temperature instability is inherent in this type of device and have overcome this by placing a Peltier cooling cell and a thermistor behind the detecting surface. This thermistor is incorporated in a bridge circuit which is used in a feedback system which drives the Peltier cell (See Fig. 3.18). The manufacturers claim to be able to maintain the temperature of the detecting surface to ±0.1°C (Optoelectronics (1978)). The temperature of the detecting surface can be changed by varying the set-point resistor in Fig. 3.18. The effects of a colder surface temperature are increased responsivity to incident radiation (which is desirable), but also increased power consumption and slower response time for the detector. Also, the power output of the cooler is limited and thus temperature instability can still occur when the difference between ambient and detector temperatures becomes too large.

In order to measure $\tau$, the transmission of the light beam, the detector is placed in a bridge circuit, the output of which is demodulated using a sample-and-hold technique (See Fig. 3.19). It was found by experiment that this method of demodulation produced significantly less noise at the chopping frequency than either the method of Hyson & Hicks (1975) or Raupach (1977). This system sampled the "light on" section of the output of the bridge (See Fig. 3.20) and held this level as output for one cycle. Any remnant noise at the chopping frequency (or higher) was removed with the low-pass smoothing filter. This means that at point B in Fig. 3.19, the voltage is a dc level directly proportional to the transmission, $\tau$, of the light beam and this transmission is related to humidity by Eqn. 3.32. This is shown
Fig. 3.19
DETECTOR CIRCUIT - MODEL NO.2 HYGROMETER
as Curve B in Fig. 3.21. To convert this to a more useful form, the final stage biases (Curve C), amplifies (Curve D) and inverts the signal to produce the final output of voltage vs. humidity (Curve E). Thus, the output is related to humidity by

\[ V = -g V_o \exp \left( -\frac{X}{1 + Y} \right) a^{-\frac{a}{2}} + b \]  

where
- \( V \) = output voltage
- \( g \) = gain of final stage
- \( b \) = bias voltage
- \( V_o \) = voltage at Point B corresponding to \( q = 0 \) g/kg

Variable resistors are provided to adjust \( V_o \), \( g \) and \( b \), and two of these terms (\( g, b \)) can be determined electronically. Optimally, it would be desirable to adjust \( b \), such that when \( q = 0 \), the \( V = 0 \), but this is somewhat difficult.

Details of the detector bias voltage supply is given in Fig. 3.22.

3.5.8 Calibration

The calibration system used for the second model is shown in Fig. 3.23. It is a 1 m³ insulated wooden box in which was placed the infrared hygrometer, a dew-point psychrometer, a heater and a fan to thoroughly mix the internal air. Inlet pipes and an external air pump allowed wet and dry air to be introduced into the chamber. This air could come from various sources, e.g. dry air cylinder, cold storeroom (~ 2°C), and above a heated water bath. This system could maintain internal air temperatures from 5°C to 45°C and humidities from 3.5 g/kg to 25 g/kg.

During a calibration run, which could take less than 2 hours, the internal air temperature and dew-point were continuously monitored, as were the voltages at points B and E in Fig. 3.19. The results of one calibration run are shown in Fig. 3.24.

During this test, the internal air temperature varied from 29.7°C to 33.5°C, and the humidity from 4.3 g/m³ to 21.2 g/m³.
Fig 3.21 OUTPUT CURVES - MODEL NO. 2 HYGROMETER
CIRCUIT FOR DETECTOR BIAS VOLTAGE
MODEL NO. 2 HYGROMETER

Fig 3.22
Fig 3.23 CALIBRATION SYSTEM FOR MODEL NO. 2 HYGROMETER
Fig 3.24  CALIBRATION CURVE - MODEL NO. 2 HYGROMETER

Transmission Curve (Eqn. 3.37) Where
\[ X = 0.0033 \text{ m}^3/\text{g} \]
\[ V_0 = 9.76 \text{ V} \]
\[ g = -6.26 \]
\[ b = 63.06 \]

High points due to condensation on lens.
The points marked + need explanation. These points were taken early in the test. At this time, the temperature of the body of the hygrometer was less than the chamber temperature. Thus when moist air was added to the chamber, condensation occurred on the lenses, thus decreasing the transmission of the light beam and increasing the output voltage. By flushing the chamber with dry air and allowing a longer equilibrium time, this effect was eliminated.

In order to determine the calibration curve shown in Fig. 3.24, $V_E$ was regressed against $V_B$ to determine $g$ and $b$ in Eqn. 3.37. Then $V_B$ and $a$ were plotted (See Fig. 3.24) and $V_0$ and $X$ were determined by a least squares analysis. The value of $X$ determined for this device was

$$X = 3.33 \text{ m}^3 \text{ kg}^{-1}$$

This is approximately one third of the value obtained by Raupach (1977) (i.e. 9.10 m$^3$ kg$^{-1}$). Consequently, when the transmission function for this device is plotted against the functions obtained by Raupach (1977) and Hyson & Hicks (1975) (See Fig. 3.25), it is evident that this device has a much weaker absorption function. This occurs because of the combination of source, filter and detector used here determines a different $c_0$ value (See Section 3.5.4) than for the other instruments.

This has two effects. Firstly, because the variation in transmission is so low, then large signal amplification is necessary to obtain reasonable sensitivity. This is seen as the reason for the poor performance of the first model when compared to the apparently identical Hyson-Hicks device. Because the absorption is approximately three times less, then three times as much electronic gain is necessary and consequently the noise (which primarily originates from the detector) is also three times greater. On the positive side, a weak absorption function produces a calibration of $V_E$ vs. $q$, which is very nearly linear. This is an advantage, particularly when analogue covariances systems are used.
Fig 3.25  TRANSMISSION CURVES FOR VARIOUS HYGROMETERS
Whether a strong or weak absorption function is desirable depends largely on the device's intended application. When it is intended to build a fluctuation hygrometer for an analogue system, a weak absorption function is desirable. This is because the result shown in Eqn. 3.35 and Eqn. 3.36 has a weak dependence on absolute humidity, which is amplified by strong absorption functions (Raupach (1978)). Thus the weak, near-linear function is desirable, yet this causes noise and calibration problems. For a digital system where \( \bar{q} \) as well as \( q' \) is required, a strong absorption function is required. The non-linearity of this function is not an issue with the digital system, but the resulting decrease in noise and increase in stability is highly desirable. Clearly this is an aspect of infrared hygrometer design which has yet to be clearly understood and yet has a large bearing on the device's performance.

3.6 SUMMARY

The eddy correlation technique is the most fundamental method of measuring evaporation, but its routine application has been hindered by a lack of suitable sensors and covariance computers. For adequate frequency response, it has become evident that a radiation-absorption hygrometer is necessary. Therefore, a fluctuation infrared hygrometer was built for this project. This device did not operate successfully, but further understanding of the nature and variation of the transmission function has been obtained. Explicit selection of the strength of this function is seen as desirable for future devices and this selection is dependent on its intended use.

A review of the vertical wind theory (See Section 3.3 (iv)) has revealed that a three dimensional wind sensor array is necessary at the Warrambine Creek site in order to avoid the measurement of the horizontal transport of heat and water vapour. The resolution of these three dimensional wind components can only be adequately performed with the introduction of a microprocessor based covariance computer.
The introduction of this device offers further advantages, namely:

(i) Correction of sensor deficiencies
(ii) Elimination of low frequency losses in flux computation
(iii) The use of non-linear sensors
(iv) Further data output (e.g. \( \bar{q}, \bar{T}, \bar{u} \))
(v) Accurate flux computations
(vi) Enhanced system flexibility

It is seen that even though developmental problems still exist, the use of microprocessors will soon become commonplace in eddy correlation systems.
4.0 INTRODUCTION

4.1 A.W.R.C. DATA REQUIREMENTS

4.2 CLIMATE STATION

4.2.1 Location of the Climate Station
4.2.2 Maintenance of the Climate Station

4.3 COMPARISONS OF UM AND BM DATA

4.3.1 Temperature Data
4.3.2 Rainfall
4.3.3 Wind Run
4.3.4 Pan Evaporation
4.3.5 Global Radiation Data

4.5 CONCLUSIONS
CHAPTER 4

INPUT DATA FOR ARBM

4.0 INTRODUCTION

The primary aim of the representative basins project is to make accurate predictions of streamflow (runoff) from various catchments throughout Australia. This is done using the Australian Representative Basins Model (ARBM) which uses as input, climatological data collected in each catchment. Accurate predictions of streamflow can only be made if:

(i) the modelling process is "correct"; and
(ii) the input data are reliable and accurate.

The modelling process will be examined in Chapter 5 but first, the quality of the input data will be examined; particularly that data required for the estimation of evaporation. The A.W.R.C. Advisory Panel, when setting up the representative basin project, stated that, to estimate evaporation accurately "the model will require high grade climatic data from at least one site in or near each representative basin". (A.W.R.C (1969)) Furthermore, the Panel went on to say that "it is essential for each representative basin to be instrumented to a certain minimum level in terms of both the elements sampled and the order of accuracy of the readings. Unless such standards are set much of the advantage to be gained from the establishment of these basins could be lost owing to unreliable data". (A.W.R.C. (1969))

It is the aim of this chapter to examine critically the climatic data collected in the Warrambine Creek basin and to ascertain whether it can be considered as "high grade climatic data". This examination has two distinct aspects;

(i) the representativeness of the data
(ii) the reliability and accuracy of the data
This examination of the quality of the data collected at the climate station was not included in the original concept of this project. However, most of the parameters measured at the climate station were also measured at the University of Melbourne site as part of the evaporation measurement programme and comparisons were initially made only to verify the University of Melbourne data. This comparison proved enlightening and, later, essential to the success of the model testing. For convenience, the University of Melbourne and Bureau of Meteorology data will be referred to from now on as UM and BM respectively.

4.1 A.W.R.C. DATA REQUIREMENTS

The A.W.R.C. lists the climatic data it requires in two categories:

(i) primary measurements and

(ii) secondary measurements

Primary measurements are the minimum requirements of the project, while secondary measurements are highly desirable, but not essential. The following is an abridged listing of A.W.R.C. requirements. (A.W.R.C. (1969))

(i) **Primary measurements**

*Temperature and Humidity*

Both wet and dry bulb temperatures should be observed hourly and this will also allow the derivation of humidity properties. In the absence of continuous wet bulb recordings, then relative humidity or dew point may be substituted.

*Wind*

Run of wind at height of two metres is the minimum requirement. Wind direction and velocity averaged over hourly intervals would be classified as secondary but desirable data.

*Sunshine*

Hours of bright sunshine can be used in estimation formulae to derive the radiation balance of the catchment. Depending
on the other equipment provided at the site, it may be more practical to measure Total Global Solar Radiation with a photometric integrator or even an integrating solarimeter.

(ii) Secondary measurements

Radiation

A desirable refinement of radiation estimation would be the recording of hourly integrals of Total Global Solar Radiation and net radiation over a typical sample of catchment vegetation.

Evaporation (pan)

Pan evaporation offers a fairly simple integration of many climatological factors. It may be used to provide a crude estimate of evaporation from the catchment, but primarily provides a useful check on other more refined estimates of evaporation.

Weather Remarks

The classification of weather situations in terms of visual observations can be most useful in guiding the analysis of instrumental data. The operation of the climatological site by an observer making regular observations is a desirable level of sophistication.

4.2 CLIMATE STATION

In order to provide the above data, the Bureau of Meteorology established in 1972 a climate station and three daily rainfall stations within the catchment. The climate station is the Warrambine Basin No.3 (No. 89094; 37°50'S, 143°53'E, Elevation 311.0 m) and is situated in the north-east corner of the catchment (See Fig. 1.1).

The data recorded at this station are:

(i) Daily rainfall
(ii) 30-minute rainfall (pluviometer)
(iii) Daily Class 'A' pan evaporation
(iv) Daily maximum temperature  
(v) Daily minimum temperature  
(vi) Dry bulb temperature at 0900 & 1500 hours  
(vii) Wet bulb temperature at 0900 & 1500 hours  
(viii) Total global radiation (since Sept. 1978)  
(ix) Wind run at 2 m (0900 to 0900)  
(x) Wind speed and direction at 0900 and 1500 hours  
(xi) Total amount of cloud coverage at 0900 and 1500 hours  

N.B. During daylight saving (i.e. last Saturday in October to first Saturday in March), measurements are made at 1000 and 1600 daylight saving time i.e. 0900 and 1500 E.S.T..

In addition to the above, observations of visibility and the occurrence of frost, fog, haze, mist and thunderstorms are made; maximum and minimum temperatures in the evaporation pan are measured; and past and present weather is coded in the international WW code.

4.2.1 Location of the Climate Station

Special mention must be made of the siting of this climate station. When selecting a site in 1972, the Bureau of Meteorology presumably had two major prerequisites. Firstly, the site should be within or close to the catchment. Secondly, somebody must be in attendance daily at 0900 and 1500 hours to make the necessary observations. One of the farmers who lived within the catchment area would be the logical choice for this person. However, as there are only nine farmers who fall into this category, the choice was fairly limited and furthermore, many farmers would be unwilling to accept such a time consuming obligation. Thus the site finally chosen for the climate station is not ideal.

The instrument enclosure is situated in a clearing between a homestead and some machinery sheds. When this particular
property was established over 100 years ago, the homestead was wisely sited on the leeward side (i.e. eastward) of one of the few hills in the area. Although this hill (Mt. Lawaluk) is only a small vent of the nearby extinct volcano, Mt. Mercer (See Fig. 1.1), it still acts as a substantial obstruction to the predominant westerly wind. Furthermore, a grove of trees has been planted around the house and surrounding buildings to provide additional shelter. Consequently, the climate station is situated in what appears to be a most unrepresentative site with respect to the rest of the catchment.

In fairness, it should be noted that the immediate site satisfies the normal exposure criteria for a meteorological enclosure. For example, for an evaporation pan, the site requirements are as follows (Bureau of Meteorology (1979)):

(i) There are no obstructions which will cast shadows on the pan.

(ii) The distance of the pan from objects which are higher than the top of the pan is not less than twice and preferably not less than four times the height of the object above the rim of the pan.

(iii) The pan is not closer than 1.5 m to any other object which is lower than the top of the pan.

That is, the site is suitable according to usual Bureau of Meteorology criteria but it may not be suitable to A.W.R.C. criteria. The farmer who records the climate data has stated that the wind is significantly reduced in the immediate area and therefore other parameters are likely to be affected. Since A.W.R.C. (1970(b)) points out that "instrument precision, such as would apply to a tank evaporimeter, for example, is quite meaningless in terms of natural evaporation, if site representativeness is not satisfactory", it is important that this aspect be examined. Data presented in later sections will demonstrate the representativeness of the site.
PLATE 4.1 Climate station (centre) looking west from the access road. Compare the vegetation density here with Plates 1.2, 1.3 and 1.4.

PLATE 4.2 General view of the climate station with the farm buildings and Mt. Lauwak in the background.
Mt. Lawaluk is shown in Plate 1.3. The climate station is located in the clump of trees immediately to the right (east) of the hill. Plate 4.1 shows the site from the access road to the farm and Plates 4.2 and 4.3 show the actual enclosure.

4.2.2 Maintenance of the Climate Station

Plate 4.3 demonstrates the second problem associated with this site, i.e. proper maintenance. When the instruments were installed, a wire fence was erected around the site in accordance with usual practice. This has long since fallen or been knocked down, thus allowing dogs and sheep to wander freely in and out of the enclosure. This presumably has little effect on the wind and temperature measurements, but interference is possible with the rain gauge and evaporation pan. The water in the pan has frequently had a yellow-green
colour, (See Plate 4.4), indicating algal growth caused by some form of pollutant. This must affect the results obtained and is symptomatic of the generally non-scientific, non-meticulous approach of the farmer to his task. This comment is not meant to be a criticism of the farmer. It only points out that when a person with no scientific background is employed to use scientific instruments, his lack of understanding of important aspects of the instruments and their operation can lead to poor data being collected. For example, it is likely that he will correctly read the numbers on the wind run counter but where some skill is required in a measurement (e.g. measuring and refilling the evaporation pan), the result is more likely to be in error. This makes the value of some, but not all, of the data questionable.

4.3 COMPARISONS OF UM & BM DATA

4.3.1 Temperature Data

Comparisons of all the various temperature parameters are shown in Fig. 4.1 to Fig. 4.5. These are daily values of the UM and BM data plotted against each other for the period October 1979 to February 1980. The regression equations for these are given in Table 4.1. In each case, the BM data was measured with a mercury-in-glass thermometer to the nearest 0.5°C. The UM data was taken from measurements made by the Bowen ratio system described in Section 2.3 and these are taken to be accurate to ±0.1°C. In the case of the maximum and minimum temperatures, the UM data is a half-hourly average and thus probably slightly underestimates and overestimates respectively the time value. In the case of the 0900 and 1500 wet and dry bulb data, the UM data again is a half-hourly average, as distinct from the instantaneous BM reading.

An alternative representation of the correlation between the two data sets is shown in Fig. 4.6. The differences between the UM and BM data are plotted as a frequency diagram. In all cases, these differences appear to be normally distri-
PLATE 4.4 Bureau of Meteorology evaporation pan. Note the colour of the water.

PLATE 4.5 University of Melbourne evaporation pan with automatic water level mechanism. This pan has no enclosure around it so interference by stock is possible.
Fig 4.1
MAXIMUM DAILY TEMPERATURE

See Table 4.2 for details of these data.
See Table 4.2 for details of these data.
Least squares linear regressions

0900  \( y = 0.078 + 0.880 \times \quad r^2 = 0.883 \)

1500  \( y = -0.348 + 0.901 \times \quad r^2 = 0.874 \)

Fig 4.5  DAILY WET BULB DEPRESSION
Fig 4.6a  Daily Maximum Temperature

Fig 4.6b  Daily Minimum Temperature

Fig 4.6  FREQUENCY PLOT OF (BM-UM) TEMPERATURE DIFFERENCES
Fig 4.6c Daily Dry Bulb Temperature

Fig 4.6d Daily Wet Bulb Temperature

Fig 4.6 FREQUENCY PLOT OF (BM-UM) TEMPERATURE DIFFERENCES
Fig 4.6e  Daily Wet Bulb Depression

Fig 4.6  FREQUENCY PLOT OF (BM-UM)
TEMPERATURE DIFFERENCES
TABLE 4.1
STATISTICAL ANALYSIS OF UM AND BM
TEMPERATURE DATA

<table>
<thead>
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<th>PROPERTY (SEE BELOW)</th>
<th>MAXIMUM</th>
<th>MINIMUM</th>
<th>DRY BULB</th>
<th>WET BULB</th>
<th>DEPRESSION</th>
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<td>85</td>
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</table>


BM = X
UM = Y

N = Total no. of data points
r² = Correlation coefficient

of Y = a + b.X

x = Mean of X - Y
s = Standard Deviation of X - Y
(See Fig. 4.6)
buted and, except for the wet bulb temperatures, the average difference is less than 0.9°C. This indicates that there is no significant difference between the UM and BM dry bulb temperatures.

For the wet bulb temperatures, Fig. 4.4 appears to show a difference between the UM and BM data. As will be shown in Chapter 5, an important parameter in evaporation estimation is the wet bulb depression (i.e. T-Tw) and when this parameter is plotted (Fig. 4.5), it is evident that the BM data underestimates the UM data by, on average, 0.5°C to 1.0°C. There are three possible reasons for this. Firstly, because the BM thermometers are mounted in a Stevenson screen, the wet bulb may not always be sufficiently aspirated and thus the full wet bulb depression is not reached. Secondly, if distilled water is not always used or if the wicks are not frequently changed then errors may occur in the BM data. Thirdly, the surrounding vegetation and buildings may cause the local environment of the climate station to be slightly more humid than the surrounding plains, thus resulting in a smaller wet bulb depression. Whatever the cause, the BM wet bulb depression is less than the UM data which is considered to be better maintained and aspirated, and more representative of the catchment. Thus, when using BM data in the model, a suitable correction should be applied.

An interesting feature of the data was that on a few occasions the BM and UM data differed significantly. The cause of some of these differences is evident from an examination of Table 4.2. When compared with all the other UM and BM data for the particular day, it appears that one BM reading is in error by 10°C. It is likely that either the farmer misread the thermometer or made an incorrect entry into the field record book. The Bureau of Meteorology (Bureau of Meteorology (1979)) places this first in its list of possible sources of error in temperature measurements. Alternatively, an error or incorrect "correction" may have been made when transposing the data from the field book to computer file. Whatever the reason, it is unlikely that this type of error will be restricted to the temperature
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<th>DATE</th>
<th>MAX. TEMPERATURE</th>
<th>MIN. TEMPERATURE</th>
<th>0900 TEMPERATURE</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>UM</td>
<td>BM</td>
<td>UM</td>
</tr>
<tr>
<td>31/10/1979</td>
<td>19.2</td>
<td>29.0</td>
<td>7.6</td>
</tr>
<tr>
<td>4/11/1979</td>
<td>27.0</td>
<td>26.0</td>
<td>15.7</td>
</tr>
<tr>
<td>10/11/1979</td>
<td>27.0</td>
<td>27.5</td>
<td>14.9</td>
</tr>
<tr>
<td>16/12/1979</td>
<td>18.0</td>
<td>29.5</td>
<td>6.5</td>
</tr>
<tr>
<td>3/01/1980</td>
<td>32.1</td>
<td>33.0</td>
<td>17.6</td>
</tr>
</tbody>
</table>

**TABLE 4.2 Examples of large differences between UM and BM temperature data**

data alone. Thus all BM data which, is recorded by hand must be expected to exhibit these spasmodic aberrations. On other days, the BM and UM data differed by 4°C to 8°C. The cause for this is unknown; it may be due to the maximum and minimum thermometers not being reset correctly; it may be due to local differences in micro-climate or it may be a 5°C reading or recording error. Surprisingly, the above type of error was not present in the 0900 and 1500 readings (See Fig. 4.3 and Fig. 4.4).

### 4.3.2 Rainfall

Although Williamson & Turner (1980) have made a study of the rainfall patterns and intensities throughout the catchment, a brief comparison is made here of the data collected at the UM and BM site. This comparison can provide further evidence of the quality and representativeness of the BM data.

Each site has a bulk gauge and a pluviometer. At the BM site, the bulk gauge is inspected daily at 0900 and 1500 and the pluviometer has its chart replaced daily at 0900. On those occasions when the pluviometer record is unavailable, the rain recorded by the bulk gauge is considered to have fallen evenly over the preceding time period. The UM pluviometer is a RIMCO 8", 0.2 mm tipping bucket pluviometer.
connected to a Summer Long Period chart recorder. This recorder has been modified to record at 2-minute intervals. It is also used to record pan evaporation. On every visit to the UM site, a record was made of the rain in the bulk gauge and the corresponding pluviometer count. The time interval between visits to the site varied between 2 to 10 days, depending on the maintenance requirements of the other equipment. The total rainfall between visits recorded by each instrument is shown in Fig. 4.7. This figure shows that the bulk gauge consistently underestimates the pluviometer totals by up to 20%. This is consistent with the data collected by Williamson & Turner (1980). The calibration of both devices was checked and no errors were found. Williamson attributes this difference to evaporation from the bulk gauge. Since the pluviometer records and discards rain as it occurs, only a small amount of evaporation (the volume of one partially filled bucket per storm, i.e. < 0.2 mm) is possible. Thus, the pluviometer data is considered to be the true representation of rainfall.

The relationship between the UM pluviometer and the BM bulk gauge is shown in Fig. 4.8. Since the BM data represents the 0900 to 0900 totals, while the UM data was totalled for midnight to midnight, some storms apparently occur on different days. However, the difference between the cumulative totals (i.e. 143.8 mm, 148.3 mm) is not significant over the 100 day period.

4.3.3 Wind Run

Wind run is an important parameter in some evaporation estimation formulae but it is not a parameter which is required for the measurement of evaporation, either with the Bowen ratio or Eddy Correlation methods. Hence, a wind speed recorder was not installed in the early part of the project. It was intended that the BM data be used in the model testing but, after an examination of the BM site and an analysis of some of the data collected there, it became apparent that in all probability, the BM wind data would be unrepresentative of the catchment as a whole. In order to confirm this
Least squares linear regression

\[ y = -0.589 + 0.785 \times \quad r^2 = 0.995 \]

**Fig 4.7  PLOT OF UM RAINFALL DATA**
Fig 4.8  CUMULATIVE RAINFALL DATA

110 days data from 22/10/1979 to 6/2/1980
hypothesis, a wind speed and direction recorder similar to that described by Sumner (1965) has been installed at the UM site. Unfortunately, it took some time for the recorder to become fully operational and no data is available yet.

4.3.4 Pan Evaporation

Evaporation is measured at the BM site with a Class 'A' evaporation pan (See Plate 4.4). This is observed daily at 0900. At the UM site, evaporation is measured with a continuously recording Class 'A' pan similar to that described by Sumner (1963). (See Plate 4.5) This data is recorded on a Sumner Long Period Recorder. Williamson (1979) installed this pan in June, 1976. Unfortunately, for the 1979-80 period, the automatic float valve mechanism which maintains the level of water in the pan was malfunctioning and consequently no useful data was collected. However, Williamson (1979) recorded pan evaporation for two years and a comparison of his data and the BM data is shown in Fig. 4.9.

Two conclusions can be drawn from Fig. 4.9. Firstly, on average, the BM is 20% less than the UM data. This is almost certainly due to the sheltered (and therefore, unrepresentative) siting of the BM pan. This data is consistent with the findings of Hanson & Rauzi (1977) who showed that sheltered sites can reduce pan evaporation significantly.

Secondly, Fig. 4.9 shows a large amount of scatter ($r^2 = 0.6$) between the two sites. This compares poorly with the correlations of rainfall and temperature ($r^2 = 0.95$). The reason for this is twofold. Firstly, as distinct from wind speed or temperature, some skill is required in making an observation at the BM site and thus some human error will be present in the data. The second source of the scatter originates from the nature of the instrument itself. It is well known that a number of factors, other than the weather, affect the performance of an evaporation pan. These include the colour of the pan; the turbidity and salinity of the water (See Plate 4.4); ventilation under and around the pan; shading of the top and sides of the pan; thermal character-
Fig 4.9  DAILY EVAPORATION FOR TWO PANS 3km APART

taken from Williamson (1979)
istics of the underlying soil and the presence of a bird guard. Also, accurate compensation for rainfall is necessary. Also, at the UM site, winds have been observed whipping up waves in the pan sufficient to cause overtopping.

Thus, it is not surprising that two pans, sites nearby, but maintained by different observers can record different amounts of evaporation, due solely to non-meteorological influences. Fig. 4.9 demonstrates this clearly. The consequences of this finding with respect to the testing of the model will be discussed in Chapter 5.

4.3.5 Global Radiation

Because of the significant correlation between the UM and BM rainfall and temperature data, it was expected that the correlation of global radiation data would be similar. When the data for the period from October 1979 to January 1980 were plotted, the results were most disappointing. (See Fig. 4.10) Obviously, one or both sets of data was in error. In order to isolate the problem, data was obtained from a third site. This was the CSIRO Division of Atmospheric Physics, at Aspendale which is 100 km east of the UM site and on the same latitude. The correlation between the UM and CSIRO data is shown in Fig. 4.11. In Fig. 4.12 data for all three sites for a part of the investigation period are plotted against time. These graphs, particularly Fig. 4.12, demonstrate the BM data to be erroneous. The BM instrument was an old model RIMCO integrating pyranometer (See Plate 4.6) installed in September 1978 (Bureau of Meteorology 1979). After informing the Bureau of the apparent problem, the instrument was examined and found to be faulty. It has subsequently been replaced. Apparently, the Bureau's staff had previously confirmed the performance of the device by checking the maximum and minimum values recorded each month. If these fell within reasonable limits (and they apparently did), the instrument was deemed to be operating correctly. The farmer who took the readings had no idea how the device should respond and was consequently unaware of any problem. A trained observer might have
noticed a problem. In the end, several months of useless data were collected. This discovery was a great disappointment as these data had been intended to be the basis of the evaporation prediction model.

PLATE 4.6 Climate station solarimeter. This device was not operating properly for the whole 1979-80 season.

4.5 CONCLUSIONS

After trying to ascertain whether the data collected at the climate station can be considered as representative of the catchment and as "reliable and high grade", the following conclusions can be drawn.

(a) All the primary climatological measurements and most of the secondary measurements recommended by the A.W.R.C.
advisory panel are measured at this climate station. One notable exception, especially with respect to evaporation modelling, is the absence of net radiation. This parameter is not usually measured at climate stations and, as will be seen in Chapter 5, good estimates of net radiation can be made from global radiation data, once a field correlation has been obtained. Also, soil temperature, atmospheric pressure and sunshine hours are not recorded.

(b) The following parameters can be considered as representative, reliable and accurate;

(i) Daily maximum temperature
(ii) Daily minimum temperature
(iii) Dry bulb temperatures at 0900 and 1500 hours
(iv) Rainfall

(c) The wet bulb temperatures appear to be a little high and as a consequence, the wet bulb depression measured at the climate station is 0.5°C to 1.0°C smaller than measurements made by the UM equipment in the surrounding grass plains.

(d) The quality of the pan evaporation data is questionable. This is because of the nature of the instrument itself, its maintenance and its location in the catchment.

(e) The quality of the global radiation data from May 1980 onwards, and the wind run data have yet to be determined. These are important parameters in evaporation estimation formulae (See Chapter 5). It is strongly recommended that these data be examined in any future project.

The data presented in this chapter illustrates how carefully one must view unsubstantiated data. It was only by chance that a spare channel became available on the data logger and the decision was made to measure global radiation. This was not in the original concept of the project as it was assumed that "accurate and reliable" global radiation data would be available from the BM site only 3 km away. If the UM data had not been recorded, the error in the BM data may
never have been detected. The implications of this with respect to future model testing can only be imagined.

From this experience, it seems reasonable to recommend that, at the commencement of any future modelling projects, the climate data to be used as input should be correlated with a second nearby site (as in Fig. 4.11) or should be confirmed by comparative field measurements (as in Fig. 4.1 to Fig. 4.5). A short period spent doing this at the start of a project could prevent months of fruitless endeavour.
Chapter 5

THE EVAPORATION MODEL

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5.9 SUMMARY
CHAPTER 5

THE MODELLING OF EVAPORATION

5.0 INTRODUCTION

It was intended in this chapter to test ARBM using an improved evaporation model. The test was to compare observed and predicted runoff and evaporation. The long term testing of ARBM was not possible for three reasons. They are:

(i) There was no global radiation data available from the climate station (See Section 4.3.5). As will be seen in Section 5.3.1, this data was intended to form the basis of a physically based evaporation prediction model. Without it, the new model could not be tested.

(ii) There was a lack of long term UM data. The missing radiation data from the climate station could have been replaced with the same data collected at the UM site. However, frequent instrument or logger failures meant that, although over 100 days of complete data were collected, there were periods of up to a week when no useful data was available. Consequently, there was no continuous global radiation data available for the catchment during the study period.

(iii) There was an initial misconception of the definition of potential evaporation. As a consequence of this, wind speed was not simultaneously measured with the Bowen ratio data. This meant that it was impossible to define a catchment characteristic relating to potential evaporation. This will be discussed in Section 5.4.1.

Therefore, the objectives of this chapter are:

(i) to propose a physically based evaporation model for ARBM

(ii) to test various components of that model with the limited data which is available.
5.1 THE AUSTRALIAN REPRESENTATIVE BASINS MODEL, ARBM

5.1.1 Deterministic Hydrological Modelling

A deterministic hydrological model is not a statistical correlation between a series of meteorological inputs and streamflow outputs. It is a numerical analogue of the whole sequence of hydrological processes which occur within a catchment. Between entering and leaving a catchment, water undergoes a number of changes both in state and place. The function of deterministic hydrological modelling is to identify and understand each of these changes and to propose a sequence of numerical calculations which will simulate them. The skill (or art) of successful modelling often is to identify which changes are important and which are insignificant and then to simulate these changes in the correct sequence.

The sequence of calculations which simulates a hydrological process or a number of processes is called a model and models can be divided into two types; physically based and empirical.

Physically based models are used for important processes where sufficient laboratory or field studies have been conducted to enable a truly representative model to be proposed. Empirical models are used when:

(i) a sufficient understanding of the actual physical process does not exist

(ii) a complete understanding of the process does exist but the physical modelling of that process is not possible due to a lack of suitable input data

(iii) overly complex and time consuming calculations are necessary to model physically a relatively insignificant process

In both physically based and empirical models there are a number of constant parameters which will be called "catchment characteristics". For physically based models it should be possible to measure these catchment characteristics
within the catchment area (although this is certainly not always the case), while for empirical models, these parameters must be fitted statistically. It is one of the aims of the representative basins project "to systematically eliminate empiricism and increase the use of measured catchment processes". (A.W.R.C. (1969)) Presumably, if all empiricism could be eliminated and the model based solely on physical analogues using measured catchment characteristics, then the model could be applied to similar basins without the need of model performance verification. This is one of the objectives of the Representative Basins project; an objective that many hydrologists would say is impossible. It is one aim of this chapter to improve the evaporation model of ARBM by eliminating empirical models where possible and replacing them with physically based models using measured catchment parameters.

5.1.2 Types of Deterministic Hydrological Models

Mathematical models of the rainfall-runoff process usually consist of a number of storage areas (stores) which, conceptually, represent the moisture holding capacity of the vegetation and soil of the catchment. The movement of water into, out of and within the catchment is simulated by transferring water into and out of these stores according to known or assumed functions which represent such physical processes as infiltration and evaporation. Overflow from this system of stores is regarded as modelled runoff and theoretically should correspond to the observed runoff from the catchment. This comparison is usually the method of model verification. Usually, spatial variation of the processes and catchment characteristics within the catchment is not taken into account and thus averaged or "lumped" values are used. Examples of deterministic rainfall-runoff models are the Stanford Model, Mark IV (Crawford & Linsley (1966)); the Boughton Model (Boughton (1968)); the Systeme Hydrologique European (SHE) (Bovon (1979)); Watsim (Aston & Dunin (1980)) and ARBM (A.W.R.C. (1969)). These models are basically similar but they differ in;
(i) the need to cater for processes which are usually neglected but for particular applications, assume greater significance, e.g. snow, cracking soils, flood routing.

(ii) the degree of "refinement". Greater refinement in a model means that the physical processes are more closely or exactly modelled but this normally involves:

   (i) greater complexity, e.g. Williamson (1979) takes account of spatial variability by introducing the concept of a "distributed" parameter

   (ii) an increased number of processes and catchment characteristics

   (iii) an increase in computing time due to more complex algorithms and shorter modelling periods

   (iv) more stringent requirements of the quantity and quality of input data

5.1.3 The ARBM

The ARBM assumes that all the water in the catchment moves between five stores. (See Fig. 5.1) The arrows in Fig. 5.1 indicate physical processes which are simulated by algorithms of various complexities. The sequence of calculations which controls the movement of water between these stores is given in Fig. 5.2. A complete listing of the present version of the model is given in Williamson & Turner (1980) while a listing of the programme variables is given in Appendix A.

A brief description of the model simulation for each time period is as follows. The terms in brackets are the programme variables. Each store is defined in terms of an equivalent depth of water (mm).

When rain (P) falls, it enters the interception (IS) and depression (DS) stores in proportions dictated by the interception area ratio (IAR). If the capacity of the interception store is exceeded, the excess enters the depression store. Water then infiltrates (INF) into the upper
Fig 5.1  WATER STORES AND FLOW PATHS OF ARBM
Start

The time interval is daily if depression store is empty, otherwise hourly, and a fraction of an hour during rainfall.

Add any rain to the interception store and the depression store.

Excess interception goes to the depression store.

Calculate water entry to the soil stores and the amount of runoff by the infiltration function.

Evaporate water from the stores.

Redistribute water between the upper and lower soil stores.

Calculate drainage from the lower soil store.

End

Fig 5.2 SEQUENCE OF CALCULATIONS IN ARBM
soil store (US) and, if after infiltration has finished, the depression store is full, then runoff (RE) occurs. Evaporation then takes place. Firstly, this occurs at a potential rate (PE) from the interception and depression stores. If the evaporation demand is not satisfied when these stores have been emptied, then evaporation from the upper and lower (LS) soil stores occurs. This is at a rate determined by the present soil moisture content. Evaporation from the lower soil store represents transpiration by deep rooted vegetation, e.g. trees. The amount of this is determined by the proportion of the catchment covered by shallow rooted vegetation (ER). When evaporation is complete, water is redistributed (RDI) between the upper and lower stores and deep drainage (LDR) occurs. The model can also simulate cracking of the soil.

The time interval chosen for each simulation needs consideration. The usual approach in mathematical modelling is to assume uniform rates of input and output over fixed time intervals. If this time interval is too long, information will be lost and the analysis will lack discrimination; if it is too short, unnecessary computation is involved with no gain in information. A.W.R.C. (1969) concluded that, for periods of no rainfall input (and if the depression store is empty), calculations at daily intervals should be adequate but otherwise an hour period should initially be tried. Aston & Dunin (1980) use a 15 minute period and Williamson (1979) used a 30 minute period.

5.1.4 The Present Evaporation Model in ARBM

The structure of the evaporation calculations as used by Williamson (1979) is shown in Fig. 5.3. The sequence of calculations is as follows.

(i) The total potential evaporation for the day (PE) is calculated. Williamson took this to be equal to pan evaporation.

(ii) If necessary, this is distributed throughout the day in shorter time intervals (PEJ).
Start

Potential Evaporation (PE) at any stage of the day is calculated.

PE=0

Yes

Evaporate at the Potential rate from the Interception store and adjust PE.

PE=0

Yes

Evaporate at the Potential rate from the Depression store and adjust PE.

PE=0

Yes

Calculate moisture limited evaporation from the upper and lower stores.

Upper and Lower soil evaporation = 0.

Define evaporation as the lesser of the remaining PE and that calculated above.

Reduce soil stores by the defined evaporation.

End

Fig 5.3 EVAPORATION CALCULATIONS IN ARBM
(iii) Evaporation from the interception store (IS) occurs at a potential rate.

(iv) If PEJ > IS, evaporation from the depression store (DS) occurs at a potential rate.

(v) If PEJ > (IS + DS), evaporation from the upper and lower soil stores then occurs at a rate determined by PEJ and the present soil moisture content.

(vi) The upper and lower soil stores are adjusted for this loss of water.

To improve this section of ARBM, the following aspects of the model must be examined:

(i) Conceptual validity: It must be determined whether the concept of the present model is a true analogue of the actual physical processes involved.

(ii) Potential evaporation: The definition and modelling of this parameter must be examined.

(iii) The diurnal variation of evaporation predicted by the model must be compared with field data.

(iv) The relationship between soil moisture content, potential evaporation and actual evaporation must be examined.

5.1.5 Conceptual Validity of the Model

To an extent, it should not be necessary to question the conceptual validity of the model since this was proposed by experienced hydrologists. However, some of the data presented in Chapter 2 neatly confirms the validity of the model and this will be discussed here.

Before doing this, it is necessary to pre-empt the discussion on potential evaporation (PE) by briefly describing PE as the evaporation rate which will occur from a wet surface. It can be shown that under these conditions, the Bowen ratio is typically less than 0.5. By this definition,
evaporation from the interception and depression stores must occur at a potential rate. For evidence of this, refer to the data presented in Chapter 2. (Example 6, 10/12/1979). On the morning of 10/12/1979, the interception and depression stores were initially empty and evaporation was occurring at a rate controlled by the current soil moisture (i.e. \( \beta = 3 \)). At 1200, a brief shower (0.4 mm) occurred and this partially filled the interception store. Evaporation then occurred at a potential rate (\( \beta < 0.2 \)) until the interception store was emptied. The Bowen ratio then returned to near pre-storm values and evaporation was again controlled by the soil moisture content. This process was repeated at 1500 when a further 0.2 mm of rain fell. Another example of this was on 10/1/1980 (Example 5). On this day, 0.4 mm of rain fell before dawn and partially filled the interception store. This was evaporated at a potential rate in the early morning (\( \beta = 0 \)) until the interception store was empty. After that, evaporation occurred at a rate determined by the current soil moisture content.

The data collected on these days is taken as evidence that the evaporation model in ARBM is conceptually correct.

5.2 POTENTIAL EVAPORATION

5.2.1 Definition of Potential Evaporation

Potential evaporation, PE, is one of the most misunderstood concepts in hydrology. Before discussing this process in detail, it is instructive to review some of the many definitions of PE found in the literature.

(i) Potential evapotranspiration is "the water loss which will occur if at no time there is a deficiency of water in the soil for the use of vegetation". Thornthwaite (1944).

(ii) Potential transpiration is "the amount of water transpired, in unit time, by a short green crop, completely shading the ground, of uniform height and never short of water". Penman (1956).
(iii) "In the interest of clarity and reproducible results, there is good reason to consider potential evapotranspiration to be equivalent to the evaporation from a free-water surface of extended proportions, but with negligible heat storage capacity. Potential evapotranspiration as defined by Thornthwaite approaches free-water evaporation provided there is complete vegetal cover, and the effects of meteorological factors on the two are sufficiently alike to be converted into actual evapotranspiration in the same manner". Linsley et al (1975).

(iv) "A definition based on the consensus of opinion of participants at this conference would consider potential evapotranspiration as the amount of water transferred from a wet surface to vapour in the atmosphere. By this definition, potential evapotranspiration represents the greatest possible mass transfer and energy dissipation from the earth's surface by vapourization of water. Potential evapotranspiration is a product of the atmospheric environment and the magnitude does not involve the type of surface cover so long as it may be considered wet." Decker (1966).

(v) "Potential evaporation can be defined for any situation in terms of the appropriate meteorological variables and the radiative and aerodynamic properties of the surface. When the surface is wet and imposes no restriction upon the flow of water vapour, the potential value is reached." van Bavel (1966).

(vi) "The concept of potential evaporation as introduced by Penman (1948) applies to short crops, fully covering the ground, whose water supply is never limiting; physiological controls by the plants are assumed negligible, so that the evaporation rate is determined by the prevailing weather." Szeicz et al (1969).

(vii) "Potential evaporation, i.e. when the liquid water supply to the effective evaporating surface is adequate for the prevailing demand." Dilley & Shepherd (1972).
(viii) "Potential evaporation is defined as the evaporation that occurs from a freely transpiring surface or from a wet bare surface." Davies & Allen (1973).

As can be seen from the above, there exists some fundamental differences in the way in which researchers have interpreted potential evaporation. A close examination reveals that two important issues are raised.

(i) Is PE dependent solely on meteorological factors as suggested by Decker (1966) or do surface properties have an effect, as suggested by van Bavel (1966)?

(ii) Is it necessary that the surface be physically "wet" before PE can take place or is it sufficient that the surface be actively transpiring?

Before answering these questions, it is first necessary to discuss the difference between the terms "evaporation" and "evapotranspiration". Evapotranspiration is a term that has been coined to describe the total loss of water from a surface as the sum of direct evaporation from the soil plus transpiration by the crop. Since hydrologists are only interested in the total loss of water from the catchment, a fine distinction between direct evaporation and transpiration is unnecessary, unless it affects the type of model proposed. Certainly, the distinction becomes meaningless for bare soil, stubble and water surfaces. Thus, the term will not be used here and evaporation will be defined as the total loss of water from the catchment by vapourization.

To answer the question of surface dependence, it is necessary to return to Chapter 1 and review the processes involved in evaporation. Evaporation is largely dependent on the amount of energy available at the surface, i.e. net radiation. However, the primary source of this energy is global radiation and this is related to net radiation via the surface reflectivity (albedo). Thus, it is immediately evident that the surface type does influence potential evaporation. Furthermore, as will be shown later, surface roughness also has
an effect. Thus, the definition given by Decker (1966) is fundamentally incorrect. The origin of the concept that PE is independent of the surface type can be found in a review of the historical development of the PE concept. Although it is unnecessary to do this, it is important to emphasis that this concept is not correct and, when proposing a model to estimate PE, it is necessary to parameterize surface properties, notably albedo and surface roughness.

In response to the question of surface wetness, it can be partly answered by defining PE as the evaporation rate which will occur when the surface is wet. This implies that the surface is saturated and can be mathematically defined by:

$$h_o = 100\%$$

or

$$D_o = 0^\circ C$$

where

- $h_o = \text{relative humidity at the surface}$
- $D_o = \text{wet bulb depression at the surface}$

An immediate consequence of this definition is that evaporation from free water surfaces must always occur at a potential rate. Also, because water stored on vegetation following rain and water ponded on the soil is modelled by water in the interception and depression stores respectively, then evaporation from these stores must occur at a potential rate. However, consider, as described in some of the PE definitions above, a crop fully covering the ground, unrestricted in water supply and actively transpiring. Under these circumstances there is no reason to believe that the effective surface humidity will be saturation. It seems likely that for some crops, the transpiration process will restrict the flow of water to the atmosphere, even when there is an unrestricted supply of water to the crop's roots. It seems reasonable to assume that plant species which have evolved in arid and humid climates will transpire at different rates. Also, the stage of growth may affect the transpiration rate. This condition of maximum
transpiration might be described as potential evapotranspiration, but a definition of potential evaporation based on a description of the state of the vegetation cannot be considered as universal. Potential evapotranspiration is sometimes very close to potential evaporation and the reason for this will be given in Section 5.2.3.

Thus, potential evaporation will be re-defined here as "the evaporation rate which will occur from a particular surface when the surface is wet and the effective surface humidity is saturation".

This definition is mathematically quantified by Eqn. 5.1 or Eqn. 5.2. Of the other definitions cited above, the definition given by van Bavel (1966) is most similar.

5.2.2 Prediction Models for Potential Evaporation

When discussing the problems associated with predicting evaporation, Wartenberg (1974) identified three main categories of models which have been used to estimate potential evaporation. They are:

(i) Empirical models
(ii) Instrumental models
(iii) Physically based models

Although the first two types of model will not be used here, they will be discussed briefly. This is because they are often used elsewhere.

(i) Empirical Models

There are a number of well known empirical formulae which have been used to estimate actual and potential evaporation. These use one or more of the meteorological factors which directly or indirectly influence evaporation and have included air temperature, relative humidity, number of daylight hours, solar radiation and precipitation. Examples of this type of formulae are Thornthwaite (1948), Blaney & Criddle (1950) and Jensen & Haise (1963). Although these formulae must be locally calibrated and often only relate to the
area in which they were developed, they have been used elsewhere indiscriminately and with poor results. As has been noted previously, the use of empirical formulae is to be avoided and no use of these formulae will be made.

(ii) Instrumental Models

In this method, the evaporation from a meteorological device called an evaporimeter (See Section 1.6.4) is related to PE by simple "pan to crop" factors or, for large water bodies, by more complex pan-coefficient formulae (A.W.R.C. (1970(a)). It has been found that Class 'A' pan evaporation adequately simulates PE from many short crops (McIlroy & Angus (1964), Campbell & Phene (1976)) and this approach was followed by Williamson (1979) when testing ARBM. He defined PE from the grassland of the catchment as being equal to pan evaporation recorded at the climate station.

5.2.3 Combination Formulae

From an understanding of the physical processes involved in evaporation (See Section 1.5), it should be possible to propose a physically based model to simulate PE. However, all of the evaporation measurement techniques cited in Section 1.6 require data which is not routinely available at a climate station. In particular, this includes surface values of temperature and humidity. To overcome this problem, Penman (1948) proposed a formula which only requires data at one level. As will be seen below, he combined the energy balance and aerodynamic formulae and thus, his formula and all its subsequent modifications have been called "Combination formulae".

Penman's derivation is as follows. Consider a saturated surface. Then the evaporation from that surface can be defined by:

\[
(i) \quad \text{the energy balance method. That is;}
\]

\[
LE = \frac{R_n - G}{1 + \beta}
\]

2.5
where
\[ \beta = \gamma \frac{\Delta T}{\Delta e} \]

\[ = \gamma \frac{T_o - T_a}{e_o^* - e_a} \]

\[ = \gamma \frac{T_o - T_a}{e_o^* - h_a e_a^*} \]

where
\( T_o \) = effective surface temperature, °C
\( T_a \) = air temperature (say, at 2 m), °C
\( e_o^* \) = vapour pressure at the surface; this being saturation for PE, mb
\( e_a^*, e_a \) = saturation; actual vapour pressure of the air at the same height as \( T_a \), mb
\( h_a \) = relative humidity of the air

(ii) the bulk aerodynamic method. That is:

\[ \text{LE} = L f(u) \ (e_o^* - e_a) \]

By defining the terms

\[ s = \frac{e_o^* - e_a}{T_o - T_a} \]

\[ = \frac{3e^*}{\beta T} \text{ at } T_a \]

and

\[ \text{LE}_a = L f(u) \ (e_a^* - e_a) \]

the above equations can be combined to derive an expression for \( \text{LE} \) as

\[ \text{LE} = \frac{s}{s + \gamma} \ (R_n - G) + \frac{\gamma}{s + \gamma} \ \text{LE}_a \]

This is Penman's original formula and is an attractive method of calculating PE. Firstly, it requires only commonly available meteorological data at one level. Secondly, it is physically based and conveniently separates the energy and aerodynamic components of the evaporation process.
An extension of this theory is that if the energy term remains constant, then the minimum PE rate occurs when $h_a = 1$. Thus,

$$\beta = \gamma \frac{T - T_a}{e^* - e_a^*}$$

$$= \frac{\gamma}{s}$$  \hspace{1cm} 5.9

Thus

$$\text{LE} = \frac{s}{s + \gamma} (R_n - G)$$

$$= \text{LEQ}$$  \hspace{1cm} 5.10

This is called "Equilibrium evaporation" and was first described by Slatyer & McIlroy (1961). This is the evaporation rate which would occur when the air overlying the surface had travelled sufficiently far over that surface so as to come into equilibrium with it. That for a wet surface, $D_a = D_o = 0$. Note that this can occur for any surface where $D_a = D_o \neq 0$.

Since 1948, a number of modifications to this formula have appeared. For example, consider the McIlroy formula (Slatyer & McIlroy (1961)) below.

Consider that the flux of heat and water vapour can be expressed by analogy with Ohm's law as being defined by a potential difference and a resistance (or conductance). Thus, for any surface, wet or dry,

$$H = \rho \ C_p \ h \ \Delta T$$

$$= \frac{\rho \ C_p}{r_a} \ \Delta T$$  \hspace{1cm} 5.11

and

$$\text{LE} = \frac{\rho \ C_p \ h \ \Delta e}{\gamma}$$

$$= \frac{\rho \ C_p}{\gamma r_a} \ \Delta e$$  \hspace{1cm} 5.13

where $h = \text{aerodynamic conductance \ (m/s)}$

$r_a = \text{aerodynamic resistance \ (s/m)}$
\[ \Delta T = T_o - T_a \]
\[ \Delta e = e_o - e_a \]

and the other terms have been previously defined.

Note also that
\[ \Delta T = \Delta T_w + \Delta D \] 5.15
\[ \Delta e = s \Delta T_w - \gamma \Delta D \] 5.16

where \( D = T - T_w \) (See Section 2.1.3).

Thus, from Eqns. 5.11 and 5.13,
\[ H = \rho \, C_p \, h \, \Delta T_w + \rho \, C_p \, h \, \Delta D \] 5.17
\[ LE = \frac{s \, \rho \, C_p \, h}{\gamma} \, \Delta T_w - \rho \, C_p \, h \, \Delta D \] 5.18

Thus, from Eqn. 2.1,
\[ R_n - G = H + LE \]
\[ = \rho \, C_p \, h \, (1 + \frac{s}{\gamma}) \, \Delta T_w \]

Therefore,
\[ \frac{\rho \, C_p \, h}{\gamma} \, \Delta T_w = \frac{R_n - G}{s + \gamma} \] 5.19

and then substituting into Eqn. 5.18, it can be shown that
\[ LE = \frac{s}{s + \gamma} \, (R_n - G) - \rho \, C_p \, h \, \Delta D \] 5.20

If the surface is saturated (i.e. PE conditions, \( D_o = 0 \))
\[ PE = \frac{s}{s + \gamma} \, (R_n - G) + \rho \, C_p \, h \, (T_a - T_{wa}) \] 5.21

This is a convenient expression for PE since it expresses the aerodynamic term in terms of wet bulb depression rather than vapour pressure deficit. This can be shown to be equivalent to Penman's formula (Eqn. 5.10). Knowing that,
\[ s = \frac{e^*_a - e^*_w_{wa}}{T_a - T_{wa}} \] 5.22
and \( e_a = e_{Tw}^* - \gamma (T_a - Tw_a) \)  \[ 2.14 \]

Then,

\[
(s + \gamma) \text{PE} = s (R_n - G) + \rho C_p h (s + \gamma) (T_a - Tw_a)
\]

\[
= s (R_n - G) + \rho C_p h (s(T_a - Tw_a) + \gamma(T_a - Tw_a))
\]

\[
= s (R_n - G) + \rho C_p h (e_{Tw}^* + e_{Tw}^* - e_a)
\]

\[
\text{PE} = \frac{s}{s + \gamma} (R_n - G) + \frac{\rho C_p h}{s + \gamma} (e_a^* - e_a)
\]

From Eqn. 5.9 and 5.10, it then follows that;

\[
f(u) = \frac{\rho C_p}{\gamma L} h
\]

\[
= \frac{\rho b}{p} h
\]

\[
= \frac{\rho b}{p} \frac{1}{r_a}
\]  \[ 5.24 \]

\[ 5.25 \]

It is interesting to note that "s" in Penman's formula (Eqn. 5.8) is calculated at the mean dry bulb temperature, while in McIlroy's formula (Eqn. 5.21) "s" is calculated at the mean wet bulb temperature. This anomaly causes little difference in "s" in a humid environment since \( T = Tw \). However, in a dry environment, larger differences can occur. Since in Penman's original formula, "s" was intended to represent the ratio of the humidity to temperature gradients, "s" should be calculated at a dry bulb temperature, i.e. \( T_a \).

The last but most significant type of combination formula proposed is that of Monteith (1965). He continued the analogy with Ohm's law and concluded that evaporation from a transpiring crop was controlled by two resistances acting in series. Thus, referring to Fig. 5.4,

\[
H = \frac{\rho C_p}{r_H} (T_O - T_a)
\]  \[ 5.26 \]
\[ \text{LE} = \frac{\rho C_p}{r_w} (e_o - e_a) \]  
5.27

\[ \text{LE} = \frac{\rho C_p}{r_s} (e_s - e_o) \]  
5.28

FIG. 5.4 Evaporation from a Stomata in a Transpiring Crop

where \( r_H \) = aerodynamic resistance to the flow of heat
\( r_W \) = aerodynamic resistance to the flow of water vapour
\( r_s \) = stomatal resistance to the flow of water vapour
\( e_s, e_o, e_a \) = vapour pressure in the stomata, leaf surface and air respectively
It can now be shown that

\[ LE = \frac{\rho C_p}{\gamma} \frac{(e_s - e_a)}{(r_s + r_w)} \]  \hspace{1cm} 5.29

Now, if the stomatal vapour pressure, \( e_s \), is saturation at the leaf temperature, \( T_o \), then;

\[ e_s = e^* \]
\[ = e_o \]

Thus, Eqn. 5.29 becomes

\[ LE = \frac{\rho C_p}{\gamma} \frac{(e^* - e_a + e^* - e_o)}{r_s + r_w} \]  \hspace{1cm} 5.30

Now, by substituting this result into Eqns. 2.5, 5.6 and 5.26, it can be shown that;

\[ LE = \frac{s (R_n - G) + \rho C_p (e^* - e_a)/r_H}{s + \gamma (r_s + r_w)/r_H} \]

If the aerodynamic resistances of heat and water vapour are equivalent and equal to \( r_a \), then;

\[ LE = \frac{s (R_n - G) + \rho C_p (e^* - e_a)/r_a}{s + \gamma (1 + r_s/r_a)} \]  \hspace{1cm} 5.31

Attempts have been made to expand Monteith's concept of surface resistance into more complex formulae (e.g. Shuttleworth (1975), Shuttleworth (1978)). However, Monteith's formula is most often used by redefining \( r_s \) as not a single stomatal resistance, but as an average "canopy resistance to the transport of water from some region within or below the evaporating surface to the surface itself, and is expected to be a function of the stomatal resistance of individual leaves". (Beven (1979)).

By re-introducing Eqn. 5.23, Eqn. 5.31 can be reduced to;

\[ LE = \frac{s + \gamma}{s + \gamma (1 + r_s/r_a)} \frac{PE}{r_a} \]  \hspace{1cm} 5.32
From this, it can be seen that evaporation from a surface will occur at a potential rate if the surface resistance, $r_s$, is zero. Furthermore, evaporation can occur at a rate which is very close to potential whenever $r_s$ is much smaller than $r_a$ although it may not necessarily be zero. This explains why the evaporation from freely transpiring crops can be very close to PE, even though the surface is not actually wet.

This relationship between $r_s$ and $r_a$ is the reason why the definition of PE has so often included a reference to transpiring crops. In many studies, the values of LE measured over actively transpiring crops often very nearly equalled PE as calculated by a combination formula. The results of McCaughey (1968) are an example of this type of study. McCaughey (1968) even noted that the evaporation rate immediately following sprinkler irrigation of his crop (that is PE) was slightly greater than the evaporation rate from the crop. However, probably because of the loose definition of PE, he was unable to distinguish between true PE and conditions when $r_s$ was much less than $r_a$. Shepherd (1972) also found evidence that evaporation rates from well watered crops could fall below PE as calculated by a combination formula.

In summary, potential evaporation only occurs when the surface is physically wet. It can be physically modelled using any form of the combination formulae. The most convenient form for ARBM is Eqn. 5.21. This equation has two components. They are the energy term and the aerodynamic term. The modelling of this will be discussed in detail in the following sections.

5.3 THE ENERGY COMPONENT OF PE

5.3.1 Physical Model of $R_n$

The various components of the radiation balance of a natural surface are shown in Fig. 5.5.
FIG. 5.5 The Radiation Balance of a Natural Surface

In Fig. 5.5,

\[ \begin{align*} 
0 & = \text{Direct-beam solar radiation} \\
D & = \text{Diffuse solar radiation} \\
R_g & = \text{Global radiation} \\
& = 0 + D \\
R_a & = \text{Atmospheric (Back) radiation} \\
R_t & = \text{Terrestrial radiation} \\
\alpha & = \text{Albedo} 
\end{align*} \]

The net radiative exchange, \( R_n \), at the surface is;

\[ R_n = (1 - \alpha) R_g + R_a - R_t \quad 5.33 \]

The estimation of the components of \( R_n \) will be discussed below.
(i) Global Radiation, $R_g$

Global radiation is the sum of the diffuse and direct-beam components of incoming solar radiation received on a horizontal plane at the earth's surface. It consists essentially of short-wave (visible) radiation in the waveband of 0.3 μm to 3 μm. At any time, the instantaneous value of $R_g$ can be calculated from:

$$R_g = \psi \cos Z I_o$$

where

$\psi$ = effective absorption of the atmosphere

$Z$ = zenith angle of the sun

$I_o$ = solar constant (1353 W/m²)

= solar radiation received on a plane perpendicular to the direction of the sun at the outside of the atmosphere.

The value of $I_o$ is known accurately. It is also possible to accurately calculate $Z$ for any location at any time (See Appendix C). However, it is only possible to accurately calculate $\psi$ for clear skies. (Idso (1970), Davies et al (1975)). For cloudy skies, $\psi$ can be reasonably accurately predicted from a detailed knowledge of the cloud type, height and amount of cover (Paltridge & Proctor (1976), Atwater & Brown (1974)). Usually, this type of data is not available. Empirical formulae which use the number of hours of sunshine or the amount of cloud cover have been developed. The most well known type is:

$$R_g = (a + b \frac{n}{N}) R_o$$

where

$a, b$ = locally determined correlation co-efficients

$n$ = actual number of hours of sunshine per day

$N$ = possible number of hours of sunshine per day (daylength)

$R_g$ = daily total of global radiation

$R_o$ = daily total radiation received on a horizontal surface at the outside of the atmosphere (extra-terrestrial global radiation).
\[ \int_{t_1}^{t_2} I_0 \cos Z \, dt \]

\( t_1, t_2 = \text{sunrise, sunset} \)

The methods described above can be used to estimate \( R_g \) for any catchment area, if locally determined values of "a" and "b" are available. Fortunately, this should not be necessary at the Warrambine Creek site as \( R_g \) is routinely measured at the climate station. (See Chapter 4).

(ii) **Albedo, \( \alpha \)**

Albedo is the short-wave reflectivity of a natural surface. This aspect of the radiation balance has been extensively investigated and is well understood. It is known that \( \alpha \) is reasonably independent of the sky clarity (cloudiness) (Polavarapo (1970), Stanhill et al (1966)). Also, \( \alpha \) not only varies with surface type but also with the angle of incidence of the radiation (i.e. \( Z \)). Consequently, \( \alpha \) varies as a time of year and a time of day. Examples of this variation are given by Paltridge (1975), Moore (1976), Nkemdirim (1972), Kalma & Badham (1972) and Idso et al (1969). Many studies have integrated this variation throughout the day to produce daily average values of \( \alpha \). Typical values for grassland are given in Table 5.1. Moore (1976) collected data over grassland in the SE corner of South Australia. This is at a similar latitude and over similar vegetation to the Warrambine Creek site. Thus this data is probably most appropriate for modelling of this catchment.

(iii) **Atmospheric (Back) Radiation, \( R_a \)**

This is the radiation emitted towards the earth's surface by water vapour, \( CO_2 \) and aerosols in the atmosphere. This is infrared (long-wave) radiation of wavelengths greater than 3 \( \mu m \). Many empirical formulae have been developed to relate this to surface properties, notably air temperature and humidity. The most well known of these for clear skies are:
<table>
<thead>
<tr>
<th>AUTHOR</th>
<th>SURFACE DESCRIPTION AND LOCATION</th>
<th>ALBEDO</th>
</tr>
</thead>
<tbody>
<tr>
<td>Monteith &amp; Szeicz (1961)</td>
<td>Grassland (England)</td>
<td>0.23 - 0.27</td>
</tr>
<tr>
<td>Stanhill et al (1966)</td>
<td>Semi-steppe perennials (Israel)</td>
<td>0.19 - 0.23</td>
</tr>
<tr>
<td>Oguntoyinbo (1970)</td>
<td>Savannah grassland (Nigeria)</td>
<td>0.19 - 0.23</td>
</tr>
<tr>
<td>Polavarapo (1970)</td>
<td>Mown grass (Canada)</td>
<td>0.20 - 0.24</td>
</tr>
<tr>
<td>Nkemdirium (1972)</td>
<td>Prairie grassland (Canada)</td>
<td>0.21 - 0.23</td>
</tr>
<tr>
<td>Moore (1976)</td>
<td>Grassland (S.E. Australia)</td>
<td>0.27 ± 0.03 WINTER</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.24 ± 0.01 SPRING</td>
</tr>
</tbody>
</table>

**TABLE 5.1** Typical Values of Albedo for Grassland Areas

(i) Brunt (1932)

\[ R_a' = (a + b \sqrt{e}) \sigma T_a^4 \]

5.36

(ii) Angstrom (1936)

\[ R_a' = (a - \beta 10^{-\gamma e}) \sigma T_a^4 \]

5.37

(iii) Brutsaert (1975)

\[ R_a' = 1.24 (\frac{e}{T_a})^{1/2} \sigma T_a^5 \]

5.38

(iv) Swinbank (1963)

\[ R_a' = 5.31 \times 10^{-13} T_a^6 \]

5.39

(v) Idso & Jackson (1969)

\[ R_a' = (1 - 0.261 \exp (- 7.77 \times 10^{-4} (273 - T_a)^2)) \sigma T_a^4 \]

5.40
where \( R'_a \) = clear sky back radiation
\( T_a \) = screen air temperature (\(^0\)K)
\( e \) = screen vapour pressure (mb)
\( \sigma \) = Stefan-Boltzman constant

Cloud cover can then be accounted for by an empirical formula such as:

\[
R_a = (b + (1 - b) \frac{n}{N}) R'_a
\]

When reviewing all these formulae, Idso (1974) concluded that for periods of a day or longer, formulae using \( T_a \) only are adequate since there is a long term correlation between \( T_a \) and \( e \). For periods of less than a day, formulae using \( T_a \) and \( e \) should be used where locally derived coefficients are available. As these coefficients vary markedly from site to site, it is probably better to use the Swinbank or Idso-Jackson formulae when no local coefficients are available. Over the temperature range, 0\(^0\)C - 40\(^0\)C, these formulae are very similar. Corrections for diurnal variations from these formulae can be made by the method of Paltridge (1970).

(iv) Terrestrial Radiation, \( R_t \)

This is the radiation emitted by the surface acting as a grey-body emitter. Like \( R'_a \), this is all longwave radiation. Monteith & Szeicz (1962) showed that this could be predicted by:

\[
R_t = \epsilon \sigma T_o^4 + (1 - \epsilon) R_a
\]

where \( \epsilon \) = emissivity of the surface
\( T_o \) = surface temperature (\(^0\)K)

It is known that \( \epsilon \) is approximately 0.98 for vegetation and 0.90 - 0.95 for bare soils. (Monteith & Szeicz (1962)). Usually it is not possible to use this formula because \( T_o \) is unknown. As a first approximation, it can be assumed that \( T_o = T_a \). Over a 24 hour period, this produces adequate results. (Paltridge (1975)).
In summary, it can be seen that an adequate knowledge of all the components of the radiation balance exists. Therefore a physically based model can be proposed to predict $R_n$. The data collected at the climate station consists of spot readings of temperature and daily totals of $R_g$. Thus, daily totals of $R_n$ can be most easily modelled. If the diurnal variation in $R_n$ is required then the inputs must be distributed throughout the day. For clear skies, the diurnal variation of $R_g$ could be calculated using Appendix C. For cloudy skies, a method similar to Fleming (1970) (See Section 5.5) could be devised. Paltridge (1975) gives the diurnal variation of $\alpha$ and Johnson & Fitzpatrick (1977) give formulae which can predict the diurnal variation of $T_a$ from the maximum and minimum values.

It is usual when using ARBM that when a physical process is modelled, certain "catchment characteristics" need to be measured. From the above discussion, it seems reasonable to attempt to model $R_n$ using variables and formulae found in the literature; i.e. without measuring any "catchment characteristics". To test this hypothesis, $R_n$ was modelled both as half-hourly averages and daily totals (See Fig. 5.6 and Fig. 5.7). Since $R_g$ was unavailable from the climate station (See Chapter 4) only UM data was used. Equations 5.33, 5.39 and 5.42 were used and it was assumed that $\alpha = 0.24$, $\epsilon = 0.98$ and $T_o = T_a$. In Fig. 5.6, half-hourly averages of $R_g$ and $T_a$ were used to model half-hourly averages of $R_n$. In Fig. 5.7, daily average values of $R_a$ and $R_t$ were calculated in W/m². These were then multiplied by the daylength to obtain daily totals in MJ/m².

5.3.2 Empirical Model of $R_n$

The most common type of empirical formula used to estimate $R_n$ is a simple linear regression between $R_n$ and $R_g$. That is:

$$ R_n = a + b \ R_g $$

5.43

This is sometimes justified as a physical model by considering that the net longwave exchange $(R_a - R_t)$ is approximately linearly related to $R_g$. That is;
$R_n$ modelled from literature review only.
Input data - $R_g$, $T_a$
Eqns. 5.33, 5.39, 5.42
$n = 0.24$
$s = 0.98$
$T_0 = T_a$

Fig 5.6  PHYSICAL MODEL OF $R_n$ - HALF HOURLY DATA
\( R_n \) modelled from literature review only.

Input data - \( R_g, T, \) Date.

\( T = \) mean of dry bulb at 0900 & 1500

Daylength calculated from Appendix C.

Eqns, 5.33, 5.39, 5.42

\( \alpha = 0.24 \)

\( \epsilon = 0.98 \)

\( T_0 = T_a \)

\[
Y = -4.822 + 1.504 \times X
\]

\( r^2 = 0.931 \)

\( S_{y,x} = 1.428 \)

**Fig 5.7**  PHYSICAL MODEL OF \( R_n \) - DAILY DATA
\[ R_a - R_t = c + d R_g \]  
and then
\[ R_n = (1 - \alpha) R_g + R_a - R_t \]
\[ = (1 - \alpha) R_g + c + d R_g \]
\[ = c + ((1 - \alpha) + d) R_g \]
\[ = a + b R_g \]  

Examples of this type of regression are numerous. Some of these are Gay (1971), Gay (1979), Davies (1966), Nkemdirium (1972), Polavarapo (1970), Moore (1976), Idso et al (1969) and Kalma (1972).

Sometimes this formula is expressed as;
\[ R_n = a (1 - \alpha) R_g + b \]  

Fritzchen (1967) showed that no benefit is gained by this inclusion of albedo.

Another empirical formula is that proposed by Monteith & Szeicz (1961). This is;
\[ R_n = \frac{(1 - \alpha)}{(1 + \beta)} R_g + L^* \]  
where  
\[ \beta = \text{heating coefficient} \]  
\[ L^* = \text{net longwave exchange when } R_g \text{ is zero} \]

Although attempts have been made to physically justify this formula, \( \beta \) is not a measureable property. Thus, this formula is only a disguised form of Eqn. 5.43.

Since Eqn. 5.43 is only a statistical correlation between \( R_n \) and \( R_g \), it is feasible to propose alternate statistical models with more degrees of freedom. Also, additional variables could be introduced. As noted in the previous section, \( R_n \) is also dependent on \( T_a \). Thus an alternate model could be;
\[ R_n = a R_g + b \sigma T_a^4 \]  

5.44
5.33
5.43
5.45
5.46
A similar approach to this was taken by Linacre (1967). Correlation coefficients for Eqn. 5.43, 5.47 and a number of other statistical models were determined using daily totals of $R_n$ and $R_g$ collected at the UM site. Predicted values of $R_n$ were regressed against measured values. It was found that no significant improvement in predictive performance could be achieved by any model over the simple linear regression. The predictive results of the linear analysis (Eqn. 5.43) are shown in Fig. 5.8.

By comparing Fig. 5.7 and Fig. 5.8, it can be seen that $R_n$ can be modelled more accurately by not using the physically based model which has no measured catchment characteristics. Clearly some catchment characteristics must be measured. However the determination of the relevant parameters for a physically based model is difficult. This would include the measurement of albedo, the determination of the relationship between $T_o$ and $T_a$ and possibly the determination of "a" and "b" in Eqn. 5.36. The determination of linear correlation coefficients is considerably easier. Only two instruments are needed; viz. global radiometer and net radiometer and the results from this model are quite acceptable (See Fig. 5.8).

Thus it is proposed to use Eqn. 5.43 as the model to predict $R_n$ using the catchment characteristics, "a" and "b" as determined by field measurements.

5.3.3 Model of G

The most usual (and convenient) method of modelling $G$ is to assume that $G$ is approximately zero over a day. This is generally true over a 24 hour period. However, when daytime values of $R_n$ are used, then $G$ becomes significant. In this case, $G$ is usually expressed as a percentage of $R_n$. This varies from 5% to 10% e.g. 7%, Rouse & Stewart (1972); ~10% Dilley (1974).

For this study, daytime values of $G/R_n$ were plotted. (See Fig. 5.9). The average value of $G/R_n$ was approximately 4%. Although this is a little lower than usually found in
Model:
\[ R_n = -0.254 + 0.522 \times R_g \quad r^2 = 0.927 \]

based on daily totals of UM data only.

Fig 5.8   EMPIRICAL MODEL OF \( R_n \) - DAILY TOTALS
Fig 5.9  DAYTIME VALUES OF $G/R_n$.
the literature, it seems to be a reasonable result. Considerable scatter exists in the result.

To see if this scatter could be reduced, it was decided to see if $G/R_n$ was dependent on some other factor. Idso et al (1975) contend that soil moisture content has an influence. Their model is;

$$\frac{R_n - G}{R_g} = a + b \theta$$

5.48

where $a,b = \text{correlation coefficients}$

$\theta = \text{volumetric soil moisture content}$

During this study, $\theta$ was measured on every visit to the investigation site. Ten soil samples were taken from the top 50 mm of the soil surface. Each sample weighed 50g - 100g. The gravimetric soil moisture content was calculated. For the intervening days, the soil moisture content was interpolated, taking note of any rainfall. When Eqn. 5.48 was plotted using data collected in this study (See Fig. 5.10), no correlation was found. When $Gl/R_n$ was plotted against $\theta$, only a slight dependence was seen. (See Fig. 5.11).

Because $G$ is a relatively insignificant component in the total PE model, it seems reasonable from the above data to propose the following model.

$$\frac{G}{R_n} = 0.06 \text{ when } \theta < 15\% \text{ D.B.}$$

$$\frac{G}{R_n} = 0.04 \text{ when } \theta > 15\% \text{ D.B.}$$

5.3.4 Model of Energy Component

The last term of the energy component to model is $s/(s + \gamma)$. As seen in Appendix B, $\gamma$ can be considered to be constant and equal to 0.66 mb/°C for the Warrambine Creek catchment. In Section 5.2.3, it was shown that "s" should be calculated at the dry bulb air temperature. The climate station records four dry bulb temperature; viz. daily maximum, daily minimum, and dry bulb at 0900 and 1500. Thus, any one or combination of these could be used to determine "s". Modelled values of $s/(s + \gamma)$ using various temperature combinations were compared with measured daily averages of $s/(\gamma + s)$. 
$R_g, R_n,$ and $G$ are daytime totals i.e. $R_n > 0$

Fig 5.10  PLOT OF $(R_n - G)/R_g$ vs. SOIL MOISTURE CONTENT
G1 and $R_n$ are daytime values (i.e. $R_n > 0$)

Fig 5.11  PLOT OF $G/R_n$ vs. SOIL MOISTURE CONTENT
The least error occurred when the mean of the 0900 and 1500 dry bulb temperatures was used.

Thus, the model proposed to estimate the energy component of PE is as follows;

(i) Input daily totals of $R_g$ and the dry bulb temperatures at 0900 and 1500
(ii) Input the catchment characteristics, "a" and "b"
(iii) Calculate $R_n$ using Eqn. 5.43
(iv) Calculate $G$
(v) Calculate $s$ using the mean dry bulb temperature
(vi) Calculate $\frac{s}{s + \gamma} (R_n - G)$

The FORTRAN listing of this model is given in Section 5.7. The comparison of the modelled values of the energy component with measured values is shown in Fig. 5.12.

5.4 THE AERODYNAMIC COMPONENT OF PE

5.4.1 Aerodynamic Resistance

The aerodynamic component of the combination formula has two components. One is the aerodynamic resistance of the atmosphere to the transfer of water vapour. The other is the saturation deficit of the atmosphere and this will be discussed in the following section.

In Penman's original derivation of the combination formula (Eqn. 5.8), he proposed an empirical formula which implicitly related windspeed to aerodynamic resistance. His formula applied to water surfaces. It is

$$f(u) = 0.26 \left(1 + \frac{U}{100}\right) \quad 5.49$$

where $f(u) = \text{wind function (mm/mb day)}$

$U = \text{wind run (miles/day)}$
see Section 5.3.4 for model details.

Data correlation

Y = 0.019 + 1.008 * X  \( r^2 = 0.963 \)
\( S_{y,x} = 0.504 \)

Fig 5.12  MODEL OF ENERGY COMPONENT OF PE
It was soon realized that it is necessary to explicitly parametrize surface roughness. By equating the eddy diffusivities of heat, momentum and water vapour, Tanner & Pelton (1960) used the work of Businger (1956) to derive an expression for $r_a$. This was done using the log wind profile in neutral stability, i.e.

$$u = \frac{u^*}{k} \ln \frac{z - d}{z_o}$$

where $u =$ windspeed at the height, $z$ (m/s)
$z_o =$ roughness length (m)
$d =$ zero plane displacement (m)
k =$ von Karman's constant
$= 0.40$ Pierson & Jackman (1975)
$= 0.41$ van Bavel (1966)
$= 0.428$ Tanner & Fuchs (1968)
$u^*$ = friction velocity (m/s)

A number of slight modifications to their original formula have been developed. The form most commonly used is that of Szeicz et al (1969), viz.

$$\frac{1}{r_a} = \frac{k^2 u}{(\ln \frac{z - d}{z_o})^2}$$

Fuchs & Tanner (1967) noted that atmospheric instability could be accounted for by introducing the KEYPS function for the diabatic wind profile (Sellers (1965)). Their formula is:

$$\frac{1}{r_a} = \frac{k^2 u}{(\ln \frac{z + z_o}{z_o} + \phi)^2}$$

where
$\phi = \int_0^z \frac{((\psi - 1)/(z + z_o)) \, dz}{(1 - 18 R_i)^{0.25}}$
$\psi = (1 - 18 R_i)^{-1.825}$
$R_i = \frac{\gamma}{\kappa} \left( \frac{z + z_o}{z} \right) \frac{T_3 - T_o}{u^2} z$

5.50
5.51
5.52
5.53
5.54
5.55
This sophisticated aerodynamic resistance formula has also been used by Davies (1972) and Nkemdirim (1976). For the ARBM, it is not feasible to use this formula since insufficient input data is available. The only relevant data available from the climate station is total daily wind run at 2 m.

However some account of atmospheric instability can be taken. Thom & Oliver (1977) closely examined the aerodynamic term of Penman's original formula. They modified it to suit surfaces other than water. They also found that, somewhat fortuitously, Penman's formula is in close agreement with theoretical curves relating the changes in $r_a$ in unstable atmospheres. Thus, they propose a formula for aerodynamic roughness as:

$$\frac{1}{r_a} = 1 + 0.54 \frac{u}{4.72(\ln \frac{z'}{z_0})^2}$$

$$= a + b \ u$$

where

$$a = \frac{1}{4.72(\ln \frac{z'}{z_0})^2}$$

$$b = \frac{0.54}{4.72(\ln \frac{z'}{z_0})^2}$$

In Eqn. 5.57, 'a' and 'b' can be determined by measuring or estimating $z_0$. Thom & Oliver (1977) suggest that $z_0$ is of the order of 0.1 h where h is the crop height. The vegetation at the Warrambine Creek site varies in height from ~50 mm in late Autumn - Winter to ~300 mm in Spring - Summer. For a windspeed height ($z$) of 2 m, this yields:

$$\frac{1}{r_a} = 0.0059 + 0.00319 \ u \quad (h = 50 \ mm)$$

$$\frac{1}{r_a} = 0.0120 + 0.00649 \ u \quad (h = 300 \ mm)$$

where $u = \text{m/s}$ and $r_a = \text{s/m}$
The more usual method of determining 'a' and 'b' is to simultaneously measure \( r_a \) and \( u \). (e.g. Dilley & Shepherd (1972)). From this, 'a' and 'b' can be determined as statistical correlation coefficients. The only way in which \( r_a \) can be determined is to measure PE. Then \( r_a \) can be deduced from the combination formula (Eqn. 5.21).

It was intended in this study to use this method. It was known that for long periods each spring, the pasture in the Warrambine Creek catchment is well supplied with water and is actively transpiring. It was therefore assumed that PE was occurring. This assumption was made using some of the PE definitions in Section 5.2.1. It was assumed that daily totals of PE could be readily attained. From these data, daily average values of \( r_a \) could be calculated. These would then be correlated with windspeed measured at the climate station. However it is now known that this would be an incorrect method of determining \( r_a \). Thom & Oliver (1977) note that "even with unlimited soil moisture, surface resistance \( (r_s) \) is never negligible, but retains a significant residual value (unless the surface is actually wet). In particular, for short vegetation of moderate roughness, \( r_s \) is often similar in size to \( r_a \)." Thus the assumption that PE would be occurring (i.e. \( r_s = 0 \)) was incorrect.

For the Warrambine Creek catchment, the surface is often physically wet. Cold fronts frequently pass over the catchment causing brief showers. However, it is not common for the surface to remain wet for 24 hours. This is clearly demonstrated by the results presented in Section 2.4. The only way in which the relationship between \( u \) and \( r_a \) can be determined is with the simultaneous measurement of PE and \( u \) for an appropriate number of half hour periods following rain. The pluviometer records (See Section 4.3.2) could be used to indicate when rain fell. PE would commence immediately following the storm.

It is then strongly recommended that for any future evaporation studies in this catchment, the equipment described in Chapter 2 should be modified to include the measurement of
windspeed at 2 m. It has already been shown in Chapter 4 that this measurement of windspeed is desirable.

The determination of 'a' and 'b' in Eqn. 5.57 can be seen as the measurement of certain catchment characteristics, viz. catchment roughness. For the present it is suggested that Eqn. 5.60 and Eqn. 5.61 be used. The FORTRAN listing of this model is given in Section 5.7. To see how these equations compare with results from other studies, see Fig. 5.13.

It is important that a future study determines the coefficients in Eqn. 5.57 accurately possibly even allowing for seasonal variations. Beven (1979) performed a sensitivity analysis on Eqn. 5.31. He found that under certain circumstances, the combination formula can be quite sensitive to errors in $r_a$. From Fig. 5.13, it can be seen that $r_a$ can change fourfold; from ~200 s/m at low windspeeds to less than 50 s/m at higher windspeeds. It can be shown that in Eqn. 5.21, the aerodynamic component usually constitutes 20% to 30% of PE. Thus a 50% error in the determination of $r_a$ can cause a 10% to 15% error in PE. This is quite significant in terms of ARBM.

5.4.2 Saturation Deficit

In Eqn. 5.21, the saturation deficit of the atmosphere is represented by the wet bulb depression, $(T - Tw)$. It has been shown in Section 4.3.1 that the wet bulb depression measured at the climate station is, on average, 10% lower than measured at the UM site. It is probable that the UM data is representative of the catchment in general while the BM data may not be equally representative, (See Section 4.2.1). Therefore it is proposed that when ARBM is tested, the data from the climate station should be increased by 10%. To approximate the average wet bulb depression for each day, the average of the 0900 and 1500 wet bulb depressions should be used. The FORTRAN listing for this is given in Section 5.7.
Fig 5.13  EXAMPLES OF $1/r_a$ vs. $u$ RELATIONSHIPS
5.5 THE DIURNAL VARIATION IN PE

The model described in the preceding sections can be used to model PE as a daily total. However, ARBM often requires evaporation data for periods as short as 30 minutes. Therefore, a model must be developed to distribute PE throughout the day. Presently an empirical model developed by Fleming (1970) is used. Fleming (1970) only used a few days of data to develop his model. He noted that "the results should be taken as preliminary, particularly as regard to cloudy days. It is hoped that other hydrologists may be encouraged to test the method with other data for a wider ranges of conditions". Fleming's model will be tested below.

The procedure is firstly to normalize the time of day in which the evaporation rate is required. The time index (DHOUR) is the time from sunrise expressed as a ratio of the daylength, i.e.

\[
DHOUR = \frac{\text{TIME} - \text{SUNRISE}}{\text{DAYLENGTH}} \quad 5.62
\]

DHOUR varies from 0 to 1 as the time varies from sunrise to sunset. Fleming assumes that no evaporation occurs at night, i.e. before sunrise and after sunset. The model then calculates the average daytime evaporation rate. Using PE, then;

\[
\text{AVDVAP} = \frac{\text{PE}}{\text{DAYLENGTH}} \quad 5.63
\]

The model then distributes AVDVAP proportionately throughout the day. On clear days, the model uses a hyperbolic function. On cloudy days, a truncated triangle function is used. When Williamson (1979) tested ARBM, the basis for selection of the clear or cloudy day model was the ratio of actual pan evaporation to the monthly maximum pan evaporation. Fleming (1970) originally suggested using the ratio of actual global radiation to either clear sky global radiation or extraterrestrial global radiation. He suggested that if the incoming global radiation, \( R_g \), was greater than 0.6 \( R_o \), where
$R_\circ$ is the extra-terrestrial radiation, then the clear day function should be used.

McCullough (1968) proposes a method of calculating $R_\circ$ as a harmonic series throughout the year. His formula is:

$$R_\circ = a_0 + a_1 \cos \theta + a_2 \cos 2\theta + b_1 \sin \theta + b_2 \sin 2\theta$$

where $\theta$ is defined in Appendix C

$R_\circ$ is in langley's (cal/cm$^2$)

McCullough (1968) calculated these constants for various latitudes. For the Warrambine Creek catchment ($\phi=-37.83^\circ$), the constants are:

$$a_0 = 693.9$$
$$a_1 = 362.8$$
$$a_2 = 1.7$$
$$b_1 = 55.0$$
$$b_2 = 1.9$$

To test Fleming's model, actual evaporation data measured at the UM site was compared with model predictions. (See Fig. 5.14a to Fig. 5.14g) The methods cited in Appendix C were used to calculate sunrise and daylength.

The results indicate that the model works very well. Clear sky predictions are excellent (See Fig. 5.14a). For cloudy days, the model is less accurate. However it would be impossible for any model to simulate the erratic evaporation rates which occur as cloud cover changes throughout the day (See Fig. 5.14f). For uniformly overcast days, the model works very well (See Fig. 5.14e and Fig 5.14g).

These results indicate that the present model for distributing evaporation throughout the day is quite satisfactory. Thus, this model is recommended for use in ARBM. A FORTRAN listing of this is given in Section 5.7.
Fig 5.14a  DIURNAL DISTRIBUTION OF LE - 31/10/1979

Date 31/10/1979
Dawn (Hr) 6.4
Daylength (Hr) 13.5
$\frac{R_g}{R_o}$ 0.75
Evaporation, (mm) 5.30

Graph showing diurnal distribution of LE with time (Hr) on the x-axis and evaporation (W/m²) on the y-axis. The graph includes a solid line labeled "Distributed LE" and a dashed line labeled "Measured LE."
Evaporation, LE

(W/m²)

Date: 26/10/1979
Dawn (Hr): 6.5
Daylength (Hr): 13.3
R_g/R_0: 0.73
Evaporation (mm): 5.61

---

Fig 5.14b  DIURNAL DISTRIBUTION OF LE - 26/10/1979
Evaporation, LE

Date                11/12/1979
Dawn (Hr)           6.0
Daylength (Hr)      14.6
R_g/R_o             0.71
Evaporation (mm)    1.99

Fig 5.14c  DIURNAL DISTRIBUTION OF LE - 11/12/1979
Fig 5.14e  DIURNAL DISTRIBUTION OF LE - 4/11/1979

- Date: 4/11/1979
- Dawn (Hr): 6.3
- Daylength (Hr): 13.6
- $R_g/R_o$: 0.54
- Evaporation (mm): 5.74

**Graph Details:**
- X-axis: Time (Hr)
- Y-axis: Evaporation, LE (W/m²)
- Solid line: Distributed LE
- Dashed line: Measured LE
Date: 21/11/1979
Dawn (Hr): 6.1
Daylength (Hr): 14.2
$R_g/R_o$: 0.46
Evaporation (mm): 2.94

Fig 5.14f  DIURNAL DISTRIBUTION OF LE - 21/11/1979
5.6 THE RELATIONSHIP BETWEEN LE and PE

In ARBM, evaporation from the interception (IS) and depression (DS) stores occurs at a potential rate. However, evaporation from the soil stores (US, LS) cannot occur at a potential rate since the evaporating surface is not wet. Thus evaporation from these stores occurs at some rate which is less than PE. The problem is to calculate the degree of reduction below PE.

This is a vexing problem which has confronted micrometeorologists ever since the PE concept was first conceived. As yet, no unique solution to the problem has been found. However the relevant controlling influences are now recognized. For example, it is known that the reduction of evaporation below PE for growing vegetation is a function of the vegetation species, stage of growth, soil moisture stress and under some circumstances, PE itself. For non-transpiring surfaces, the reduction depends on the soil type, moisture content and the amount of dead vegetation covering the surface.

The exact physical processes which restrict the flow of water into the atmosphere are not yet fully understood. Therefore many studies have tried to derive an empirical relationship between the reduction below PE and some other relevant parameter.

The most common relationship developed is between LE/PE and some measure of the soil moisture content. This measure has variously been expressed as a moisture content or deficit using either moisture content or potential units. Examples of this type of study are Denmead & Shaw (1962), Davies & Allen (1973), Nkemdirim & Haley (1973), Ritchie (1973) and Nkemdirim (1976). The basic shape of the relationships obtained is similar. Davies & Allen (1973) proposed an equation to relate LE/PE. The general form of their equation is:

\[
\frac{\text{LE}}{\text{PE}} = a \left(1 - \exp \left(-b \frac{\theta}{\theta_T}\right)\right)
\]

\[5.65\]
where \( a, b \) = regression coefficients
\( \theta \) = surface soil moisture content
\( \theta' \) = field capacity moisture content of
surface soil

These studies usually assumed that when \( \theta = \theta' \) then LE = PE. Thus from Eqn. 5.65 it can be shown that;

\[
a = (1 - \exp(-b))^{-1} \quad 5.66
\]

Thus the selection of 'b' defines 'a'. Eqn. 5.65 is shown plotted in Fig. 5.15 for various values of 'b'. Davies & Allen (1973) determined 'b' to be 10.563 for their situation.

Their approach is often used but if a physically based model is required then it is fundamentally incorrect. Once again the problem lies in the incorrect definition of PE. As has been noted previously, when \( \theta = \theta' \), LE does not necessarily equal PE. If the contention that for short vegetation with unlimited water supply, \( r_s = r_a \) (Thom & Oliver (1977)), then LE/PE can be calculated from Eqn. 5.32. This yields

\[
\begin{align*}
LE &= 0.69 \text{ PE at } T = 10^\circ C \\
LE &= 0.76 \text{ PE at } T = 20^\circ C \\
LE &= 0.82 \text{ PE at } T = 30^\circ C \\
LE &= 0.87 \text{ PE at } T = 40^\circ C
\end{align*}
\]

Clearly the assumption that LE/PE = 1 is open to question. For a physically based model to be used, LE must be related to PE using Eqn. 5.32. Thus it is \( r_s \) which should be modelled.

Many studies have measured \( r_s \). For example, Stewart & Thom (1973), Davies (1972), Fuchs & Tanner (1967), Szeicz & Long (1969) and van Bavel (1967) all have measured the diurnal variation of \( r_s \). Shepherd (1973) and Shepherd (1975) related \( r_s \) to various measures of plant moisture content. He found best correlation between leaf relative water content (RWC) and \( r_s \). Correlations of this type are of little value to ARBM since the input parameter (e.g. RWC, LAI) is not modelled. Szeicz & Long (1969) correlated \( r_s \) with the soil
Figure 5.15  EMPIRICAL RELATIONSHIP BETWEEN LE/PE AND $\theta/\theta'$

The equation for the relationship is:

$$\frac{\text{LE}}{\text{PE}} = a(1 - \exp(-b \frac{\theta}{\theta'}))$$
moisture potential in the top soil layer. They found that $r_s$ had a residual value of 26 s/m at zero moisture deficit for a grass-clover crop.

It was intended in this study to develop a relationship between $r_s$ and soil moisture. However, for the reasons cited in Section 5.4.1 it was impossible to derive a method of accurately calculating PE on those occasions when LE was less than PE. Consequently $r_s$ could not be calculated. In a future study when the relationship between $r_a$ and $u$ is determined, then it will be possible to determine PE and $r_s$.

The more difficult problem will be how to relate $r_s$ to some parameter or parameters which are used in ARBM. For example, Dunin et al (1978) models $r_s$ as a function of solar radiation and incorporates the influences of soil moisture supply and stage of plant development. Perhaps this model could be used in ARBM. Another aspect to consider is the need for separate evaporation models for the vegetated and bare soil areas of the catchment. An example of this is the model developed by Saxton et al (1974).

The determination of the $r_s$ model can be regarded as the determination of another catchment characteristic. Beven (1979) noted that the accurate determination of $r_s$ is important since evaporation estimates are quite sensitive to $r_s$. It is indeed unfortunate that it was not possible to determine $r_s$ in this study. However any future study should be able to concentrate on this aspect now that the equipment and the theory has been developed.

5.7 PROPOSED EVAPORATION MODEL

5.7.1 Programme Variables

A list of most of the ARBM variables used by Williamson and Turner (1980) is given in Appendix A. To their list have been added the new variables proposed in this study. These new FORTRAN variables are listed below with the equivalent symbol used in the text.
(i) Input data from the climate station

RG  Daily total of global radiation (MJ/m²), R_g
T9  Dry bulb temperature at 0900 (°C), T
TW9 Wet bulb temperature at 0900 (°C), Tw
T3  Dry bulb temperature at 1500 (°C), T
TW9 Wet bulb temperature at 1500 (°C), Tw
U   Wind run (km/day), u

(ii) Empirical parametric data

B1, B2, B3, B4,                  Parameters in Eqn. C-2
B5, B6, B7
E1, E2, E3, E4, E5  Parameters in Eqn. C-6
R1, R2, R3, R4, R4  Parameters in Eqn. 5
S1, S2, S3, S4,            Parameters in Eqn. B-20
S5, S6, S7

(iii) Catchment characteristics

AR, BR   "a" and "b" in Eqn. 5
CR       ratio of G to R_n
WR, UR   "a" and "b" in Eqn. 5
AE, BE   as yet undefined parameters in r_s model

(iv) Variables

D2   mean wet bulb depression (°C), T-Tw
E    equation of time (radians), E
G    soil heat flux (MJ/m²), G
H    hour angle (radians), H
IDAY day number, see Eqn. C-3
O    day angle (radians), see Eqn. C-3
Q    ratio of actual global radiation, R_a, to extra-terrestrial global radiation, q_R_o
RA   aerodynamic resistance (s/m), r_a
RGX extra-terrestrial global radiation (MJ/m²), Rₕ
RN1 net radiation (MJ/m²), Rₙ
RN2 Rₙ - G, (MJ/m²)
RN3 energy component of PE (MJ/m²), \( \frac{s}{s + \gamma} (Rₙ - G) \)
RN4 aerodynamic component of PE (MJ/m²)
\( \rho Cₚ (T - T_w)/rₐ \)
RS surface resistance (s/m), rₛ
SL slope of the saturation vapour pressure
versus temperature curve (mb/°C), s
T mean dry bulb temperature (°C), T

5.7.2 Subroutine "DAY"

This subroutine has been developed to calculate DAWN, DAYLEN and RGX. The formula used are given in Appendix C and Section 5.5. The FORTRAN listing is shown in Fig. 5.16. These parameters are slowly varying throughout the year. It is probable that little error will be introduced if this subroutine is only executed once in every 5, 7 or 10 days of simulation. This would save some computer time. This aspect should be investigated in some future study.

5.7.3 Subroutine "PEVAP"

This subroutine calculates daily totals of potential evaporation, PE. It uses data collected at the climate station and the formulae developed in Sections 5.2, 5.3 and 5.4. The FORTRAN listing is given in Fig. 5.17.

5.7.4 Subroutine "EVAPO"

This subroutine determines the amount of potential evaporation, PEJ, which occur in the current time interval. PEJ is zero before and after sunrise and sunset respectively. During the day, PEJ is determined using the model developed by Fleming (1970). (See Section 5.5) The FORTRAN listing of this is given in Fig. 5.18.
SUBROUTINE DAY
COMMON /A/ ............... 
DIMENSION IK (12)
DATA (IK(L),L=1,12)/0,31,59,90,120,151,181,212,243,
               273,304,334/
DATA R1,R2,R3,R4,R5 / 693.9,362.8,1.7,-55.,1.7/
DATA B1,B2,B3,B4,B5,B6,B7/6.918E-3,-.399,7.026E-2,
               -6.758E-3,
C9.07E-4,-2.697E-3,1.48E-3/
DATA E1,E2,E3,E4,E5/7.5E-5,1.868E-3,-3.21E-2,-1.46E-
               2,-4.085E-2/
IDAY=ID+IK(IK)
O=0.01721*(IDAY-1)
B=B1+B2*COS(0)+B3*SIN(0)+B4*COS(0*2)+B5*SIN(0*2)+B6
   *SIN(0*3)+
CB7*SIN(0*3)
EQN=E1+E2*COS(0)+E3*SIN(0)+E4*COS(2*0)+E5*SIN(2*0)
H=3.8197*ACOS(TAN(B)*0.7766)
DAYLEN=2.*H
DAWN=12.4-H-EQN*3.8197
RGX=R1+R2*COS(0)+R3*COS(2*0)+R4*SIN(0)+R5*SIN(2*0)
RGX=RGX/23.0
RETURN
END

FIG. 5.16 SUBROUTINE "DAY"
SUBROUTINE PEVAP
COMMON /A/ ............
INTEGER U
DATA S1,S2,S3,S4,S5,S6,S7/.4438,2.857E-2,7.938E-4,
1.2152E-5,
C1.03656E-7,3.5324E-10,-7.0902E-13/
DATA AR,BR/0.046,0.538/
DATA CR/0.04/
DATA WR,UR/0.0156,8.41E-3/
IF(IM.GE.10 .OR. IM.LE.3) GO TO 229
AR=1.145
BR=0.467
CR=0.06
WR=0.0588
UR=3.18E-3
CONTINUE
T=(TD3+TD9)/2.
SL=S1+T*(S2+T*(S3+T*(S4+T*(S5+T*(S6+T*S7))))))
RN1=AR*BR*RG
G=CR
RN2=(1.-G)*RN1
RN3=SL*RN2/(SL+0.66)
DZ=((TD9-TW9)+(TD3-TW3))/2.
DZ=DZ*1.1
RA=1/(WR+UR*U/(24*3.6))
RN4=1.21*3.6*DZ*DAYLEN/RA
PE=RN3+RN4
PE=PE/2.45
RETURN
END

FIG. 5.17 SUBROUTINE "PEVAP"
SUBROUTINE EVAPO
COMMON /A/ .......... 
ICLEAR=0 
Q=RG/RGX 
IF(Q.GE.0.6) ICLEAR=1 
AVDVAP=PE/DAYLEN/60.
IF(TIME.LT.DAWN.OR.TIME.GT.(DAWN+DAYLEN)) GO TO 2905 
DHOUR=(TIME-DAWN)/DAYLEN 
IF(ICLEAR.EQ.0) GO TO 1930 
AAA=31.4*DHOUR*DHOUR-33.6*DHOUR+2.23 
EVI=2.78-SQRT(7.73+AAA) 
IF(EVI.LT.0.) EVI=0.
GO TO 2010
1930 IF(DHOUR.GE.0.26) GO TO 1980 
EVI=6.75*(DHOUR-0.06) 
IF(EVI.LT.0.) EVI=0.
GO TO 2010
1980 EVI=1.35-6.75*(DHOUR-0.8) 
IF(EVI.GT.1.35) EVI=1.35 
2010 PEJ=EVI*AVDVAP*TI 
GO TO 2910 
2905 PEJ=0.
2910 CONTINUE 
RETURN 
END

FIG. 5.18 SUBROUTINE "EVAPO"
5.7.5 Evaporation Model

This is the section of the model shown as a flow chart in Fig. 5.3. Where possible, the same variables and line numbers as Williamson and Turner (1980) are used. It is intended that this section of programme could replace the equivalent section in the present ARBM. The FORTRAN listing of this is given in Fig. 5.19. As yet, the form of the equation defining RS is unknown.

5.8 THE RELATIONSHIP BETWEEN PE AND PAN EVAPORATION

Williamson and Turner (1980) modelled PE by simply equating it with pan evaporation measured at the climate station. It would be desirable to test their hypothesis.

As mentioned previously, PE cannot be accurately determined yet because the aerodynamic resistance, \( r_a \) is unknown. However it is possible to approximate \( r_a \) using Eqn. 5.60 with the wind speed data collected at the climate station. Using this data, PE can be calculated using net radiation, soil heat flux and wet bulb depression as measured at the study site. It is probable that the aerodynamic term will be in error. However the term is likely to be consistently larger or smaller than the true value. Thus, this method will either consistently over-estimate or under-estimate PE. In any event, the comparison of this estimate of PE with pan evaporation will indicate the degree of correlation. This is shown in Fig. 5.20. As can be seen, considerable scatter between daily totals exists.

The usefulness of pan evaporation as a model for PE should be further investigated. It should be noted that this is an empirical model and no catchment characteristics are explicitly contained in the model. Therefore, hydrological changes caused by land use changes cannot be readily examined.

5.9 SUMMARY

ARBM is a physically based hydrological process model. The sequence of calculations in the evaporation model were
Estimated PE - see below (mm/day)

\[ \text{PE} = \text{LEQ} + \text{LEA} \]

\[ \text{LEQ} = \text{measured totals at UM site} \]

\[ \text{LEA} = \frac{C_p (T - T_w)}{r_a} \]

\( (T - T_w) \) = mean of wet bulb depression at climate station

\[ \frac{1}{r_a} = 0.0059 + 0.00319 \times U \]

\( U \) = wind run at climate station

Priestly & Taylor, \( \eta = \frac{\text{PE}}{\text{LEQ}} \)

for 49 daily averages, \( \bar{\eta} = 1.27 \) and \( \sigma = 0.134 \)

Fig 5.20 PLOT OF PAN EVAPORATION vs. PE
2290 IF(PEJ.EQ.0.) GO TO 2560
2295 IF(PEJ.GT.IS) GO TO 2340
2300 IS=IS-PEJ
2305 PEC=PEJ
2310 GO TO 2560
2320 PET=PEJ-IS
2325 PEC=IS
2330 IS=0.
2335 IF(PET.GE.DS) GO TO 2390
2340 PEC=PEC+PET
2345 DS=DS-PET
2350 GO TO 2560
2360 PET=PET-DS
2365 PEC=PEC+DS
2370 DS=0.
2380 RS=..........
2390 FU=PET*(SL=0.66)/(SL+0.66*(1+RS/RA))
2400 UE=ER*FU
2405 PEC=PEC+UE
2410 UEC=UEC+UE
2415 US=US-UE
2420 LE=(1.-ER)*FU
2425 PEC=PEC+LE
2430 LEC=LEC+LE
2435 LS=LS-LE
2440 GO TO 2565
2450 LE=0.
2455 UE=LE

FIG. 5.19 EVAPORATION SECTION OF ARBM
examined. It was found that this is a conceptually correct model of the physical processes involved.

An important parameter in this model is potential evaporation, PE. The definition of PE has been examined in detail. PE has been re-defined and formulae have been derived to predict this term. A model of PE is proposed for ARBM. This model includes certain catchment characteristics which must be determined for each catchment. These characteristics can be modified to simulate land use changes in the catchment.

The model which simulates the diurnal variation in evaporation was examined. It was found that predictions from this model closely followed the diurnal variation of evaporation as measured in the field.

The most important segment of the evaporation model has yet to be determined. This is the model of surface resistance, \( r_s \). This model controls the rate of drying of the catchment.
Chapter 6

CONCLUSIONS

6.0 SUMMARY AND CONCLUSIONS

6.1 RECOMMENDATIONS
6.0 SUMMARY AND CONCLUSIONS

The original aim of this thesis was to measure evaporation from a typical area within a representative basin and to propose and test a physically based prediction model compatible with ARBM. All of these objectives were not attained. Perhaps the scope of the work was too broad for one thesis. However some conclusions were obtained. The most pertinent of these are:

(i) The energy balance method is a reliable and convenient technique for the measurement of evaporation in a pastoral representative basin. The system is compact and can be made mobile. Thus spatial variations in evaporation can be studied. Importantly, many relevant meteorological parameters are measured.

A detailed error analysis of this method was performed. This analysis revealed that for evaporation to be accurately determined, $R_n$, $\Delta T$ and $\Delta Tw$ must be carefully measured. The analysis also revealed that under dry surface conditions, erratic results can be expected. This was confirmed by experimental evidence.

The data collected with this type of equipment can be used to test and propose changes to the evaporation model in ARBM.

(ii) The eddy correlation method is the most fundamental method of measuring evaporation. However its routine application for long term studies is not yet possible. Several aspects need further development.

In particular it is seen as desirable to replace the present analogue covariance computers with microprocessors. The use of a microprocessor offers many advantages, viz. greater system flexibility; enhanced computational accuracy; use of non-linear sensors, collection of additional data and the elimination of flux loss due to low frequency filtering. This study developed the modifications to conventional eddy correlation theory to make it compatible with the characteristics of a microprocessor.
A detailed review of the theory indicates that a three-dimensional wind sensor is necessary at the Warrambine Creek site. This is necessary to determine the "vertical" component of the wind. The microprocessor can be used to resolve the wind components.

The use of an infrared hygrometer to measure the humidity component is recommended. However, further development is required. This study has provided improvements in the physical and electronic design of the hygrometer. Also, further understanding of the performance of apparently identical hygrometers has been obtained.

It has been discovered that the strength of the transmission function can vary markedly from one instrument to another. Further work on the strength of the transmission function is required.

(iii) The predictive performance of ARBM depends on the physical soundness of the model and the reliability and representativeness of the input data. An examination of the catchment's climate station indicated that its siting may not be representative of the catchment as a whole. Thus, the data collected at the climate station was compared with data collected at the study site.

It was shown that dry bulb temperature data and rainfall data correlated well. Differences in wet bulb temperatures were noted but these can be largely explained by the lack of aspiration of the climate station's thermometers. However, pan evaporation correlated poorly. The climate station's data appears to underestimate pan evaporation in the catchment. This is probably due to the sheltered site. Unfortunately, comparisons of global radiation data only revealed that the instrument at the climate station was faulty. No comparison of wind speed data was possible.

(iv) One aim of the Representative Basins project is to develop a physically based hydrological model. Where possible, empiricism should be eliminated. Data collected during this study indicated that the present model is concept-
ually correct. A review of the physical process of evaporation showed that a physically based model using measured catchment characteristics can be proposed.

A central part of the evaporation model is the parameter called potential evaporation. A detailed review indicated that there are substantial differences in the interpretation of this parameter. Consequently, potential evaporation was re-defined in more exact physical and mathematical terms. It was shown that using this definition, the combination formulae can be derived to predict potential evaporation. Unfortunately an initial misconception of the definition of potential evaporation meant that the aerodynamic and surface resistance teams could not be determined. Many previous studies have been equally incorrectly conceived. Their results must be open to question. A model to predict potential evaporation was developed. It requires catchment characteristics defining the radiative and roughness properties of the catchment.

The present model for the diurnal distribution of evaporation was examined. It was found to be quite satisfactory providing that sunrise and daylength are accurately determined. Subroutines to determine sunrise, daylength and the diurnal variation of evaporation were developed.

The important catchment characteristic, viz. the relationship between surface resistance and soil moisture content, has not been determined.

(v) The original objectives of this study were not attained. This occurred for many reasons. The main problem was the extremely lengthy period of development of electronic equipment. However, the theory and equipment has now been developed. This will allow the rapid attainment of the original aims in some future study in the Warrambine Creek catchment.

6.1 RECOMMENDATIONS

Certain recommendations are offered for the measurement and modelling of evaporation in a Representative Basin.
These recommendations can be grouped into two categories; those relating to the catchment and equipment described in this thesis and those relating to the Representative Basins project in general.

A. Recommendations for extensions of this study:

(i) It is recommended that the data logger used in the energy balance method be modified to halt the integration of the $\Delta T$ and $\Delta Tw$ channels for two minutes following each sensor interchange. (See Section 2.2.7) Also, wind speed at 2 m should be measured. This could be achieved using the existing data logger by measuring soil heat flux on one channel only.

(ii) The development of the eddy correlation system should continue. The microprocessor system could be made more useful by modifying it to measure additional parameters. The more important of these would be net and global radiation, soil heat flux wet bulb depression and wind speed at 2 m. With the energy balance system remaining at the present site, the eddy correlation system could then be moved about in the catchment. The spatial variation of many meteorological parameters could then be assessed. This data could then be used in the distributed parameter model used by Williamson and Turner (1980). Since there are presently a number of spare channels on the analogue-to-digital converter, the measurement of additional parameters would only involve a programme modification. Other recommendations for the microprocessor are given in Section 3.2.6.

(iii) The reliability and representativeness of the global radiation and wind speed data measured at the climate station should be assessed.

(iv) The measurement and modelling of surface resistance and its related parameters should be examined in detail.

(v) A sensitivity analysis of Eqn. 5.31 should be undertaken using the methods given in Section 2.3. This analysis should indicate the degree of measurement and
modelling accuracy required for each parameter; notably $R_n$, $r_a$ and $r_s$.

(vi) One aim of the Representative Basins project is the prediction of the effect of land use changes in a catchment. One change which often occurs in parts of this catchment is bush-fire. Areas of the catchment are burnt intentionally and unintentionally each year. The relevant catchment characteristics of a burnt surface should be measured. Thus, the hydrological results of an extensive bush-fire could be modelled.

(vii) It has been shown that global radiation is the primary variable used in the modelling of evaporation. However, global radiation has only recently been measured at this basin's climate station and is generally not available at other sites. Few measurement stations for global radiation presently exist. However, the reasonable correlation between data collected at the study site and data collected over 100 km away (See Fig. 4.11) raises an interesting possibility.

It seems reasonable to assume that global radiation will be less affected by the local environment than many other meteorological factors. For example, local topography could affect wind speed and direction and this coupled with altitude would influence temperature. However local topography and altitude should not affect global radiation to the same degree. Therefore global radiation may be relatively homogenous over large areas particularly if variations in latitude are compensated. The main factor influencing global radiation is cloud cover. This may be similar over areas affected by the same weather systems. Thus it might be possible to use global radiation data from a station some distance from the catchment. This would enormously expand the range of application of the few existing global radiation measurement sites.

To test this hypothesis, global radiation data from the study site could be compared with data from the Laverton station. This is approximately 60 km east of the catchment.
If the correlation is acceptable, previous Laverton data could be used to test the model for the same years when Williamson and Turner (1980) studied the catchment.

B. Recommendations for the Representative Basins project:

(i) At the start of any modelling study in a Representative Basin, the data from the basin's climate station should be assessed to determine whether it is representative and reliable. This would involve a subjective assessment of the climate station. Points to note are the location within the catchment, local environment and method of data collection. Also, the data collected at the station should be correlated with data from the nearest Bureau of Meteorology station. As most climate data is now stored on computer files, this would not be an overly difficult task. However, it could quickly give an indication of the reliability of the data. It would be more desirable to compare the data with alternate data collected in the catchment. Such a procedure could prevent a considerable waste of effort in testing a model with poor data.
REFERENCES


HARRISON, L.P. (1965(b)). "Some fundamental considerations regarding psychrometry." Symp. on 'Humidity and Moisture', R.E. Ruskin (ed.), Reinhold, Vol. 1, (2) 71-


APPENDIX A

LIST OF SYMBOLS

Listed below are the symbols used in the text of this thesis along with the value of constants, where appropriate. The S.I. units for each symbol are also given. The symbols in brackets behind some definitions are alternate symbols for that parameter which often appear in the literature. A small number of symbols used in the text are not listed. These symbols have limited usage and are defined in the text.

<table>
<thead>
<tr>
<th>SYMBOL</th>
<th>DEFINITION</th>
<th>UNITS</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>Psychrometric constant (See Appendix B)</td>
<td>-</td>
</tr>
<tr>
<td>a</td>
<td>Absolute humidity</td>
<td>kg m(^{-3})</td>
</tr>
<tr>
<td>a, ..., g</td>
<td>Various correlation coefficients</td>
<td>-</td>
</tr>
<tr>
<td>b</td>
<td>Ratio of the molecular weight of water to the molecular weight of air, 0.62198 ((\epsilon, \epsilon))</td>
<td>-</td>
</tr>
<tr>
<td>C</td>
<td>Heat capacity</td>
<td>J kg(^{-1}) (^{\circ})K(^{-1})</td>
</tr>
<tr>
<td>(C_p)</td>
<td>Specific heat of air at constant pressure, 1010</td>
<td>J kg(^{-1}) (^{\circ})K(^{-1})</td>
</tr>
<tr>
<td>c</td>
<td>Mass of absorbent per unit area</td>
<td>kg m(^{-2})</td>
</tr>
<tr>
<td>D</td>
<td>Wet bulb depression</td>
<td>(^{\circ})C</td>
</tr>
<tr>
<td>d</td>
<td>Zero plane displacement (D)</td>
<td>(^{\circ})C</td>
</tr>
<tr>
<td>E</td>
<td>Equation of time</td>
<td>min.</td>
</tr>
<tr>
<td>E</td>
<td>Flux of water vapour, evaporation</td>
<td>kg s(^{-1}) m(^{-2})</td>
</tr>
<tr>
<td>e</td>
<td>Vapour pressure of water</td>
<td>mb</td>
</tr>
<tr>
<td>f</td>
<td>Normalized frequency</td>
<td>-</td>
</tr>
<tr>
<td>G</td>
<td>Soil heat flux ((S, G_n))</td>
<td>W m(^{-2})</td>
</tr>
<tr>
<td>g</td>
<td>Acceleration due to gravity, (9.81)</td>
<td>m s(^{-2})</td>
</tr>
<tr>
<td>SYMBOL</td>
<td>DEFINITION</td>
<td>UNITS</td>
</tr>
<tr>
<td>--------</td>
<td>------------------------------------------------</td>
<td>--------------</td>
</tr>
<tr>
<td>H</td>
<td>Sensible heat flux (A)</td>
<td>W m⁻²</td>
</tr>
<tr>
<td>H</td>
<td>Hour angle</td>
<td>rad.</td>
</tr>
<tr>
<td>h</td>
<td>Turbulent transfer coefficient</td>
<td>m s⁻¹</td>
</tr>
<tr>
<td>h</td>
<td>Relative humidity</td>
<td>-</td>
</tr>
<tr>
<td>I₀</td>
<td>Solar constant (1353)</td>
<td>W m⁻²</td>
</tr>
<tr>
<td>K_H</td>
<td>Eddy conductivity or eddy diffusivity of sensible heat</td>
<td>m² s⁻¹</td>
</tr>
<tr>
<td>K_M</td>
<td>Eddy diffusivity of momentum or eddy viscosity</td>
<td>m² s⁻¹</td>
</tr>
<tr>
<td>K_W</td>
<td>Eddy diffusivity of water vapour (Kᵥ, Kₑ)</td>
<td>m² s⁻¹</td>
</tr>
<tr>
<td>k</td>
<td>Thermal conductivity</td>
<td>W m⁻² °C⁻¹</td>
</tr>
<tr>
<td>k</td>
<td>von Karman's constant</td>
<td>-</td>
</tr>
<tr>
<td>LAI</td>
<td>Leaf area index</td>
<td>-</td>
</tr>
<tr>
<td>L</td>
<td>Latent heat of vapourization of water (λ), See Appendix B</td>
<td>J kg⁻¹</td>
</tr>
<tr>
<td>LE</td>
<td>Latent heat flux, actual evaporation</td>
<td>W m⁻²</td>
</tr>
<tr>
<td>LEQ</td>
<td>Equilibrium evaporation</td>
<td>W m⁻²</td>
</tr>
<tr>
<td>LEₐ</td>
<td>Term in Penman's formula (See Chapter 5)</td>
<td>-</td>
</tr>
<tr>
<td>m</td>
<td>Mass</td>
<td>kg</td>
</tr>
<tr>
<td>N</td>
<td>Possible number of hours of sunshine per day, daylength</td>
<td>hrs.</td>
</tr>
<tr>
<td>n</td>
<td>Actual number of hours of sunshine per day</td>
<td>hrs.</td>
</tr>
<tr>
<td>P</td>
<td>Precipitation</td>
<td>mm</td>
</tr>
<tr>
<td>P</td>
<td>Photosynthesis</td>
<td>W m⁻²</td>
</tr>
<tr>
<td>PE</td>
<td>Potential evaporation (E_p, E_o)</td>
<td>W m⁻²</td>
</tr>
<tr>
<td>p</td>
<td>Pressure (P)</td>
<td>mb</td>
</tr>
<tr>
<td>q</td>
<td>Specific humidity</td>
<td>kg/kg</td>
</tr>
<tr>
<td>R*</td>
<td>Universal gas constant (8.31432) J⁻¹ K⁻¹ mole⁻¹</td>
<td>-</td>
</tr>
<tr>
<td>SYMBOL</td>
<td>DEFINITION</td>
<td>UNITS</td>
</tr>
<tr>
<td>--------</td>
<td>------------</td>
<td>-------</td>
</tr>
<tr>
<td>( R_a )</td>
<td>Atmospheric radiation (L+)</td>
<td>W m(^{-2})</td>
</tr>
<tr>
<td>( R_g )</td>
<td>Global radiation, insolation (K, ( Q_s ), S)</td>
<td>W m(^{-2})</td>
</tr>
<tr>
<td>( R_i )</td>
<td>Richardson's number</td>
<td>-</td>
</tr>
<tr>
<td>( R_n )</td>
<td>Net radiation (( R ), ( Q_n ), ( Q^* ), ( K^+ ))</td>
<td>W m(^{-2})</td>
</tr>
<tr>
<td>( R_o )</td>
<td>Extra-terrestrial radiation</td>
<td>W m(^{-2})</td>
</tr>
<tr>
<td>( R_t )</td>
<td>Terrestrial radiation (L+)</td>
<td>W m(^{-2})</td>
</tr>
<tr>
<td>( s )</td>
<td>Slope of the saturation vapour pressure vs. temperature curve (( \Delta ), S)</td>
<td>mb ( ^\circ )C(^{-1})</td>
</tr>
<tr>
<td>( T )</td>
<td>Temperature</td>
<td>( ^\circ )C, ( ^\circ )K</td>
</tr>
<tr>
<td>( Tw )</td>
<td>Wet bulb temperature</td>
<td>( ^\circ )C</td>
</tr>
<tr>
<td>( Td )</td>
<td>Dew point temperature</td>
<td>( ^\circ )C</td>
</tr>
<tr>
<td>( t )</td>
<td>Time</td>
<td>s</td>
</tr>
<tr>
<td>( U )</td>
<td>Mean horizontal wind</td>
<td>m s(^{-1})</td>
</tr>
<tr>
<td>( u^* )</td>
<td>Friction velocity</td>
<td>m s(^{-1})</td>
</tr>
<tr>
<td>( u, v, w )</td>
<td>Components of the total wind vector in the x, y and z directions respectively</td>
<td>m s(^{-1})</td>
</tr>
<tr>
<td>( v )</td>
<td>Volume</td>
<td>m(^3)</td>
</tr>
<tr>
<td>( V )</td>
<td>Voltage</td>
<td>V</td>
</tr>
<tr>
<td>( x, y, z )</td>
<td>Directional co-ordinate system with z perpendicular to the surface and x in the direction of the mean horizontal wind</td>
<td>-</td>
</tr>
<tr>
<td>( z_o )</td>
<td>Roughness length</td>
<td>m</td>
</tr>
<tr>
<td>( Z )</td>
<td>Zenith angle</td>
<td>rad.</td>
</tr>
<tr>
<td>( \alpha )</td>
<td>Albedo, reflectivity (( r, \lambda ))</td>
<td>-</td>
</tr>
<tr>
<td>( \beta )</td>
<td>Bowen ratio</td>
<td>-</td>
</tr>
<tr>
<td>( \beta )</td>
<td>Heating coefficient</td>
<td>-</td>
</tr>
<tr>
<td>( \gamma )</td>
<td>Psychrometric constant, see Appendix B</td>
<td>mb ( ^\circ )C(^{-1})</td>
</tr>
<tr>
<td>SYMBOL</td>
<td>DEFINITION</td>
<td>UNITS</td>
</tr>
<tr>
<td>--------</td>
<td>------------</td>
<td>-------</td>
</tr>
<tr>
<td>δ</td>
<td>Solar declination</td>
<td>rad.</td>
</tr>
<tr>
<td>ε</td>
<td>Emissivity</td>
<td>-</td>
</tr>
<tr>
<td>φ</td>
<td>Latitude (L)</td>
<td>rad.</td>
</tr>
<tr>
<td>ρ</td>
<td>Density (unless subscripted, density of air)</td>
<td>kg m(^{-3})</td>
</tr>
<tr>
<td>θ</td>
<td>Potential temperature</td>
<td>(^{\circ})C</td>
</tr>
<tr>
<td>θ</td>
<td>Day angle of the year, see Appendix C</td>
<td>rad.</td>
</tr>
<tr>
<td>θ</td>
<td>Soil moisture content</td>
<td>-</td>
</tr>
<tr>
<td>σ</td>
<td>Stefan-Boltzmann constant</td>
<td>W m(^{-2})(^{\circ})K(^{-4})</td>
</tr>
<tr>
<td>τ</td>
<td>Transmission</td>
<td>-</td>
</tr>
<tr>
<td>τ</td>
<td>Shearing stress, vertical flux of horizontal momentum</td>
<td>kg m(^{-1}) s(^{-2})</td>
</tr>
<tr>
<td>ψ</td>
<td>Absorption of the atmosphere</td>
<td>-</td>
</tr>
<tr>
<td>φ</td>
<td>Diabatic wind function</td>
<td>-</td>
</tr>
<tr>
<td>ω</td>
<td>Frequency</td>
<td>Hz</td>
</tr>
<tr>
<td>Β</td>
<td>Dry diabatic lapse rate</td>
<td>(^{\circ})C m(^{-1})</td>
</tr>
</tbody>
</table>

**Subscripts**

- **a**: Air
- **c**: Crop
- **s**: Soil
- **w**: Water (except Tw)
- **2,1**: At heights 2 and 1 (z\(_2\) > z\(_1\))
- **0**: At height, z = 0 i.e. surface level

**Others**

- **x̄**: The mean value of x
- **x'**: The fluctuating component of x and defined by x = x̄ + x'
- **Δx**: The difference between x\(_1\) and x\(_2\)
- **δx**: A small change (error) in x
- **x***: The saturation value of x, e.g. e*, q* (except R*, u*)
- **f(x)**: Some function of x
LIST OF ARBM PROGRAMME VARIABLES

Most of the variables used by Williamson and Turner (1980) and all the new variables proposed in this study are given below. Most of these variables are real unless the initial letter is I, J, K, L, M or N which are integer. Exceptions are noted. For a complete listing of ARBM, See Williamson and Turner (1980).

<table>
<thead>
<tr>
<th>VARIABLE NAME</th>
<th>EXPLANATION</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>Saturation hydraulic conductivity for upper soil store</td>
</tr>
<tr>
<td>AAA</td>
<td>Parameter in Fleming (1970) model</td>
</tr>
<tr>
<td>AE</td>
<td>Factor for evaporation from soil stores</td>
</tr>
<tr>
<td>AL</td>
<td>As for A, but lower soil store</td>
</tr>
<tr>
<td>ALO</td>
<td>Predicted unsaturated conductivity for lower store</td>
</tr>
<tr>
<td>AL3</td>
<td>As for A3, but lower soil store</td>
</tr>
<tr>
<td>AO</td>
<td>Predicted unsaturated conductivity for upper store</td>
</tr>
<tr>
<td>AR</td>
<td>Catchment characteristic, Eqn. 5.43</td>
</tr>
<tr>
<td>AVDVAP</td>
<td>Average daytime potential evaporation rate</td>
</tr>
<tr>
<td>AO0</td>
<td>Second parameter of Philip infiltration equation</td>
</tr>
<tr>
<td>Al</td>
<td>Ratio of upper and lower store capacities (USC/LSC)</td>
</tr>
<tr>
<td>A2</td>
<td>Bias term for redistribution between soil stores</td>
</tr>
<tr>
<td>A3</td>
<td>Exponent for decrease of unsaturated hydraulic conductivity with decreased moisture content for upper store</td>
</tr>
<tr>
<td>B</td>
<td>Solar declination</td>
</tr>
<tr>
<td>BE</td>
<td>Similar to AE</td>
</tr>
<tr>
<td>BR</td>
<td>Similar to AR</td>
</tr>
<tr>
<td>VARIABLE NAME</td>
<td>EXPLANATION</td>
</tr>
<tr>
<td>--------------</td>
<td>-------------</td>
</tr>
<tr>
<td>B1,B2,B3,B4</td>
<td>Parameters in Eqn. C-2</td>
</tr>
<tr>
<td>B5,B6,B7</td>
<td></td>
</tr>
<tr>
<td>CR</td>
<td>Ratio of soil heat flux to net radiation, G/Rn</td>
</tr>
<tr>
<td>CRK</td>
<td>Moisture limit below which the upper soil store is cracked</td>
</tr>
<tr>
<td>CRKL</td>
<td>As for CRK but lower soil store</td>
</tr>
<tr>
<td>DAWN</td>
<td>Time (hours) at which sunrise occurs</td>
</tr>
<tr>
<td>DAYLEN</td>
<td>Daylength (hours)</td>
</tr>
<tr>
<td>DHOUR</td>
<td>Normalized time, Eqn. 5.62</td>
</tr>
<tr>
<td>DS</td>
<td>Current content of depression store</td>
</tr>
<tr>
<td>DSC</td>
<td>Capacity (maximum allowable content) of depression store</td>
</tr>
<tr>
<td>DZ</td>
<td>Mean wet bulb depression</td>
</tr>
<tr>
<td>ER</td>
<td>Proportion of evaporation from upper soil (proportion of catchment covered by shallow rooted vegetation)</td>
</tr>
<tr>
<td>EVI</td>
<td>Parameter in Fleming (1970) model</td>
</tr>
<tr>
<td>E1,E2,E3,E4,E5</td>
<td>Parameters in Eqn. C-6</td>
</tr>
<tr>
<td>FL</td>
<td>Predicted evaporation from lower store</td>
</tr>
<tr>
<td>FU</td>
<td>As for FL, but upper soil store</td>
</tr>
<tr>
<td>G</td>
<td>Soil heat flux/net radiation</td>
</tr>
<tr>
<td>GN</td>
<td>Groundwater recession factor</td>
</tr>
<tr>
<td>GR</td>
<td>Groundwater runoff</td>
</tr>
<tr>
<td>GS</td>
<td>Current content of the groundwater store (unlimited)</td>
</tr>
<tr>
<td>H</td>
<td>Hour angle at sunrise and sunset</td>
</tr>
<tr>
<td>IAR</td>
<td>Interception area ratio (REAL), the proportion of the surface covered by vegetation to collect interception</td>
</tr>
<tr>
<td>ICLEAR</td>
<td>Flag to indicate clear or cloudy day</td>
</tr>
<tr>
<td>ID</td>
<td>Day of month</td>
</tr>
<tr>
<td>IDAY</td>
<td>Day of year</td>
</tr>
<tr>
<td>VARIABLE NAME</td>
<td>EXPLANATION</td>
</tr>
<tr>
<td>---------------</td>
<td>-------------</td>
</tr>
<tr>
<td>IK</td>
<td>Array to give cumulative day of year</td>
</tr>
<tr>
<td>IM</td>
<td>Month of year</td>
</tr>
<tr>
<td>INF</td>
<td>Infiltration (REAL) in time period, TI</td>
</tr>
<tr>
<td>IS</td>
<td>Current content of interception store (REAL)</td>
</tr>
<tr>
<td>ISC</td>
<td>Capacity of interception store</td>
</tr>
<tr>
<td>KG</td>
<td>Groundwater recession constant</td>
</tr>
<tr>
<td>LDR</td>
<td>Amount of drainage from lower store (REAL)</td>
</tr>
<tr>
<td>LE</td>
<td>Amount of evaporation from lower store (REAL)</td>
</tr>
<tr>
<td>LEC</td>
<td>Daily total of LE (REAL)</td>
</tr>
<tr>
<td>LS</td>
<td>Current content of lower soil store (REAL)</td>
</tr>
<tr>
<td>LSC</td>
<td>Capacity of lower soil store (REAL)</td>
</tr>
<tr>
<td>O</td>
<td>Day angle</td>
</tr>
<tr>
<td>P</td>
<td>Precipitation in this time interval</td>
</tr>
<tr>
<td>PE</td>
<td>Potential evaporation for current day</td>
</tr>
<tr>
<td>PEC</td>
<td>Total modelled evaporation for current time period</td>
</tr>
<tr>
<td>PECC</td>
<td>Total modelled evaporation for current day</td>
</tr>
<tr>
<td>PEJ</td>
<td>Potential evaporation in current time interval</td>
</tr>
<tr>
<td>PET</td>
<td>Remaining evaporation potential</td>
</tr>
<tr>
<td>PI</td>
<td>Equivalent to P</td>
</tr>
<tr>
<td>PIN</td>
<td>Proportion of PI entering the interception store</td>
</tr>
<tr>
<td>PT</td>
<td>Proportion of PI entering the depression store (through fall)</td>
</tr>
<tr>
<td>Q</td>
<td>Ratio of actual global radiation to extra-terrestrial global radiation</td>
</tr>
<tr>
<td>RA</td>
<td>Aerodynamic resistance</td>
</tr>
<tr>
<td>VARIABLE NAME</td>
<td>EXPLANATION</td>
</tr>
<tr>
<td>---------------</td>
<td>-------------</td>
</tr>
<tr>
<td>RD</td>
<td>Amount of water redistributed between soil stores</td>
</tr>
<tr>
<td>RDI</td>
<td>Direction decision for redistribution</td>
</tr>
<tr>
<td>RE</td>
<td>Rainfall excess (runoff)</td>
</tr>
<tr>
<td>RG</td>
<td>Daily total of global radiation</td>
</tr>
<tr>
<td>RGX</td>
<td>Daily total of extra-terrestrial global radiation</td>
</tr>
<tr>
<td>RN1</td>
<td>Net radiation, $R_n$</td>
</tr>
<tr>
<td>RN2</td>
<td>$(R_n - G)$</td>
</tr>
<tr>
<td>RN3</td>
<td>Energy component of PE</td>
</tr>
<tr>
<td>RN4</td>
<td>Aerodynamic component of PE</td>
</tr>
<tr>
<td>RS</td>
<td>Surface resistance</td>
</tr>
<tr>
<td>R1, R2, R3, R4, R5</td>
<td>Parameters in Eqn. 5.64</td>
</tr>
<tr>
<td>S</td>
<td>Current predicted value of sorptivity</td>
</tr>
<tr>
<td>SO</td>
<td>Dry soil sorptivity</td>
</tr>
<tr>
<td>SL</td>
<td>Slope of the saturation vapour pressure versus temperature curve (mb/°C)</td>
</tr>
<tr>
<td>S1, S2, S3, S4, S5, S6, S7</td>
<td>Parameters in Eqn. B-20</td>
</tr>
<tr>
<td>T</td>
<td>Mean dry bulb temperature</td>
</tr>
<tr>
<td>TD3</td>
<td>Dry bulb temperature at 1500</td>
</tr>
<tr>
<td>TD9</td>
<td>Dry bulb temperature at 0900</td>
</tr>
<tr>
<td>TF</td>
<td>Current time interval as a fraction of a day</td>
</tr>
<tr>
<td>TI</td>
<td>Current time interval (minutes)</td>
</tr>
<tr>
<td>TIME</td>
<td>Current time (hours)</td>
</tr>
<tr>
<td>TWF</td>
<td>Time since last infiltration (time wetting front)</td>
</tr>
<tr>
<td>TWFMAX</td>
<td>Time before S is recalculated</td>
</tr>
<tr>
<td>TW3</td>
<td>Wet bulb temperature at 1500</td>
</tr>
<tr>
<td>TW9</td>
<td>Wet bulb temperature at 0900</td>
</tr>
<tr>
<td>VARIABLE NAME</td>
<td>EXPLANATION</td>
</tr>
<tr>
<td>--------------</td>
<td>-------------</td>
</tr>
<tr>
<td>U</td>
<td>Daily total wind run (INTEGER)</td>
</tr>
<tr>
<td>UE</td>
<td>As for LE, but upper store</td>
</tr>
<tr>
<td>UEC</td>
<td>As for LEC, but upper store</td>
</tr>
<tr>
<td>UR</td>
<td>Catchment characteristic, Eqn. 5.57</td>
</tr>
<tr>
<td>US</td>
<td>Current content of the upper soil store</td>
</tr>
<tr>
<td>USC</td>
<td>Capacity of the upper soil store</td>
</tr>
<tr>
<td>WFINF</td>
<td>Total water infiltrated to calculate synthetic start time</td>
</tr>
<tr>
<td>WR</td>
<td>Similar to UR</td>
</tr>
</tbody>
</table>
APPENDIX B

PHYSICAL PROPERTIES OF AIR AND WATER

Throughout this thesis, constant use is made of certain physical properties of air and water and, in particular, the psychrometric properties of the air-water vapour mixture. For clarity and completeness, all the relevant definitions and values of constants are included in this appendix.

A. DEFINITIONS

These definitions are taken from Harrison (1965(a)).

Consider a volume, \( v \), of moist air. Then

\[
\begin{align*}
m_v &= \text{mass of water vapour} \\
m_a &= \text{mass of dry air} \\
m_{ma} &= \text{mass of moist air} \\
&= m_v + m_a
\end{align*}
\]

The air-water vapour mixture can be considered as a perfect gas and will follow the perfect gas equation

\[
\begin{align*}
p \cdot v &= n \cdot R^* \cdot T \\
&= \frac{m}{M} \cdot R^* \cdot T
\end{align*}
\]

where

\[
\begin{align*}
p &= \text{pressure} \\
v &= \text{volume} \\
m &= \text{mass of gas in, } v \\
n &= \text{no. of moles in, } v \\
M &= \text{molecular weight} \\
T &= \text{temperature (}^\circ\text{K)} \\
R^* &= \text{universal gas constant} \\
&= 8.31432 \text{ J } ^\circ\text{K}^{-1} \text{ mole}^{-1}
\end{align*}
\]

Furthermore, the total pressure of the mixture, \( p \), can be expressed as the sum of the partial pressures of dry air, \( p_a \), and water vapour, \( e \). Thus

\[
p = p_a + e
\]
The following properties can be defined.

(i) Density of Dry Air, \( \rho_a \)

\[
\rho_a = \frac{m_a}{v} \quad \text{(kg/m}^3\text{)} \quad \text{B-4}
\]

\[
= \frac{p_a m_a}{R^* T} \quad \text{B-5}
\]

\[
= \frac{(p - e)}{T R_a} \quad \text{B-6}
\]

where \( R_a = \frac{R^*}{m_a} \)

\[= 2.87 \text{ mb m}^3 \text{ k}^{-1} \text{ kg}^{-1} \quad \text{(Anon (1979))} \]

\[= 2.927 " " " " \quad \text{(Agrawal & Rao (1974))} \]

(ii) Mixing Ratio, \( r \)

This is also referred to as the humidity ratio or moisture content.

\[
r = \frac{m_v}{m_a} \quad \text{(kg/kg)} \quad \text{B-7}
\]

\[
= \frac{e m_v}{p_a m_a} \quad \text{B-8}
\]

\[
= \frac{b e}{(p - e)} \quad \text{B-9}
\]

where \( b = \frac{m_v}{m_a} \)

\[= 0.62198 \]

(iii) Specific Humidity, \( q \)

\[
q = \frac{m_v}{m_{ma}} \quad \text{(kg/kg)} \quad \text{B-10}
\]

\[
= \frac{m_v}{m_a + m_v} \quad \text{B-11}
\]

\[
= \frac{r}{r + 1} \quad \text{B-11}
\]

\[
= \frac{b e}{p_a} \left( \frac{b e}{p_a} + 1 \right) \quad \text{B-11}
\]

\[
q = \frac{b e}{(p - (1 - b) e)} \quad \text{B-11}
\]
(iv) Absolute Humidity, \( a \)
\[
\begin{align*}
a &= \frac{m_v}{v} \quad \text{(kg/m}^3\text{)} \\
&= \frac{e m_v}{R \cdot T} \\
&= r \cdot p_a
\end{align*}
\]

(v) Relative Humidity, \( h \)
\[
\begin{align*}
h &= \frac{e}{e^*} \times 100\% 
\end{align*}
\]

where \( e^* \) = saturation vapour pressure
\( e \) = actual vapour pressure

B. THE WET BULB TEMPERATURE

The most common method of measuring the psychrometric properties of air is to use wet and dry bulb thermometers. In this method, two similar thermometers are used; one (the dry bulb) measures the air temperature, \( T \); the other (the wet bulb) is covered by a wet wick and measures the wet bulb temperature, \( T_w \).

By undertaking an elaborate thermodynamic examination of the wet bulb (Harrison (1965(b)), it can be shown that
\[
\begin{align*}
e &= e^* - A \cdot p \cdot (T - T_w) 
\end{align*}
\]

where \( e \) = actual vapour pressure at, \( T \) (mb)
\( e^* \) = saturation vapour pressure at, \( T_w \) (mb)
\( p \) = pressure
\( (T - T_w) \) = wet bulb depression, \( D \) (°C)
\( A \) = \( A_o \cdot (1 + 0.00115 \cdot T_w) \) (°C⁻¹)
\( A_o \) = "psychrometric constant" (°C⁻¹)

This is known as the Psychrometric Equation.

The value of \( A_o \) is dependent on many factors including the ventilation rate of air over the wet bulb, the size of the bulb and the density, viscosity, specific heat, etc. of the air. However, for average size thermometers and ventilation
rates near 4 m/s, a value of 0.000653 is usually experimentally obtained (Bindon 1965). This value closely agrees with a theoretical prediction where:

\[ A_0 = \frac{C_p}{b/L} \]  

B-16

but, to quote Bindon, "to what extent this idealized theory may be related to the phenomenon occurring in the vicinity of a real wet bulb is a moot point". The term \( 1 + 0.00115T_w \) is incorporated to take account of the variation of \( L \) with temperature.

From the psychrometric equation, it can be seen that air is completely saturated with water vapour (\( e = e^* \)) when \( T = T_w \). This leads to another common method of humidity measurement. Air of constant mixing ratio and pressure can be cooled until \( T = T_w \). At this time, condensation occurs and this temperature is known as the dew-point temperature, \( T_d \). Note that \( T_d = T_w = T \) only when air is saturated. Usually \( T_d \) does not equal \( T_w \).

C. **CALCULATION OF PSYCHROMETRIC PROPERTIES**

The air-water vapour mixture is essentially a system with three degrees of freedom (Agrawal & Rao (1974)) and all of the above properties can be determined given any three intensive properties, e.g. \( (T, T_w, p); (T, T_d, p); (T, h, p) \). When all these properties are plotted on one graph, this is called a psychrometric chart.

Most of the calculations can be done using the formulae cited above, but some properties can only be related through empirical formulae. The most common empirical formula necessary is the relationship between \( e^* \) and \( T \) (usually over water). Over the years, many formulae have been proposed and these can be divided into two categories, exponential formulae and polynomial formulae. An example of each type is given below.

(a) Murray (1967)

\[ e^* = 6.1078 \exp \left[ \frac{17.269 T}{(T + 237.3)} \right] \]  

B-17
\( \text{(b) Lowe (1977)} \)

\[
e^* = a_0 + T (a_1 + T(a_2 + T(a_4 + T(a_5 + a_6 T))))
\]

where \( e^* \) = saturated vapour pressure (mb)

\( T \) = temperature (\( ^\circ C \))

and \( a_0 = 6.107799961 \)
\( a_1 = 4.436518521 \times 10^{-1} \)
\( a_2 = 1.428945805 \times 10^{-2} \)
\( a_3 = 2.650648471 \times 10^{-4} \)
\( a_4 = 3.031240396 \times 10^{-6} \)
\( a_5 = 2.034080948 \times 10^{-8} \)
\( a_6 = 6.136820929 \times 10^{-11} \)

These formulae are both quite accurate and should be considered as complementary. Murray's formula is useful for a small number of calculations using a pocket calculator, while Lowe's formula has specifically been developed to minimize computing time on large computers.

As has been noted in Chapters 2 and 5, it is often necessary to calculate, \( s \), where

\[
s = \frac{3e^*}{5T}
\]

To obtain this, the above equations have been differentiated to obtain

\( \text{(i) Dilley (1968)} \)

\[
s = \frac{25029}{(T + 237.3)^2} \quad \exp \quad \frac{17.269T}{(T + 237.3)}
\]

\( \text{(ii) Lowe (1977)} \)

\[
s = a_0 + T(a_1 + T(a_2 + T(a_3 + T(a_4 + T(a_5 + a_6 T))))))
\]

where \( s \) = mb \( ^\circ C^{-1} \)

\( T \) = Temperature, \( ^\circ C \)
and \[ a_o = 4.438099984 \times 10^{-1} \]
\[ a_1 = 2.857002636 \times 10^{-2} \]
\[ a_2 = 7.93805404 \times 10^{-4} \]
\[ a_3 = 1.215215065 \times 10^{-5} \]
\[ a_4 = 1.036561403 \times 10^{-7} \]
\[ a_5 = 3.53242181 \times 10^{-10} \]
\[ a_6 = 7.090244804 \times 10^{-13} \]

D. **Sundry Physical Properties**

(i) **Latent Heat of Vapourization of Water, L**

\[ L = 2502.5 - 2.386 \, T \quad \text{ (Anon (1979)} \quad \text{B-21} \]

where \[ L = \text{latent heat, kJ/kg} \]
\[ T = \text{temperature (°C)} \]

(ii) **Gamma, \( \gamma \)**

In the use of the Bowen ratio and the combination formulae, the term \( \gamma \), is often used. This term is often called the psychrometric constant, but is not the same as the psychrometric constant, \( A \). It is unfortunate that both these terms are given the same name, particularly when \( \gamma \) is not even a constant.

From the derivation of the Bowen ratio, it is seen that

\[ \gamma = \frac{A \, p}{b \, L} \quad \text{B-22} \]

\[ = \frac{C \, p}{L} \quad \text{B-23} \]

This is often taken as constant, but as has been pointed out by Storr & den Hartog (1975), Ripley (1976) and Stigter (1976), gamma is a function of both temperature and pressure.

For typical conditions at Warrambine Creek, i.e. \( T = 15^\circ \text{C} \), \( p = 1000 \text{ mb} \),

\[ \gamma = A \, p \]
\[ = 0.000653 (1 + 0.00115T)p \]
\[ = 0.66 \text{ mb } 1^\circ \text{C}^{-1} \]
or \[ \gamma = \frac{c_p P}{b L} \]

\[ = \frac{1.01 \times 1000}{0.62198 \times 2467} \]

\[ = 0.66 \text{ mb} \, ^\circ \text{C}^{-1} \]

This is the "constant" value of gamma which will be used throughout the thesis.
APPENDIX C

CALCULATION OF THE POSITION OF THE SUN

For the evaporation model of ARBM, it is necessary to be able to determine the position of the sun at any time. From this, the daylength and the time of sunrise and sunset can be determined.

A. THE POSITION OF THE SUN

The position of the sun at any time can be defined as in Fig. C1.

FIG. C1 The Solar Co-ordinate System

In Fig. C1, θ is the zenith angle and θ is the azimuth.

It can be shown (Spencer (1965)) that the zenith angle can be determined from:
\[ \cos Z = \cos \phi \cos \delta \cos H + \sin \phi \sin \delta \quad \text{C-1} \]

where \( \phi = \) Latitude
\( \delta = \) Solar declination
\( H = \) Hour angle

(i) **Latitude, \( \phi \)**
The latitude of the climate station in the Warrambine Creek basin is 37° 50' S. This is -37.83°.

(ii) **Solar Declination, \( \delta \)**
This is the angle of inclination of the earth to its ecliptic plane. It varies between ±23° 27' throughout the year. Spencer (1971) gives a Fourier series approximation of \( \delta \) in radians as a function of the time of year. The formula is:

\[ \delta = 0.006918 -0.399912 \cos \theta + 0.070257 \sin \theta \\
-0.006758 \cos 2\theta + 0.000907 \sin 2\theta \\
-0.002697 \cos 3\theta + 0.001480 \sin 3\theta \quad \text{C-2} \]

where \( \theta = \frac{2 \pi d}{365} \quad \text{C-3} \)

\( d = \) day number
\( = 0 \) on 1 January
\( = 364 \) on 31 December

(iii) **Hour Angle, \( H \)**
The hour angle is defined as the time from noon expressed as an angle i.e. local noon equals 0°; local 0600 equals -90° and local 1800 equals 90°. In determining \( H \), it is necessary to use local time rather than clock (zone) time. A correction of 4 minutes per 1° longitude is needed for the correction. A further correction, called the Equation of Time, \( E \), is needed to compensate for the eccentricity of the earth's orbit. Thus:

\[ H = \frac{2\pi}{24} (t_l - 12) + E \quad \text{C-4} \]

where \( t_l = \) local time (hours)
\( E = \) Equation of Time (degrees)
and \[ t_1 = t_2 - (\ell_1 - \ell_2) \times \frac{4}{60} \quad \text{C-5} \]

where \( t_2 \) = zone time (E.S.T.)
\( \ell_1 \) = longitude of time zone
  = 150° for E.S.T.
\( \ell_2 \) = longitude of site
  = 143° 53' for Warrambine Creek climate station

(iv) **Equation of Time, \( E \)**

Spencer (1971) gives a Fourier series approximation of \( E \) in radians. Using \( \theta \) as defined in Eqn. C-3, the formula is:

\[
E = 0.000075 + 0.001868 \cos \theta - 0.032077 \sin \theta \\
- 0.014615 \cos 2\theta - 0.040849 \sin 2\theta 
\]

\[ \text{C-6} \]

**B. SUNRISE, SUNSET, DAYLENGTH**

The time when sunrise and sunset occurs and the daylength can be determined using the above equations by noting that sunrise and sunset occurs when \( Z = \pm 90^\circ \). Thus, \( \cos Z = 0 \).

From Eqn. C-1,

\[ \cos H = \frac{-\sin \varnothing \sin \delta}{\cos \varnothing \cos \delta} \]

Thus

\[ H = \cos^{-1} \left( \frac{-\sin \varnothing \sin \delta}{\cos \varnothing \cos \delta} \right) \quad \text{C-7} \]

The zone time at which sunrise occurs is then:

\[ t_3 = 12 + (\ell_1 - \ell_2) \times \frac{4}{60} - (H - E) \frac{24}{2\pi} \quad \text{C-8} \]

and for sunset,

\[ t_4 = 12 + (\ell_1 - \ell_2) \times \frac{4}{60} + (H - E) \frac{24}{2\pi} \quad \text{C-9} \]

The daylength is the time between sunrise and sunset. Thus:

\[ t_5 = t_4 - t_3 \\
= 2 \left( H - E \right) \frac{24}{2\pi} \quad \text{C-10} \]
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