ASPECTS OF A FOREST MICROCLIMATE

by

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PREFACE

This thesis contains the results of climatological observations taken during the summers of 1963 and 1964 in a deciduous forest at Mont St. Hilaire, Quebec. I wish to express my gratitude to those who have aided me during the course of this project.

Special acknowledgements go to Professor F.K. Hare, University of London, who first stimulated my interest in climatology and who made this study possible, and to Professor S. Orvig, McGill Meteorological Department, who agreed to act as my supervisor after the project was well underway and whose criticism has been indispensable. I wish to thank Colonel P.D. Baird who aided me in many ways during my field work. Further acknowledgements go to Dr. P.F. Maycock, Botany Department, McGill, who kindly allowed reference to unpublished vegetation maps of Mont St. Hilaire and offered valuable advice on the botany of the mountain, Mr. L.S. Chia, an energetic companion and stimulating colleague, Mr. Blane Coulcher for providing daily meteorological data from St. Hubert Air Base and the National Research Council of Canada for its continuous support during my field studies.

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Chapter I Introduction

A. General Statement of Problem

In recent years the application of energy and water-balance methods has become firmly established as a working matrix which facilitates detailed exploration of the climate at the earth's surface. This has been a particularly profitable approach for microclimatic investigations of agricultural crops. For example, Curry (1952) reports on a system of growth units which was derived from the work of Thornthwaite (1952). Under such a system, a crop needs a particular number of growth units to achieve maturity. The growth units in turn are directly related to the energy and water balance of the crop. Brouwer (1956) achieved a good correlation between radiation intensity and rate of transpiration in the Netherlands. The development of corn crop in Iowa in relation to evapo-transpiration was traced by Denmead and Shaw (1959). Penman and Long (1960) in a detailed study of a wheat crop in England, traced the movements of water vapour and heat energy and calculated the upward vapour flux at different levels in the wheat crop. Through the use of lysimetric techniques, Graham and King (1961) explored the fraction of net radiation which was utilized in evapotranspiration and the energy balance. Krogman and Lutwick (1961) studied the consumptive use of water by forage crops in the Upper Kootenay River Valley and determined the irrigation amounts which give highest yields in terms of the evaporation of alfalfa and grass. Monteith and Szeicz (1961) took measurements of the radiation balance over bare soil, grass covered areas, and spring wheat and sugar beet crops for a three year period.

Energy and water-balance studies have also been applied with considerable success to grassland and water surfaces. In an important early paper, Thornthwaite and Holzman (1942) developed an aerodynamic method for calculating the vapour flux over grassland and water surfaces during conditions of neutral stability. Penman (1948) combined energy-balance and aerodynamic concepts into an equation which allowed the calculation of evaporation from open water, bare soil and grass. The relation between evapotranspiration and net radiation for a blue grama natural grass area was explored by Halstead (1954), and House, Rider and Tagwell (1960) noted that for a short grass area in England the energy for evapotranspiration essentially equalled the daily net radiation. The Great Plains Turbulence Program (Lettau and Davidson (ed) - 1957) which was carried out over the natural grassland in Nebraska gives a comprehensive picture of summertime values of solar and net radiation, soil heat flow and soil temperature profiles, and air temperature and humidity profiles. The determination of evaporation from open water surfaces has spawned a host of empirical equations. Thornthwaite and Holzman (1942) note that Dalton (1802) was the first to point out that evaporation is proportional to the difference between vapour pressure of air at the water surface and that of the overlying air. Rohwer (1931) presented and discussed a number of evaporation formulae. Meyer (1942) and Penman (1948) have each presented methods for calculating the evaporation from water. Anderson (1954) and Harbeck, Kohler and Koberg (1958), have reported respectively on evaporation investigations of the Lake Hefner and Lake Mead studies which were sponsored by the United States Geological Survey. Blaney and Corey (1955) calculated evaporation from water surfaces in California. McKay and Stichling (1961) applied various evaporation formulae to a water reservoir in the Canadian Prairies and compared the results to floating pan and water budget measurements, and McKay (1962) added computations of the energy budget over the reservoir. Williams (1961) presented radiation and evaporation measurements for a small lake at Ottawa. He compared the energy and water-balances which would be expected from a shallow, water-filled tank; a small lake, and a large deep lake.

Areas with a forest cover have not been subjected to integrated energy and water-balance studies to the same extent as the surfaces which have been discussed above. It is the water-balance aspects of a forest which have received the greatest attention, especially in terms of the water yield of forested as compared to non-forested areas. Layton (1960) raised the problem of decreased water yield over afforested terrain and Schneider and Ayer (1961) concluded that reforestation on New York watersheds decreased water yield by 24 percent as a result of increased evapotranspiration. Trimble and Reinhardt (1963) found that water yield increased proportionately to the amount of timber cutting on a watershed while a study of forest regrowth on abandoned farms in the Alleghany Plateau by Muller (1963) pointed to a 25 percent decrease in water yield on small watersheds. In a United States Committee Report on Water Resources and Evapotranspiration Reduction (1960) it was estimated that the thinning of forested areas in the western states would increase the water yield in certain areas by as much as 20 percent as a result of decreased evapotranspiration from the forest cover. One of the most frequently used methods for determining water movements within the forest involves the measurement of changes in soil moisture. Both Bates (1923) and Potzger (1939) studied the influence of slope aspect on forest evapotranspiration. Stoeckeler and Curtis (1960) measured the soil moisture depletion on south-facing forested slopes to be twice as great as on their north-facing counterparts. The soil moisture utilization of various species was studied by Shear and Stewart (1934) and Kramer (1952). In the first study the most rapid depletion of soil moisture occured at the time when new foliage was produced. Kramer found that the winter time evapotranspiration from conifer trees was almost as low as that from the bare branches of deciduous trees. Although the maximum soil moisture depletion under mature stands of pine and oak was found by Urie (1959) to occur at different times during the growing season, for the whole growing period the total evapotranspiration for each forest type was the same. Fletcher and Lull (1963) compared soil moisture changes for bare and forested plots during both wet and dry conditions and observed noteworthy differences which will be treated in some detail in Chapter VII. By isolating a patch of bare forest soil and comparing its evaporation to that of bare soil in the open, Kittredge (1954) was able to show that evaporation from the forest soil closely approached that of the open bare soil plot for the year. The effect of the leaf litter on rainfall interception has been measured by Curtis (1960) and Helvey (1964). Thames and Stoeckler (1955) and Fraser (1957) report on the annual and seasonal variations in soil moisture under hardwood stands for Wisconsin and Ontario respectively.

The use of lysimetric methods for the calculation of the forest's water balance is discussed by Patric (1961) who seriously questions the reliability of such methods, and by Sartz (1963) who reports on 6 years of measurements of the hydrologic balance in lysimeters containing hardwood seedlings. An amelioration of microclimate within a sugar maple forest in comparison to a surrounding grassland was noted by Rice (1960), (1962). The fate of water which is intercepted by the forest canopy is discussed in papers by Silina (1955), Rakhmanov (1958), Slayter (1961) and Penman (1963). The energy regime for a forested layer has been considered less extensively. The classic treatment is by Geiger (1950), though much significant work was accomplished by Angstrom (1925),(1937), at an earlier date. Miller (1955) has treated the energy regime of a coniferous forest in the Sierra Nevadas in some detail, particularly during the winter season. Cantlon (1950) in a paper discussed by Biehl (1961) studied differences in exposure microclimate for the tree, shrub and moss levels, on north and south slopes in New Jersey. Thornthwaite and Hare (1955) correlated the thermal and moisture indices of Thornthwaite's (1948) classification with forest zones in North America. Relatively complete energy balances for a spruce plantation in Germany and a mixed hardwood forest in Russia have been presented by Baumgartner (1956) and Dzerdzeevskii (1963) respectively. Denmead (1964) used an energy-balance technique for computing moisture and heat fluxes within a forest canopy and Philip (1964) discusses various aspects of the energy movements in air layers occupied by vegetation.

Although the various investigations into aspects of a forest climate and microclimate are by no means exhausted in the above resume, the diversity of the measurements is readily appreciated. Notably lacking are studies which integrate water-balance and energy-balance measurements for periods covering the whole growing season. The hydrologic regime of various forested regions has received much more thorough study, and much knowledge of the vertical energy regime within a forest has been gained especially in the last few years. The microclimatic differences which are created in a forested layer due to topographical variation have been considered in a few isolated studies, but many more investigations into the effects of exposure are needed. Philp (1964) notes that studies of energy movements in a vegetated layer have been pursued mainly in one-dimension (i.e. considering the vertical distribution of energy terms at one place only) and that although these apparently have been successful, the vegetated layer is three-dimensional. When studying a mature forest it is desirable to have both continuous measurements, and values averaged over area, the former to bring out both diurnal and seasonal trends, and the latter to counteract the uneven distribution of forest trees as compared to agricultural crops, grass and water surfaces. A very real problem of any investigation into the forest microclimate, is presented by the physical difficulty of taking measurements. Most field workers have resorted to the construction of semi-permanent towers on which sensors are mounted. This is in some ways undesirable, especially as it disturbs the natural forest environment and does not give a sampling over area. Although it proves a rather arduous and time-consuming task, it is possible to suspend instruments from trees satisfactorily with a minimum of scaffolding, and to move them from place to place regularly.

This thesis is a study of the available energy as it varies horizontally and vertically within the forest, and how this energy is utilized, the movement of water above and within the forest, the problems of instrumentation in forest research, the effects of exposure on some phases of the forest microclimate, and some of the ways in which the forest as an active surface layer differs from other ground covers. An attempt is made to correlate the observations which are presented within this thesis with other similar studies, and to assess their geographical significance.

B. Physical Nature of the Study Site

All of the measurements contained in this report were taken at Mont St. Hilaire. The mountain which is of volcanic origin lies about 20 miles east of Montreal, and presents a sudden topographic deviation from the St. Lawrence Lowlands. It is composed of seven distinct peaks, the highest of which rises to 1400 ft., a height which is roughly 1300 ft. above the surrounding lowlands. The mountain has a total area of 75 acres. The outer slopes are very steep and in places form almost vertical cliffs. This perimeter is, however, broken by four small valleys which lead inward to the interior basin, and the slopes of this latter basin are less drastic than those of the outer margin.

Mont St. Hilaire is one of a series of landforms known collectively as the Monteregian Hills. General theory (Adams 1903 , Dresser 1906), places these hills as igneous intrusives which have been exposed through differential erosion. Evidence of Pleistocene glaciation is found in both extensive layers of till and sand and in the shallow depths of soil in the summit areas. The existence of beach terraces at the 575 ft. level points out the likelihood that the tops of the Monteregians formed islands during the blockage of meltwaters from the northward-retreating continental ice sheet.

Geologically, the type of bedrock can be divided into two quite distinct sections. A north-west area is dominated by essexites, while in the south-east, nepheline-syenites form the major rock type. Soils vary greatly with topographic condition, but have commonly developed on glacial till or sand deposits. The soil forming process is a podzolic one, and the texture of the soils is usually of a sandy-loam or gravelly-loam nature. Alluvial muck deposits are found in depressions, stream valleys, and in areas adjacent to the lake. It is difficult to generalize about the topographic nature of the mountain. If on a contour map one draws two lines which intersect at right angles in the centre of Mont. St. Hilaire, and sums all of the upward and downward height changes for the total distance covered by both lines, he can calculate a mean slope of 16 percent. A similar value is acheived no matter in what directions the perpendicular intersecting lines are oriented. The north-south oriented hill on which many of the radiation measurements were taken showed a slope of 18 percent on each side where the instruments were installed.

C. Vegetation

Extensive work on the vegetation of Mont St. Hilaire has been carried out under the leadership of Maycock (1959), (1961). Figure 1, presents a vegetation map which was drawn both from ground survey and aerial photographys by O. Maryniak. By plainimetering the area which was mapped the following percentage breakdown into various cover types is achieved.

Cover	Percent
Mainly maple	15
Various combinations of maple and beech	16
Combined maple and oak	28
Scrub oak	14
Coniferous trees (hemlock and pine)	5
Wet,rocky,burned, highly disturbed, and other areas	18
Lake Hertel	4
Total	100

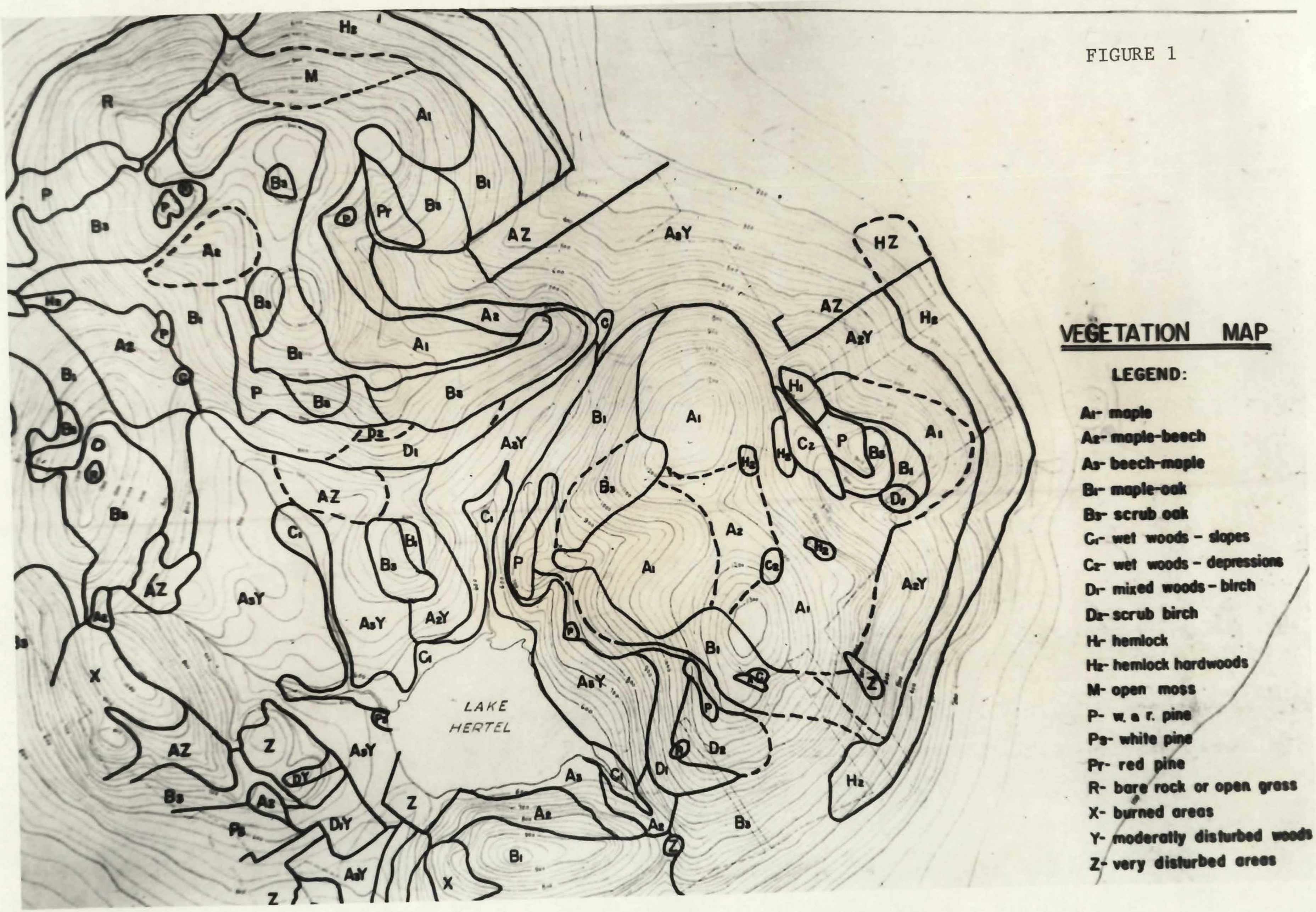

LAKE HERTEL
VEGETATION MAP
LEGEND:
A_1- maple
A_2- maple-beech
A_3- beech-maple
B_1- maple-oak
B_3- scrub oak
C_1- wet woods - slopes
C_2- wet woods - depressions
D_1- mixed woods - birch
D_2- scrub birch
H_1- hemlock
H_2- hemlock hardwoods
M- open moss
P- w. & r. pine
P_3- white pine
P_r- red pine
R- bare rock or open grass
X- burned areas
Y- moderatly disturbed woods
Z- very disturbed areas

FIGURE 1

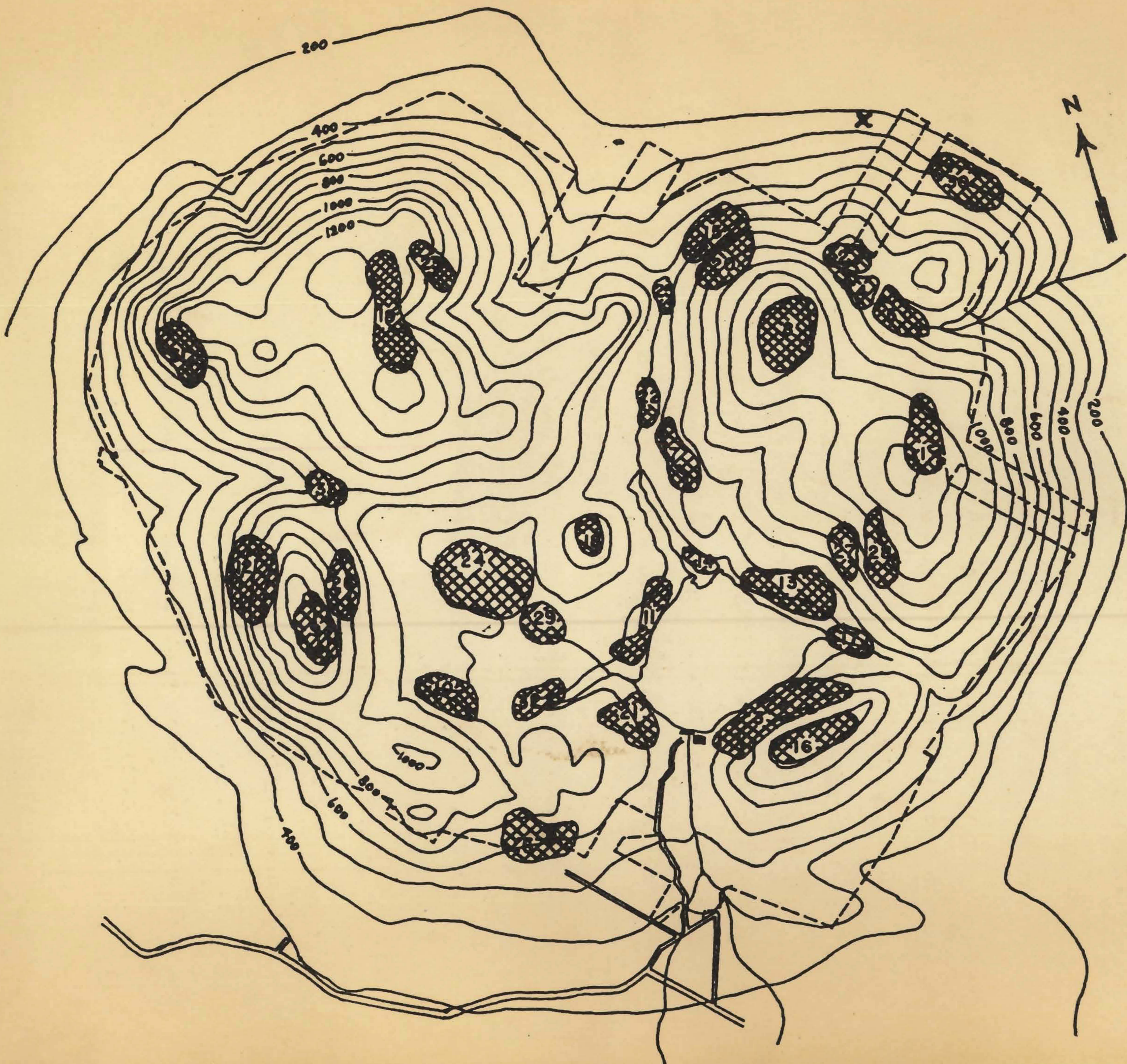
From the above figures it can be calculated that mature hardwood forests of maple, beech and oak species compose 58 percent of the total area, and if scrub oak is included, this total is brought to 73 percent for the total area of hardwood forest.

In a more detailed survey, Maycock chose 34 different stands whose locations are shown in Figure 2, and sampled them quantitatively. In this quantitative sampling, each tree type was rated according to its numerical frequency, density, and dominance and was given an overall importance number which sums the aforementioned three characteristics. For example in stand 8, which according to Figure 1 lies in a predominantly maple forest, sugar maple (Acer saccharum) has an importance of 60 percent, with beech (Fagus granifolia) a poor second at 23 percent. In an area of mixed beech and maple as found in stand 21, the relative importances for the two dominant species are 49 and 36 percent respectively. Similarly stand 16 in an area of maple and oak (Quercus rubra) gives importances of 51 and 32 percent respectively.

Maycock points to the decided tendency for pure deciduous forest to grow on the north and east-facing slopes and coniferous or mixed coniferous-deciduous types on south and west-facing slopes. Maple-beech forest may occur to an altitude of 1350 ft. wherever the soil mantle is sufficiently thick and moist. Where soils on the summits are shallow and conditions are relatively dry, oak-maple or oak-pine are found. Moist depressions favour trees such as elm, ash, birch, hemlock and balsam fir, but nowhere is the environment suitable for bog forest. Maycock adds, that on the summits of the more inaccessible peaks, of the eastern half of the mountain, there are hardwood stands which display no

FIGURE 2

Locations of stands sampled quantitatively by P. Maycock



SAINT-HILAIRE MOUNTAIN

QUEBEC

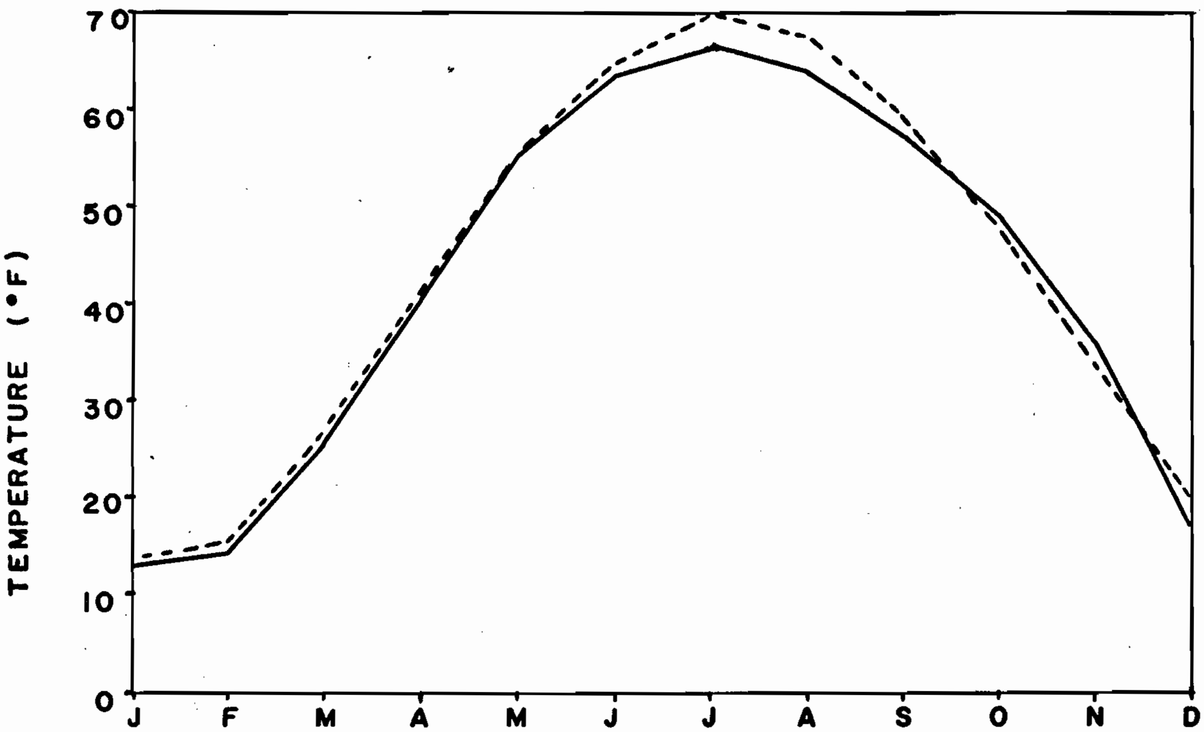
Scale - feet :

0 600 1200 2400 3600

human disturbance of any kind. However, trees of great age and size are not characteristic, because of the severe environmental conditions inherent in their situation. The unfavourable environmental conditions include topographic, pedologic, and climatic limitations. The many steep slopes with their accompanying thin soils and unprotected locations, expose the forest trees to periodic drought, due to the low moisture capacity of the soils, to root damage by wind and cold where the roots must abide in shallow soils, and to occasionally damaging winds. The length of the growing season between the last spring frost and first in the fall averages around 170 days at lake level and is probably considerably shorter at higher elevations. This period makes up less than one-half of the total year and the trees must survive through a long period of dormancy. In the 5 complete years of temperature records which have been recorded between 1960 and 1964, the lowest minimum temperature was - 28.2^{o}F. at lake level and -22.7^{o}F at the 1300 ft. summit level. The changes in temperature to which the trees are subjected must create a strong physiological stress. For example in the fall of 1961, during a period of 11 weeks the average maximum weekly temperature variation was 32^{o}F, always through a range from below to above freezing. Moreover, during this interval there was no protective snow cover. Maximum weekly temperature ranges of 54^{o}F were recorded in the middle of February and end of April 1962. During both occasions the minimum and maximum temperatures were below and above freezing respectively, and on the latter occasion there was no snow cover to provide protection to the lower parts of the tree.

D. General Climate

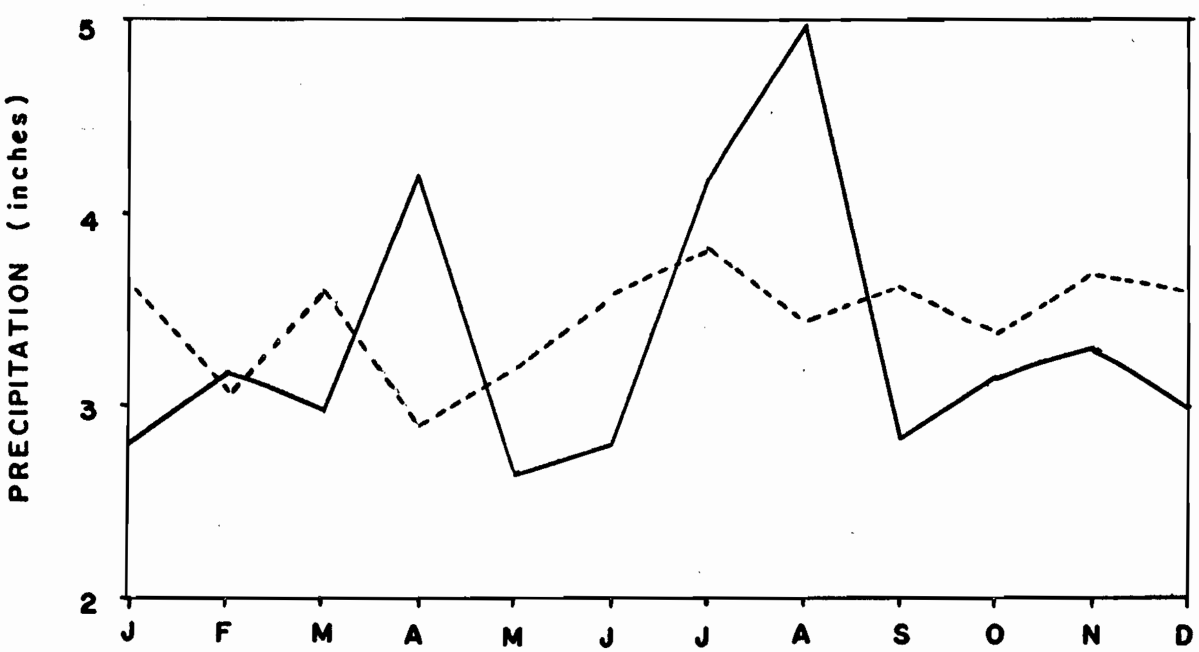
There are five full years of daily temperature and rainfall measurements, which have been recorded at the Gault House, which lies at the southern side of Lake Hertel. The elevation of the station is 570 ft. above sea level. In Figure 3, the monthly mean temperatures and precipitation are plotted for Montreal and Mont St. Hilaire. For the former station the data represent a period of 75 years, which have been sumarized by Longley (1951). The latter station includes only 5 years of records. The similarity of monthly mean temperatures for records of such different duration is striking. Such correspondence may be fortuitous, but may represent the fact that 5 years of temperature records can be sufficient to define a normal temperature regime in this area. The same argument can not be advanced for monthly mean precipitation. As is shown in Figure 3, the 5 years of data for Mont St. Hilaire show strong monthly fluctuations, whereas the longer period of measurement in Montreal has smoothed out periodic fluctuations, so that a more even trend is observed. The annual mean totals bear a closer similarity, giving 41.5 inches for Montreal and 40.0 inches for St. Hilaire. One can state with some confidence that a normal temperature regime can be established from the limited data available at St. Hilaire, and that the results of the long period of temperature measurements in Montreal can be treated as representative for St. Hilaire. Precipitation on the other hand shows a greater diversity, and it would not be accurate to extrapolate the Montreal records to incorporate St. Hilaire. With these facts in mind, the average climate in terms of temperature is summarized in the following table.



MONTHLY MEAN TEMPERATURES

—— St. Hilaire (5 Years)

---- Montreal (75 Years)



MONTHLY MEAN PRECIPITATION

—— St. Hilaire (5 Years)

---- Montreal (75 Years)

FIGURE 3

Annual mean temperature	41.9°F
Extreme minimum temperature	-28.2
Extreme maximum temperature	97.0
Maximum possible temperature range	123.2
Earliest date of last spring frost	March 28
Latest date of last spring frost	May 24
Latest date of first fall frost	November 9
Earliest date of first fall frost	September 29
Maximum possible frost free period	216 days
Minimum possible frost free period	128 days

As well as its large-scale climatic characteristics, Mont St. Hilaire will display a special type of mesoclimate due to its topographic variations. Because a thermograph has been in operation since 1960 on one of the hilltop locations, it is possible to compare temperatures between the 570 and 1320 ft. levels. Figure 4 shows the highest weekly maxima and lowest weekly minima temperatures as recorded at the hilltop, and the lake levels, for a period between August 12 and December 30, 1962. The only clear cut trend, which can be observed in Figure 4, is that temperatures are higher at the lake station, than on the hilltop, which would certainly be expected. For the whole period the average range between lowest minima and highest maxima was 32.8°F at the lake, a figure slightly greater than the 31.6°F at the hilltop station. The greatest weekly temperature range at the lake was 47°F in contrast to 44°F at the hilltop station. For a 2 week period in December the lake station's minima temperatures were lower than those at the higher elevation. This feature might suggest nightime cold air drainage from the surrounding hills into the central basin. It must be concluded, however, that from the data which are available, there are no unexpected differences in temperature due to location. For the 20 week period which has been considered, the mean temperature change with height is 2.6°F/1,000 ft.

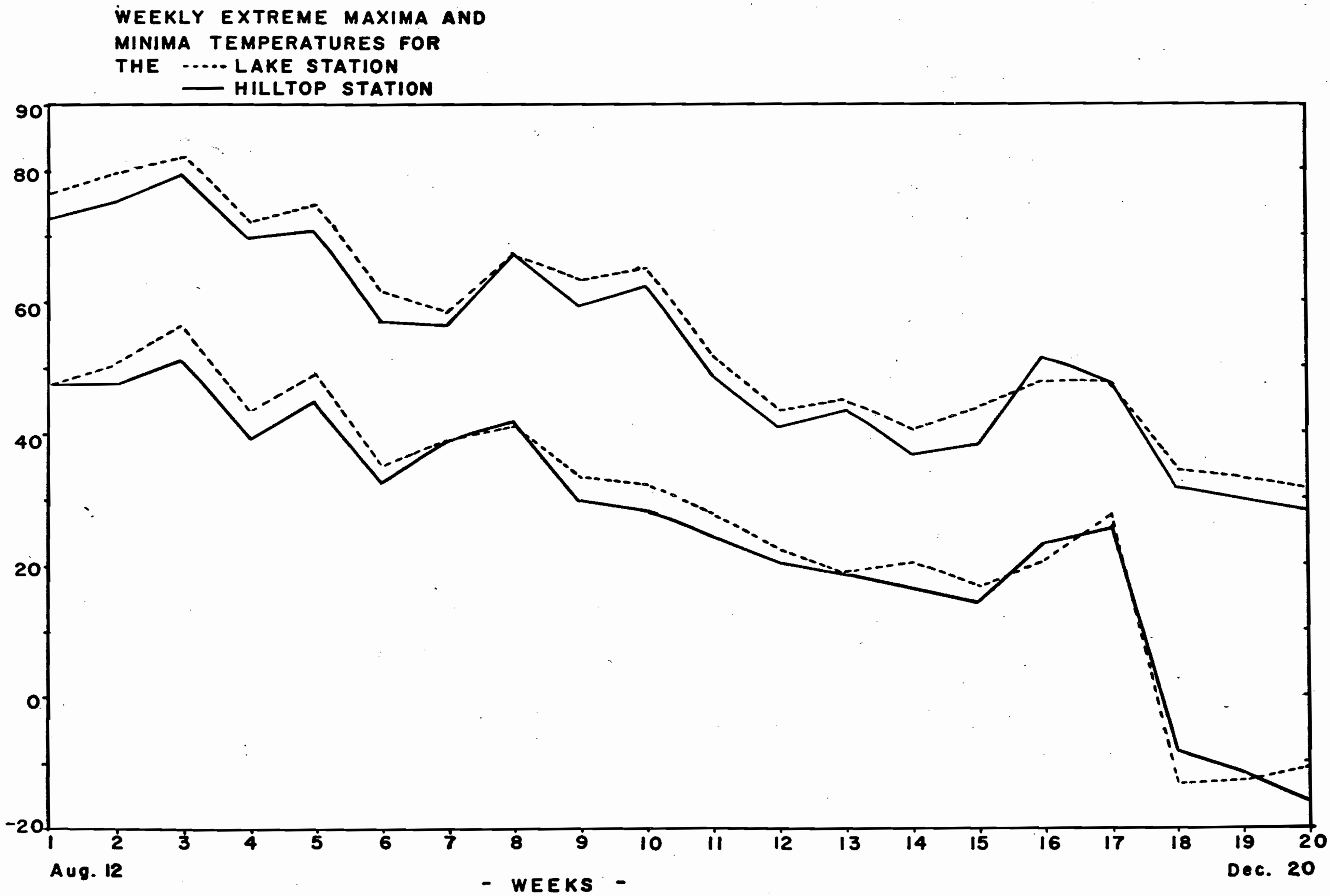

WEEKLY EXTREME MAXIMA AND MINIMA TEMPERATURES FOR THE
LAKE STATION
HILLTOP STATION
90
80
60
40
20
0
-20
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19
20
Aug. 12
Dec. 20
- WEEKS -

FIGURE 4

E. Duration of Study

The total observation period for the measurements which are presented in this thesis spreads over the better part of two years, from spring 1963 to winter 1964. The summer of 1963 was mainly spent in assembling and installing various instruments, and in carrying out preliminary surveys. The first measurements were not made until September-October of 1963. The bulk of the data were collected between the beginning of May and end of September 1964, an uninterrupted period which covered all of the growing season. The climate for this latter period can be compared to the 5 year averages for Mont St. Hilaire. The 5 month mean temperature of 61.1^{o}F compares favourably with the 5 year average mean of 61.3^{o}F. Moreover, each monthly mean was very close to the average mean temperature for that month. The rainfall of 13.3 inches was 23 percent less than the average of 17.4 inches. The months of June and September were particularly dry, in each case yielding less than one-half of the average values for those months. Some of the specific results of this study, therefore, apply to a year that was drier than is normally the case.

F. The Gault Donation

Since 1914, Mont St. Hilaire has been the property of Brigadier A. Hamilton Gault and upon his death in 1959 he bequeathed the property to McGill University, in order that the youth of Canada could pursue studies in the sciences in an undisturbed natural preserve. The bequest has been administered by Colonel P. D. Baird since that date. Under Baird's impetus, the two climatic stations were put into operation, and have been kept running continuously to the present. In 1963 a student laboratory and residence was established by McGill University in order that research could be pursued with proper facilities. This writer has been most fortunate in having such facilities at his disposal and in the cooperation and guidance which he has received. The foregoing analysis of climate has been based on the processed data which Colonel Baird has kindly provided.

Chapter II Theoretical and Empirical Background

(A) Energy-Balance Concepts

This study is primarily concerned with the microclimate of a forest layer but the background material cannot be limited exclusively to considerations of the forest environment. Many of the more valuable studies have been pursued in the realm of agricultural climatology, and their findings often have direct relevance when considering the forest habitat.

The most simplified form of the energy-balance equation at the surface usually reads as follows:

$$Rn = E + A + S \qquad (1)$$

where Rn is net radiation, E is evapotranspiration, A represents the sensible heat flux, and S is soil heat flow. The measurements are usually expressed in cal/cm^2, min or the equivalent expression langleys/minute (ly/min). The latter term will be employed generally in this paper. It is not uncommon to see the net radiation and evapotranspiration terms of the surface energy-balance equation expressed in the water equivalent of the energy values. In this form, the energy of net radiation and that which is used in evapotranspiration, is divided by the latent heat of vapourization of water at mean air temperatures, to give the depth of water which could be, or is evapotranspired (usually in mm/unit time). This latter unit of expression is often used when water movement is the prime consideration of a study.

The energy-balance in its simplified form can prove inadequate for the study of a fully-developed vegetative layer, because it omits several processes involving energy exchange, some of which may assume quantitative importance at a particular time or place. To appreciate this, one can consider results from agroclimatology. King (1961), in his studies of mature corn crops at Guelph, Ontario, notes that for a typical area of land, the evaporating surface is three-dimensional. The surface layer extends vertically from the soil surface to the extreme height of vegetation, and evaporation can take place at any level within this surface layer. This statement holds true not only for evaporation but also for other energy transformations which take place at the earth's surface. Budyko (1958) suggests that it might be more expedient in certain cases to use the concept of an active layer instead of an active surface when considering energy exchange, and the present author will use the term "active surface layer" to describe the forest condition.

King (1961) presents a complete energy-balance equation for a three-dimensional active surface layer, and illustrates it in a drawing which is reproduced in Figure 5. A simplified form of the equation gives:

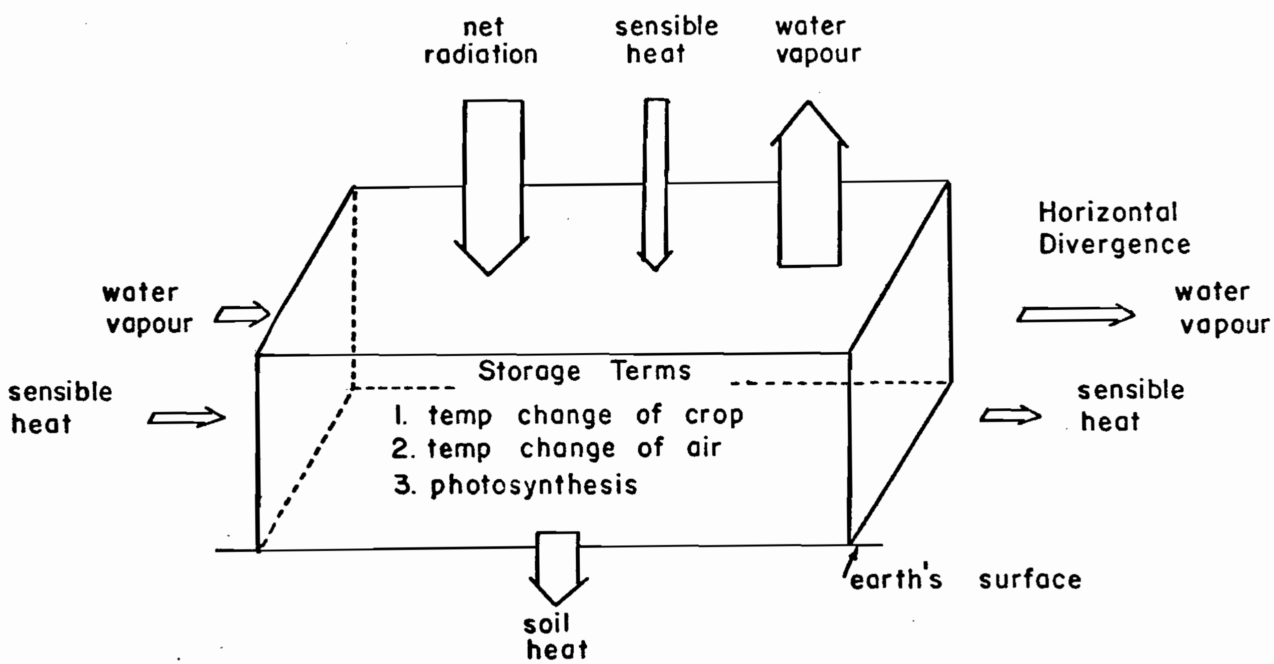
$$Rn + \text{div. } A + \text{div. } E = S + A + E + N + \Delta T \text{ air} + \Delta T \text{ crop} \qquad (2)$$

where div. A and div. E refer to the horizontal divergences of sensible and latent heat respectively, S, A and E represent the same processes as in equation (1), N is the energy stored in photosynthesis, Δ T crop represents the changes in heat storage in the crop, and ΔT air is the change in the heat content of the air.

It is well at this point to note that the employment of symbols to represent various processes is very diverse. Throughout this thesis

Figure 5

Energy Balance for a Land Surface
(after K. M. King)



the author will use a single set of symbols and will standardize the different symbols which are employed by various authors. Terminology also varies, but to a lesser degree. "Net radiation" is the most widely accepted term for the difference between the radiant energy directed downward to the earth's surface, and that which is directed upward into the atmosphere. The Russian school, however, prefers the term "radiation balance" Budyko (1958). "Evapotranspiration" denotes the movement of all water from the earth's surface into the atmosphere. It includes water which is evaporated from moist soil surfaces, the evaporation of water which has been intercepted by plant leaves, and water which is transpired from the aerial parts of plants. When the movement of evapotranspired water through the lower atmosphere is considered, it may be referred to as a "water vapour flux" or a "latent heat transfer". In this study, the most commonly-used terminology will be employed to describe various processes.

By examining the various terms in equation (2), one can arrive at some estimate of their relative importance.

(a) Net Radiation

Net radiation which is the single most important quantity in energy-balance considerations, is commonly represented by the following equation.

$$Rn = Rsi\ (1-\alpha) + Rli - Rlr \qquad (3)$$

where Rsi is downward directed short-wave radiation from sun and sky, and α is the albedo, which represents that fraction of Rsi which is reflected by the earth's surface upward into the atmosphere. Rli and Rlr represent downward directed and upwarded directed long-wave radiation, respectively.

The incoming short-wave radiation (Rsi), includes the portion of the radiation spectrum between 0.3 and 2.0 microns. This band incorporates the visible portion of the spectrum between the wavelengths of 0.36 and 0.76 microns, and the longer wavelengths in the infrared band. It represents the solar radiation which will penetrate a clear flint glass of the type used in most pyrheliometers and solarimeters. The albedo often stays approximately constant for a given type of surface. Investigations which have been made of the albedo values for a mature mixed hardwood forest, however, show a fairly wide range. Early measurements, using photometers in aircraft, were taken by Kimball and Hand, Towsey and Hurlbut, and Luckiesh (all quoted in Smithsonian Meteorological Tables, 1958, 442). They gave values of 3-6, 4-10 and 3-5 percent respectively. Kondrat'ev (1954), from surface measurements obtained values of 18,14 and 10 percent, respectively, above an oak, pine and fir tree. Budyko (1958) gives an average value for deciduous forests of 10 to 15 percent. An albedo of 18 percent for oak forest was measured by Angstrom (1925), who used pyrheliometers. It is clear that above-canopy measurements of albedo give significantly higher values than airborne measurements and are more pertinent to microclimatic considerations. The incoming long-wave radiation (Rli) represents dark heat rays, which generally reach maximum intensities in the region of 15 microns at normal atmospheric temperatures. The outgoing long-wave radiation (Rlr) is also termed terrestrial radiation. Its intensity can be determined from the Stefan-Bolzman law;

$Rl\,r = c\,\sigma\,T^4$, where σ is the Stefan-Bolzman constant, T is the temperature at the earth's surface in degrees Kelvin, and c is a correction for the departure of a surface from a perfect black body. Brooks (1959) gives a value of c for oak woodland and pine forest of 0.90. Penman, Angus and van Bavel (1964) note that both Rli and Rlr are large quantities, with Rlr invariably being the larger.

(b) Divergence of Sensible and Latent Heat

The two terms of div. A and div. E appear on the left side of equation (2), and represent an addition to, or subtraction from, the available energy at any given spot. They are independant of net radiation and respond to the movement of horizontal wind. King (1961), presents the following equations for the measurement of the divergences of sensible and latent heat.

$$\text{div. A} = \int_0^z Cp \; \nabla_H \; (\rho \, uT) \; \delta z \qquad (4)$$

$$\text{div. E} = \int_0^z \frac{L\epsilon}{R} \; \nabla_H \; \frac{(ue)}{T} \; \delta z \qquad (5)$$

Cp is the specific heat of air at constant pressure and ρ is air density. T, u, and e, are air temperature, horizontal wind speed, and vapour pressure, respectively. L is the latent heat of vapourization, ϵ a constant which represents the ratio of molecular weights, water to air, and R is the specific gas constant for air. ∇ H symbolises $\partial/\partial x + \partial/\partial y$, and gives the divergence of the variable quantities in equations (4) and (5), in the x-y plane. The integration takes place between the zero plane where u, the wind speed, becomes zero and the height z which is taken as the top of the chosen layer. The importance of the sensible and latent heat divergences will vary with the area under consideration, and with the wind direction. King states that the divergence terms will be most significant near the borders of cropped fields and for small plots or individual plants. Horizontal divergence of energy usually shows up as more sensible heat or less latent heat going in the upwind side of the box (Figure 5), than that coming out the downwind side. He found from measurements made in a corn field that there was a decrease in wind speed within the crop cover with distance downwind, and that this allowed a significant amount of latent heat, relative to the net radiation, to be carried in bulk flow out through the top of the crop. One would expect a similar result near the border areas of a forest cover, but would feel safe in subscribing a minimal role to the divergence of sensible and latent heat when dealing with a large relatively homogeneous forest mass. However, when investigating a forested area which slopes sharply away from a lake - a lake which is open to wind jets which are focused through narrow valleys - a considerable import of latent heat to the surrounding forest might be expected. Mont St. Hilaire has this somewhat unusually physical aspect, and although it has not been included in the present study, it constitutes a problem worthy of special examination. As will be discussed in Chapter IV the import of sensible heat energy may be large enough to make results from evapotranspiration tanks in an open field and a forest border area unreliable, to the extent that potential evapotranspiration may at times exceed substantially the estimated available energy over periods of a week.

(c) Soil Heat Flow

Heat movement within the soil may take place in both horizontal and vertical directions. Budyko (1958) notes that the mean horizontal gradients of temperature in the soil are very small so that, for land, the heat exchange between the soil and atmosphere is directly proportional to the vertical flow of heat within the soil. The rate of soil heat flow in the vertical will respond to the temperature gradient which is developed within the soil, the soil type, amount of soil moisture, and the latent heat exchange which occurs during periods of evaporation and condensation, and of freezing and thawing. In middle latitude areas the thermal gradient shows both a seasonal and diurnal cycle. The seasonal pattern displays a spring-time warming of the soil from above, which reaches a peak in mid-summer, and cooling from above which occurs through late summer and fall and reaches a maximum in the early stages of winter. Thus, there will be a net upward flow in late summer and fall. Due to the fact that, for the average year, the upper layers of soil are neither warmed nor cooled, one must assume that there is no yearly net storage of heat energy in the soil. Superimposed on the seasonal cycle is a diurnal pattern in which the upper soil layers receive heat from above in the morning and early afternoon periods and give off heat in the late afternoon and evening. Different soil types show great variations in their heat capacities. In a study of frost hazards in Wisconsin, Wang (1963) noted that in the peat soils of bog areas, the daytime heat storage was low, and as a result, the night-time movement of heat from the soil to the atmosphere was small. Because of this the frost occurance was always greater in areas of peat soil than in surrounding regions. Williams (1965) in studies of the heat balance over saturated sphagnum moss at Ottawa found that because moss is a poor thermal conductor, the relatively small amount of heat stored in the moss during the day cannot be readily returned to the air during night-time cooling. The result is that the ground surface and the adjacent air layers cool rapidly. In addition, evaporation from wet moss would tend to keep the moss surface temperature lower during the day than the temperature of drier sand or grass surfaces. Williams concluded that under similar atmospheric conditions, the night-time surface temperature of grass or sand-covered terrain would not fall as much as that of a sphagnum surface cover. It is apparent that the moisture content of a soil strongly influences heat flow because of the high- specific heat of water in comparison to most soils. Evaporation from soil surfaces will use heat energy which is then not available for heating the soil. Conversely, the condensation of dew or fog on the soil will release heat. Freezing and thawing will similarily create a heat exchange within the soil. The type of vegetation which covers a soil is most important in influencing the amount of heat it receives or loses. Measurements of soil heat flow in a grassland at O'Neil, Nebraska, showed that up to 16 percent of the energy of net radiation flowed downward into the soil during noon hour periods in August (Lettau and Davidson, 1957). In contrast Baumgartner (1956) and Dzerdzeevskii (1963) observed maximum daytime heat flows downward into forest-covered soils of 2.5 and 5 percent respectively during the period of active growth. Although it can be assumed that for any year the net soil heat storage approaches zero, a similar assumption cannot be applied for any daily or weekly period. Thus, Baumgartner found that the mean diurnal soil heat flow during a 2 week period around the first of July gave a net downward heat movement which equalled 0.5 percent of the average daily net radiation for that period. The relatively small soil heat flow under forest vegetation was due both to the low net radiation at the forest floor, and the protective cover of forest duff which provides a layer of insulation above the mineral soil.

If λ is a coefficient of soil thermal conductivity, and $\partial T / \partial z$ the temperature gradient in the upper soil layers, the soil heat flow in the vertical can be expressed as follows:

$$S = \lambda \, \partial T / \partial z \qquad (6)$$

The coefficient λ is difficult to determine, as it varies continously with soil type and moisture content. Soil heat flow can, however, be measured directly and a discussion of this will follow in the next chapter.

(d) Sensible and Latent Heat Flux

It is convenient to discuss the sensible and latent heat fluxes together, since in many respects they respond to the same forces. The following two equations express the vertical movement of heat and of water vapour.

$$E = - \rho \, K_w \frac{\partial \bar{e}}{\partial z} \qquad (7)$$

$$A = - \rho \, C_p \, K_h \frac{\partial \bar{\theta}}{\partial z} \qquad (8)$$

In these equations, E and A are the latent heat and sensible heat fluxes respectively. The density of the air is given by ρ , and C_p is the specific heat of air at constant pressure. The mean vertical vapour pressure gradient and mean vertical potential temperature gradient are given by $\partial \bar{e} / \partial z$ and $\partial \bar{\theta} / \partial z$. K_w and K_h are coefficients which denote the eddy diffusivities of latent and of sensible heat. These coefficients have proved difficult to determine, and they may vary over several orders of magnitude. Moreover, the eddy motion which contributes to each flux is not the same for heat and for water vapour. The departures of K_h/K_w from unity depend upon the thermal stratification of the lower atmosphere and are discussed by Pasquill (1949), Rider (1954), Sheppard (1958) and Priestly (1959). The general conclusion has been reached, that because of increased buoyancy under unstable conditons in the lower atmosphere, K_h/K_w is then larger than for neutral or stable conditions.

Although the sensible and latent heat fluxes have been determined using the eddy correlation technique under experimental conditions, which are summarized by Priestly (1959), this approach has not proved amenable to sustained measurements under field conditions. The equations for determining the sensible and latent heat fluxes from the eddy correlation approach become:

$$E = \rho \overline{e' \, w'} \qquad (9)$$

$$A = Cp \, \rho \overline{\theta' \, w'} \qquad (10)$$

where ρ and Cp are the same as in equations (7) and (8), and e', θ' and w', are the variances of vapour pressure, potential temperature, and vertical wind velocity, as determined from Reynold's Resolution, where s is the instantaneous value of any moving property which undergoes turbulent motion, $\overline{s}$ is the time-average value or mean motion, and s' represents the variance about the mean. Munn (1961) lists the instrumental difficulties in such an approach as follows:

(a) the separation between the instruments measuring the two variances must be smaller than the smallest eddies contributing to the flux;

(b) the sensing instruments must have the same response time; otherwise spurious covariances will result;

(c) very short response times are necessary to record the flux contributions of the very small eddies.

The covariance technique also assumes that there is no net flux due to the mean vertical motion of the air (i.e. that all updrafts are balanced by all downdrafts over the measuring period). As was noted by Thornthwaite et al. (1961), this is not likely to be the case at any particular location due to the development of standing eddies, which are created by surface roughness or thermal effects.

To avoid the problems which are presented by the coefficients of turbulent mixing K_h and K_w, early workers led by Thornthwaite and Holzman (1939) (1942), attempted to use the horizontal wind profile as a function of logarithmic height, to determine the influence of forced convection. These formulations which are generally labelled aerodynamic equations appear as follows:

$$E = \frac{\rho k^2 \ (\partial e/\partial z) \ (\partial u/\partial z)}{(\ln z_2/z_1)^2} \qquad (11)$$

$$A = \frac{C_p \rho k^2 \ (\partial \theta/\partial z) \ (\partial u/\partial z)}{(\ln z_2/z_1)^2} \qquad (12)$$

In the aerodynamic equations, k is Von Karman's constant; $\partial e/\partial z$ and $\partial \theta/\partial z$, are the vapour pressure gradient and potential temperature gradient between heights z_1 and z_2; and $\partial u/\partial z$ is the vertical gradient of the horizontal wind speed. Such equations are only adequate around neutral thermal stability and are difficult to use with very stable or unstable temperature gradients. Moreover, extremely accurate wind profile data are needed during a neutral period.

A method which has been employed frequently for estimating the evaporative and sensible heat fluxes, involves the determination of E and A within the surface energy-balance equation. In equation (1) $Rn = S + A + E$; the net radiation and soil heat flow can be measured readily. Use is made of the Bowen Ratio β, to proportion the remaining energy between E and A.

$$\beta = \frac{A}{E} = \gamma \frac{K_h}{K_w} \frac{\theta 1 - \theta 2}{e1 - e2} \tag{13}$$

where γ, the psychrometer constant equals $Cp\ P/L\varepsilon$, with P the air pressure and ε a constant. The assumption is made that K_w equals K_h and the equation reduces to

$$\beta = \gamma \frac{\theta_1 - \theta_2}{e_1 - e_2} \tag{14}$$

and combining equations (1) and (14) gives

$$E = \frac{Rn - S}{1 + \beta} = \frac{Rn - S}{1 + \gamma \frac{\theta_1 - \theta_2}{e_1 - e_2}} \tag{15}$$

In theory, the simultaneous measurement of the vapour pressure and potential temperature gradients should allow a determination of the division of available energy between the latent and the sensible heat transfers. Problems arise, however. There is the question which has been discussed above, of assuming the unity of K_h and K_w. King (1963) states that there is little problem in estimating E when the vapour flux is large and important, because during such periods, the Bowen Ratio is generally close to zero. However, for periods of large sensible heat flux, the mean temperature gradient may not indicate the magnitude of that flux. Penman, Angus, and van Bavel (1964) note that during periods when stability is far from neutral, a correction is needed. The most common correction employs the Richardson number

$$Ri = \frac{g}{T} \frac{(T_2 - T_1)\ (z_2 - z_1)}{(u_2 - u_1)} = 3.4\ \frac{(T_2 - T_1)\ (z2 - z1)}{(u_2 - u1)} \qquad (16)$$

where g is 981 cm/sec^2 and T is 290^o K. This evaluates the relative importance of free or buoyant convection and forced or mechanical convection. The correction factor as employed by Penman, Angus and van Bavel, takes in a constant $\sigma' = 10$, the factor being either $(1 + \sigma' Ri)^{-1}$ or $(1 - \sigma' Ri)$. To keep the factor to the minimum and to avoid absurdly large or negative corrections the authors' found it best to divide by (1 + 10 Ri) when Ri is positive and to multiply by (1 - 10 Ri) when Ri is negative. In an example developed for lapse conditions by day, a multiplication factor of 1.07 was used. Such a correction is of course purely empirical and has been successfully applied for a particular location.

Direct measurements and indirect calculations of the sensible and latent heat fluxes over land vary widely in design, but have the common qualities of being few in number and encompassing short periods of time. Thornthwaite and Holzman (1942) used aerodynamic methods to measure the vapour flux over a grassland cover at Arlington, Virginia. The measurements spanned a full year but were not complete for the months of February and September. For a complete 10 month period they measured 12.0 inches of evapotranspiration from a total rainfall of 26.3 inches. There was at this early date no attempt to fit the results into an energy-balance format. Thornthwaite et al. (1954) published energy-balance calculations for a few days over a grass surface in the Nebraskan prairie. Their results which are shown in Table 1 indicate that when the soil is moist, grassland is an efficient evaporator and can use over 80 percent of the net radiation for evapotranspiration.

Table 1
Energy-budget over grassland. Values are in cal/cm^2 for the total period of measurement. Thornthwaite et al. (1954).

Date	Rn	E	A	S	E/Rn	A/Rn	S/Rn
Aug. 13-14	463	377	56	30	0.81	0.13	0.06
Aug. 18-19	342	287	59	-5	0.84	0.17	-0.03
Aug. 22	333	216	98	19	0.65	0.29	0.06
Aug. 25	355	132	181	42	0.37	0.51	0.12
Aug. 31	315	45	242	28	0.14	0.77	0.09

Thornthwaite and Hare (1955) in commenting on the results of Table 1, note that forest vegetation can do little better, for even if all the energy is used for evapotranspiration, the total flow can be only 25 percent greater than it is for a grassland. They also comment that those forms of forest having a lower albedo than grassland will absorb more energy, and will as a result have more available for evapotranspiration, but the difference will be small. It is noteworthy in Table 1, that on August 31, 2 weeks after the moist period, the available energy which was used in evapotranspiration had fallen to 14 percent due to a rapid drying of the soils. Baumgartner (1956) used both aerodynamic methods and evaporimeters in the calculations of the mean daily energy-balance for a young spruce forest, which are shown in Table 2. His measurements extended over an 8 day period, and it can be seen that two-thirds of the net radiant energy was returned to the atmosphere as latent heat.

Table 2
Heat-balance of a young spruce forest. Values in ly/day.
Baumgartner (1956).

Term	ly/day	Percent of Rn
Net radiation	586	
Latent heat flux	386	65.9
Sensible heat flux	197	33.6
Soil heat flow	3	0.5

Using an approach similar to that of Baumgartner, Dzerdzeevskii (1963) calculated the evaporative and convective heat movements over a mixed deciduous-coniferous forest for the four different seasons given in Table 3.

Table 3.
Energy-balance (ly/day) for different seasons of the year. Period (a)-time of leaf formation in spring; (b) mid-summer period; (c) beginning of leaf fall in autumn; (d) period after defoliation.

Periods	(a)		(b)		(c)		(d)	
	ly/day	(%) Rn	Ly/day	(%)Rn	ly/day	(%)Rn	ly/day	(%)Rn
Rn	362		310		208		91	
E	258	71.5	229	74.0	140	67.5	34	37.4
A	101	28.0*	82	26.5	81	38.8	51	56.0*
S	2	0.8	4	1.3	6	2.4	10	11.0

*Values do not add up to 100% because the measurements do not balance.

In Dzerdzeevskii's study it is apparent that the latent heat flux utilizes a large amount of the available energy in spring and early summer periods, but that the evapotranspiration decrease rapidly toward the end of the growing season. Such a decrease may be a function either of less available soil water, or decreased transpiration efficiency on the part of the trees; or a combination of both factors. Graham and King (1961) used both weighing lysimeters and the anergy budget method (employing the Bowen Ratio) to compute evapotranspiration from a corn crop at Guelph, Ontario, and achieved a close correspondence in results from the two techniques. For periods when the corn crop was mature (covering 80 percent of the ground) and the area was uniformly moist, the energy used in evapotranspiration made up 79 percent of the net radiation. However, when the corn had been irrigated and the surroundings were dry, the evapotranspiration used up to 12 percent more energy than was supplied by net radiation. This latter result, the authors concluded, was due to the horizontal divergence of sensible heat within the irrigated area. Mather and Thornthwaite (1956), (1958) calculated heat balances for a moist site of grass and bare soil at Point Barrow, Alaska, during a period which extended from July 1956 to September 1958. They measured net radiation and soil heat flow directly and used aerodynamic techniques to calculate the evaporative and convective heat fluxes. For 13 different days during July-August 1957, the energy used in the evaporative, convective and soil heat fluxes was calculated to be 37, 51 and 12 percent of net radiation respectively. In view of the wet nature of the surface, the small latent heat flux is surprising. The authors attributed it to the occurence of permafrost but did not elaborate further. Penman, Angus and van Bavel (1964) detail two unpublished energy-balance studies by Monteith and by Fritschen and van Bavel respectively. Monteith whose results appear in Table 4, determined the energy budget for a short-cropped grassland in England for period (a) when the soil was dry after drought conditions; and period (b) when the soil was wet after a rain. For the former period, the energy used in the sensible, latent and soil heat fluxes equalled the net radiation, but in the wet period, negative sensible and soil heat flows provided extra heat energy so that the evaporative flux consumed more energy than was available from the net radiation.

Fritschen and van Bavel calculated the heat

Table 4
Energy budget calculations over grassland, Harpenden, England. Values in ly/day for Period (a) - dry soil and Period (b) - wet soil. Monteith (unpublished).

	Period (a)		Period (b)	
	ly/day	(%) of Rn	ly/day	(%) of Rn
Rn	234		210	
E	122	52	230	109 *
A	102	44	-12	
S	10	4	-7	

* Downward convective heat flux and upward soil heat flow provide extra heat energy for evapotranspiration.

balance for a one day period of measurement over a field of sudan grass in Arizona. The measurements were made three days after the field had been irrigated. The results resemble those of period (b) in Monteith's experiment, except that the negative flux of sensible heat was greater. The energy which was used in evapotranspiration was 59 percent greater than the energy available from the net radiation. Three-quarters of the extra heat supply came from sensible heat which was extracted from the air (i.e. A was negative) and the remaining one-quarter was gained from the upward directed soil heat flow. Denmead (1964) measured the available energy at various heights in the canopy of pine forest and partitioned the available energy into its component fluxes of sensible and latent heat. Among other things he concluded that, at levels within the canopy where the energy used in the latent heat flux is greater than the net radiation at those levels, the higher latent heat flux is promoted by the vertical movement of sensible heat from other levels in the canopy (usually from higher to lower levels), by a net gain of heat from the overlying atmosphere and an upward flow of heat from the soil.

(e) Heat Exchange in the Air Layer and Wood Mass of the Forest

The energy exchange within the three-dimensional layer employed by King includes Δ T crop and ΔT air, the last two terms in equation (2). The exchange of heat in these two media mainly follows a diurnal pattern, and any gain or loss of heat from season to season is insignificant in comparison to the overall energy-balance.

During the daily period the air and plant matter in the vegetative layer will store heat as warming takes place through the early morning hours, and will liberate a similar amount of heat during the evening period of cooling. There will be a short lag in the heating and cooling of the forest wood mass behind similar processes in the forest air. The heat exchange in the air varies with the depth, pressure, temperature and relative humidity of an air column in the active surface layer. For example, an air column 16M deep and 1 cm. square, at a pressure of 1000 mb., with temperature and relative humidity of 10°C and 75 percent respectively, would require 0.0008 cal. to raise its temperature by 1°C. To calculate the heat exchange in the wood mass, the density of the wood and its specific heat must be known. For a mature forest 16 M in height with a specific heat for the wood mass of 0.005 cal/g of wood mass/°C temperature change, a 1°C increase in temperature of the total wood mass would require 0.08 cal/cm^2. Since the change in tree temperature which responds to the air temperature change effects only the outer parts of the tree, the actual wood mass which undergoes significant temperature change during daily periods will involve less heat exchange than that which has been calculated above. Table 5 gives Baumgartner's (1956) calculations for the heat exchange in the forest air and wood mass during different hours of the day.

Table 5
The movement of heat energy in the air and wood mass of a spruce forest layer in comparison to net radition (Baumgartner 1956). Energy values in cal/cm^2, min.

Hour	6	12	18	24
Net Radiation	0.30	1.17	0.20	-0.06
Heat Movement in the air Layer	0.007	0.003	-0.006	-0.001
Heat Movement in the Wood Mass	0.045	0.025	-0.045	-0.015

(+) energy used for heating
(-) energy liberated during cooling

It can be seen that during morning and evening hours the heat which is used or liberated in the heating and cooling processes may be equivalent to 16 percent of the energy of net radiation, but during the high sun period this fraction is reduced to less than 3 percent. For the mean daily values calculated by Baumgartner from 8 days of records, the net heat exchange in the forest wood mass and air layer was nil.

The author concludes that for hourly periods during the day, the heat energy which is used or liberated in the cooling of the forest wood mass and air, can form a significant part of the total heat balance of the forest layer. For periods longer than a day, such heat exchange becomes negligible in comparison to the major terms in the energy-balance equation.

B Forest Hydrology

The traditional hydrologic-balance formulation which consists of a downstroke (rain), overland or underground flow, and upstroke (evapotranspiration), forms a convenient frame of reference within which to discuss forest hydrology.

(a) Rain

Rain as an outcome of atmospheric processes is a meteorological problem and is not considered in this study. However, the disposition of rain as it enters the active surface layer is of some concern. The amount of water which penetrates through a forest canopy varies with both intensity and duration of the rainfall, and with the density of the forest crown. Some rain falls through openings in the canopy without touching leaves or stems, but most is intercepted, at least temporarily, by the tree crown. At the beginning of a storm, the interception is nearly complete, but gradually the rain drops begin to seep through the canopy. The water which drops and runs to the ground is not distributed uniformly, but rather is concentrated below drop points and near the base of stems. Chapman (1948) states that water drops from the canopy are large, and if they fall more than 25 feet, they have a greater striking force than all but the largest raindrops. The litter layer absorbs this drop impact with no consequence to the soil, but where the litter is scarce or absent as a result of fire or other disturbance, canopy drip may cause considerable soil disturbance. Penman (1963) summarizes the empirical results of canopy interception from some North American sources. Forests of Douglas Fir apparently give the greatest interception at 40 percent of normal rainfall. Coniferous forests in general offer greater protection than do their hardwood counterparts. Hoover (1962) notes that canopy interception is of major importance only in smaller storms. For storms giving a rainfall greater than 2 inches, the interception would be less than 0.2 inches. In addition to interception in the forest canopy, rainfall may be prevented from reaching the mineral soil through interception by the leaf layer on the forest floor. For oak forest in Tennessee, Blow (1955) measured litter interception at 2 percent of the total rainfall. Similar results were achieved by Curtis (1960) and Helvey (1964) who measured a 3 to 4 percent interception of the total rainfall for the the leaf litter of hardwood forests in Wisconsin and the southern Appalachians, respectively. Intercepted water is not available to trees as a nutrient-carrying medium, since it never penetrates to the mineral soil. It does, however, enter into the evapotranspiration term of the energy-balance equation, since it stores energy when being evaporated from the forest canopy. Rakhmanov (1958) by weighing wetted tree branches and dry ones in glass jars, found the transpiration from the dry branches to be about 25 percent greater than from the wetted branches. This result would be expected from energy-balance considerations.

The amount of water which reaches the ground by flowing along tree branches and stems, the so-called stemflow, varies both with the total amount of rainfall, and with the shape and density of the forest crown. The percentage of total rainfall which reaches the forest floor as stemflow is, as a result, highly variable. For a desert woodland in central Australia, where the trees were spaced about 2 M apart and were from 5 to 6 M in height, Slatyer (1961) found that 40 percent of the water reaching the ground came from stemflow. In contrast, Ovington (1954) measured the stemflow for 12 plantations of different tree species in England, and found that red oak gave the greatest stemflow at only 0.32 percent of total rainfall, and that tamarack and cedar trees gave the least at 0.05 percent.

The measurement of rainfall above and within forests presents some problems. If measurements are made in forest clearings, there is the chance that special turbulence effects will make the readings non-representative of the above-canopy rainfall. The same problems may be encountered for measurements near forest border areas, where air movement may be predominantly upward or downward. Geiger (1950) carried out measurements in clearings of different sizes located in a mixed pine-beech forest, which averaged 25 M in height. In comparing his measurements to open site measurements which were upwind from the forest, he calculated ratios of (P clearing/P open) which ranged from 0.87 for a clearing 12 M in diameter, to 1.05 for a clearing of 28 M in diameter. In order to obtain representative rain measurements under the tree crown it is necessary to place a dense network of gauges within a small area because of the variable density of the forest canopy. Mention can be made of the very early work of Hoppe (1896) quoted in Geiger (1950). Hoppe arranged 20 rain gauges on the forest floor in two rows which crossed one another at right angles, and found an average difference between gauge readings of around 25 percent. For a single fall, however, the difference might exceed 100 percent.

Penman (1963) reviews the effects of horizontal interception of rain and fog by forests. The evidence, although not conclusive, suggest that in areas of prevalent fog such as mountain tops, fog may indeed by an important source of moisture. In more normal topographic situations, however, it seems safe to conclude that it is only the forest border areas which are likely to have their water supply much augmented from this source.

(b) Infiltration and Runoff

With the exception of permanent ponding in marshes or swamps, which can be considered a special situation, all the rainwater which reaches the forest surface is either stored in the leaf litter, a process which has already been noted; penetrates into the mineral soil; or runs off directly as surface flow. Hoover (1962) distinguishes between the processes of infiltration and percolation. The former is defined as the flow or movement of water downward through the soil itself. The degree to which incoming precipitation is converted to overland runoff is determined by conditions at the immediate soil surface. Soils of course texture or with fine particles clumped together into stable large aggregates have high infiltration rates. Duley (1930) found that whereas over bare soils the action of flowing water plugs pore spaces with fine particles the litter layer under a forest filters out fine particles to prevent the stoppage of the large pore spaces. In addition to the physical character of the soil, the soil surface tension is an important factor in influencing infiltration rates. Penman (1963) states that the higher the amount of evapotranspiration at the soil surface, the drier the soil surface layers become, and as a result a strong moisture tension is built up. After a dry period, the first rainwater which reaches the surface is literally "sucked" into the soil. When the whole of the profile is rewetted, the tension is neutralized, and a steady rate of percolation is attained, determined by the permeability of the soil.

Numerous studies have shown higher infiltration rates in forest floors than in corresponding areas of cultivated crops or pasture lands. Zonn and Mina (1949) quoted in Penman (1963) found a 7:1 ratio in infiltration rates which favoured a 160 year old oak forest over an open steppe. Mather (1953) describes the value of the high infiltration rates of forest soils for commercial purposes in New Jersey. Here the industrial effluent from canning factories was irrigated into forested areas at the rate of 400 to 600 inches/year, with no resulting surface runoff. Hoover (1962) claims that infiltration rates are seldom a concern under tree covers, where the forest floor layer is well-developed. He notes, however, certain conditions which may decrease the infiltration capacity at the forest floor. Occasionally, hardwood leaves may pack tightly to form a shingle-like surface over which water flows rapidly. In addition, dry organic layers are difficult to wetand heavy rains on such a surface may result in large quantities of runoff. Experiments by Delfs et al (1958) show that the type of leaf litter has some influence on runoff. For forests on a 40 percent gradient, they found that runoff from a broadleaf-humus-covered floor was 4 percent of the total rainfall, but that with a coniferous needle litter, it fell to 1 percent. These values for woodland soils compared to 17 percent runoff over bare soils of similar slope.

(c) Storage and Water Movement in the Forest Soil

Soil consists of a mixture of solid particles, air and water. Hoover (1962) states that 50 to 60 percent of the soil is composed of pore spaces, and it is the size, shape, distribution, and continuity of the soil pores which determines the rate at which water will move, and the amount which will be held. Water movement is influenced by the action of both gravity and capillary pull. The smallest pores hold water against gravitational forces and are emptied by evapotranspiration. In contrast, the water in larger pores moves primarily downward under the influence of gravity. The former water does not contribute directly to stream flow and is labelled "retention" storage by Hoover, so as to distinguish it from the "detention" or temporary storage within the large pore spaces. The upper limit of retention storage is field capacity — the maximum amount of water which a soil can retain against the pull of gravity. The lower limit is the permanent wilting point — the moisture content, below which roots no longer extract water.

When retention storage in the smaller pore spaces is filled, water begins to fill the larger voids. It is in these larger pores that water moves rapidly during rainfall. Such movement may be both parallel to the slope and downward in the large pore spaces. In forest soils, root channels may be the principal pathway for water movement, and Gaiser (1952) found more than 4,000 vertical channels per acre in a hardwood forest growing on silt loam soils. Such channels were formed by root decay and radiated outward from the central core. Gaiser states that in the course of one forest generation, several thousands of vertical channels must be formed.

Whether such channels are hollow, filled with organic matter, or contain loose mineral soil, they are very efficient in distributing water throughout the deeper soils. An interesting experiment by Hursh and Hoover (1941) illustrates the importance of lateral movement on sloping lands. The authors dug a pit in a hillside in order to catch laterally-moving ground water. By artificially applying water, they found that 2.5. percent was caught in the forest litter layer, but that 12.5 percent moved downslope within the top 12 inches of soil. When water enters the soil at a rate faster than it can move into lower layers, it is stored in the large pore space above the limiting layer. Hoover (1962) notes the importance of such detention water. It both prolongs the period during which water can percolate into underlying strata and provides emergency storage for water which would otherwise be forced to move over the soil surface.

(d) Evapotranspiration

The transfer of water vapour to the atmosphere from a vegetated surface represents the most complex part of the hydrologic cycle. During the discussion of energy-balance theory, some attention was paid to the latent heat flux, because it represents an important term in the heat balance equation. The discussion, however, considered only a single aspect of evapotranspiration, albeit an important one -- namely the turbulent flux of water into the atmosphere. At this point a thorough discussion of the overall process of evapotranspiration is in order.

The term evapotranspiration is a well-accepted one. It encompasses all the losses of water from the earth's surface to the air, including direct evaporation from soil surfaces, the evaporation of rainwater which has been intercepted by plant leaves, and the transpiration of water through the stomata of leaves. Thornthwaite and Hare (1964) dissect the evapotranspiration process into five parts as follows: (1) the movement of water within the soil towards the soil surface, or into the zone of absorption around each active root system; (2) the movement of water into roots, and thence up through the plant tissues to the green stem and leaf surfaces; (3) the vapourization of this water either at the soil surface or at the stomata of the plants, with a large conversion of energy into latent heat; (4) the vapourization of rainwater or snow resting on the outer plant surfaces; (5) the turbulent removal of the evaporated water by the eddy motion of the lower part of the planetary boundary layer. Apart from process (5) which has already been discussed in some detail and process (4) which represents the physical transformation of liquid water to the vapour state, it will be useful to consider the other movements of water in the vegetative layer.

Water normally moves from the soil into the root in response to pressure differences. In a thoroughly wet soil such pressure differences are the main control over the rate of water absorption into the roots. When a moisture deficit develops in the soil, however, the volume of water in the soil, the size of the absorbing root surface, and the stress which is set up in the water stream due to transpiration from the shoot, all become important factors in determining the absolute rate of water absorption. Slavik (1964) states that the horizontal extent of the root system is mainly determined by competition, whereas the vertical extant is conditioned by the depth of the water supply, and the state of aeration in the soil. Slavik points out that in moist soil, far fewer root hairs are produced. If water content is low but there is enough to insure growth, roots are stimulated to greater development. Weaver and Clements (1929) point out the example of corn, which is grown in moist rich soil with a water content of 19 percent by weight, and the same crop growing in a drier soil of 9 percent water. In the former situation, the total area of the roots averaged 1.2 times the total area of the aerial parts, whereas for the drier soil the root area was 2.1 times as great. With a high water table, plants root shallowly in response to a lack of aeration. In bogs, plant roots and rhizomes commonly form a mat which is only a few inches thick above water level.

Although plants transpire essentially through their whole surface, the most intensive transpiration takes place in the leaves. Water vapour moves by diffusion out through the stomata of the leaves, so long as the vapour pressure of the atmosphere is less than that of the intercellular spaces. As well as stomatal transpiration, some water evaporates directly from the exterior surfaces of the leaf and stem. Meyer and Anderson (1952) believed the amount to be less than 10 percent of the total transpiration. The morphology of the leaves permits a ready intake of carbon dioxide, and their shape and orientation permits the exploitation of incident sunlight. Transpiration and convection over the relatively large leaf surface causes the leaf temperature to be near ambient air temperature. However, in strong insolation, the leaf temperature is normally a few degrees higher than the surrounding air temperature, and in some plants, intensive transpiration will reduce leaf temperature below that of the air. Wang claims that leaf temperatures may rise up to 25^{o}F higher than the surrounding air, under conditions of bright sunshine and light winds.

Waggoner and Shaw (1952) in measurements on potato and tomato plants showed that on a clear day upper exposed leaves were warmer than the air and lower sheltered leaves were cooler than the air. At night, upper exposed leaves became cooler and the sheltered leaves warmer than the surrounding air. Shaw (1954) measured the air temperature of tomato leaves to be 1.8^{o}F lower than that of the air 1 cm. above the leaves during clear calm night-time periods. Loomis (1965) noted that leaves in sunlight are heated from a few degrees above air temperature for thin leaves up to 30^{o}C or more for thick leaves before reaching a steady temperature. Moreover he found that the heating and cooling rates for wilted leaves are not significantly different from those of transpiring leaves. Leaves in sunshine are cooled quickly toward air temperature by winds of 5 miles/hour or more.

The opening and closing of the stomata is the most effective control which a plant has over transpiration. Hoover (1962) points out that even though the stomata occupy only 1 to 3 percent of the surface area of the epidermis their capacity to carry water vapour is much larger than is required for the maximum transpiration rates. Penman and Schofield (1951) state that the resistance to flow from stomata, is more important than the structure and number of stomata, and they claim that for practical purposes, all that matters is whether stomata are open or closed. Konis

(1950) found that transpiration increases rapidly with increasing leaf temperature, and as has been noted above, during the daytime period the temperature of leaves is commonly higher than that of the air about them.

The cooling of the leaves by transpiration prevents excessive heating and is, therefore, one of the most important benefits of transpiration. If transpiration is reduced by inadequate water supply, overheating can cause irreversible heat injury even in the temperate zone. Thornthwaite and Hare (1964) when considering the artificial control of evapotranspiration by using ingested substances to inhibit transpiration, note that the Bowen Ratio will be altered. This will create a corresponding change in leaf temperatures which might be injurious to the plant. Like most higher plants, trees control their water balance to a degree by lowering transpiration or increasing absorption. As well as a possible seasonal water deficit, Slavik (1964) pointsto a regular daily deficit due to a temporary midday excess of transpiration over absorption. This leads to a closure of the stomata and a drop in transpiration rate, until the water balance starts to improve again. This is important, in that it indicates a possible reason why, even when there is abundant moisture in the soil, the latent heat flux rarely equals the heat energy which is available from net radiation.

Evaporation from the forest soil is dependant on meteorological factors, on soil properties, and on the moisture content of the soil. The presence of the plant cover greatly reduces evaporation from the soil surface. Both the crown canopy and the forest floor are effective in reducing evaporation. Kittredge (1948) states that the evaporation from forest soil is 10 to 80 percent of that from bare soil. In a later study, Kittredge (1954) carried out simultaneous measurements of evaporation from bare soil and forest soil in California. He found that under the litter and shade of pines, the soil evaporation was about two-thirds that of the bare soil in the open from March to August, but that for September through February this ratio increased to five-fourths. The seasonal differences in these ratios reflects the fact that for the second period, more moisture for evaporation was available in the forest soil than in the bare soil.

The foregoing discussion on the part played by vegetation in the evapotranspiration process, leads one into a consideration of the concept of potential evapotranspiration. Thornthwaite (1944) defined potential evapotranspiration as the water loss off a moist soil tract which is completely covered by vegetation, and which is large enough for oasis effects to be negligible. Penman (1956) gave a similar definition for "potential transpiration" as the amount of water transpired by a short green crop which completely covers the ground and has an abundant water supply. Thornthwaite's definition does not specify any particular type of vegetation, but only demands a complete ground cover, whereas Penman is somewhat more specific in detailing a short green crop. The whole idea of potential evapotranspiration involves a climatological control in the nature of available energy, rather than any physiological control on the part of the plants. If there is no advective energy input and the heat storage in the soil and plant layer is small, it might seem reasonable that with an abundant water supply the evapotranspiration should equal the energy available from net radiation. However, as was pointed out by Thornthwaite and Hare (1964), this correspondence involves a zero eddy heat flux which in nature is not often realized. In many circumstances the field investigator is probably safe in assigning to the type of vegetation little role in determining the rate of potential evapotranspiration. However, some types of vegetation may lose transpiring efficiency or possess an unusual environmental adaption. In the first category one can note the work of Lemon, Glaser and Satterwhite (1957) who found that irrigated cotton showed a marked reduction in evapotranspiration after maturation in September. Similarly Denmead and Shaw (1959) found a reduction in evaportranspiration as a corn crop matured, but this latter evidence was opposed by that of King (1961) who found no decrease in the transpiration of a corn crop until after the first frost. Thornthwaite and Hare (1964) maintain that for natural vegetation, any effect of maturation is minimized because of the variation in development among members of the association. In the category of unusual environmental adaptation, the work of Nebiker and Orvig (1958) provides a good illustration. The authors found that because a lichen is primarily independant of the soil surface, a lichen cover in the subarctic did not use more than one-third of the energy of net radiation for evapotranspiration. When water is non-limiting it is probably safe to assume that the role of most different types of natural vegetation is not very important in determining the rate of evapotranspiration.

When a water deficit develops, the type of plant can become quite important in determining the amount of actual evapotranspiration.

Celjniker (1957) quoted in Penman (1963) examined the transpiration from oak and ash trees in relation to the changes in the soil moisture content in the top layers of a chernozem soil in Russia. She found that there was little change in transpiration rate until the soil water content fell beneath 75 percent of field capacity. There was then a sharp drop in the transpiration rate, after which it became constant at a lower rate for the drier soils. The depth of the water tables in clay soils under spruce and beech forest during two years of measurement at the Danish Experimental Station were reported by Holstener-Jorgensen (1959). Because the water table rose to the same level at the end of the growing season as it held at the beginning, it was apparent that total evapotranspiration was equal to total rainfall during the growing season. Holstener-Jorgensen's results are summarized in Table 6.

Table 6
Two years of evapotranspiration from spruce and beech forests in Denmark (Holstener-Jorgensen 1959).

Year	Spruce	E	Beech	E
1956	April 19-Dec.14	481 mm	Apr.15 - Dec.15	471 mm
1957	April 10-Dec.28	417 mm	Feb.28 - Dec.28	469 mm

The data indicate that during 1956 the difference in transpiration from the two tree types was negligible, but for 1957 the evapotranspiration from the beech forest exceeded that of the spruce forest by 11 percent. It is noteworthy that the evapotranspiration from the hardwood forest was the same each year. Urie (1959) compared the rates of soil moisture depletion between pine plantations about 25 years old and oak stands of 60 to 70 years of age in Michigan. He found that for the whole growing season both forest types ultimately utilized most of the stored moisture in sandy soils, and found also that there was no major difference in total evapotranspiration between the two forest types. At different times in the season, however, the rate of withdrawal did vary. The pine stands started using soil moisture in April and May and evapotranspired at a uniform rate until the moisture reserves in the rooting zone were nearly exhausted in late July and early August. In contrast, the oak trees evapotranspired very slowly until full-leaf development was reached in June. During July the rate of moisture depletion was very rapid and the soil moisture reserves reached the same low level as those under pine by August 1. By mid-October, the soil moisture under both forest types had recharged to the same levels.

A very important problem involves the relation of the magnitude of forest evapotranspiration in comparison to other cover types. Hoover (1962) states that forest trees develop to use as much sunlight as possible and probably deplete soil moisture more thoroughly and more deeply than other types of vegetation in the same climatic zones. He cites the following hypothetical example.

> "Assume that climatic conditions are such that 0.25 inches of water can be evaporated per day and that soil moisture is at field capacity. It is desired to estimate the difference in soil moisture after 14 days under pine forest, grass, and bare soil conditions. Pine roots extend to depths below 6 feet and grass to only 30 inches. For the first few days, the soil at each site dries rapidly but after 4 days, evaporation on the bare site decreases from 0.25 per

> "day to 0.04 inches per day, and this rate is maintained for the rest of the period. At the end of 14 days, 1.4 inches of water is removed on the bare site, and nearly all of this is from the surface 15 inches of soil. At the grass plot, a loss rate of 0.25 inches per day continues for 12 days to remove 3 inches of moisture from the 30-inch root zone and to reduce it to the wilting point throughout. The rate of loss then becomes negligible and the total loss in 14 days is 3 inches . Under the pine forest, 3.50 inches of water will be removed by the end of 14 days. If at the end of the 14-day interval, a heavy storm occurs, the retention storage capacity will be 1.4 inches on the bare site, 3.0 inches under grass, and 3.5 inches under forest. The emptying of soil moisture storage provides room for storm rainfall and influences flood peaks during the growing season."

It is important to determine the extent to which Hoover's assumptions are substantiated by field measurements. From soil moisture measurements on forested and grassland areas in northern Wisconsin, Thames, Stoeckler and Tobisaki (1955) found that forest depleted soil moisture more rapidly than a grass cover. In 1960 an United State's Water Resources Committee report to the Senate concluded that the greater evapotranspiration from forested areas decreased the water yield into streams. The committee predicted, that selective cutting in forested areas could increase the water yield as much as 20 percent by reducing interception and evaporation losses. Schneider and Ayer (1961) studied reduction in streamflow for three watershed areas in central New York which were partially reforested with pine and spruce trees between 1932 and 1958. They found there was an average reduction in streamflow of 24 percent, with seasonal maximum reduction of 37 percent between May and October. They felt that the increased canopy interception of rain accounted for a major part of the reduction as well as a higher total evapotranspiration, the latter due in part to the greater soil moisture capacity in forest soils. Some interesting observations were obtained by Fletcher and Lull (1963) when measuring soil moisture under undisturbed forest plots, plots where the trees were cut down but the leaf litter left in situ, and plots where the trees were cut and litter burned. It was found that when the soil moisture was at field capacity, the bare soil areas had drying rates which were two-thirds those of the undisturbed forest soil. During June, July and August of a dry year, the undisturbed forest soils averaged 2.8 inches less soil moisture in 40 inches of soil than the open burned soil plot, and 5.8 inches less than the open litter-covered soils. Flectcher and Lull noted that during a period of wet years the foliage density of hardwood stands tended to become excessive, but in dry years the density diminished. Muller (1963) observed the effects of reforestation on water yield in the Alleghany Plateau area of New York, and found a 25 percent greater evapotranspiration loss from woodland areas than from nonforested regions. He felt that it was created by a lower albedo and a greater availability of soil moisture in the reforested zones. The results of evapotranspiration from soil-block lysimeters which were planted in mulched and unmulched hardwood seedlings, pine seedlings, and annual grains, are given by Sartz (1963) for two 3 year periods between 1936-38 and 1939-41. His overall measurements indicate that the evapotranspiration from the pine seedlings exceeded that from the mulched hardwoods, unmulched hardwoods and annual grains by 14,25, and 22 percent respectively. Sartz stated that it would be presumptuous to draw inferences on the relative effects of forests and agricultural crops on different elements of the water cycle from the data which he presented. Finally Trimble and Reinhardt (1963) compared the results of four degrees of tree cutting ranging from commercial clearing to a light selective cutting on stream runoff. They found that the increase in water yield from the watersheds was roughly proportional to the degree of cutting.

Various methods have been proposed for estimating the water balance of vegetation-covered surfaces and some have been widely accepted. Of these, one group of methods employs easily obtained meteorological data to determine potential evapotranspiration, a second group measures potential evapotranspiration instrumentally, and a third measures actual evapotranspiration directly in the field. Each of these types of calculations will be discussed briefly in the following pages.

(1) Estimates from Meterological measurements

The most widely-used systems for determining the water balance from meteorological data are associated with the names of Thornthwaite and Penman. The two systems have in common, a means of computing potential evapotranspiration, the determination of actual evapotranspiration from a knowledge of potential evapotranspiration, rainfall and soil moisture changes, and a means of determining soil moisture variations.

Thornthwaite (1944) first introduced the concept of potential evapotranspiration (PE), and elaborated more fully in a later paper, Thornthwaite (1948). The author expressed PE as a function of mean temperature in the following empirical equation:

$$PE = 1.6\ (10\ t/I)^a \qquad (17)$$

where t is mean monthly temperature in degrees C, I is a heat index and equals $\sum_{1}^{12} (t/5)^{1.514}$, and a is a cubic function of I. Although the equation is complex and non-linear, actual values for PE are readily calculated from tables which find their most refined form in Thornthwaite and Mather (1957). The expression is thoroughly empirical since it regresses potential evapotranspiration on mean temperature, from data obtained in watershed and lysimeter measurements in continental United States. Pelton, King and Tanner (1960) in an evaluation of the Thornthwaite system note that mean temperature cannot be relied upon for general use in estimating evapotranspiration during short periods. In measurements at Wisconsin they found that thermal storage in the soil, the effect of variable moisture on the Bowen Ratio, and the sometimes large influence of thermal advection, could seriously influence the correlation between net radiation and mean temperature and hence the potential evapotranspiration.

Once the potential evapotranspiration is determined it is relatively straightforward to derive actual evapotranspiration from a knowledge of rainfall and assumptions as to soil moisture capacity and soil moisture storage change.

Penman's approach to potential evapotranspiration involves a marriage of the energy balance and aerodynamic methods of calculating the latent heat flux. The derivation of Penman's equation can be obtained from Penman (1948, 1956) and Penman, Angus and van Bavel (1964).

The most recent form of the equation reads as follows:

$$PE = \frac{\Delta/\gamma \ Rn + Ea}{\Delta/\gamma + 1/XD} \qquad (18)$$

where Δ = e1 - e2/T1 - T2, the slope of the saturation vapour pressure curve, γ is the psychrometer constant, Rn is net radiation and $Ea=B_T \ (e_a - e_2)$. B_T represents an empirically determined function of wind speed, crop height and roughness, and $e_a - e_2$ is the saturation deficit above the surface. X and D are a stomatal factor and a daylength factor respectively. Penman's method for determining potential evapotranspiration necessitates measurements of vapour pressure and temperature gradients, net radiation, and wind speed. It is a technique which will be confined to areas where these measurements can be taken or estimated.

(2) Determination of potential evapotranspiration instrumentally.

Various mechanical devices have been employed to measure potential evapotranspiration in the field. Some rather closely simulate the natural surface to which their values are applied, while others bear little resemblance. The types of instruments can be divided into the three broad categories of pans, atmometers, and lysimeters.

The use of evaporating pans is more widespread than any other method for estimating evaporation. Mukammal (1961) presents a thorough critique on the use of evaporating pans. He states that they are by no means ideal evaporation instruments and their use is at present mainly limited to open water evaporation (i.e. they are not suitable for measuring potential evapotranspiration from land surfaces). Mukammal and Bruce (1960) did find, however, that for the pan the relative importance of radiation, humidity and wind was in the ratio of 80:6:14

which does show that radiation is the most important evaporating agent as it is with a transpiring vegetation-covered land surface. Thornthwaite and Hare's (1964) remarks can be accepted as a fair criticism:

> "Evaporation pans standing on a land surface break the cardinal rule of evaporation measurements, that the observations should come from a small area of surface undistinguishable from its environment. Pans in effect measure the evaporation off pans."

The same criticism as above can be made of the various types of atmometers. There are various varieties of atmometers, but they all have in common a small disc which is constantly wetted by water from a storage cylinder. This disc is meant to represent the evaporating surface and the change in the storage water represents the amount which is evaporated. As noted by Mukammal (1961) the greatest response of these instruments is to wind. Prescott and Sterk (1951) found that it is possible to use the Piché atmometer as a simple measure of the evaporating power of the air, but that this value could not be appropriately regarded as an index to evaporation from soil or vegetation. Unlike the pan, radiation is not the most important factor in evaporation from the atmometers. Mukammal and Bruce (1960) for the importance of radiation, humidity and wind, derived rations of 41:7:52: Because they are so sensitive to wind and have energy-balances quite different from those of green transpiring vegetation, it is unlikely that such devices can ever be relied upon to give useful indications of potential evapotranspiration.

Potential evapotranspiration can be measured directly through the use of evapotranspirometer tanks or lysimeters on which a plant cover indistinguishable from the surroundings is established. These tanks may be either fixed in position or of a weighing type. The weighing lysimeters are either directly weighed on a large balance built into the field site, or are weighed by the displacement of a liquid. The principle of all lysimeters is the same. A column of soil is enclosed on all sides except the top by a waterproof container. From the base of the container the percolation water is drained off to be measured. The tanks are irrigated regularly except during periods of abundant rainfall so that the soil is always kept near field capacity. The difference between irrigation water plus rainfall and the percolation gives potential evapotranspiration. In the stationary type of tank, water is supplied by sprinkler irrigation and runoff passes from the bottom of the tank to a stilling well. These can be used only for measuring potential evapotranspiration. Harold and Dreibelbis (1958) discuss the monolithic lysimeters which employ direct-balancing mechanisms for weighing. In such instruments the change in weight of the soil body gives the potential evapotranspiration. The floating lysimeter employs an inner soil-filled tank floating in an outer tank which is connected to a stilling-well. With a change in weight of the soil body the water level in the stilling-well changes through liquid displacement. King, Tanner and Suomi (1956) give a thorough discussion of the floating lysimeter, its record, and its problems.

(3) Instrumental Measurement of actual evapotranspiration.

The weighing lysimeter can be used for continuous measurement of actual evapotranspiration. No irrigation water is applied so that the lysimeter receives and evapotranspires the same amount of water as the surrounding environment. Although such instruments are difficult and expensive to construct and require maintenance by skilled field workers, when operating properly their record is very valuable. The following points raised by Pelton (1961) can be applied to the use of all lysimeters.

(a) the ideal lysimeter should contain a representative undisturbed profile.

(b) the lysimeter itself should be indistinguishable from the surrounding area because only then is it a true sample of prevailing field conditons. To realize this uniformity the lysimiter should be in the centre of a large area planted, watered, fertilized and managed in exactly the same manner as the lysimeter itself.

(c) the heat storage and transfer in the lysimeter walls should not be different from that in the surrounding soil.

Pelton notes that few lysimetric installations meet the requirements set forth for the adequate study of evapotranspiration. The use of lysimeters for measuring forest evapotranspiration creates the major problem of the time needed in order to grow a mature tree. Patric (1961) states that lysimeters have found little use in forestry in contrast to agriculture. He notes problems such as the trees which are grown in lysimeters being smaller than their neighbours. The root systems cannot expand normally and drainage is retarded at the lysimeter bottom especially in small tanks, and this creates an artificial water table. Patric concludes that quantitative results from lysimeters should not be applied directly to natural forest conditions.

(C) Distinguishing Features of a Forest Microclimate

Heat and water-balance factors form an integral part of the forest microclimate. It is, however, other microclimatic factors such as temperature, insolation, humidity and wind, which have the longest history of investigation. Kittredge (1962) writes:

> "The influences of the forest in modifying climate under the trees are matters of common experience. In the woods the eyes are relieved from the glare of the sun. A strong wind in the open becomes a light breeze in the forest. In varying degree, the forest affects light and solarradiation, air and soil temperature, wind, atmospheric humidity, precipitation, evaporation and transpiration; and the weather in the forest is called the microclimate because it is sufficiently distinct from the weather in the open."

To date, any discussion of the forest microclimate has tacitly assumed that the forest is a homogeneous surface type. Forested areas are, however, not homogeneous in their microclimatic characteristics. The influence of the forest varies with the species of trees, their size and age, and the density of the stands. Differences in site, exposure, slope, latitude, altitude and time of day and year also exert a strong influence. The distinguishing features of a forest microclimate can be discussed within the general topics of vegetative influence and topographic influence.

(1) Vegetative influence

Geiger (1950) notes that "forest climate" was considered to be "trunk-space climate" by the early investigators, but that this is only one portion of forest climate. He goes on to state that no conclusion as to the effect of the forest on the climate can be drawn from a comparison between open country climate and trunk-space climate near the ground. Kittredge (1962) states that the forest exerts its influences through the crowns of the trees which form canopies and screens or barriers of varying density.

The influence of the canopy on solar radiation is most strongly felt. When the radiation is measured below the crowns of the trees at or near the ground, a remarkable range of variation is found within short intervals of time and space. At one extreme where sunlight passes through an opening between the crowns, the intensity may be almost 100 percent of that which arrives above the forest. At the other extreme beneath a dense canopy of two or more stories, the intensity may be reduced to less than 1 percent of that above the forest. Geiger (1950) summarizes some of the earliest studies of illumination within the forest carried out by Lauscher and Schwabl (1934), Sauberer and Trapp (1937) and Nagelli (1940). All of these investigators employed different varieties of photocells and quote the light under the forest as a percent of that above. In an investigation in Sweden, Wallen (1932) found that when the radiation above the forest was 1 ly min^{-1}, that under oak forest it was about 0.06, under pine, 0.04, and under dense spruce, only 0.008 ly min^{-1}. Baumgartner (1956) plotted some interesting graphs showing profiles of net radiation in a spruce forest over summertime diurnal periods. His results are important from the point of view of the present study and will be summarized in Chapter VI. Ovington and Madgwick (1955) carried out measurements in England comparing light intensities in foot-candles from beneath the canopies in various forest plots to those in the open. They found that their values varied considerably with the position of the sensors under the forest cover. The following quote is noteworthy. "Random readings taken under a canopy which looks reasonably uniform show a wide range of light intensities, and the distribution of these intensities is characteristic of the time of year and the particular canopy." Some very interesting measurements of radiation in hemlock and balsam forests are reported by Berger (1953) in the mountains of Oregon. It was found that as the crown coverage increased the net short-wave radiation decreased but the net long-wave gain from forest increased while the net long-wave radiation from the forest floor to the sky decreased. The total net radiation gain started at 0.12 ly/min when the canopy was very open, dropped to almost nil when the canopy coverage increased by 20 percent, and climbed to 0.06 under complete coverage. It is apparent that the trend of the net radiation is quite different from that of the short-wave radiation. Observations by Tanner, Peterson and Love (1960) for the radiation distribution in a corn field must be similar to the situation which develops beneath a forest canopy. The author's found that 80 percent was made up of long-wave radiation from the corn itself. A large portion of the outgoing radiation from the soil surface is intercepted by the corn and only a small portion escapes to the sky. This creates what is virtually a closed system for long-wave radiation exchange between the soil and the overlying vegetation. Angstrom (1925) one of the early workers in studies of forest microclimate had an acute insight into the energy and heat exchange within a forest. He notes that in the case of ground which is covered by trees and plants, it is a very weak part of the incident energy which reaches the ground in the form of radiation. Rather, Angstrom claims, the major movement of heat energy is in the form of a long-wave energy interplay between the vegetative layer itself, the vegetation and ground, and the vegetation and sky. Vezina (1962) found when measuring the transmission of solar radiation through spruce plantations that the percentage of transmission decreased as the solar altitude became smaller and smaller between April and October. He also found that the clear day transmission was lower than that on cloudy days. Dealing with the same problem, Dzarnowski and Slomka (1959) noted that radiation penetration is not only a function of the mass of forest foliage but also of the sun's zenith angle. They calculated theoretical values of penetration at different zenith angles based on a knowledge of age, height, diameter, number of trees and their basal area in a scots pine forest, and compared their calculations to field measurements. These calculations are given in Table 7, where Rsio is the incident radiation above the forest and Rsi is the radiation which penetrates to the ground.

Table 7 Theoretical and measured values of the penetration of solar radiation through a scots pine forest for different zenith angles of the sun. Czarnowski and Slomka (1959).

Zenith Angle	Experimental Rsi/Rsio	Theoretical Rsi/Rsio
0	38	38
5	NV	36
15	NV	33
25	27	27
35	19	18
45	9	10
55	4	4

N.V. - No Values

Augstrom (1951) notes that the radiation which is measured by instruments is seldom the radiation which is effective in the biological processes. One cannot gauge accurately the radiation which is intercepted by a natural surface from the record given by a horizontally-mounted sensor nor that of a sensor which is placed perpendicular to the solar beam.

Daytime and maximum air temperatures vary with the forest cover in the same directions as solar radiation, for the sun is the chief source of heat both for the air and for the surface layer of the soil, from which the convectional currents of heated air arise. In the same way, the minimum air temperatures at night reflect the varying intensities of outgoing radiation from vegetation and soil. Mean temperatures will obscure the influence of the forest, for the trees both reduce the maxima and increase the minima (Kittredge 1962). Geiger (1950) found that in the space above the sunny crown of a forest, the air was warmer than at the same height in the open, but that the heat energy which was gained there was lost in the trunk space. Nocturnal outgoing radiation proceeded exclusively from the upper surface of the tree crowns. In a mature forest stand the night-time temperature was close to isothermal. During the day there was marked temperature maximum in the crown space. Evergreen forest was more uniform in its temperature range than was deciduous forest. The trunk-space of the coniferous forest was at no time shielded so little as the deciduous forest before the leaves came out, nor so much as when the latter were in full leaf. Baumgartner (1956) took simultaneous measurements of temperature of the air, tree wood mass and forest soil. He found that for the tree wood mass the greatest temperature increase occurred just after sunrise, whereas for the air it was about two hours before noon and for the soil just after the noon hour period. Similarly the tree mass was the first to cool at night, followed by the air and the soil in turn. De Percin (1960) pursed a study of climatic differences between a spruce forest and a clearing in central Alaska for the months of June and December. In his profile measurements of air and soil temperature he found differences in air temperature ranging around 3^{o}F for June when comparing the open and woods stations, but the soil temperatures showed differences greater than 15^{o}F at all levels, being much lower in the woods than in the open. The change of air temperature with height in the forest is small in comparison to a corresponding open site. Kittredge (1962) published the temperature profiles for an open and forested site during a day in October at 3 p.m. at Berkely, California. These profiles which are shown in Table 8 indicate the large scale differences which can develop between two different surfaces.

Table 8 Vertical gradients of temperature over bare dry soil and under forest. Kittredge (1962)

Height (cm)	Temperature (^{o}C) Open	Forest
305	27.2	25.0
213	27.8	25.6
152	27.2	25.3
122	27.5	25.3
91	27.2	25.0
61	27.8	25.3
46	27.8	25.3
30	29.4	25.3
15	30.0	25.6
6	31.7	25.6
-0.9	35.6	25.6

The forest influences the diurnal trends of temperature. On clear days, the forest influence in reducing the maximum temperature is greatest. Its influence in increasing the night minimum temperature according to Kittredge is much smaller and sometimes negligible -- even reversed.

The decrease of wind speed within the forest increases with the density of the tree layer. Geiger (1950) found that the reduction of wind speed in a forest occured principally in the crown space. From the lower limit of the crown down to just above the ground, there prevailed a uniform gentle air movement. The greater part of the wind's kinetic energy is, therefore, consumed at the canopy roof. Kittredge (1962) states that a mixture of tolerant species with long wide crowns has more effect in reducing wind speeds than a pure stand of an intolerant narrow-high-crowned species. Hursh (1948) traced the effects of deciduous vegetation by following the annual cycle of wind speed reduction in an oak forest in Tennessee in comparison to an open site. He found that the mean forest wind speed in comparison to the mean open wind speed varied from 2 percent in the summer months to 12 percent during the winter period. The reduction to 2 percent during the summer represents an extreme value.

The influence of the forest on the humidity of the air has most often been expressed as relative humidity. The relative humidity is governed partly by the water output of the leaves of the crown space. The forest floor evaporates water to an extent dependant on the degree of development of the ground flora and the openess of the stand. The lack of air movement in the trunk-space retains the water vapour so that high humidity is the most characteristic feature of its microclimate. According to Geiger (1950) drying out occurs only from the top of the forest layer, where the higher daytime temperatures are favourable to a lower ralative humidity. Baumgartner (1956) found for a spruce forest that, above 2 M in the high sun period the relative humidity was constant upward through the forest, but below this level it increased substantially to the forest floor. Kittredge (1962) points out the disadvantages of relative humidity as an indicator of atmospheric moisture in the forest layer as compared to an open site. Since relative humidity varies inversely with temperature and the temperatures in forest and open are usually different, the relative humidities will be different even if the amount of water vapour in the atmosphere is the same. If the amount of water vapour is different, the relative humidities give no evidence to distinguish the part of the difference attributable to water vapour and the part attributable to temperature. Kittredge proves his point with data showing profiles of vapour pressure and relative humidity for open and forested sites in California. In this example, the relative humidity increased strongly with height in the open and showed no change with height in the forest, whereas the vapour pressure decreased with height in both the open and the forested sites. It is clear that the interpretation of records of relative humidity in forested sites must be pursued with caution. Hursh (1948) compared records of monthly mean saturation deficits in the forest were always lower than in the open, reaching a summer peak and wintertime minimum.

(2) Topographic Influence.

It has long been recognized that the variable topography of the earth's surface exerts a strong influence on climate. In fact as noted by Thornthwaite (1953), climate to the ancient Greeks meant simply slope, and referred to the mathematical subdivisions of the earth's surface. It is the relation of steepness and direction of exposure of various slopes to the sun's rays, which produces the strongest topographic influence on microclimate. For example, unvegetated north-facing and south-facing slopes (in future called north and south slopes) will receive quite different radiation intensities especially during clear sky periods, and these differences will vary with the position of the sun during different seasons of the year. Table 9 shows the intersection angle of the sun's direct rays with north and south slopes having 20° slope angles and with a level surface, during the summer and winter solstices, and the fall and spring equinoxes, at 45.5°N. Latitude, for the noon hour period. The figures are based on charts which are given in the Smithsonian Meteorological Tables (1958).

Table 9. Angle at which the sun's direct rays intersect various slopes during the noon hour period at 45.5°N. Latitude.

Time of Year	North Slope (20°)	Level	South Slope (20°)
December 22	1.5	21.5	41.5
March 21 and September 23	24.5	44.5	64.5
June 22	49.0	69.0	89.0

If 1 langley/min, is received on level ground for different sun angles, Table 10 gives the radiation intensity on north and south hillsides of 20° slope during periods of clear sky radiation and satisfies the equations

$$Rsi_N = Rsi_L \; (\phi_N / \phi_L) \qquad (19)$$

$$Rsi_S = Rsi_L \; (\phi_S / \phi_L) \qquad (20)$$

where Rsi_N, Rsi_S and Rsi_L are the intensities of solar radiation on north-facing, south-facing and level ground respectively; and ϕ_N, ϕ_S and ϕ_L are the angles at which the direct solar beam intersects north-facing, south-facing and level ground respectively. It is clear

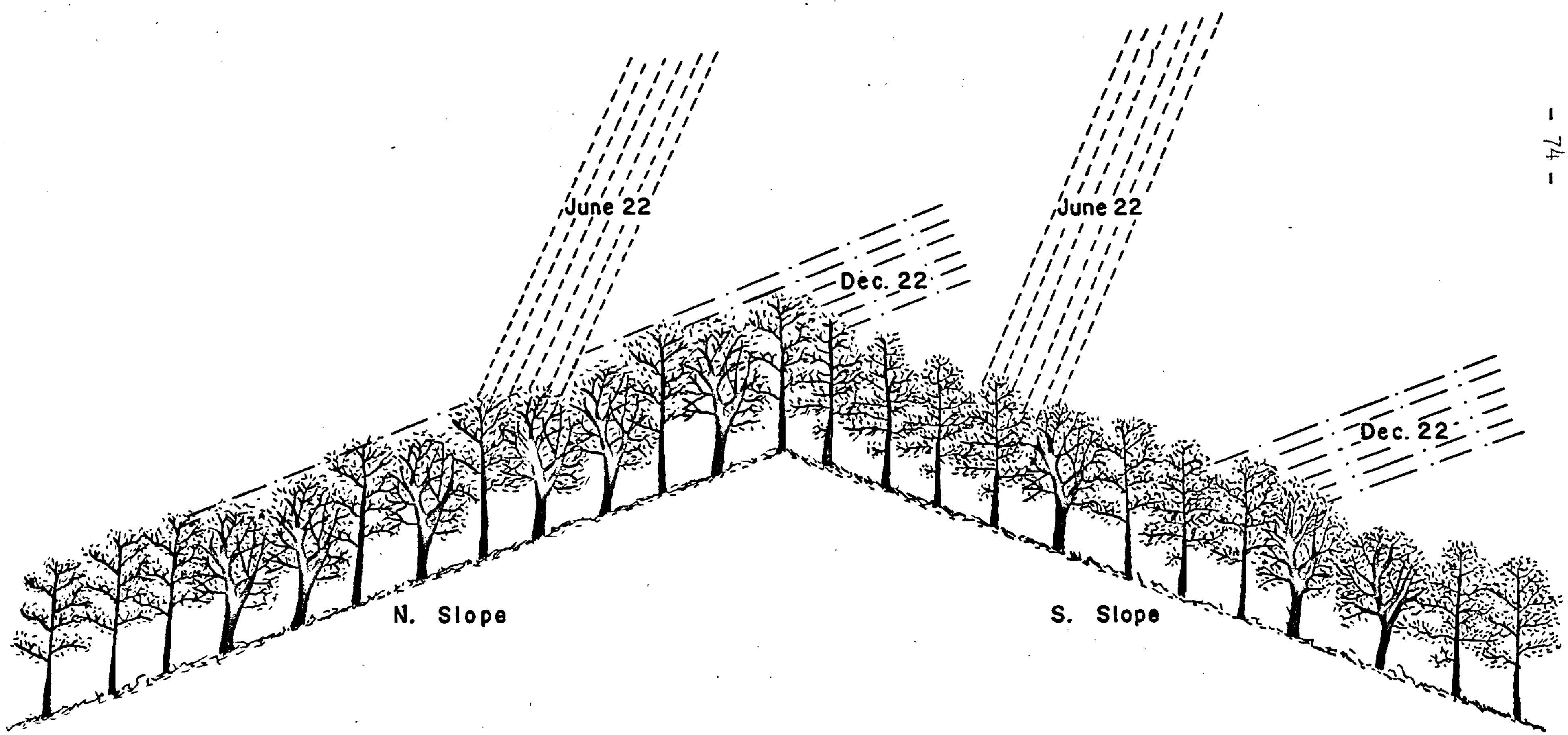
Table 10. Radiation intensity on north and south slopes when 1 ly/min. is the radiation intensity on level surfaces during various seasons of the year at 45.5°N. Latitude.

Time of Year	North Slope	Level	South Slope	Percentage Difference between north and south slopes.
Dec. 22	0.07	1.00	1.93	186
March 21 Sept. 23	0.55	1.00	1.45	90
June 22	0.71	1.00	1.29	58

from Table 10, that the differences between clear sky radiation on north and south slopes is least during summer and greatest in winter. So far the discussion has referred to unvegetated hillslopes. Figure 6 shows what might be expected to happen on forest-covered north and south slopes. The rough surface presented by the top of the forest canopy is quite different from the surface of the ground beneath. Tree leaves take up many different orientations with respect to the direct solar beam. With reference to Figure 6 it can be seen that during the summer solstice there is little liklihood that the top of the forest canopy on the south slope will present an overall surface which is more perpendicular to the direct solar beam than the top of the canopy on the north slope. Indeed, given a random orientation of tree leaves one would expect the same interception at the top of the forest for both slopes. This being the case, the top of the canopy on each slope should heat up at the same rate and outgoing black body radiation will be similar. Since the cloud cover over each slope will be the same for most periods, one would expect a similar net radiation above the forest on each slope. The situation will be quite different at ground level. For the north slope the direct radiation must penetrate a greater volume of tree vegetation before reaching the ground than for the south slope. If more solar radiation is absorbed in the canopy of the north slope one would expect greater heating there and as a result a greater flux of infra-red radiation downward to the ground. This phenomena may also increase the outgoing radiation and slightly reduce the net radiation above the forest canopy on the north slope.

At the winter solstice different conditions will prevail . Even at noon hour the sun's angle of interception with a 20^{o} north slope is only 1.5^{o} for this latitude. On a bare surface this would allow a very diffuse light to strike the ground. As is evident in Figure 6, a forest is likely to block out all of the direct radiation before it reaches ground level. In contrast, for the south slope the bare tree branches will present some vertical faces to the solar beam. The ground surface will receive less direct radiation than if it were unvegetated and some heating within the crown zone might be expected. This turn would result in a greater long-wave flux to the ground surface. With a wintertime snow cover the ability of the dark coloured tree branches and trunks to absorb solar

FIGURE 6. The influence of forest cover on the intensity of solar radiation for south and north hillsides of 20° slope. Solar radiation is shown for the noon hour period during summer and winter solstices.



radiation will lessen the surface albedo.

The discussion above has been concerned with direct solar radiation under clear sky conditions. With cloudy skies and diffuse radiation the importance of topography on the amount of insolation which is received becomes negligible, since solar radiation reaches the surface from many different directions. If the foregoing discussion is correct it means that for periods around the summer solstice a forest cover strongly diminishes microclimatic differences arising from topography during all meteorological conditions. Measurements will be presented in Chapter VI which tend to support this conclusion.

In opposition to the above arguments, there are a number of studies which point to major microclimatic differences between north and south slopes during the growing season. Cottle (1932) carried on a study over three growing seasons for vegetation on north and south slopes in Texas. He found the vegetation on north slopes to be twice as dense while the soil moisture was up to 16 percent greater than on the south slope. Potzger (1939) observed substantially more soil moisture on north slopes than on south slopes in central Indiana. In an excellent study, Cantlon (1950) investigated the vegetation and microclimatic differences between the north and south slopes of a ridge in New Jersey which was 700 ft. in elevation, 600 ft. above the surrounding lowland. Cantlon noted that there was a significant difference in the insolation-radiation balance (not measured) on the two slopes and that the daytime climatic differences were greater than at night. Temperature profiles in the lower two meters showed daytime lapse conditions developing on the south slopes during sunny weather, while isothermal profiles were characteristic of cloudy periods. In contrast, the north slope showed persistant inversion conditions. Soil temperatures at the 4 cm. depth showed an average 7°F difference in January between south and north slopes, but significantly, in July the differences vanished. Cantlon found the greatest differences in air temperature between slopes in the fall, just after leaf drop, and in the spring just before canopy closing. By measuring changes in the radial dimensions of lower story vegetation, the author found that the south slope vegetation showed a more marked response to drought. Geiger (1951) states that it is only the direct radiation which produces microclimatic differences due to the direction of exposure, while diffuse radiation from the sky affects all slope orientations almost equally. The fact that north slopes produce better timber than south slopes is according to Stoeckler and Curtis (1960) caused by a difference in solar radiation which affects the rate of evapotranspiration, and this in turn influences the supply of available moisture. The authors measured soil moisture on north and south slopes in Wisconsin during the year 1957. They found that the north slope showed a gradual increase in soil moisture from the top to the bottom of the hill, but on the south slope, there were no clear cut trends. However, on the average, the south slope showed only one-half as much soil moisture as its north-facing counterpart. The results of this difference were displayed in timber volumes which were in a ratio of 2:1 for cordwood and 2.6:1 for sawn timber in favour of the north slope. Stoekler and Curtis' drawings show a steeper grade on the south slope than on the north, which may partially explain the major differences which they observed. In studies of a ravine forest in Illinois, Fritts (1961) found substantially lower maximum temperatures on north than on south slopes. He attributed the differences in maximum temperature to differences in the energy balance. According to Fritts, on the south slope with its more open canopy, a large quantity of solar radiation enters the canopy and is ultimately absorbed by the lower sapling and herbaceous layers where it heats the air. McHattie and McCormack (1961) observed the effects of topography on microclimate at Petawawa, Ontario. They measured air and soil temperatures and Piche evaporation on the north and south slopes of both a cleared and a wooded ridge between May 1 and September 30, 1955. The following results are most noteworthy. Temperature differences due to aspect and elevation were much lower on the wooded than on the cleared ridge. The maximum temperatures on the north slope of the wooded ridge averaged 4^{o}C lower than on the north slope of the cleared ridge. This according to the authors was the result of greater shading and higher evapotranspiration from the forested site. Interestingly, for the wooded ridge McHattie and McCormack found that although soil temperatures on the north slope were substantially lower than on the south slope the maximum air temperatures differed in the reverse order a finding opposed to that of Fritts. Their interpretation of this phenomenon was that a smaller proportion of direct beam solar radiation reached the soil surface on the north slope than on the south slope. Owing to the more acute angle the solar beam makes with the ground surface on the north slope a higher percentage should be intercepted by vegetation. This serves to increase the amount of heat which goes into the air and decreases that which enters the soil. Nash (1963) proposed a method for evaluating the effects of topography on the soil water balance. He used the average solar radiation values for the growing season for various slopes and aspects to compute the receipt of solar radiation as percentage differences of the amount received on a horizontal surface. He then used these percentages to correct Thornthwaite's calculations of potential evapotranspiration. Nash computed water balances for 21 combinations of slope and aspect for wet, normal and dry precipitation years. Among his results he calculated up to a 100 percent difference in moisture deficits for north and south slopes during a dry season.

The reader is faced with the inevitable conclusion that major microclimatic differences occur between slopes of different exposures, and in particular between north and south slopes. However, such differences may not be created wholly during the growing season. Cantlon found that in the high sun period the microclimatic differences between wooded slopes were small. This was particular true of soil temperatures, a finding which was not supported by the investigations of McHattie and McCormack. Several investigators point to much drier conditions on south than on north slopes. This may in part be due to a moisture deficit which begins in the early spring. It is common experience that winter snow cover lasts longer on north than on south slopes (up to 2 or 3 weeks). This certainly will exert an influence on the available water in ground storage on the different slopes. This writer suggests that the microclimatic differences between forested north and south slopes are least during the growing season, and are most significant in fall, winter and spring. Measurements in support of this hypothesis will be presented in Chapter VI.

(D) Summary

The background material which has been discussed in this chapter promotes the following general observations into the nature of the forest microclimate.

Successful measurements of the major elements of the surface energy-balance of a forest are possible. Measurements of net radiation and soil heat flow are most easily obtained. Calculations of the turbulent fluxes of latent and sensible heat have relied mainly on aerodynamic equations, the use of the Bowen Ratio within the energy-balance framework, or the estimation of the convective heat flux as a residual term after net radiation, soil heat flow, and evapotranspiration have been measured. According to measurements by Baumgartner and Dzerdzeevskii the soil heat flow within a forest accounts for less than 2.5 percent of the available energy of net radiation over a period of days or weeks during the growing season.

Grass-covered surfaces can evapotranspire at a rate which consumes more than 80 percent of the energy of net radiation when the soils are naturally moist. Under dry conditions, the latent heat flux from grasslands may incorporate 14 percent or less of the available energy from net radiation. Forests presumable are as efficient in returning moisture to the atmosphere as are grasslands. Under conditions of moist soil, those forests with a lower albedo than grasslands may yield more moisture into the atmosphere because their net radiation is greater. More important, tree roots are able to tap a large volume of moist soil. It is not known whether forest evapotranspiration can diminish to the extent that only 14 percent or less of net radiational energy is converted into latent heat.

When irrigation is employed during dry periods, the vapour flux from an irrigated plot may consume more energy than that which is supplied by net radiation. The degree to which the latent heat flow exceeds net radiational energy depends on the dryness of the surrounding terrain. The extra heat energy which stimulates the increased evapotranspiration may be provided by a negative convective heat flux, a negative soil heat flow, or an import of advective sensible heat to the irrigated area.

Dzerdzeevskii's measurements under normal moisture conditions for a mixed forest in middle latitudes showed that the rate of evapotranspiration decreased from spring and mid-summer periods through to the end of the growing season. Average values for two surface energy-balance calculations over forest during the mid-summer period, one for a spruce plantation and the other for a natural mixed forest, showed that the latent heat energy which was involved in evapotranspiration was equivalent to 70 percent of the energy available from net radiation.

The vertical variations of transpiration within a forest canopy are not proportional to the vertical variations in net radiation. Rather, the vertical movement of sensible heat may increase transpiration at one level and decrease it at another so that the vapour flux within the canopy is more evenly distributed through the whole leafy layer.

Hourly values of heat movement in the forest wood mass and air layer constitute important components of the total heat balance. For periods longer than a day these heat movements are insignificant in comparison to other energy movements.

Forest hydrology is conveniently discussed within the hydrologic balance concept. Forest vegetation exerts a strong influence on the disposition and effects of rain water in the active surface layer as a result of interception by the canopy and leaf litter layer. The special nature of forest soil has a profound influence on the infiltration, storage and runoff of rain water. Evapotranspiration within the forest layer is a process of many parts. For its study the investigator must consider climate, soil and vegetation and the complex interactions of all three. With an abundant water supply the evapotranspiration from forests probably does not differ substantially from that of other vegetation-covered surfaces. When the water supply is a limiting factor, however, there is evidence which points to a greater evapotranspiration from forests than from other surfaces.

The crown layer is most important in modifying the forest climate. Measurements of trunk-space climate do not give a true indication of the forest microclimate. The importance of the forest canopy depends on its interception of sunlight. The degree of such interception depends on the nature of the canopy, the angle at which the direct solar beam intersects the earth's surface, and the cloud cover which influences the diffuseness of solar radiation.

Beneath the forest canopy large horizontal differences in radiation intensity are common. Under a dense canopy the long-wave radiation exchange is more important than the short-wave radiation in influencing net radiation. In comparison to an open site, the forest environment develops lower maximum and higher minimum air temperatures. Differences in soil temperature between forest and open are greater than those of air temperature. The tree density and the shape of the forest canopy determines its influence on wind speed. Wind speeds within a forest can be reduced to only 2 percent of their open site values. Relative humidity, because it varies with temperature, gives an unreliable indication of humidity conditions in a forest.

The topographic influence on microclimate occurs as a result of the variable angles of intersection of the ground surface and the direct solar beam. A forest cover may act in such a manner that it decreases variations in the insolation received by different slopes in the summer. A considerable body of evidence points to major differences which can develop between north and south slopes. Such differences increase as one moves downward within the forest layer. The microclimatic differences between north and south slopes must be interpreted in light of the year round climatic regime. The large moisture storage differences which develop between the slopes are not necessarily wholly created during the growing season but may have their roots in significant variations in moisture storage prior to, or in the early stages of the growing season.

Chapter III. Instrumentation, Calibration, and Data Evaluation.

A Instruments and their Location

(a) Solar Radiation

Solar radiation was measured with two kinds of instruments, one being used at a permanent control site, and the other as a mobile field instrument.

The permanently-mounted sensor was located at the laboratory as shown in Figure 7. A Kipp solarimeter was used mounted atop a section of television aerial which in turn was affixed to the top of a tall pine tree (Illustration 1). This gave a horizon free from any obstruction from north-east through south to north-west, and as a result, the solarimeter was fully-exposed to the sun during all seasons and all hours of the day. The solarimeter measures total radiation of the sun and sky between the wave lengths from 0.3 to 2.0 microns. The 14-element constantan-manganin Moll thermopile, mounted under two concentric hemispheric glass domes, is built in a holder on a metal base provided with levelling screws. The rectangular sensitive surface measures 12 by 11 mm., and is place exactly in the centre of the domes. A white lacquered screen prevents the base from being heated by radiation and a small tube, connected to a drying bottle from the bottom, is used to prevent condensation within the solarimeter.

Solar radiation was recorded on a Leeds and Northup Speedomax, H, continuous-recording potentiometer. Scale and chart calibration is in langleys per minute thus facilitating rapid evaluation of the record.

The shielded cable from solarimeter to recorder measured 200 feet in length thus creating no electrical resistance problems. An almost continuous record was gathered between May 1, 1964 and May 1, 1965.

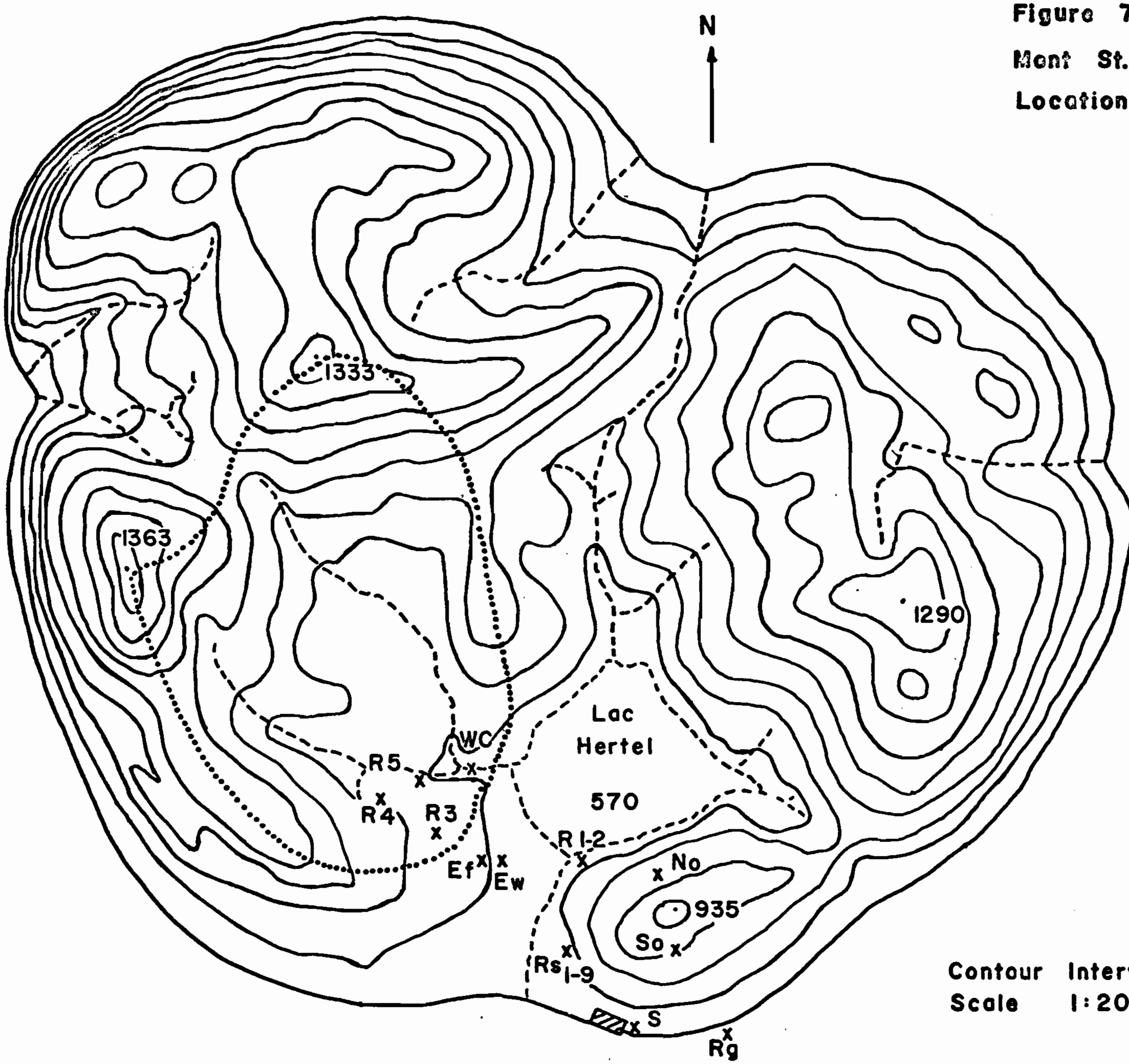
Illustration 1

Permanent site of solarimeter

Figure 7

Mont St. Hilaire

Location of Instruments



Legend

- laboratory
- x S mounted solarimeter
- x No north slope site
- x So south slope site
- x R1-5 raingauges
- x Rs1-9 raingauges stemflow site
- x Rg raingauge gravel pit
- x WC west creek stream gauge
- x Ef evapotranspirometer tank (field)
- x Ew evapotranspirometer tank (woods)
- area of west creek watershed
- streams

Contour Interval 100 ft.

Scale 1:2000

Illustration 2. Bimetallic Actinograph

A self-contained bimetallic actinograph , manufactured by the Belfort Instrument Co., (Illustration 2) served as mobile field instrument for recording solar radiation. The measuring element of the instrument consists of two identical bimetallic strips, one of which is blackened, while the other is highly polished and shielded. The blackened strip is exposed to both the ambient temperature and the radiant energy of the sun, while the shielded strip responds to ambient temperature but not to solar radiation. The two strips connect to a pen arm and the pen responds to the different temperature of the black and white plates. Recording is on a circular chart which rotates once every 24 hours or once a week if the latter speed is desired, being powered by a clock mechanism. The manufacturer claims an accuracy of plus or minus 5 percent, and a recording precision of 0.1 langley per minute. The main drawback of the instrument is its response lag of 2 minutes. This compares unfavourably with the very rapid response of the Kipp and Zonen solarimeter (a few seconds only) and the 5 second interval for full-scale deflection on the recorder. However, for values which are averaged over the hour, the large lag does not introduce much error.

The actinograph was used in many locations, usually as an accompaniment to the net radiometers. Its record allowed ratios of direct radiation penetrating through the canopy, to that which impinges above the forest crown to be determined. The actinograph was moved to different heights in the forest when radiation profiles were being determined. In order to obtain mean values representative of the forest as a whole, the actinograph was moved to a new location each day, and at each height level. Its height was altered by means of a platform which could be joined to the steps of an extension ladder. The ladder in turn was leaned against various trees. Beyond the reach of the ladder, it became necessary to mount the platform and actinograph on tree branches within the forest crown.

(b) Net Radiation

Four net radiometers were employed to measure net radiation in the field. Each consists of a thermopile transducer mounted between two half-hemispheric polythylene covers, and is given the name of Minature Net Radiometer by the manufacturers, C.W. Thornthwaite, Associates. The thermopile unit is made by winding constantan wire about a rectangular glass slide 1/16 inches in thickness. One-half of each winding is protected by wax while the other is copper-plated in an electrolytic copper sulphate solution. This gives an equal number of thermojunctions on top and bottom surfaces, and a unit which effectively measures temperature difference between each surface. After being embedded in an epoxy resin disk, thin aluminium plates are fastened to each side to allow even heat flow over the whole transducer surface. The ployethylene hemispheres are pumped to a pressure above that of the surrounding atmosphere and sealed. They serve the purpose of shedding rain water and preventing ambient temperature influences. They can be purged if moisture accumulates on the inside. The net radiometers are similar in design to the one described by Fritschen (1963) except that the latter design uses polystrene material rather than polyethylene for the hemispheres. The transparency of the polyethylene to various wavelengths in the solar and terrestrial radiation spectra is shown in Figure 8, according to tests carried out by the manufacturer.

The output of the net radiometers is in the region of 300 microvolts per langley, and response speed is rapid. The signals were measured on continuously-recording microvolt recorders which employ a galvanometer feed-back principle. Twelve volt storage batteries provided the power source for the recorders, thus allowing much needed portability for widespread field measurements. Since the recorders have a single channel, and only two recorders were available, stepping switches were used, whereby more than one signal could be fed into each recorder. Continuous stepping of the switch was achieved by using a D.C. motor which was also powered by a storage battery.

Field mounting was achieved in various ways, When measuring radiation above the forest canopy for a considerable period of time in one location, a section of television tower was attached in the tree crown so that it reached above the general forest level. From this a wooden T.-bar was raised with a horizontal aluminium extension carrying the net radiometer on its end. Illustration 3 shows a net radiometer positioned above the forest canopy. Such a device allowed the net radiometer to be rotated through a 360 degree circle with a 15 foot radius, and in this manner to sample a relatively large area of undisturbed forest crown. The major problem involved the levelling of the net radiometer. It was necessary to be very certain all components of the mast rigging were perfectly vertical or horizontal according to their function. The aluminium extension was drawn in, and while in a controlled horizontal position, the net

PERCENT TRANSMISSION OF POLYETHYLENE (4.0 MIL)

AS A FUNCTION OF WAVELENGTH

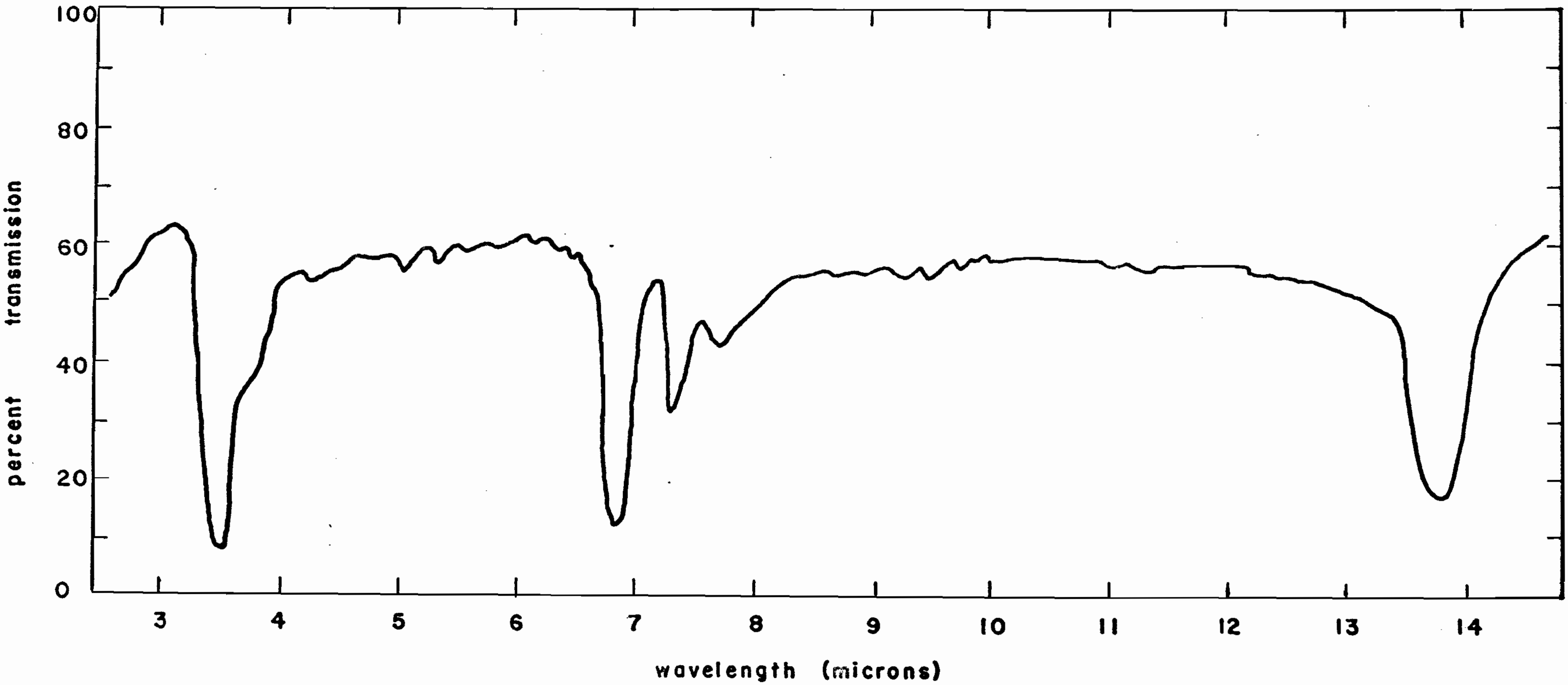


FIGURE 8

radiometer was levelled with it. With a strong wind the mounting unit swayed somewhat with the tree, but observation of the continuous record during such periods seemed to indicate that no serious measuring errors were produced.

Illustration 3. Above-canopy net radiometer.

Illustration 4. Giraffe Mast.

A second method of mounting was from a giraffe-like adjustable mast as shown in Illustration 4. Three of these masts were employed, and each allowed a sampling of all heights between the ground and the 5 M. level. The masts were readily rotated through a full circle. For heights above 5 M. the masts were strapped to the trunks of trees and from any fixed position could sample a full 5 M height interval. Levelling presented few problems as the field worker could usually climb above the net radiometer in order to level it. Only when the masts were used above the tree crowns was instrumental levelling impractical. In such cases sight-levelling was necessary. Any serious error in the sight-level became quickly evident by comparison to the net radiometer which was mounted on the T-mast, as described in the previous paragraph.

It is felt that the advantages of mounting radiation sensors in this way rather outweight the disadvantages. By mounting three net radiometers and comparing the results to a semi-permanent radiometer fixed above the forest crown, a sampling over area as well as with time is achieved. There is little disturbance to the natural forest as compared to that created by a rigid scaffolding. The main disadvantages are the time it takes to change position of the net radiometers, the aforementioned problem of levelling, and the danger of damaging the instruments while climbing the trees.

(c) Soil Heat Flow

Thermopile discs very similar to those used in the net radiometers were employed to measure vertical heat movement within the top centimeter of soil. Usually three thermopiles, each giving an electrical output in the range of 900 microvolts per langley, were connected in series, and placed at different spots on the forest floor. Each unit was buried 0.5 cm. beneath the soil, and covered with leaf litter, to reassume as nearly as possible the undisturbed condition of the forest floor. At least two weeks were allowed for equilibrium with the surrounding soil to be reached, before measurements were taken. The continuous recording of soil heat flow was at sporadic intervals when recording equipment was free from other uses.

(d) Rainfall

The positions of all the rain gauges which were used during the field season are shown in Figure 7. At the shore of Lac Hertel, rain was measured in both a standard raingauge (R1) of the type employed by the Canadian Meteorological Service, and in a tipping-bucket recording rain gauge (R2). At all other stations plastic wedge-shaped gauges of the type shown in Illustration 6 were used. Rain beneath the forest canopy in the area of stemflow measurement was measured in 9 gauges (Rs 1-9) which were spaced randomly. The positions of each of these gauges were changed following each rain which was greater than 0.25 inches in amount. The value of rainfall reaching the forest floor in the stemflow area was compared with a rain gauge in a nearby gravel pit (Rg).

(e) Stemflow

The flow of rain water down the trunks of four trees, 2 beech and 2 maple, was caught in large plastic laundry bags, after being intercepted near the base of the trunk. Interception was accomplished by using a split hose of one inch diameter. The hose was tacked in a spiral around the trunk base, so that the whole circumference was effectively blocked,

Illustration 5. Rain gauge mounted above the forest.

Illustration 6. Stemflow site.

Illustration 7. Measuring stemflow on beech tree.

(Illustration 7). Plastic wood filler was used to seal the hose tightly to the tree bark, and the unsplit end of the hose led into the measuring bags. This device for measuring stem flow worked exceedingly well. From visual observation, the catch of water flowing down the stem was 100 per-cent, while the catch of free-falling rain drops was negligible. The plastic bags would overflow after 0.50 inches of rain, so that during a heavy storm it became necessary to empty them while the rain still fell. On two nightime occasions, this was not done, and the record was spoiled. Total canopy catchment in the stem flow measuring area as well as the in-dividual area covered by each tree crown is shown in Figure 9. Illustration 6 pictures the stemflow measuring devices on a maple tree in the foreground, and a beech in the background while Illustration 7 gives a closer view of the stemflow intercepter on a beech.

(f) <u>Stream Flow</u>

Mont St. Hilaire has the advantage of possessing a completely enclosed drainage system whereby the complete circle of land surrounding Lac Hertel drains into the lake. Moreover, except during the spring runoff season, all drainage is handled by the north and west creeks as shown in Figure 7. The north creek drains the smallest portion of land, and its runoff is most difficult to measure. The stream divides shortly above its outlet and below this the water flows through an area of swamp. Thus gauging becomes impractical and although measurements were attempted during the summer of 1963 they were suspended the following year. The west creek, however, proved more amenable to runoff determination.

During the summer of 1963, measurements on the west creek were taken from a wooden weir with a rectangular notch, 12 inches deep and 18 inches

Illustration 8. View from downstream side of weir and stream gauge.

ABOVE-CANOPY VIEW OF STEMFLOW SITE

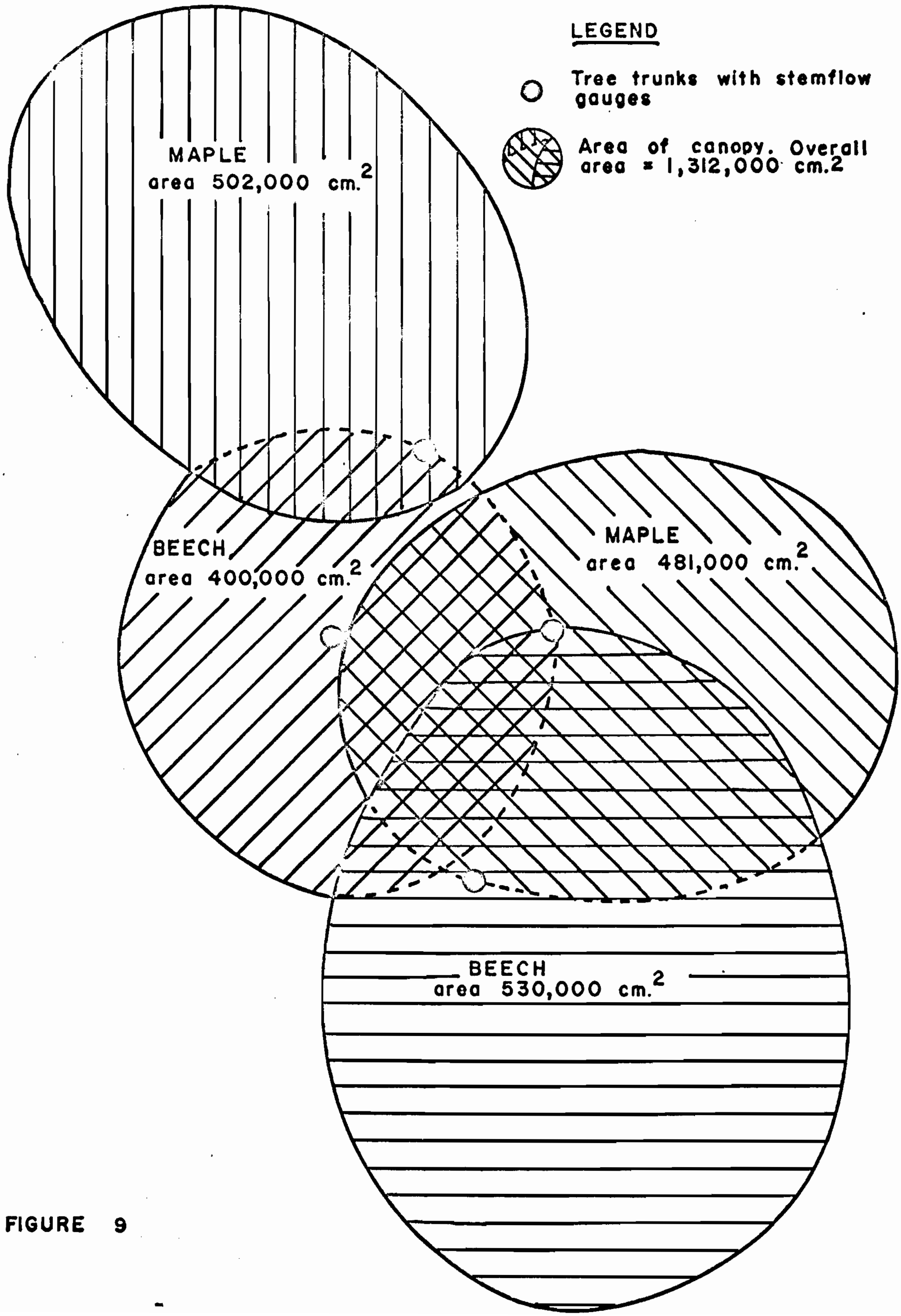


FIGURE 9

wide. These measurements are discussed by Sackeyfio (1964) who took twice daily readings during rainless periods, and more frequent measures during periods of rain. The weir was located about half-way along the 500 feet between the stream mouth and the first stream divide. Being near a public routeway, the weir was exposed to frequent damage, and its large-sized rectangular notch did not accurately record small-scale fluctuations in water level.

For the field season of 1964 the measuring site was moved 100 feet upstream to a place where the banks were well fortified by tree roots. A concrete dam 3 feet deep, 6 inches wide and 15 feet long was placed normal to the stream flow, An iron plate with a 90^{o} V-notch was bolted across the 18 by 12 inch notch in the dam. The plate was designed with bevelled edges facing the upstream side. Continuous recording was achieved in the following manner. At the place of greatest water depth behind the dam, a stilling well was placed. A float within the stilling well was attached to a thin right-angled copper rod which attached to the pen arm of a modified continuous-recording hygrometer. The length of the copper connecting rod was such that a 10 inch vertical movement of the float gave a 3 inch full-scale deflection on the hygrometer chart drum. The chart gave a full week's runoff record between changings. Figure 7 delineates the area of the west creek watershed and shows the position of the weir. Figure 10 (a) shows diagramatically the operating principles of the continuous-recording stream gauge. Illustration 8, gives a view of the entire ensemble from the downstream side.

(g) Potential Evapotranspiration

During the spring of 1963 two stationary evapotranspirameter tanks were installed on Mont St. Hilaire. One was located in the centre of a rectangular field of dimension 150 by 500 feet, which was surrounded by forest on three sides and an orchard on the fourth. The second tank was placed within the nearby wood about 50 feet from the forest edge. The evapotranspirometer tanks were of a simple design, similar to the types employed by Mather and Thornthwaite (1958) Sanderson (1950), and Nebiker (1957). Two oil drums, each measuring 4 feet in depth and 2.5 feet in diameter, served as the tanks. From the base a polyethylene tube led to an overflow tank located 5 feet away. The overflow tank consisted of an ordinary hot water heating tank from which the top had been cut. A 20 percent gradient was maintained on the polyethylene hose connecting the two tanks. Each of the oil drums was sunk to within an inch of its rim. In the case of the field tank, bed rock was reached at about 3.7 feet so that the turf around the tank had to be built up to within an inch of the rim. The soil layers were replaced within each tank in such a manner as to simulate as closely as possible the natural soil profile. Turf was replaced on the field tank but the woods tank was left bare except for a sprinkling of leaf litter. Figure 7 shows the location of the two evapotranspirometer tanks, Figure 10 (b) details the style of construction, and Illustrations 9 and 10 show the surface appearance of field and wood tanks respectively.

During 1963 an attempt was made to measure actual evapotranspiration from the tanks by just leaving them open to natural rainfall.

CROSS SECTION OF STREAM GAUGE INSTALLATION

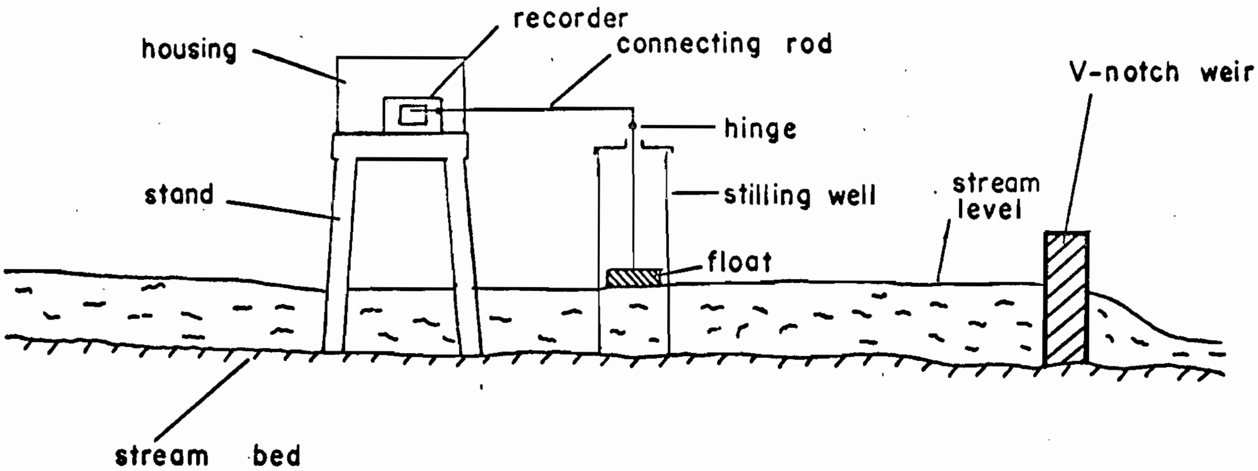


FIGURE 10(a)

CROSS SECTION OF EVAPOTRANSPIROMETER TANK

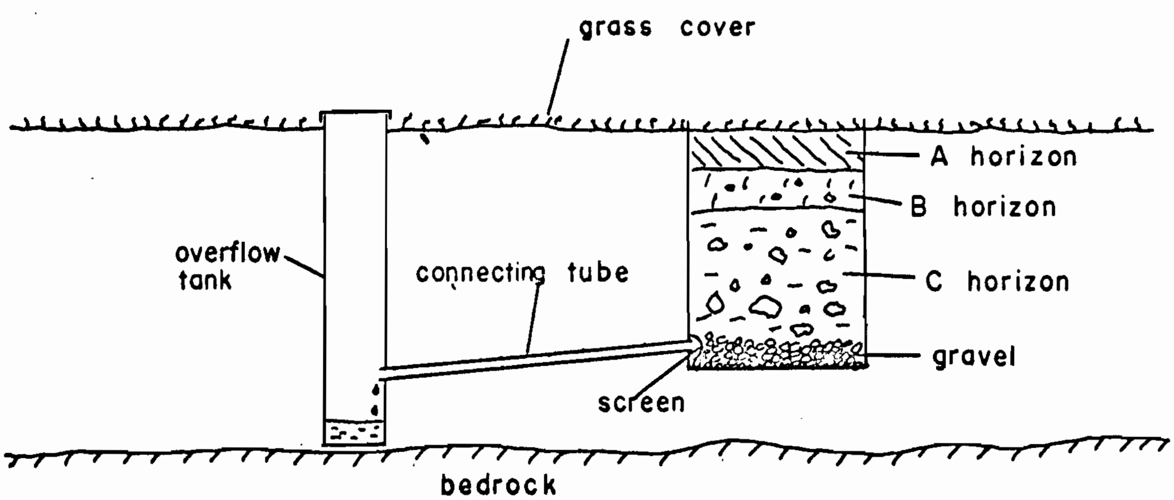


FIGURE 10 (b)

This proved quite unsuccessful because throughout the summer season there was no runoff from the tanks thus making it quite impossible to say when and how much actual evapotranspiration occured. For the 1964 field season, potential evapotranspiration was measured. This was accomplished by irrigating each tank with 0.50 inches of water for each day when rainfall was less than 0.25 inches. It was not possible to irrigate a large area surrounding the tanks in order to subdue the oasis affect, and as will be shown later, although mechanically the tanks functioned well, the values for potential evapotranspiration which they gave are misleading.

Illustration 9. Evapotranspirometer tank in field

Illustration 10. Evapotranspirometer tank at forest floor.

B. Calibration

Along with the problem of keeping continuously recording equipment in repair and operation, the necessity for careful and accurate calibration looms large in microclimatic research. Such need, particularly with the net radiometers, was very apparent during the study period.

(a) Solar Radiation

Since comparisons were to be made between values given by the Kipp Solarimeter and the bimetallic actinograph, it was important that their responses should be close to one another. For a period of three relatively clear days, the two sensing units were exposed side by side, and the calibration of the solarimeter was carefully adjusted, so that its curve corresponded as closely as possible to that given by the actinograph. The calibration of the solarimeter was then changed to that given by the manufacturer. Fortunately, this created only a 1.5 percent difference between the responses of the instruments which is well within the range of error in reading the charts. Moreover, by virtue of having such close calibrations, periodic checks of the instruments would show a calibration drift.

Latimer (1962) points to several features in the response characteristics of the Kipp Solarimeter which may lead to measurement errors. Non-linearity of response may range from -2.1 to -5.4 percent (i.e. the increase in electrical response from the thermopile is less than the increase in solar radiation). In addition, Latimer found that the dependance of response on temperature ranged from -.0.04 to -0.02

percent/degree F., depending on the individual instrument. To test the inaccuracies in the Kipp, a post-season calibration check was pursued by comparing simultaneous records from the Kipp and a recently-calibrated Eppley when they were mounted side by side at College Jean-de-Brebeuf in Montreal. Table 11 gives the records of 14 days during the winter of 1965 along with the daily mean temperatures and the percent differences between the two instruments.

Table 11. Comparison of solar radiation as recorded by the Kipp Solarimeter and Eppley Pyrheliometer.

Date	Solar Radiation (ly/day)		Daily Mean	Percent
	Eppley	Kipp	Temperature (^{o}F)	Difference
Jan 29	253	253	-3	Nil
30	250	247	13	-1.0
31	236	230	15	-2.7
Feb. 2	228	230	15	0.9
3	186	181	10	-2.9
4	164	154	9	-6.3
5	131	122	8	-6.9
6	144	135	16	-6.3
7	100	91	35	-8.6
9	289	271	30	-6.1
11	294	289	38	-1.9
13	309	292	35	-5.5
14	318	311	9	-2.0
15	239	229	27	-4.2
	3,141	3,035	Average	-3.4

The Kipp recorded an average of 3.4 percent less solar radiation than did the Eppley. It is evident from Table 11 that the differences between Eppley and Kipp are non -systematic. They can be correlated effectively neither with temperature change nor with non-linearity of response. A linear correlation of percent difference in radiation on daily mean temperature yields a correlation coefficient of only 0.06, whereas a similar correlation between radiation intensity and percent difference gives a correlation coefficient equally low at 0.10. It appears from this test that the differences between the Eppley and Kipp are on the average within ± 5 percent, and that no suitable correction can be applied for non-linearity of response or temperature factors.

(b) Net Radiation

The net radiometers proved rather tempermental in their response characteristics. Original calibration of the radiometers by the manufacturer is quite thorough. By being placed in a black box with a voltage-controlled light source, they are checked to assure a response from each side of the thermopile within 3 percent of that given by the other. Each radiometer is then mounted alongside an Eppley Pyrheliometer in a position normal to the incident rays of the sun. By shading out the direct beam, the depression of the net radiometer output is compared to that of the Eppley, and a calibration thereby achieved. The process is repeated for the opposite side of the thermopile to assure that response is still within 3 percent.

During the autumn of 1963 the author accepted the manufacturer's calibration for radiometers No. 192 and 196, which were all he had in use at the time. Calibrations were, however, checked once prior to the 1964 field season, and three times thereafter. Calibration was carried out on a clear day with the solarimeter and net radiometers mounted horizontally side by side over a uniform black tar rooftop. The direct solar beam was then blocked out, and the depression of the radiometers compared to that of the solarimeter. The experiment was repeated on the reverse side. The radiometers were then run continuously side by side to determine if their responses to all radiation, both incoming short-wave and terrestrial long-wave, were the same. The results are rather interesting, and are detailed in Table 12. For the calibration carried out prior to the 1964 field season, the results for all instruments were

Table 12 Net Radiometer Calibration Results

All values in microvolts/langley
C is the calibration
Δ C is the change from the previous calibration.

Calibration	Net Radiometer No. 192		No. 196		No. 200		No. 219	
	C	ΔC	C	ΔC	C	ΔC	C	ΔC
Manufacturer	300	---	222	---	225	---	329	---
April 30	300	0	232	+10	223	-2	329	0
June 2	436	+136	236	+4	225	+2	201	-128
July 17	442	+6	240	+4	238	+13	208	+7
August 15	438	-4	238	-2	236	-2	212	+4
Sept. 12	438	0	235	-3	238	+2	215	+3

very close to those given by the manufacturer. Moreover, once a calibration had been determined, by shading out the direct beam from both the solarimeter and the radiometer, that calibration held true for all wavelengths including the long-wave terrestrial radiation at night. Most importantly, the measurements from each side of the thermopile of the net radiometer remained within 5 percent over the whole measuring period.

However, the radiometers changed their calibrations during the summer of 1964, and some of these changes were very great. The most noteable drift occured with No. 192, which between April 30 and June 2, increased its output from 300 to 436 microvolts/langley. This was not a gradual change, but rather occured very suddenly. By checking through the records it became apparent that on the night of May 17, No. 192 drastically changed calibration. During the same period, No.219 fell from 329 to 201 microvolts/langley. The time of change could not be pinpointed in this case.

Neither the manufacturer nor the author can suggest reasons why such a sudden calibration change should have occured. It is quite noteworthy when one considers that the output from each side of radiometers Nos. 192 and 219 remained with 5 percent, and that subsequent calibrations on July 17, August 15 and September 12 upheld the new calibration figures with little variation.

(c) Soil Heat Flow:

Figure 11 shows the metal tank apparatus which was used to calibrate the soil heat flow discs. The discs are fixed in a vertical position on Plate B, which is fitted directly into the centre of Tank A. Tank A is then filled with gelatin which has a thermal conductivity the same as that of water, and this is allowed to solidify. Tank D is filled with an ice bath solution and C with warm water, both mixtures of which are kept constantly stirred. The temperatures T1 and T2 at each side of Tank A are measured as a temperature difference T = T1 - T2 by thermocouples taped to each wall. At the same time the electrical output

TANK FOR CALIBRATING SOIL HEAT FLOW PLATES

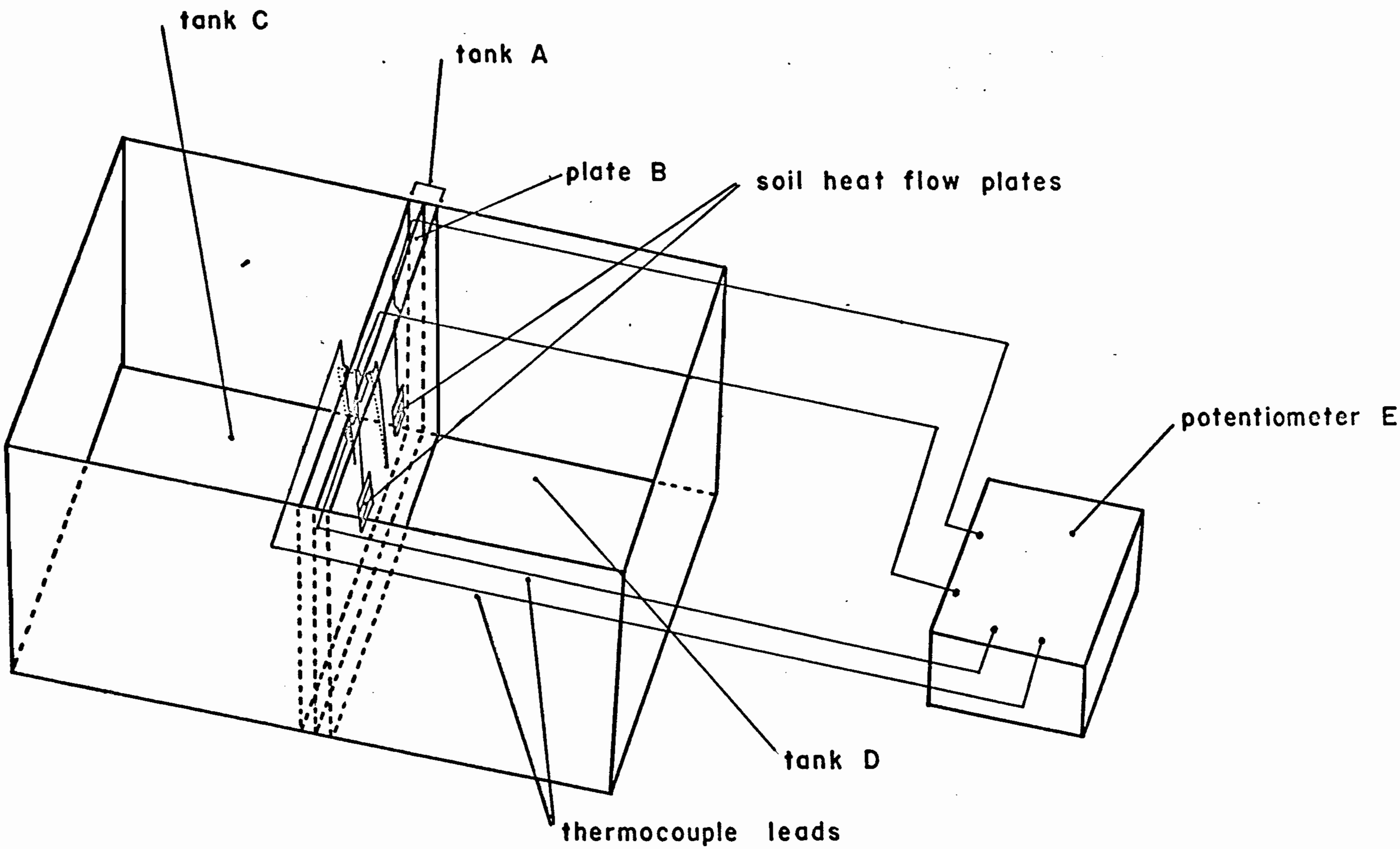


FIGURE 11

from each heat-flow disc is recorded on potentiometer E. After calibrating the disc for several temperature ranges, the ice bath is changed from Tank D to C, Tank D being filled with warm water. Calibration is then carried out for a heat flow in the opposite direction. Only the discs showing a sensitivity on each side within 5 percent were kept for use in the field. The output of soil heat flow discs averaged around 900 microvolts/langley.

(d) Stem Flow

The determination of depth equivalent of stem flow values is readily achieved. The area of the total canopy cover of the four trees used in the experiment was estimated visually, and merely divided into the total volume of flow from the trees after each rainstorm.

(e) Stream Flow

Volume of stream flow from the west creek watershed was calculated from an empirical formula for a 90 degree V-notch which reads as follows:

$$Q = 0.305 \quad H^{5/2} \tag{21}$$

where Q is the volume of stream flow in cu ft/minute and H is the height, in inches, of the water head above the vertex of the notch.

For daily periods equation (1) becomes:

$$Q = 439\ H^{5/2} \quad \text{cu ft./day} \tag{22}$$

By dividing runoff volume by the total area of the watershed the depth equivalent of runoff is obtained. The development of equation (21) is given by Gibson (1957). The conditions specified for this equation are that the lip of the notch should be 1/16 inch thick, the head should be measured at a point upstream, equal to at least four times the head, the base of the notch should not be less than 12 inches above the bottom of the channel, and the width of the channel shall be not less than 4 times the head. All these conditions were met by the installation at St. Hilaire.

(f) Evapotranspiration

Two calibrations of the evapotranspirometer tanks were carried out; the first prior to installation in 1963, and the second at the end of the 1964 field season. The first consisted simply of pouring a known volume and depth of water into the tank to be filled by soil, and determining the depth given by the same quantity in the overflow tank. The actual depths could readily be computed, but some variation in symmetry, especially in the overflow tank, made the direct approach more accurate. Once the equivalent depths were calculated, a dip stick was marked off, which when inserted into the overflow tank gave the equivalent depth of water in the soil directly. The second check was made to determine if any leaks had occured during field use. The two soil tanks were thoroughly saturated with water and sealed with a polyethylene cover. After a period of several days, when all of the gravitational water had drained off, a known amount of water was added to each tank, and it was again sealed. For each evapotranspirometer, the runoff into the overflow tank equalled the water amount added, showing that no leaks had occured.

C. Data Evaluation

(a) Solar Radiation, Net Radiation and Soil Heat Flow

The time unit for all radiation and heat flow measurements was the hour. The continuous strip charts were analysed by drawing a straight line which averaged the trace for the hourly interval. Such a line is drawn by balancing the area between it and the segment of curve above it, and the area between the line and curve segment beneath. Daily and weekly values of radiation and heat flow were obtained by summing all the average hourly values.

(b) Rainfall, Stemflow, Stream flow, and Evapotranspiration.

Values for stemflow and rainfall in gauge (Rg) in the gravel pit were determined after each major rainfall. All other rain gauges, were read once per 24 hour period between 0900 hours E.S.T.

Stream flow was evaluated both for the individual storm and for each daily period between 0900 hours. For the latter, the rapid peaking of the curve during a storm became part of the 24 hour average.

Evapotranspiration was also determined for each 24 hour period. For stationary tanks this is too short an interval to make the value for each measuring period meaninful, and negative evapotranspiration would often be determined for a daily period, due to a lag in the runoff water reaching the overflow tanks. It is only when the daily values are added into weekly values that the data become meaningful.

CHAPTER IV Water-Balance During The Active Growth Period

(a) Introduction

Measurements of above-canopy rainfall, rainfall interception and penetration in the canopy, runoff, soil moisture change, and evapotranspiration are all necessary in order to detail the complete water-balance in a forest-covered area. All of these measurements were taken for the period from May 1 to October 1, 1964, with the exception of soil moisture change.

Weekly intervals were chosen as the optimum period for the presentation and evaluation of results. Although daily readings were determined for all the moisture parameters, a day represents too short an interval for the measurement of quantities such as runoff and evapotranspiration. Water balance data for the 22 weekly periods are presented in Table 13.

(b) Active Growth Period

The season from May 1 through to October 1 includes nearly all of the interval during which active growth is occuring. Before May 1 the leaves of the deciduous trees have barely begun to show, while by October 1 most of the deciduous tree leaves have changed colour, denoting the cessation of active transpiration.

For the presentation of results, this time interval is most convenient. It can be said with some certainty that the forest soils were at field capacity on May 1. Evidence comes from the fact that winter snows had only completely disappeared from the ground some 3 weeks earlier, Lac Hertel was still emptying through its high-water runoff channel, due to its great intake of water from the surrounding land areas, and the fact that when water was added to the evapotranspirometer tanks (which had been exposed throughout the winter, as any other piece of soil) and evapotranspiration was prevented, the yield in runoff was equal to the water added.

Evidence is equally strong that in the year 1964, the soil moisture supply had been depleted to very near its hygroscopic level by the middle of September. Stream runoff from the North Creek had ceased, and it was merely an unmeasurable trickle in the West Creek. Moreover, a total rainfall of 2.25 cm during Week 19 (Sept. 4 - 10) did not serve to increase stream runoff; in fact, the runoff decreased slightly from the previous week, showing that the soil was absorbing all the rainwater, even during heavy showers. The level of water in Lac Hertel had dropped almost 1 M from its mid-summer level and reached its lowest level in 5 years of measurement. Finally, many of the wells of residents living on the outer hill slopes and in the surrounding lowland went dry during the month of September.

(c) Above-Canopy Rainfall

Rainfall results, as presented in Table 13, represent an average of the two gauges in the orchard and the one above the forest, which were described in Chapter III. Total rainfall for the growing season amounted to 33.1 cm. It was unequally distributed in time, with the very low value of 1.8 cm for the four week period from May 29 to June 25, and an equally high value of 10.8 cm between August 7 and September 3. The total rainfall of 33.1 cm compares to a mean of 46.5 for the previous 4 years of record, so that 1964 presented a considerably drier period of active growth than is usual.

Table 13. Water movements and the water-balance.
(all values in cm.)

Week	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	Totals
Above-canopy rainfall	0.00	1.34	2.08	1.98	0.74	0.46	0.48	0.08	1.34	5.56	1.70	0.23	2.80	0.10	4.90	1.35	4.14	0.36	2.25	0.05	0.00	1.14	33.08
Rainfall interception	0.00	0.54	0.90	0.46	0.13	0.18	0.19	0.06	0.66	1.18	0.62	0.08	0.68	0.07	0.42	0.55	1.20	0.21	0.89	0.05	0.00	0.36	9.43
Stemflow	0.00	0.00	0.01	0.01	0.01	0.00	0.00	0.00	0.01	0.12	0.01	0.00	0.05	0.00	0.05	0.00	0.03	0.00	0.00	0.00	0.00	0.01	0.31
Runoff	0.38	0.37	0.40	0.41	0.38	0.30	0.28	0.20	0.16	0.38	0.14	0.11	0.10	0.07	0.10	0.08	0.12	0.06	0.05	0.01	0.00	0.00	4.10
Mean Temperature (°C)	15.8	14.6	12.0	15.3	11.3	15.7	15.7	21.4	19.9	18.2	20.6	21.3	20.7	16.1	16.4	14.5	16.4	17.7	16.2	10.6	14.4	10.1	16.1 (Mean)
Field tank evapotranspiration	2.80	2.86	2.84	2.10	2.90	3.33	3.82	4.50	4.42	1.62	3.02	4.45	3.80	3.10	4.40	4.37	3.86	NV	NV	NV	NV	NV	58.19 (17 weeks)
Woods tank evapotranspiration	1.00	1.02	0.38	0.79	2.20	1.32	1.97	1.90	2.64	1.52	1.09	1.42	0.79	0.89	3.64	4.63	4.60	NV	NV	NV	NV	NV	49.42 (17 weeks)
Thornthwaite potential evapotranspiration	2.16	1.96	1.64	2.20	1.60	2.33	2.36	3.26	2.98	2.70	3.02	3.06	3.03	2.66	2.24	1.87	2.14	2.25	1.96	1.25	1.64	1.11	49.42
Thornthwaite soil moisture storage	12.90	12.40	12.84	12.70	11.90	10.50	9.20	7.50	6.60	9.46	9.70	7.10	7.10	6.00	7.59	7.40	9.40	8.30	8.59	8.00	7.10	7.13	
Thornthwaite actual evapotranspiration	2.10	1.84	1.64	2.12	1.54	1.86	1.78	1.78	2.24	2.70	2.46	1.83	2.80	1.20	2.24	1.54	2.14	1.46	1.96	0.64	0.90	1.11	39.88
Thornthwaite moisture deficit	0.06	0.12	0.00	0.08	0.06	0.47	0.58	1.48	0.74	0.00	0.56	1.23	0.23	1.46	0.00	0.33	0.00	0.79	0.00	0.61	0.74	0.00	9.54

NV = No Value

(d) Rainfall Interception, Penetration and Stemflow

Rainfall interception, penetration and stemflow was determined from the experimental area described in Chapter III, and was applied to the water balance of the whole forest. It is appreciated that such values will vary from one section of forest to another with variations in exposure and variations in forest morphology, including tree types. However, the maple-beech assemblage, which formed the vegetation of the experimental area is very typical of the many mixtures of maple and beech which are found on Mont St. Hilaire and which are delineated in Figure 1.

As seen in Table 13, the 9.4 cm of water intercepted by the tree foliage forms a sizeable proportion of the total rainfall. The stemflow of 0.31 cm, on the other hand, does not constitute an important part of the water which penetrates to the soil surface. For the growing season of 1964, only 71 percent of above-canopy rainfall pentrated to the soil surface. Nor was all of this water eligible to enter into the evapotranspiration from soil surface and plant vegetation, since some was lost as surface runoff.

The investigator is thus brought face to face with an important fact in the hydrologic and energy regime of a mature forest cover. The forest crown, through its ability to hold water, becomes an active agent in determining the utilization of solar energy as a result of its very morphology, as well as by its biologic functions.

(e) Runoff

The movement of water which reaches the forest soil is a complex problem and is treated in its very simplest form in this study.

That portion of the rainfall which immediately escapes to the stream through overland flow is usually designated as surface runoff. Kittredge (1948) details surface runoff as the way in which water reaches the stream most rapidly, compared with ground water flow. Fig. 12 shows a typical two-day hydrograph which has been reproduced from Kittredges' book and is based on the work of Horton (1935). The hydrograph is roughly divided into water supplied from surface runoff, channel storage, and ground water flow. The total area under the curve is equal to total stream flow. The rise of the stream and early part of the falling stage include the surface runoff. The termination of the surface runoff, according to Horton, comes at the point on the time scale where the descending limb of the hydrograph changes from convex to concave upward. The portion of the hydrograph below the curve at the extreme right, where the rate of decrease is very slow and where the trends after different storms coincide, represents ground water flow. Both surface runoff and ground water flow are lost to the forest vegetation.

In the water-balance table the total runoff, including both surface and ground flow, has been calculated for the total growing period. It can be seen that the runoff decreases rapidly from the first 5 weekly periods, when it held a fairly constant value at 0.40 cm per week, to a very low rate of about 0.02 cm per week during the last 5 weeks. It can be noted further that, during Week 1 and Week 8, there was a net loss of ground water because runoff was substantially higher than the rainfall received. Fig. 13 shows the weekly values of runoff and rainfall for the total period of study. One can surmise from this graph that the soil

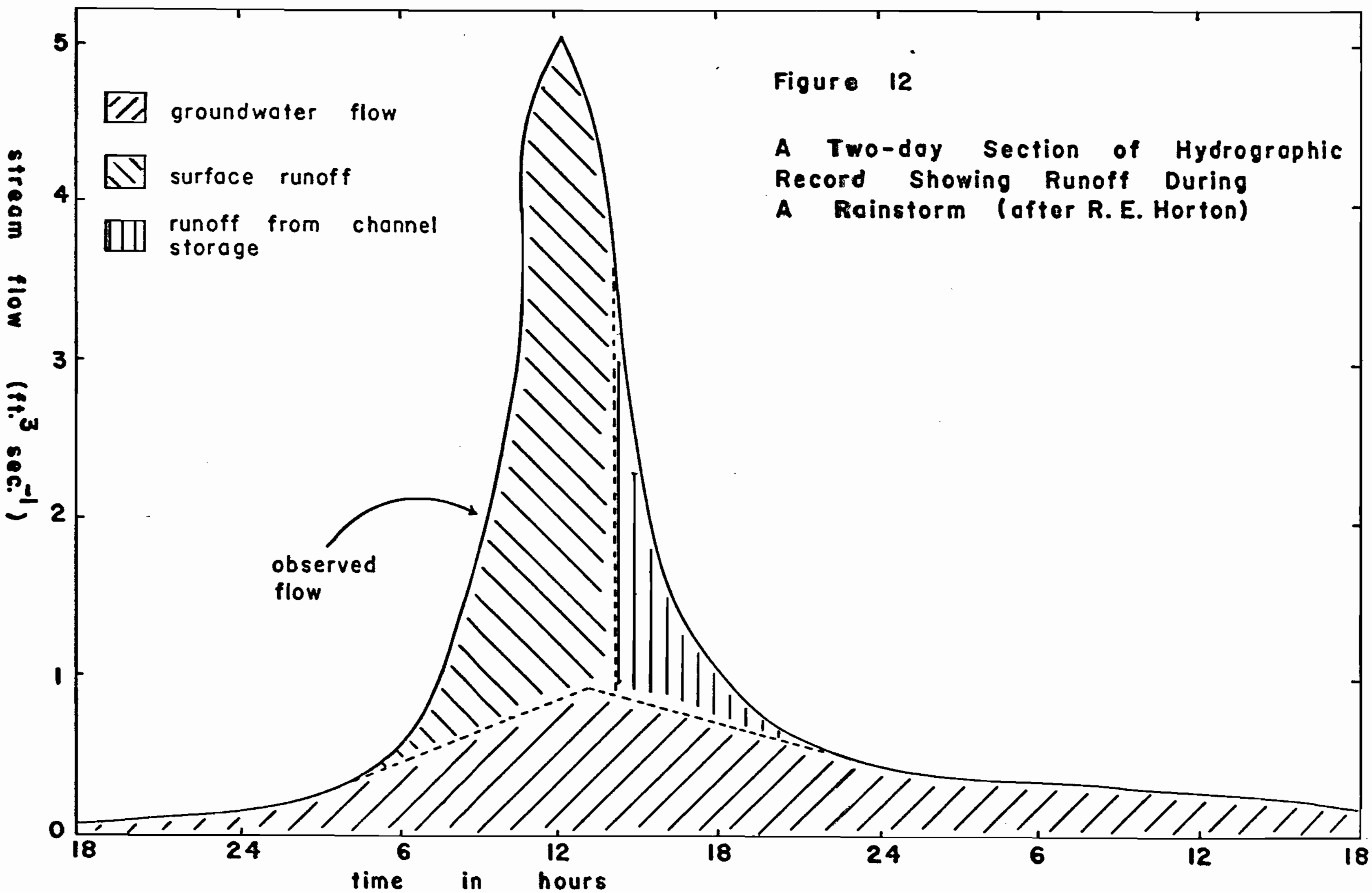


Figure 12

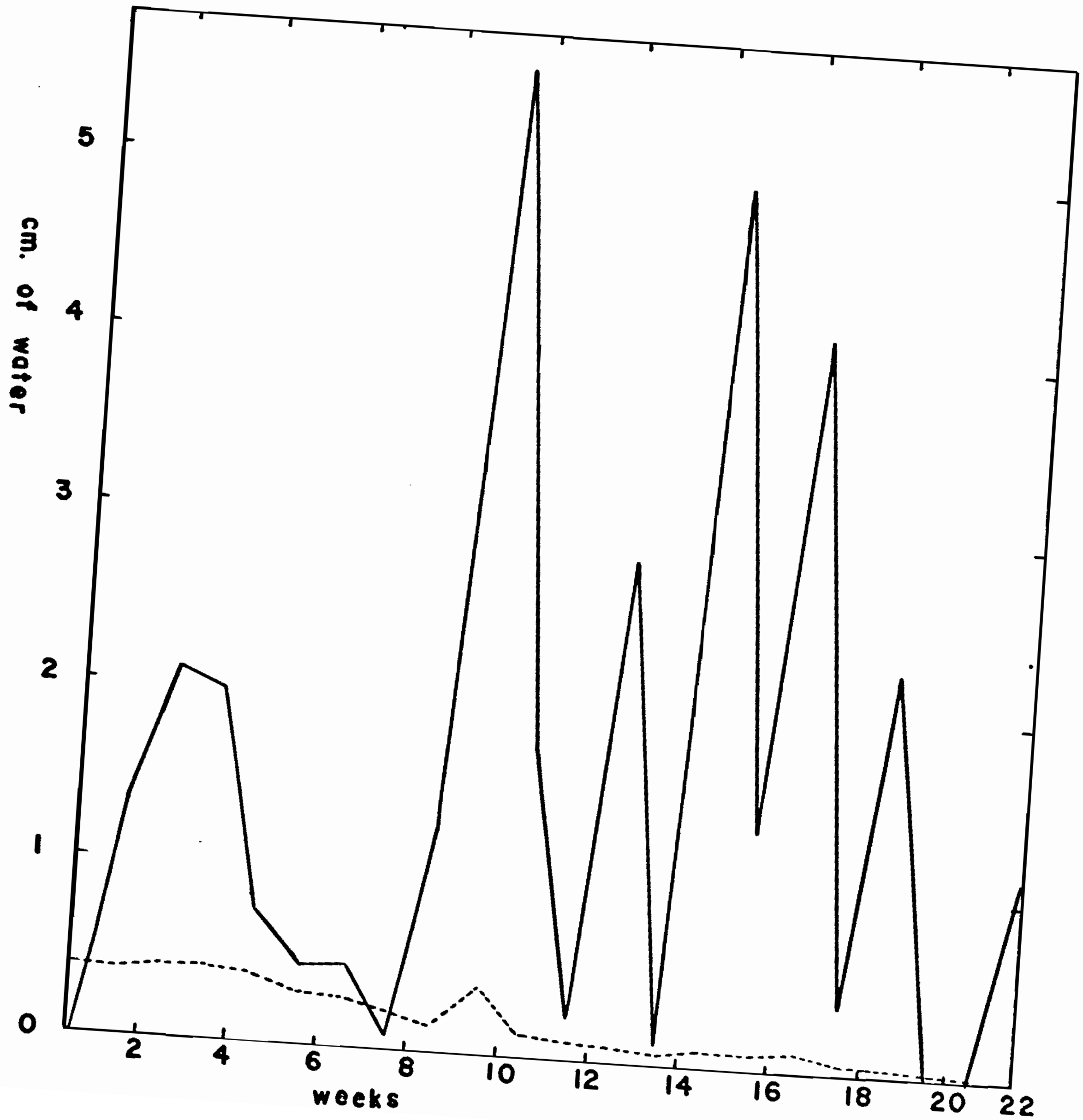
A Two-day Section of Hydrographic Record Showing Runoff During A Rainstorm (after R. E. Horton)

Figure 13

Rainfall and Runoff During the Growing Season

——— above-canopy rainfall

- - - - - - - runoff



became steadily drier as the growing season progressed. As a result of the increase in the moisture storage capacity of the soil, the runoff steadily decreased in volume during the latter two-thirds of the growing season, in spite of the increased rainfall during this period.

(f) Changes in Soil Moisture

The greatest weakness in this development of the water-balance lies in a lack of data on soil moisture changes. Direct measurements would have proved a major undertaking and would be worthy of a complete study in themselves. One of the problems is a lack of any instrument which is capable of measuring soil moisture change accurately, rapidly and continuously. Even more important, the determination of a soil moisture change representative for the whole area is a formidable task. Soil depths on the mountain vary from a minimum of 10 cm, over bedrock on steep slopes, to several metres in areas overlain by a fairly thick cover of till. It is fairly certain judging from the sample augur borings carried out as a part of this program, that the soil exceeds 2 M in thickness above bed rock in only a few places.

It is necessary to make estimates of soil depths and soil types since the soil moisture withdrawal, whether large or small, will play an important part in the calculations of total evapotranspiration in the forest. The author's assessment is that the soil types on Mont St. Hilaire most closely resemble a silt loam. It is further estimated that the average rooting depth of the forest trees is only 75 cm. By referring to tables of soil water-holding capacities, as outlined by Thornthwaite and Mather (1957), a figure of 15 cm soil moisture capacity is derived.

Since most of the soil moisture was used in evapotranspiration during the growing season, errors in estimating soil moisture capacity will directly affect the calculation of evapotranspiration, when one employs the hydrologic-balance method. If soil moisture capacity is only 10 cm, the total evapotranspiration for the growing season will be overestimated by some 12 percent. A moisture capacity of 20 cm would lead to a similar underestimation. This writer is reasonably confident that the above estimates are realistic and that the error in determining evapotranspiration should lie within plus or minus 12 percent.

(g) Evapotranspiration

Three different methods were used to calculate the evapotranspiration from the forest. These include direct estimates from the hydrologic-balance, calculations based on measuring potential evapotranspiration from soil-filled tanks, and the mean temperature method of calculating potential evapotranspiration as presented by Thornthwaite (1948) and revised by Thornthwaite and Mather (1953).

(i) Hydrologic-balance. A figure for full-season evapotranspiration can be determined from the equation

$$E = P - Q + St. \tag{23}$$

where E is evapotranspiration, P is above-canopy precipitation, Q is runoff, and St is the soil moisture withdrawal. For the 1964 growing season, equation (23) yields E = 33.08 - 4.10 + 15.00 = 43.98 cm.

It is important to note that complete withdrawal of soil moisture is assumed, and that the value for total above-canopy rainfall is used. It should be emphasized that the 9.43 cm of water that is intercepted by the canopy is part of evaporation within the total evapotranspiration process.

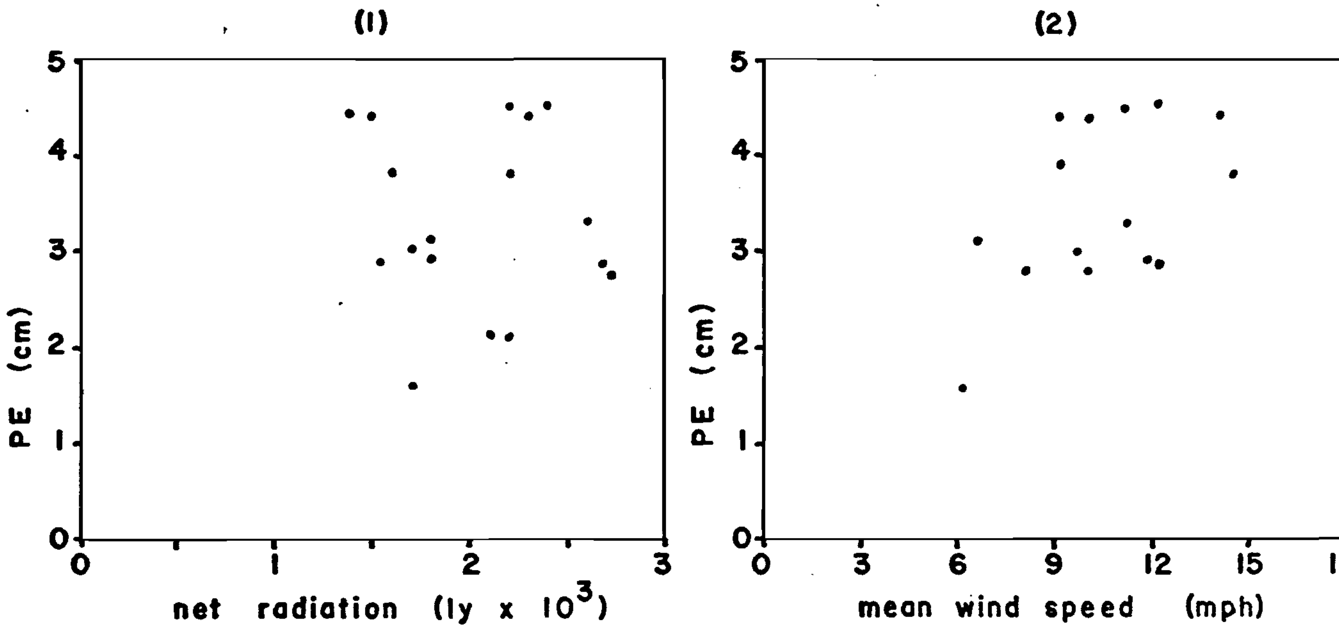
Limits can be defined for the amount of water which enters into evaporation and transpiration respectively. As will be shown in more detail in Chapter V, the ratio of net radiation at the forest floor to the above-canopy value does not exceed 0.05 for the full-leaf state in a virgin stand. By determining the moisture equivalent for net radiation at the forest floor, for the period under study, evaporation should not exceed 3.30 cm. The idea that the net radiation which is received at the forest floor places an effective limit on evaporation from the forest soil is physically reasonable. In a thorough study of the heat balance of a corn field, Tanner, Peterson and Love (1960) determined the ratio of soil evaporation to total evapotranspiration by comparing the transpiration from a plot with a mulch-covered soil to the evapotranspiration from a natural plot. They found that soil evaporation depended on the net radiation at the soil surface when the soil surface was wet, but as the soil dried the capillary property of the soil (i.e., its ability to move water upward to the surface) became important. The authors concluded that the maximum possible latent heat which was used in evaporation from the soil equalled the available energy of net radiation since the soil received virtually no advective heat. Kittredge (1962) gives measurements from California which show the dependance of evaporation from the forest soil on the density of the forest crown. By setting open pan evaporation in a large clearing equal to 100 percent, he found that in an old irregular stand of pine the amount of open pan evaporation at the forest floor was 83 percent, in a partially cut stand of balsam and pine it was 69 percent, and that it fell to 17 percent in a dense stand of mature balsam. Thornthwaite and Hare (1964) state that in dense forest or under thick, tall grass, the shading of the soil and the shielding against eddies almost eliminates evaporation from the soil, and its surface remains moist long after each rain.

The evaporation in terms of the energy available from net radiation at the forest floor will be limited to 3.30 cm of water. By adding the amount of water intercepted by the canopy to possible soil evaporation, the maximum possible evaporation would be 12.73 cm, thus leaving 31.25 cm of water to be used in transpiration. These values give a ratio for evaporation/total evapotranspiration of 0.29.

(ii) <u>Potential Evapotranspiration from Tanks.</u> Attempts to measure potential evapotranspiration from tanks were both disappointing and confusing; disappointing because the results appear to be unrepresentative of the forest condition, and confusing because there is no ready explanation for the nonrepresentative measurements. Up to the end of Week 17, when tank measurements were suspended, a total potential evapotranspiration of 58.2 cm and 31.5 cm was obtained from the Field and Woods Tanks respectively. If it is presumed that all of the net radiation was used in evapotranspiration the potential evapotranspiration from the Field Tank would consume about 108 percent of the net radiational energy, while the Woods Tank would need many times more energy than was available from the net radiation beneath the forest canopy.

In order to attempt an explanation for the evapotranspiration from the tanks being so high, Figs. 14(a) and 14(b) plot weekly values of potential evapotranspiration against weekly values of net radiation, mean wind speed, total rainfall, rainfall intensity and time, for the Field and Woods Tanks respectively. Weekly rainfall intensity was determined in the following manner. Any day of the week which had from 0.50 to 1.00 cm of rain was assigned an intensity of 1. Similarly, ranges from 1.00 to 1.50, 1.50 to 2.00, 2.00 to 2.50 were assigned intensities of 2,3,4. All the daily intensity values were then added, to give a weekly total. Such a procedure is arbitrary, but it does serve to give a rough indication of rainfall intensity. From Fig. 14(a) it can be seen that the scatter for all plots is quite random. Surprisingly, net radiation and potential evapotranspiration from the Field Tank show no correlation. Only mean wind speed and time, when plotted against potential evapotranspiration, show any trends, and those are slight. In Fig. 14 (b) an even wider scatter is shown, and any trends between potential evapotranspiration, and mean wind speeds or time are even less apparent. These particular parameters were chosen for the following reasons. Net radiation and potential evapotranspiration should be related since the former represents the amount of energy which is available for the vapourization of water. Graham and King (1961) derived an excellent correlation between net radiation and potential evapotranspiration from a corn field. Both total rainfall and rainfall intensity were plotted to investigate the possibility that with large amounts of rain or with very heavy falls,

Figure 14 (a)
Weekly values of PE (field tank) plotted aginst weekly values of
1. net radiation
2. mean wind speed
3. total rainfall
4. rainfall intensity
5. time



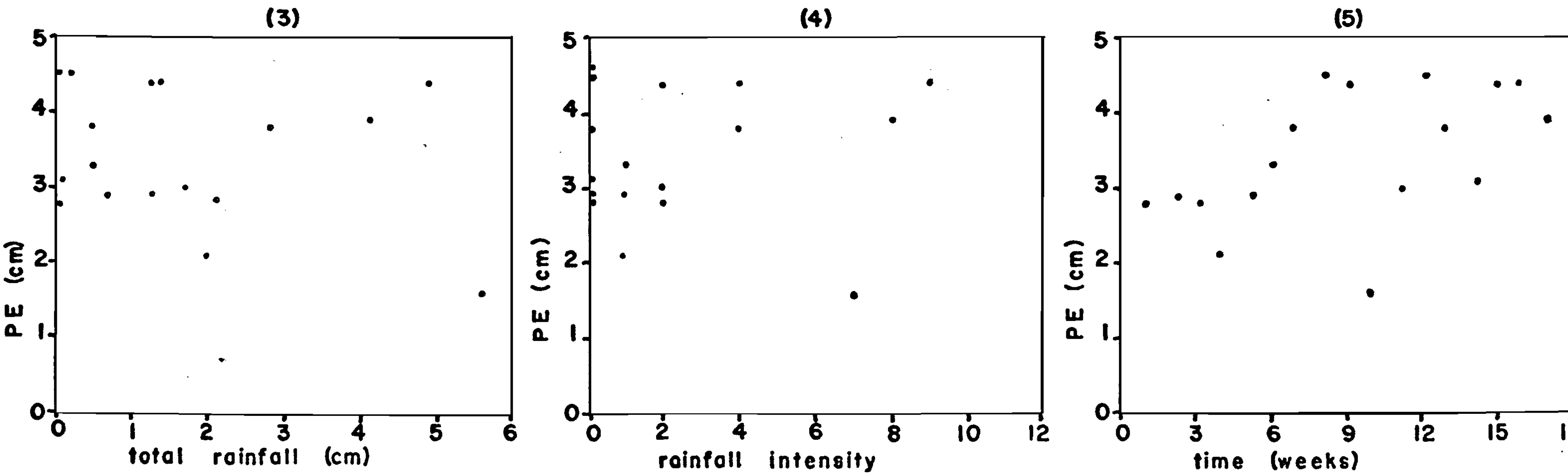
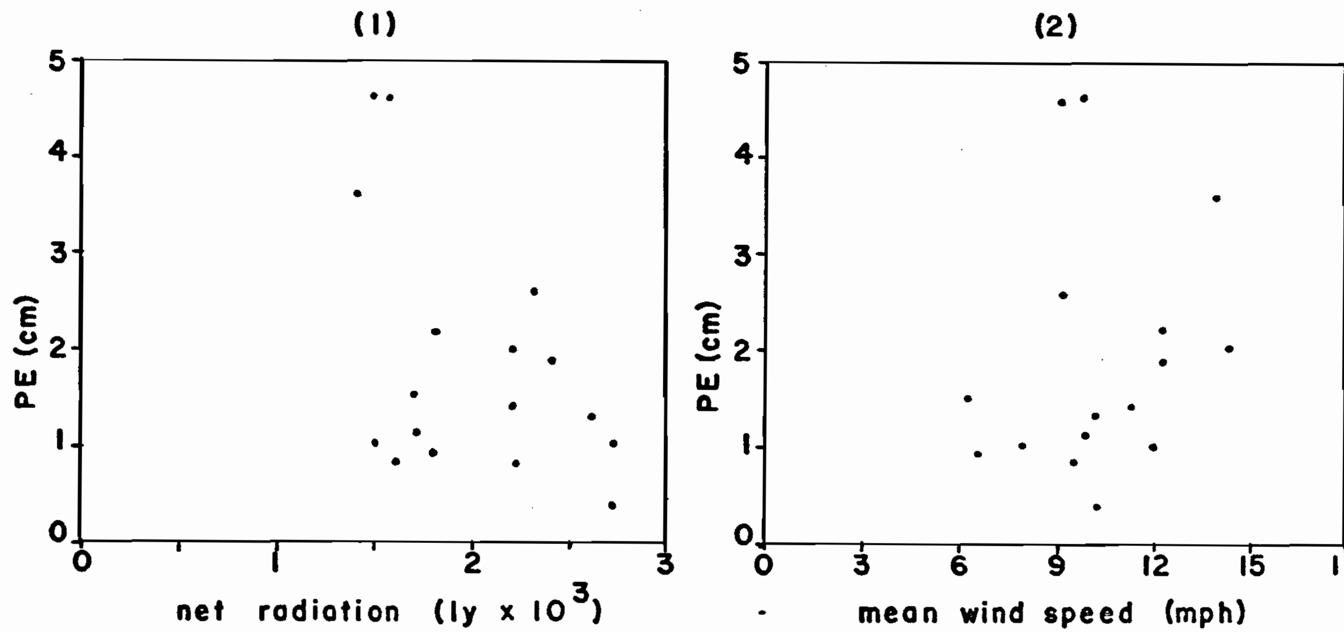
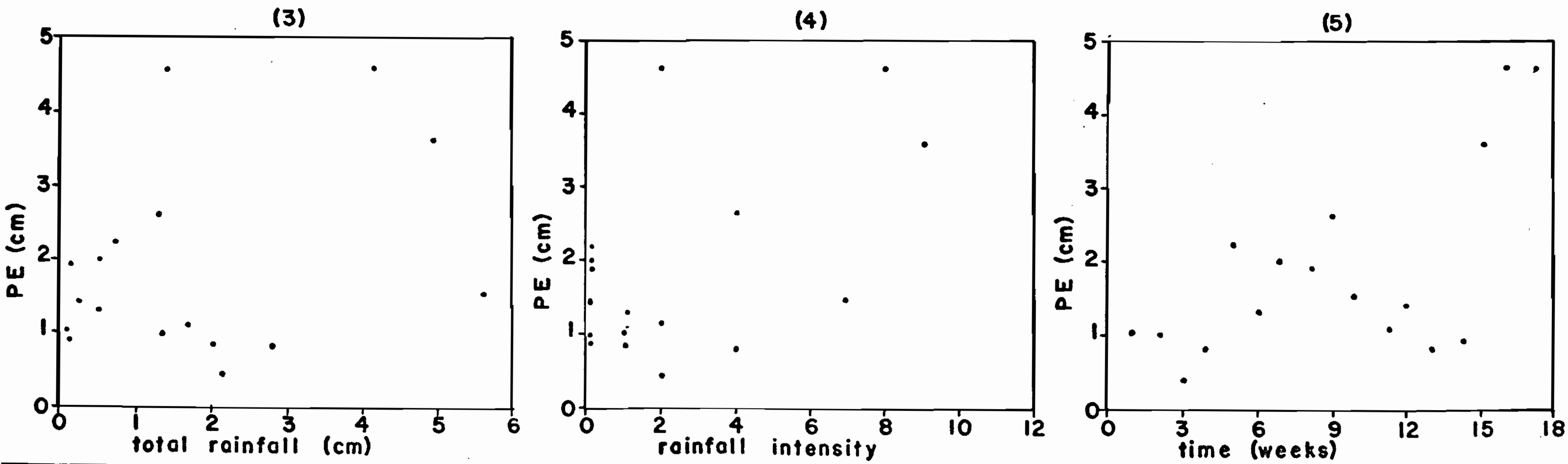


Figure 14 (b)
Weekly values of PE (woods tank) plotted against weekly values of
1. net radiation
2. mean wind speed
3. total rainfall
4. rainfall intensity
5. time





the tanks might be flooded at the surface, so that much water escaped over the rims of the tanks. If this occured, the potential evapotranspiration would be overestimated. Mean wind speed should play a part in oasis effects. If the ground surrounding the tank is drier than the tank itself, and strong winds carry warm dry air over the moistened surface of the tank, evapotranspiration will be increased. Time as a factor also involves the oasis effect. Since the ground water decreases as the growing season proceeds, the area surrounding the tank will become drier with time, and oasis effects will be accentuated. Perhaps it is significant that the three weekly periods which showed exceptionally high evapotranspiration from Field and Woods Tanks, especially the latter, occured after August 9 during the latter part of the growing period.

If the tanks were leaking, the runoff which was measured in the overflow tanks would be less and, as a result, the potential evapotranspiration would be overestimated. The tanks were tested for leaks in the following manner. Each tank was saturated with water and then covered with polyethylene sheets to prevent evapotranspiration. After all of the gravitational water had percolated to the overflow tank and a steady state condition been achieved, a known amount of water was added to each tank. In each case the amount of overflow equalled the known amount of water which was added, thus showing that the tanks did not leak during the test.

A trend might be obtained from Figs. 14 (a) and 14 (b) if the various parameters could be weighted and the potential evapotranspiration plotted against some function of these variables, but the form of such a function is not known. It is concluded that the measurements from the evapotranspirometer tanks cannot be taken as representative of conditions in an undisturbed forest since their response characteristics may be influenced by local microclimates which are peculiar to the particular sites.

(iii) Potential and actual evapotranspiration from mean temperatures data. Thornthwaite's application of temperature data for the calculation of potential evapotranspiration was outlined in Chapter II. With a knowledge of rainfall and assumptions as to soil moisture capcity and the rate of withdrawal of soil moisture, all of which are outlined in Thornthwaite and Mather (1957), a simple book keeping system allows the calculation of actual evapotranspiration, water deficit, and water surplus. This calculation was applied to the mean temperature and rainfall data for the 22 week growing period at St. Hilaire. A value of 15 cm field capacity, the same value as was used to calculate the hydrologic balance was used in the Thornthwaite calculations. Table 13 gives the potential evapotranspiration, soil moisture storage, actual evapotranspiration, and moisture deficit as calculated using the Thornthwaite approach. There was no moisture surplus. The actual evapotranspiration of 39.88 cm for the full period represents 91 percent of the evapotranspiration of 43.98 cm that was calculated from the hydrologic-balance. However, because the rate of soil moisture withdrawal lessens with increasing moisture deficiency when using the Thornthwaite system, there was still 7.13 cm of water in the soil at the end of the growing season. This contrasts to the approach when calculating the hydrologic-balance, where a complete soil moisture withdrawal was postulated. Another difference in the two approaches involves runoff. In the Thornthwaite calculations there is no runoff until field capacity is reached, whereas the field measurements which were used in the hydrologic-balance showed a runoff of 4.10 cm. If the Thornthwaite calculation is altered by adding to the actual evapotranspiration the amount of water remaining in the soil (7.13 cm), and subtracting the runoff which was measured in the field (4.10 cm), a value of 42.91 cm is derived for actual evapotranspiration. This comes within 5 per-cent of the actual evapotranspiration, as calculated from the hydrologic-balance. Thus, for the 1964 growing season at Mont. St. Hilarie, there is good agreement between the Thornthwaite system of calculating actual evapotranspiration and the field calculations from the measured and estimated elements of the hydrologic balance.

(h) Seasonal Water Balance for Mont St. Hilaire.

By using the actual values measured for rainfall, interception, stemflow and runoff, evapotranspiration values as determined from the hydrologic-balance, and a soil moisture capactiy of 15 cm, the following table of the forest water balance for 1964 emerges.

Table 14 Water Receipts

Rainfall	33.08 cm.			
Soil Moisture	15.00 cm.			
Available Moisture		48.08		100%
Water Expenditures				
Runoff		4.10		8.5%
Evaporation	12.73		29.1%	
Transpiration	31.25		70.9%	
Evapotranspiration		43.98		91.5%
Total		48.08		100%

(i) Comparison to other studies

Pierpoint and Farrar (1962) studied the water balances of the University of Toronto Forest in Haliburton County, Ontario, by using the Thornthwaite mean temperature technique. The forest lies at an elevation of 1000 feet, and consists mainly of sugar maple and yellow birch trees with an admixture of hemlock and pine. The soil averages 4 ft. in depth above bedrock and to it the authors assigned a moisture capacity of 10 cm. In temperature, precipitation and altitude the Haliburton station resembles very closely that at Mont St. Hilaire. Pierpoint and Farrar calculated the annual water balance for the years 1951 to 1957. As the figures below show, the year 1951 in Haliburton closely resembled the year 1964 at Mont St. Hilaire, the main difference being that the former station received about 6 cm more rainfall. The evapotranspiration, as calculated using Thornthwaite's formula, was very similar to that calculated for Mont St. Hilaire from the hydrologic-balance during a drier year. If the authors had used a

Table 15

Month	Mont St. Hilaire		Haliburton		
	T°C	P(cm)	T°C	P (cm)	*AE(cm)
May	14.0	5.6	11.1	7.2	7.3
June	17.3	3.0	16.5	7.6	10.0
July	20.3	10.2	18.9	9.0	10.0
August	16.4	11.5	17.8	6.7	8.0
September	13.1	3.5	13.8	9;2	7.5
Totals		33.8		39.7	43.4

*AE refers to actual evapotranspiration

field capacity of 15 cm instead of 10 cm, their calculations of actual evapotranspiration would have been somewhat higher. It appears that the Thornthwaite technique accurately estimates evapotranspiration from a forest in middle latitudes, if the writers calculations for Mont St. Hilaire and his comparisons to the Haliburton data give a true indication.

Kittredge (1948) reported on a Swiss stream flow experiment which was carried out by Engler (1919). He measured out of a total precipitation of 62.6 cm, runoff equal to 37.1 cm, 11.8 cm of transpiration and 4.6 cm of evaporation, with a canopy interception of 9.1 cm. His ratio of evaporation to transpiration worked out to 0.29 which compares to the 0.33 calculated in the present study. Bates and Henry (1928) carried out an experiment on two catchment areas at Wagon Wheel Gap, Colorado. Out of 21.2 in of rain, they measured runoff at 6.2 in, interception at 3 in., and transpiration, and evaporation amounts of 5 and 7 inches respectively. The fact that one of the catchment basins was severely denuded of trees may have led to a greater evaporation of water in comparison to transpiration, and to a lower canopy interception than would be usual in more natural forest environments. Meyer (1932) concluded that for the northern and central parts of the United States the normal annual transpiration of deciduous forest is 8 to 12 inches, and of conifers it is 4 to 6 inches. Minckler (1939), in a study of a 55 year old beech-maple forest in New York State, estimated annual transpiration at from 4 to 5 inches. Kittredge (1948), in summing up a number of studies, states that annual transpiration for most forests in the United States is probably between 5 and 15 inches (12.7 to 30.8 cm) and may approach 35 inches (89.0 cm) for large dense stands on the best sites.

Croft and Manniger (1953) used a hydrologic-balance approach to compare a bare soil plot with an area covered by poplar trees. For the bare plot without cover, the evaporation was 28.4 cm for the growing season. The evapotranspiration from the poplar forest was 47.5 cm. Kittredge (1962), in commenting upon this experiment, notes that under these conditions the transpiration would be greater than the evaporation. In other words, it would be more than 23.7 cm but less than 47.5 cm. Such an approach, Kittredge states, may lead to a very rough approximation of transpiration, but the result lacks precision. Penman (1963) comments in a similar vein. He notes that while some of the water comes directly from the soil (evaporation), and some comes indirectly from the soil through the plant (transpiration), few of the separations of evaporation and transpiration in published water budgets have any rational basis to justify their apparent precision. Penman's comment must be taken seriously. The author feels, however, that a fairly accurate separation can be made in a detailed investigation. Certainly, the water which is intercepted by the leaf layer of plants will be evaporated if the leaves are not to remain permanently wet. The evaporation of water from the soil offers more problems. The net radiation at the forest floor may not place an effective limit on the possible amount of evaporation, because of an extra energy supply created by a downward movement of sensible heat energy from the crown layer. If such a flux is important then it should be evident in daytime temperature profiles which show inversion conditions in the tree trunk area. A number of studies show that this is the case during daytime as well as nightime conditions. Hales (1949) measured a temperature increase of 2°C. between 0.6 and 46 M heights beneath the canopy of a tropical jungle during the noon hour period. As discussed in Chapter II, Cantlon (1950) found that the temperature profile beneath the forest canopy varied with slope exposure. Inversion conditions persisted on the north slope but on the south slope lapse or isothermal profiles were found during the day with inversion conditions at night. Kittredge (1962) shows measurements from California where isothermal conditions were maintained at 1500 hours beneath a forest having only a 60 percent crown coverage. Denmead (1964) measured available energy at various heights in the forest canopy and at the same time took profile measurements of temperature and humidity for a pine forest. He found that sensible heat transfer from the site of maximum absorption of net radiation to lower parts helped to even out the transpiration load within the forest canopy. His computations did not extend beneath the canopy layer. Geiger (1965) shows daily temperature profiles for an oak forest 25 M in height. On the average the period between 0600 and 1800 hours showed a temperature increase of 2.5°C between a level at 3 M above the forest floor and the lower canopy level at 21 M. Geiger also quotes measurements for a dense fir plantation which show as a daytime average a 3.0°C temperature increase between the 0.6 M and 3.0 M levels. To the extent that evaporation is hastened due to this heat source (i.e. a downward convective heat flux), errors will arise in the separation of the evaporation and transpiration terms, and such an error may appear in Table 14. The magnitude of this type of error should be small in comparison to the total amounts of water which are involved in the evapotranspiration process.

(j) Summary

The growing season at Mont St. Hilaire was concentrated between May 1 and October 1 in 1964. Measurements during this period were summarized for weekly intervals, there being 22 weeks in all. The growing season with a rainfall of 33.1 cm was much drier than usual. 29 percent of this rainfall was intercepted by the forest canopy, 1 percent reached the ground via stemflow, and the remaining 70 percent penetrated to the ground as throughfall. Of the total rainfall, a little over 12 percent was lost to the forest through runoff. Soil moisture changes were not measured, and had to be estimated from the soil type and rooting depths of the trees. A soil moisture capacity of 15 cm was assigned for Mont St. Hilaire.

Attempts were made to determine actual evapotranspiration from the hydrologic-balance, from the measurement of potential evapotranspiration in tanks, and by using the Thornthwaite mean temperature technique for calculating potential evapotranspiration and book-keeping computations for calculating actual evapotranspiration. The hydrologic-balance and Thornthwaite calculations agreed closely. The tank measurements appeared to give readings which were influenced by the microclimates of their particular sites. Possible reasons for this influence are presented, but no conclusive evidence can be cited.

All water receipts and water expenditures were calculated for the growing season. The evapotranspiration process was divided into the evaporation from the tree canopy which was determined from the canopy interception of rainfall, evaporation from the forest floor which was estimated from the net radiation at the forest floor, and transpiration which became the remaining amount of water which was transferred to the atmosphere.

Chapter V Radiation and its Energy Transformations during the Growing Season

(a) Introduction

The results discussed in this chapter apply to the overall energy-balance of the forest during the growing season. The detailed variations of the radiation and moisture regimes within the forest are treated in Chapter VI. This writer has not been able to measure continuously all the quantities necessary for a complete energy-balance description during the present study and has found it necessary to rely, to some extent, on other investigations in developing the sections on soil heat flow, heat flow in the biosphere, and heat exchange in the forest air. It is stated, with confidence, that each of these quantities becomes quite small over periods of weeks, months, and over the total growing season when compared to the much larger fluxes of net radiation, latent heat and sensible heat. All of the energy quantities which have been calculated for the twenty-two weeks of the growing season from field measurements are presented in Table 16.

(b) Global Radiation and Albedo

For the active growth period of 1964, a total global radiation of 66,918 ly impinged on the forest cover. The weekly values ranged from a low of 1909 ly during Week 22, the last week of the growing season, to a high of 4127 ly for Week 1, at the beginning of May, a period when cloud cover was very low.

A few measurements of albedo were taken during the summer by inverting the Kipp solarimeter above various tree types, and comparing the reflected radiation to the incoming radiation as measured on the actinograph. These measurements gave the following mean albedo values.

Tree	Albedo
Oak	0.16
Beech	0.19
Maple	0.19
Pine	0.14
Hemlock	0.15
Birch	0.17

Such values agree well with values of 0.18 and 0.14 for oak woodland and pine forest respectively, which were determined by Angstrom (1925), and with albedo values given by Budyko (1958), which show a range from 0.15 to 0.20 for deciduous forest in various geographical locations. The mean forest albedo for Mont St. Hilaire can be placed at 0.18.

(c) Net Radiation

(i) Total. The net total radiation has been divided into the positive values which were accumulated mainly during the daylight hours and the negative ones which were largely measured during the night-time period. The term net total radiation represents the difference between the positive and negative measurements. It is this amount of radiant energy which is available to do work at the earth's surface, and which enters mainly into the latent and sensible heat fluxes. During the growing season, the net radiation constituted 58 percent of the incoming solar radiation. Fig. 15 shows the weekly variations in the net radiation, the absorbed solar radiation, the long-wave radiation balance and the nocturnal radiation.

(ii) Absorbed solar radiation and the long-wave radiation balance. Values of radiation were measured by the solarimeter and actinograph, and the absorbed solar radiation was determined by assuming a mean albedo of 0.18 for the forest, and subtracting the albedo value from the incoming radiation. The data for the long-wave radiation balance were derived from

Table 16. Radiation Balances. (all values in langleys)

	Week	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	Totals
(a)	Global Radiation	4127	2537	4004	3612	2740	3806	3879	3766	3583	2611	3083	3730	2804	3218	2455	2244	2727	2767	2071	2699	2546	1909	66918
(b)	Reflected radiation (albedo)	743	456	720	650	494	687	700	678	645	470	556	672	506	580	440	405	492	498	374	486	460	344	12056
(c)	Absorbed radiation =(a)-(b)	3384	2081	3284	2962	2246	3119	3179	3088	2938	2141	2527	3058	2298	2638	2015	1839	2235	2269	1697	2213	2086	1565	54862
(d)	Daytime Net radiation	3010	1774	2982	2446	1944	2941	2545	2730	2530	1844	2007	2564	1890	1951	1647	1769	1824	1663	1451	1528	1496	1321	45857
(e)	Nocturnal radiation	-330	-245	-321	-248	-169	-327	-323	-293	-203	-131	-343	-338	-289	-144	-219	-287	-354	-416	-383	-677	-519	-363	-6922
(f)	Net total radiation =(d)+(e)	2680	1529	2661	2198	1775	2614	2222	2437	2327	1713	1664	2226	1601	1807	1428	1482	1470	1247	1068	851	977	958	38935
(g)	Long-wave radiation balance =(f)-(c)	-704	-552	-623	-764	-471	-505	-957	-651	-611	-428	-863	-832	-697	-831	-587	-357	-765	-1022	-629	-1362	-1109	-607	-15927
(h)	Energy used to evaporate intercepted water	0	318	530	272	77	106	112	35	390	695	365	47	403	41	247	326	710	124	525	29	0	212	5564
	Ω	0.08	0.05	0.05	0.13	0.00	0.01	0.21	0.08	0.01	-0.03	0.23	0.16	0.13	0.20	0.08	-0.12	0.19	0.35	0.14	1.04	0.64	0.13	Average 0.17

Figure 15

Radiation Regime During the Growing Season

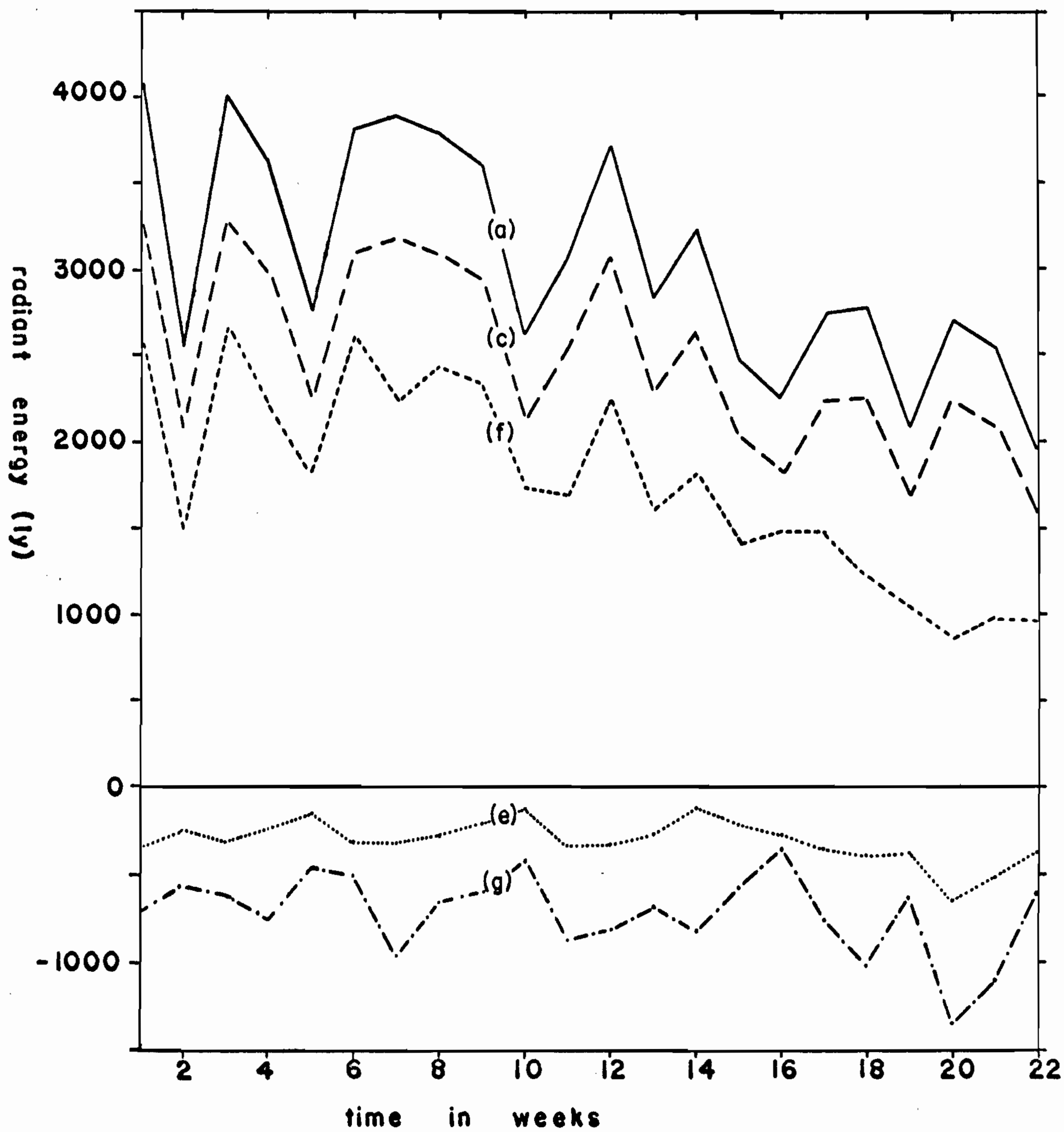
(a) global radiation

(c) absorbed radiation

(f) net total radiation

(e) nocturnal radiation

(g) long-wave radiation balance



the difference between the absorbed solar radiation and the net total radiation - both measured parameters. Nocturnal radiation was measured directly.

In an effort to evaluate the role of the radiational fluxes, this writer applied a modified form of an approach used by Monteith and Szeicz (1961), in their study of the radiation balance of soil and vegetation. Monteith and Szeicz noted the importance of net radiation in agricultural meteorolgy, and determined the dependence of net on solar radiation under clear skies for different surface covers. They plotted linear regressions of net radiation (Rn) on the absorbed solar radiation ($Rsi\ (1 - \alpha)$) for cloudless days to give the regression formula

$$Rn = a\ (1 - \alpha)\ Rsi + b \tag{22}$$

The correlation coefficients were high, varying from 0.95 for net radiation over bare soil, to 0.99 for net radiation over spring wheat. Equation (22) can satisfy

$$Rn = (1 - \alpha)\ Rsi + R_L \tag{23}$$

where $R_L = Rli - Rlr$ the long-wave radiation balance, only where R_L is a linear function of Rn having the form

$$R_L = b/a - Rn\ (1 - a)/a = b/\ a - \Omega\ Rn \tag{24}$$

thus defining Ω. Because R_L is negative, Ω is the increase in net long-wave loss per unit increase of net radiation income. As an increase in net radiation is associated with an increase in surface temperature to which the net long-wave loss is very closely related, Ω may be regarded as a heating coefficient. For various cover types and time periods, Monteith and Szeicz derived Ω values as listed in Table 17. The r values in this table show the linear correlation between Rn and Ri $(1 - \alpha)$.

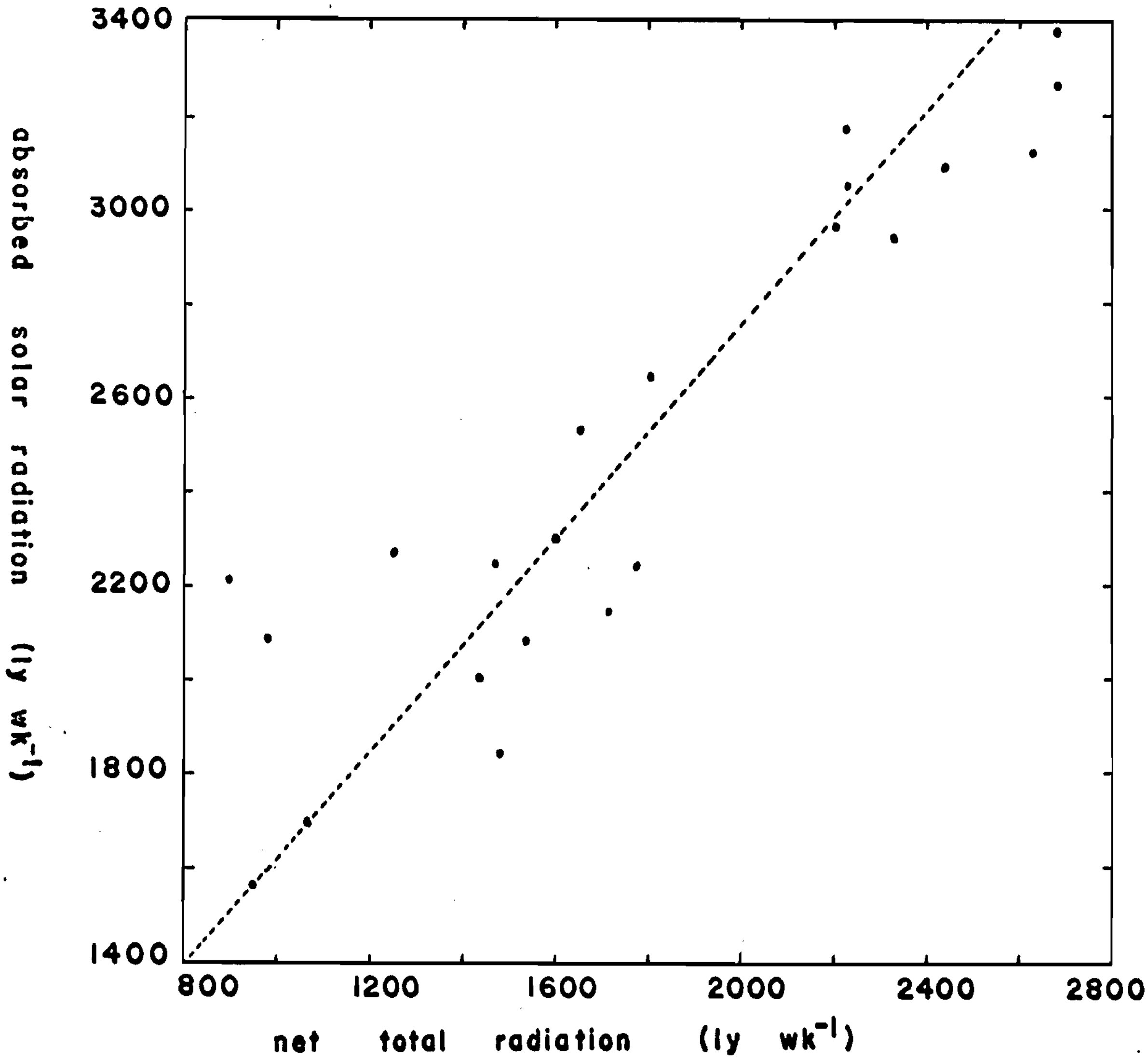
Table 17. Ω Values for cloudless days (After Monteith and Szeicz)

Surface	Date	Ω	r
Spring wheat	May 25-July 27, 1957	0.15	0.99
Sugar beet	July 3-September 27, 1958	0.21	0.97
Grass	July 23-August 27, 1959	0.22	0.98
Bare soil	April 14-April 30, 1957	0.35	0.98
Bare soil	May 1-June 18, 1958	0.39	0.95
Bare soil	June 13-June 20, 1959	0.41	0.98
Bare soil	September 7-September 20, 1959	0.37	0.97

In the present study the weekly values of net radiation were regressed on the weekly values of the absorbed solar radiation to give a correlation coefficient r of 0.92. It is important to recognise that this regression which is shown in Fig. 16 includes all sky types, not just cloudless days. Values of Ω were then calculated for each week during the growing season, and are shown in Table 16. In this case Ω may not be considered a heating coefficient because, as well as being influenced by surface temperature, it is also under the influence of cloud conditions. What it does denote is the importance of the long-wave components of the net radiation. Hence during Week 20, the highest value at 1.04 denotes a low net radiation which occured in spite of a high level of solar radiation, because of a large negative long-wave balance. In contrast, the lowest Ω at -0.12 occured during Week 16 when a relatively high net radiation was maintained in spite of low amounts of solar radiation, because of a small negative long-wave balance. In the former example, the weekly mean cloud cover was 4.1/10, whereas in the latter case it was almost doubled at 7.8/10. With one exception, these represented the lowest and highest cloud covers which occured during the growing season. The average

Figure 16

Linear Regression of Net Total Radiation on Absorbed Solar Radiation



value of Ω for the growing season was 0.17.

Further, Ω was computed for all of the cloudless days which occured during the growing season in a calculation directly comparable to that of Monteith and Szeicz. The linear regression of net radiation on short-wave balance gave a correlation of coefficient r of 0.97 and Ω was calculated at 0.24. When this value is compared to Table 17, it can be seen that for the year 1964 the forest cover at Mont St. Hilaire had a heating coefficient that was slightly greater than grass, sugar beet and spring wheat, but less than bare soil according to measurements made in England during various years.

(d) Energy Transformations

The net radiation, which represents the energy available at the earth's surface, must eventually be returned to the atmosphere. The mechanisms involved in this return must operate within a surface layer of considerable depth in the case of a forest. This creates unique microclimatic conditions which will be considered in Chapter VI. The present discussion of energy transformations applies to the average conditions which prevailed over the forest cover during 1964.

(i) Latent Heat Flux. The movement of energy in the form of latent heat occurs in the evaporation of water which is intercepted by the tree canopy, the evaporation of soil moisture, and the loss of water through transpiration. The energy which is used to evaporate intercepted water for each weekly period is shown in Table 16. With the limitations of the present study, the energy consumption for soil water evaporation and for transpiration can only be given for the total period of active growth.

By using a latent heat of vapourization of 590 cal cm^{-1} for a mean seasonal temperature of 15°C, the values which are given in Table 18 were calculated. These values are based on the hydrologic-balance measurements and assumptions which were presented in Chapter IV. The latent heat energy which was involved in the total evapotranspiration process

Table 18. Amount of latent heat energy used in the various components of the evapotranspiration process.

Energy used in evaporating intercepted water	5564 cal.
Energy used in evaporating soil water	1298 cal.
Energy used in transpiration	19087 cal.
Total	25949 cal.

for the growing season made up 67 percent of the net radiational energy.

(ii) Soil heat flow and heat storage in the forest wood mass and forest air. No consistent measurements of any of these parameters were taken in the present study. Their quantitative significance can in part be estimated from other studies. Baumgartner (1956), in nine-day measuring period in a young spruce forest from the end of June through the first of July, found that the soil heat flow and the heat storage in the forest wood mass and forest air amounted in total to only 0.7 percent of the net radiation (i.e., 0.7 percent of the energy from net radiation was stored as heat in the biosphere or moved downward into the forest soil). Dzerdzeevskii (1963) measured the same quantities in a deciduous forest for the period when the leaves were just beginning to sprout in the spring, in mid-summer, for a period just before leaf-fall, and in an interval just after the leaves had fallen. His measurements showed soil heat flow and heat exchange in the biosphere as equalling 0.8, 1.3, 2.4 and 11.0 percent of the net radiational energy, respectively, for each period. It is interesting that for every period, Dzerdzeevskii's measurements showed a net heat gain by the biosphere and forest soil. This is suprising since the forest layer cannot store heat indefinitely. In the present study, only two sets of soil heat flow measurements were taken. These are presented in Table 19. The first set was gathered during the period after leaf-fall in the autumn of 1963. The second set covered a three-week period before leaf fall in 1964. During twelve days of the canopy-bare period in 1963, there was a net storage of heat in the soil which equalled 6.6 percent of the above-canopy net radiation. Before leaf fall in 1964 the soil was liberating heat over a three-week period in an amount equivalent to 7.0 percent of the net radiation. The first set of measurements compare reasonably well with those taken by Dzerdzeevskii under similar conditions (both show a net gain of heat by the soil after leaf-fall). The second group shows the forest soil losing heat where, during a similar period, the Russian experiments showed a heat gain.

It is necessary to apply reasonable values to the heat exchange in the forest wood mass and air, and the forest soil. The former exchanges will be small in comparison to the soil heat flow, and should approach zero over the whole growing season. The soil heat flow, according to Baumgartner and Dzerdzeevskii's measurements, also becomes small over the period of growth. However, the few measurements taken by the present author showed a negative soil heat flow of considerable magnitude which endured for over 3 weeks. Recognizing that the soil does heat up in the spring and does

Table 19. Daily measurements of net radiation and soil heat flow for 2 periods at Mont St. Hilaire. (all values in ly day $^{-1}$).

1963		
Date	Rn	S
Oct. 8	200	-25.7
9	249	-6.6
15	168	22.2
16	145	15.6
17	205	10.2
18	166	22.8
19	122	10.2
21	188	-28.8
22	222	10.8
23	187	7.2
24	140	-7.2
25	91	-17.4
Totals	2083	138

S/Rn (%) = 6.6

1964		
Date	Rn	S
Aug. 26	280	-9.6
27	308	-7.2
28	212	-8.4
29	29	-4.2
30	328	-4.2
31	185	4.8
Sept. 1	193	-30.0
2	192	-16.8
3	120	-18.6
4	9	-10.2
5	113	-21.6
6	118	-19.8
7	242	-15.0
8	91	-12.6
9	274	-15.6
10	123	-3.6
11	14	-28.0
12	175	-57.7
19	162	-6.6
20	152	1.2
21	132	3.0
22	210	6.0
23	193	14.4
Totals	3855	-260.3

S/Rn (%) = -7.0

cool down in the fall, the magnitude of the energy storage within the soil should fall within $\pm$ 5 percent of the energy of net radiation.

The limited measurements of soil heat flow, which are shown in Table 19, suggest certain patterns. Toward the end of the growing season when the forest was still in the full-leaf state, and the soil was shaded from the direct solar radiation, the heat flow was directed primarily upward from the soil into the atmosphere. However, this flow was reversed after leaf-fall, at least for a time. Such a feature was probably due to the increased penetration of solar radiation through the open canopy to the forest floor.

(iii) Photosynthesis. The consumption of energy for the production of plant food and the construction of dry matter represents energy which is not returned to the atmosphere for long periods. It may represent appreciable energy amounts for hourly intervals during periods of most active plant growth, as was observed by Lemon et al (1957) for corn, but over the whole growing season will become unimportant in comparision to other energy amounts. An example of a calculation of yearly energy consumption for photosynthesis in a forest was provided by Miller (1955). He took volume, yield, and stand tables for some of the principal timber species of British Columbia, and calculated that 600 trees/acre of average trunk diameter 5.5 inches/tree would give 3000 ft^3 of wood per acre. With an average annual increment of wood of 0.01 gm cm^{-2} of forest land, and a heat of combustion of 5000 cal gm^{-1}, the yearly increment in dry weight would use 50 cal cm^{-2} of energy. If one assumes a similar amount of energy is used for forest photosynthesis at Mont St. Hilaire, it would only use 0.13 percent of the net radiation.

(iv) Sensible heat flux. The other large energy transformation involves the eddy flux of sensible heat. For mid-latitude forests at least, it appears that there is always some eddy heat flow, and that potential evapotranspiration does not equal the energy available from net radiation. As a residual quantity in the present study, the return of sensible heat to the atmosphere involved 13127 $\pm$ 1974 calories for the period of active growth, a quantity which constitutes 33 $\pm$ 5 percent of the net radiational energy. Since the sensible heat flux has been calculated as a residual term, and the soil heat flow and heat storage in the biosphere are terms which can vary in quantity from plus to minus 5 percent of the net radiation, it is necessary to set a range within which the sensible heat flux can vary. If the soil heat flow has a net positive value during the growing season (i.e., the soil increases its heat storage), then the sensible heat flux must be decreased by an equivalent amount. In contrast, a net negative soil heat flow during the growing season would yield extra heat energy to the forest layer, and the sensible heat flux would be increased.

(e) Comparison of Results

The results of the present study are compared to those of Baumgartner and Dzerdzeevskii in Table 20. The calculations of the latent and sensible heat fluxes at Mont St. Hilaire for the full growing season, and those at Munich for the mid-summer period correspond to one another within 10 percent. If one takes an average of the first 3 sets of measurements at Zagorsk (i.e., those covering various stages of the growing season), and

Table 20. Comparison of energy-balances for forest covers as calculated for a young spruce forest at Munich, a mixed deciduous and coniferous forest at Zagorsk, and for Mont St. Hilaire.

	Mont St. Hilaire	Munich
Investigator	Rouse	Baumgartner
Latitude	45°N	47°N
Forest type	Mainly deciduous	Young spruce
Net radiation	100%	100%
Latent heat flux	67%	71%
Sensible heat flux	33 ± 5%	28%
Heat flow in soil and biosphere	±5%	1%

	Zagorsk			
Investigator	Dzerdzeevskii			
Latitude	55°N			
Forest type	Mixed			
	Beginning of Leaf sprout	Mid-summer	Beginning of leaf-fall	After leaf-fall
Net radiation	100%	100%	100%	100%
Latent heat flux	71.5% *	74.0% *	67.5% *	37.4% *
Sensible heat flux	28.0%	26.5%	38.8%	56.0%
Heat flow in soil and biosphere	0.8%	1.3%	2.4%	11.0%

* Percentage do not add up to 100% due to measurement error or to advective heat fluxes.

compares them to those of Mont St. Hilaire, the results also agree within 10 percent.

It would be unrealistic to presume that an average condition for the heat-balance of a forest cover in middle latitudes involves latent and sensible heat fluxes, which utilize two-thirds and one-third of the energy available from net radiation, respectively. However, it is reasonable to suggest that, because a deciduous forest returns large amounts of water to the atmosphere, such a large flux is a necessary ingredient in the life processes of hygrophilous deciduous trees in middle latitudes. As such, these forests will be established only where sufficient water is available under average conditions. In other words, forests need abundant water in order to grown. As a corollary, under average conditions in a mid-latitude deciduous forest, the latent heat flux during the growing season will always use a large part of the available energy from net radiation.

(f) Discussion

The energy-balance, as it has been presented in this chapter, is limited to the movement of energy in the vertical direction. It does not consider the advective fluxes of sensible or latent heat. Such advective fluxes have been omitted, not because the author deems them unimportant, but because of the impossibility of measuring them during the present study. It is likely that importation of energy is significant both on the flanks of the mountain and on the slopes leading away from Lac Hertel. Moreover, it is likely to have more influence on a site such as Mont St. Hilaire than would be expected on a forest site covering a relatively flat area.

Another quantity which has been omitted is the water and energy exchange due to dew condensation. There is no way in which dew fall can be measured readily in a forested area, except as an aerodynamic downward vapour flux. Its importance to the energy-balance over long periods of time will not be significant because the latent heat energy which is liberated during night-time condensation will be balanced by the energy used for evaporation the following day.

(g) Summary

Measurements and estimations of the heat-balance over the forest during the growing season are presented. Global radiation was measured, and a mean albedo of 0.18 for forest was used, the difference giving the absorbed solar radiation. The long-wave balance was determined from the difference between the net radiation and the absorbed radiation. The various radiation measurements, as well as the energy used to evaporate the water intercepted by the tree canopy, were all plotted for each of the 22 weeks of the growing season.

By using a modified form of a calculation presented by Monteith and Szeicz, a Ω value was derived which showed the relative magnitude of the absorbed solar radiation and the long-wave radiation balance during various intervals of the growing season. The value of Ω is influenced by the surface temperatures at the forest canopy and the cloud cover.

A study of the processes into which the energy of net radiation entered was pursued. The latent heat flux used approximately two-thirds of the net radiational energy during the growing season, and the sensible heat flux about one-third. The net heat storage in the biosphere and in photosynthesis is negligible over the growing season, but soil heat flow may represent $\pm$ 5 percent of the net radiational energy according to the few measurements which were taken.

The energy-balance, which was calculated for Mont St. Hilaire, agreed closely with two similar studies of the heat-balance over a forest in Germany and Russia. The magnitudes of the energy transformations which were calculated probably resemble closely the mean conditions which persist over a middle-latitude deciduous forest during the growing season.

Chapter VI The Forest as an Active Surface Layer

A. Net Radiation Profiles

An intensive survey of variations in net radiation with height and with area in the forest has yielded most interesting results which are summarized in Figs. 17, 19 and 20 and in Appendices E (i) and E (ii).

(a) Experimental Design

The apparatus for mounting the net radiometers is described in Chapter III. Data were collected between June 5 and June 23 for the North Slope profiles, and between June 24 and July 16 for those on the South Slope. The measurements thus evenly straddle the period of maximum solar altitude and daylength. For each site, the experimental design was identical. One net radiometer was used to record continuously the net radiation at a fixed spot above the forest crown (Rn_{ac}). The remaining three radiometers were mounted initially at a height of 0.3 meters above the forest floor, after which they were raised to successive levels of 1,2,4, 6,8,10, 11,13, and 14 meters, the last height being above the general tree-top level. The time during which the sensors were left at any one level depended on the weather conditions, it being necessary to record, at each level, a full day's net radiation (Rn_{uc}) under relatively clear sky conditions.

(b) Presentation of Results

Fig. 18 shows the morphological structure of the forest on north and south slopes. It was plotted by employing a modified form of the widely-used method first proposed by Dansereau (1951). Whereas Dansereau's technique shows tree heights in groups which include an increasing height range as one progresses upward toward the forest crown, the present writer has mapped and plotted the vertical structure in categories which have a constant height range. Thus the vertical scale in Fig. 18 is linear. The total breadth of the tree crowns in a particular layer represents the portion of the ground, in percent, which is covered by the trees at each height. Thus for the North Slope diagram, trees having their crowns between 7 and 13 meters cover approximately 75 percent of the ground beneath. Triangular and circular-shaped crowns represent coniferous and deciduous trees respectively, whereas the trapezium-shaped structures which hug the ground correspond to low bushes and shrubs. Fig. 17 which shows the diurnal pattern of isopleths of net radiation is used as an overlay on Fig. 18. This gives a simultaneous view of a climatological parameter and forest site to which it applies. A similar overlay of isopleths of solar radiation is used on Fig. 18, later in this chapter.

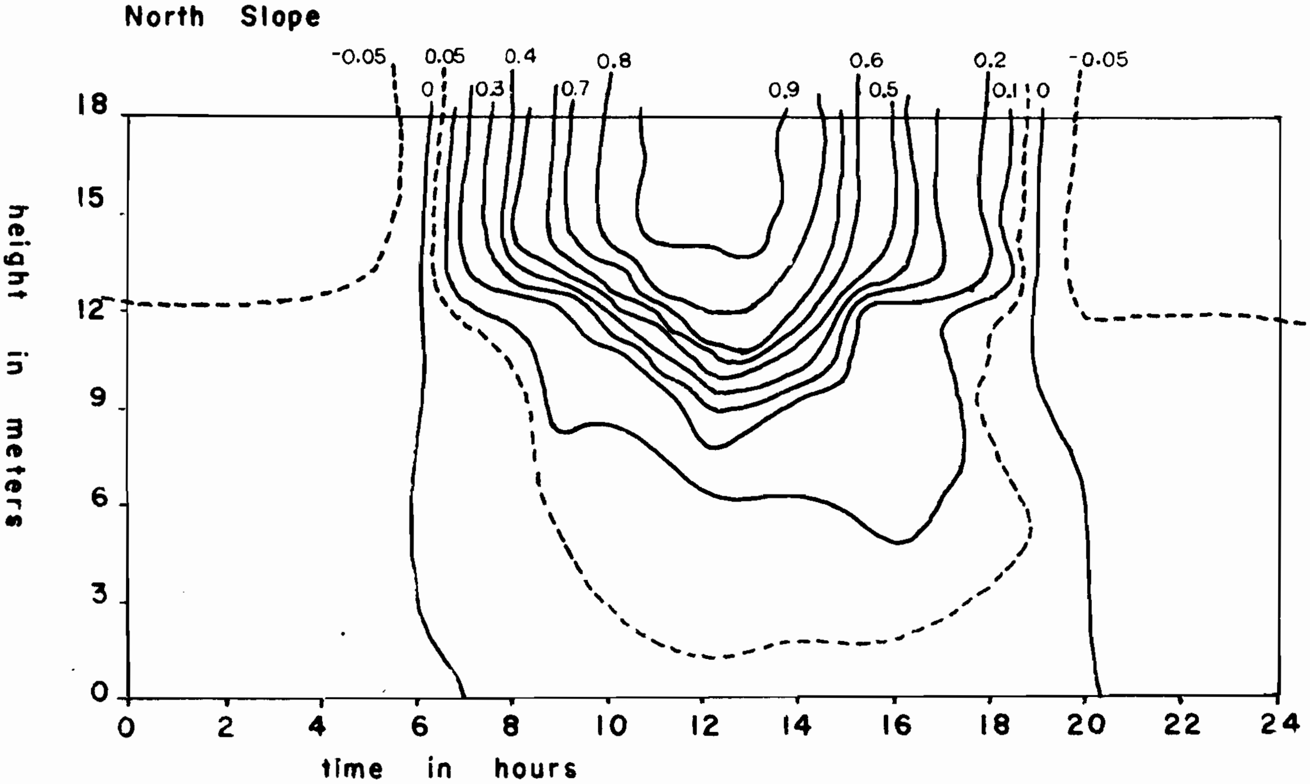
The methods of data presentation in Figs. 17 and 19 differ. For the former, the above-canopy values of net radiation were averaged for the total recording periods of 18 days on the North Slope and 22 days on the South Slope. These average values were plotted at the 14 M level. The measurements of the three net radiometers at lower heights in the forest were averaged for the full recording period at each height. This means that two or three day averages for the beneath-canopy measurements are plotted with the longer period averages of the above-canopy net radiation. Such a technique lessens the precision of the isopleths which are shown in Fig. 17. Also, because of different intensities of radiation during the recording periods, the average above-canopy net radiation on the North Slope was greater than on the South Slope by a factor of 1.30. As a result, the South Slope values have all been increased by a constant of 1.30 so that they can be compared readily to their North Slope counterparts. Fig. 19, in contrast, plots the average daytime ratio of the net radiation at a height within the forest to the net radiation above the forest. Both measurements are taken over the same interval of time. Fig. 19 is precise but does not show variations with time.

(c) Results

A comparison between the North and South Slope distribution of net radiation gives some interesting insights into topographic variations in radiation. Higher values of daytime net radiation extend to greater depths in the woodland of the South Slope. At a given height, a certain value of net radiation is reached earlier in the morning, and extends later into the evening on the South Slope than on the North. Nocturnal radiation on the South Slope is greater than that on the North. The differences in nocturnal radiation are not as great as Fig. 17 might appear to indicate. Maximum values on the South Slope reached -0.11 ly min^{-1} whereas on the North Slope they fell just below the isopleth value at -0.09. The most noteworthy result, in comparing the two sites, is the similarity in their net radiation regimes. This result is expected from the discussion in Chapter II concerning the topographic influence on the radiation over a forest during the summer solstice. Such a similarity is helped by the particular character of the vegetation. As is apparent in Fig. 18 the North Slope has a slightly thinner canopy cover, particularly

Figure 17

Diurnal pattern of net radiation above and within the forest canopy (all values in ly. min.$^{-1}$)



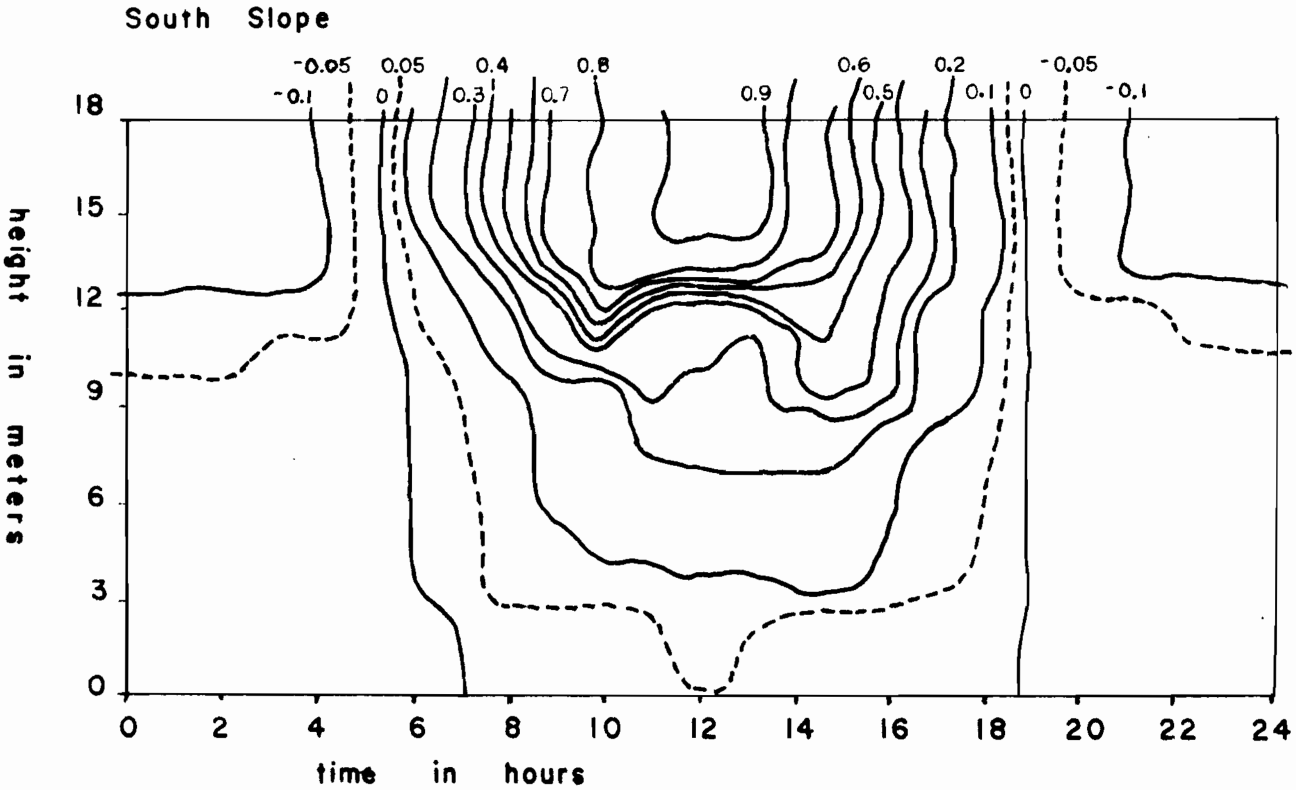


Figure 18

Morphological Structure of the Forest at the North Slope and the South Slope Sites

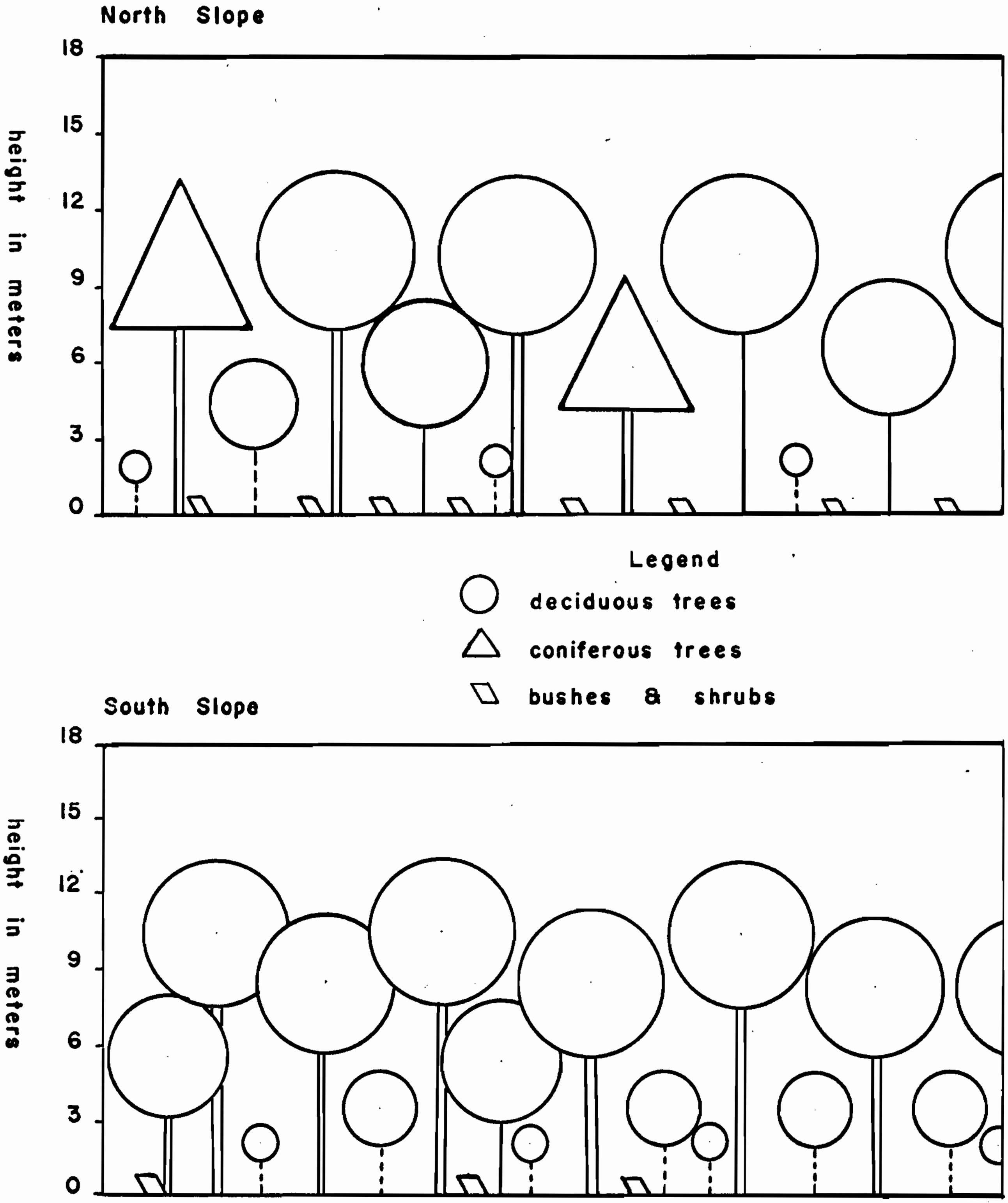


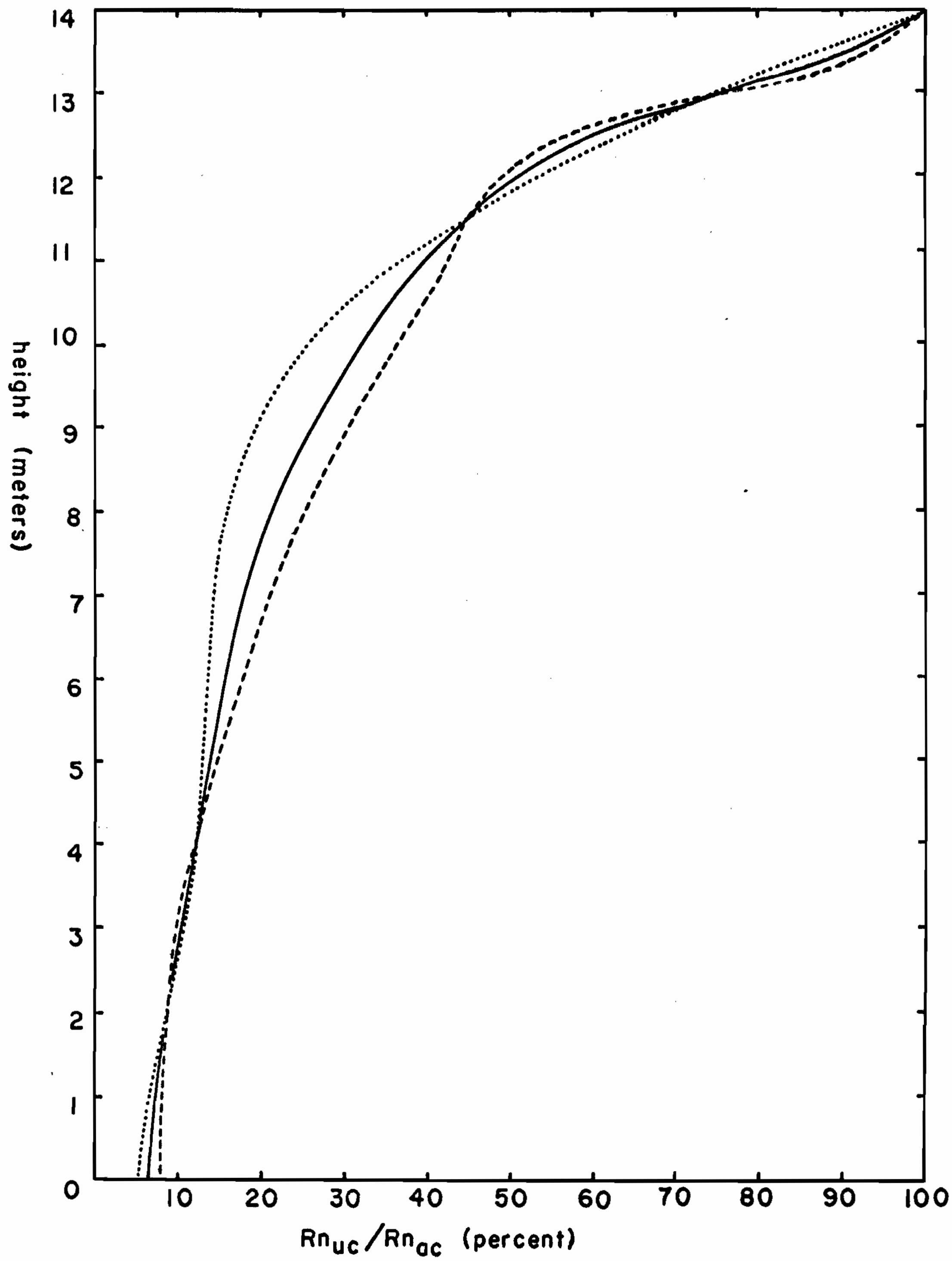
Figure 19

Mean Daytime Values of Rn_{uc}/Rn_{ac} at Different Heights Above the Forest Floor for North and South Slopes

.... North Slope

----- South Slope

—— Average Trend for Both Slopes



at heights closer to the ground. This serves to allow a greater relative penetration of sunlight than does the heavier cover on the South Slope but in turn, the thinner vegetative cover near the ground may be the result of less solar radiation at the lower levels.

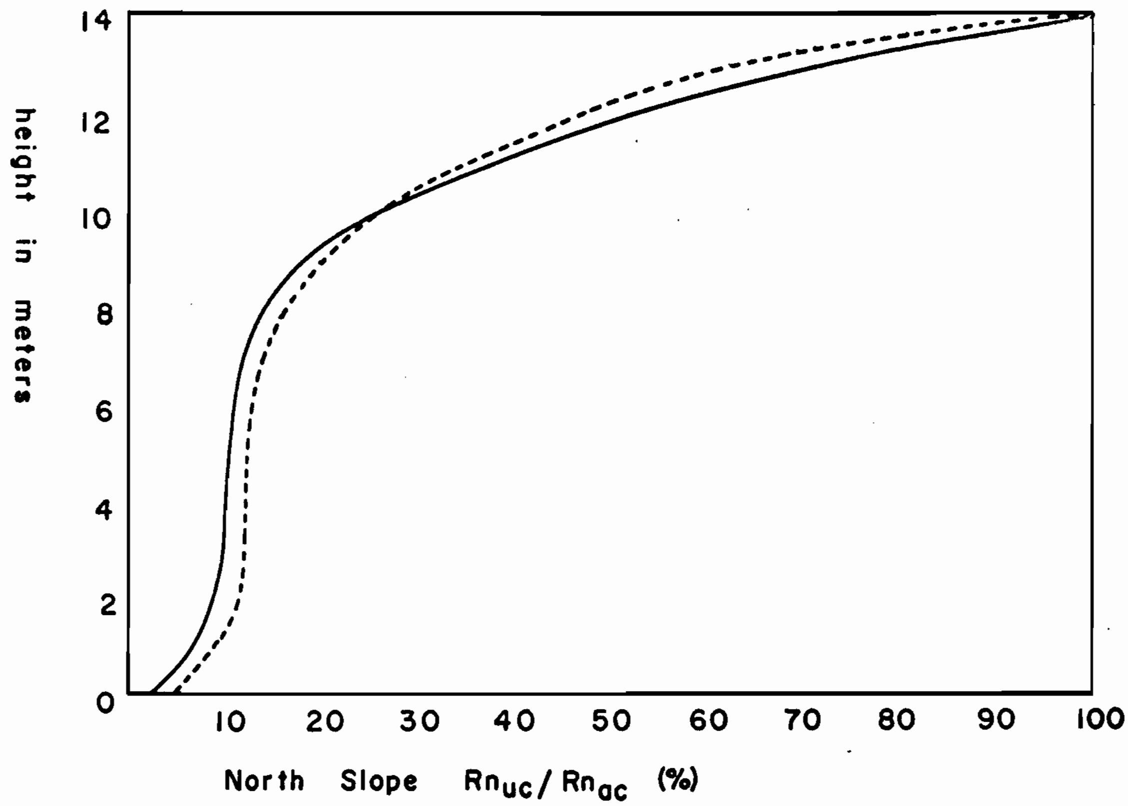
In Figure 19, the higher net radiation between the 4 and 11 M heights on the South Slope is readily seen. More interesting is the pattern achieved when the mean trend for net radiation change with height is drawn (dotted line). There is a well-defined change in slope around 11 meters. Above this point the net radiation increases rapidly to the forest top, whereas below it there is a more gradual decrease to the forest floor. In the top three meters of forest crown (11 to 14 M), there is a concentration of 60 percent of the net radiation. Such a result means that 60 percent of the energy available to do work is concentrated initially in this shallow zone. By referring back to Fig. 18 it can be seen that the most active layer in the whole forest is found in the upper one-half of the crown zone of the tallest trees.

Fig. 20 shows the effects of direct and diffuse solar radiation on the net radiation for each slope. In order to divide the solar radiation into direct and diffuse, the radiation was considered to be primarily direct on days for which the cloud cover averaged less than 4/10, and mainly diffuse on days when the cloud cover was greater than 6/10. The actual division resulted in an average cloud cover of 2.5/10, and 8.0/10 for days of direct and diffuse radiation respectively. Figure 20 shows that daytime net radiation was higher at lower heights on both slopes during periods of mainly diffuse solar radiation. These differences are most pronounced for the South Slope between the 2 and 11 meter levels.

Figure 20

Rn_{uc}/Rn_{ac} at Different Heights Above the Forest Floor for North and South Slopes Under Conditions of Diffuse and Direct Solar Radiation

------- diffuse
——— direct



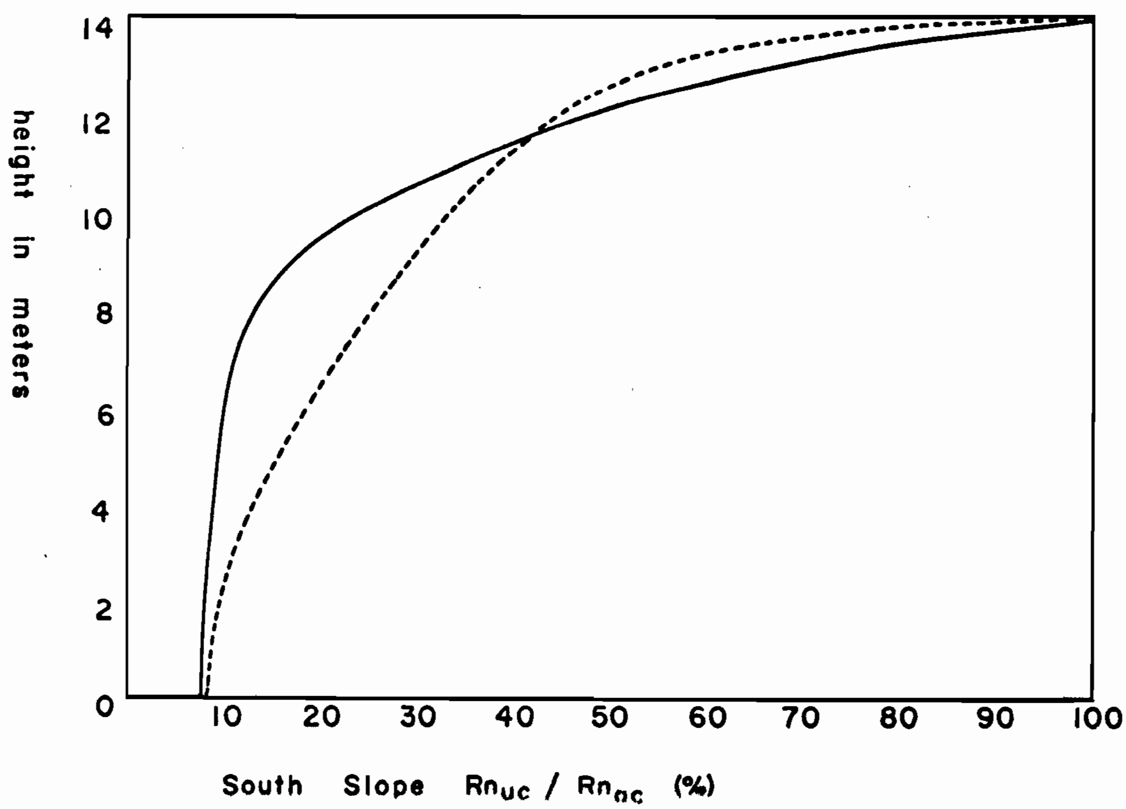


Table 21 shows the net radiation at the 1 M level relative to the above-canopy net radiation for different hours of the day. It represents an average condition for both North and South slopes. The table shows that there is little variation in the ratio until mid-afternoon when the net radiation at 1 M increases substantially relative to that recorded above the forest. This is caused by differences between tree canopy temperatures and ground temperatures which become greatest in the middle afternoon when the downward infrared radiation flux reaches its highest value relative to the upward long-wave radiation from the ground. Geiger (1965) plots the air temperatures for the canopy area of an oak-beech forest and a point 3 M above the forest floor. In Geiger's study, the maximum temperature differences occured in the mid-afternoon, when the canopy temperatures were 8 to 9^{o}C higher than those at the 3 M level. This compared to a temperature difference of 4^{o}C at 0900 hours.

Table 21 Rn_{uc}/Rn_{ac} at the 1 M level for various times of day.

(Rn_{uc} is the beneath-canopy net radiation and Rn_{ac} the above-canopy net radiation).

Hour	7-8	8-9	9-10	10-11	11-12	12-13	13-14	14-15	15-16	16-17
Rn_{uc}/Rn_{ac}	0.06	0.06	0.07	0.07	0.07	0.06	0.08	0.11	0.14	0.15

It was noted in Chapter II that various investigators have pointed to the large variations in solar radiation which can be expected from place to place beneath the forest canopy. The extent to which such variations are evident in the net radiation which is recorded beneath the canopy is shown in Table 22. This table is based on a sample of all the hourly

Table 22. Spatial variations of net radiation at levels below 8 M during the daytime period.

Location	Number of Places Sampled	Hours of Observation	M Mean Net Radiation ($ly\ hr^{-1}$)	MD Mean Deviation ($ly\ hr^{-1}$)	MD/M
North Slope (Radiation mainly diffuse)	18	42	3.72	0.86	0.23
North Slope (Radiation mainly direct)	24	52	8.88	2.58	0.29
South Slope (direct and diffuse radiation)	18	34	2.76	0.66	0.24

values of net radiation recorded by the three net radiometers for all their positions from the ground to 8 M. Above-canopy direct and diffuse solar radiation was divided on the basis of less than 4/10 cloud cover and greater than 6/10 cloud cover, respectively. For the South Slope, there were only 12 hours of measurements during periods of direct, and 22 hours during periods of diffuse radiation, insufficient to provide a reliable sample, so no division was made. Table 22 shows that there is a slightly greater spatial variation of net radiation for direct than for diffuse radiation on the North Slope. There is little difference between the variations on North and South slopes.

The spatial variation of net radiation in the canopy is influenced by the solar altitude. This is seen in Table 23, which shows values of MD/M

Table 23. Spatial variations in net radiation at levels below 8 M during different hours of the day.

Hours	7-8	8-9	9-10	10-11	11-12	12-13	13-14	14-15	15-16	16-17	17-18
MD/M	0.18	0.23	0.22	0.28	0.30	0.28	0.20	0.19	0.19	0.16	0.16

for all hours of the day as an average for both slopes, and for conditions of average cloud cover. The variation in net radiation is least during the early morning and late evening hours. It climbs steadily to the noon hour period. The lower variations in the morning and evening result from the low solar altitude, and the longer path which the sunlight must follow through the forest foliage. Such conditions serve to spread the incoming radiation more evenly in space.

(d) Results of Other Investigations

A study by Trapp (1938), which is quoted in Geiger (1965), showed the penetration of light within a stand of beeches. Trapp's measurements showed a 90 percent decrease in the light penetration within the crown area for sunny periods. On cloudy days, however, the decrease in sunlight was more gradual, and more light penetrated to greater depths in the forest. These observations agree in general with the graphs plotted in Fig. 20. Baumgartner measured net radiation within a spruce woodlot of young evenly-spaced trees of a maximum height of 6.5 meters. The measurements were taken simultaneously at heights of 0.2, 1.7, 3.0, 4.1, 6.3 and 10 meters of forest there was a 50 percent decrease in net radiation, but in the top 3 meters this had risen to 95 percent. Thus for spruce trees, the crown-space or layer where the crown flares out at the base, is an important active layer. On the basis of these results, one can assign a shallower layer of maximum energy exchange to spruce forest than is found in deciduous forests of the type investigated on Mont St. Hilaire. Conversely, if the results of Baumgartner's work and the results contained in this study are representative, net radiation is higher at greater depths in a deciduous forest than in a spruce woods. Another interesting feature of Baumgartner's measurements is that, in the bottom 2 M of forest, the net radiation was always positive except for a couple of hours in the early morning. During the present study, positive net radiation at night was never recorded except for the hour at sundown, though below 6 M it always fell to zero.

Vezina (1962, 1965), in a study of coniferous forest, found that the penetration of solar radiation within the canopy, relative to that measured in the open, was always greater on cloudy days due to the increased ratio of diffuse to direct solar radiation. Similar results were achieved by Gast (1930) who found, during measurements of the radiation penetration of a white pine stand in Massachusetts, that the percentage transmission of radiation through the canopy lessened as the intensity of radiation above the forest increased. This, he concluded, was because skylight is more effective in penetrating the forest canopy than is direct radiation. Tanner, Petterson and Love (1960) derived the ratios of net radiation beneath the canopy to that above for cloudy and for clear periods for a corn field. For east-west orientated rows, the ratio was 27 percent greater under cloudy conditions than under clear skies, but for north-south rows there was no difference. Miller (1955) summarized various measurements of light penetration through the canopy according to stand density. The penetration ranged from 7 percent in dense forest, to 95 percent in parkland. Wide ranges for the penetration of light in different deciduous stands are given by Geiger (1965). It is clear, from the summaries of Miller and Geiger, that the forest at Mont St. Hilaire is dense, and that the penetration of radiation is small. Ovington and Madgwick (1955) noted that average measurements of light which penetrates through the canopy tend to mask place-to-place differences, which may vary widely. They did not express such differences quantitatively.

B. Solar Radiation Profiles

The height and time variations of solar radiation in the forest were measured in the same locations and at the same time as net radiation. Measuring heights were at 11, 8, 4, 2, 1 and 0.3 meters. The results are presented in the same manner as the net radiation, so Fig. 21 is directly comparable to Fig. 17, and Fig. 22 to Fig. 19.

There is a larger differential in the amounts of solar radiation penetrating the forest canopy on South and North Slopes than was the case in net radiation. A comparison of North Slope and South Slope values for the 6 M level at 1200 hours (Fig. 21) shows that on the South Slope the radiation is 0.30 ly min^{-1}, whereas on the North Slope it is less than 0.20 ly min^{-1}. Moreover, at ground level on the South Slope, an incoming radiation of 0.05 ly min^{-1} or more is received for almost 9 hours during the day, whereas on the

North Slope the period is reduced to 5 hours.

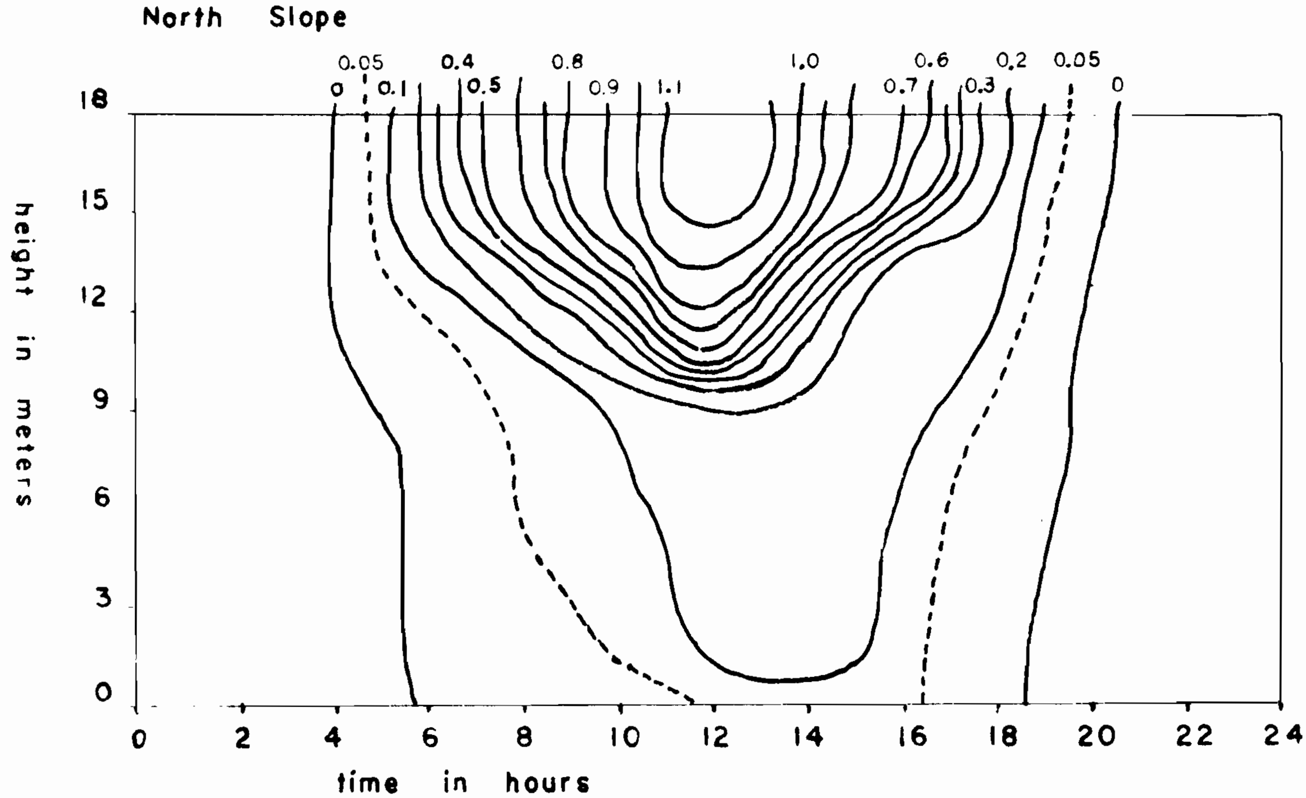
By following the mean curve for the two slopes in Fig. 22 it can be seen that there is, on an average, a 55 percent depletion of solar radiation in the top 3 meters of the forest canopy. For the North Slope, the depletion is greater at 60 percent, and it is lesser on the South Slope at 50 percent. The depletion of solar radiation is virtually equal to the rate of decrease of net radiation at all levels within the forest canopy.

The smoother isopleths in Fig. 21, and more regular curve in Figure 22, compared to Figs. 17 and 19, are a result of fewer measurements at higher levels in the forest due to the physical difficulties of mounting the actinograph. No observations were taken above 11 M. The two diagrams of solar radiation thus represent the actual situation less accurately than those showing the net radiation regime.

C. Solar and Infrared Radiation at Different Levels in the Forest

By choosing different levels within the forest and extracting the short-wave and long-wave components of net radiation, one can determine the relative importance of the various radiation exchanges above and within the forest layer. In order to do this, it is necessary to know the albedo beneath the forest. Above the 9 M level, where the forest canopy is concentrated, the albedo is placed at 0.18 the same value as above the forest. Beneath this level, however, it falls to 0.14 due to the influence of dark tree trunks and the dark ground surface. This latter value represents an average of several spot checks taken with an inverted solarimeter and using the simultaneous incoming solar radiation

Figure 21
Diurnal pattern of solar radiation above and within the forest canopy (all values in ly. min.$^{-1}$)



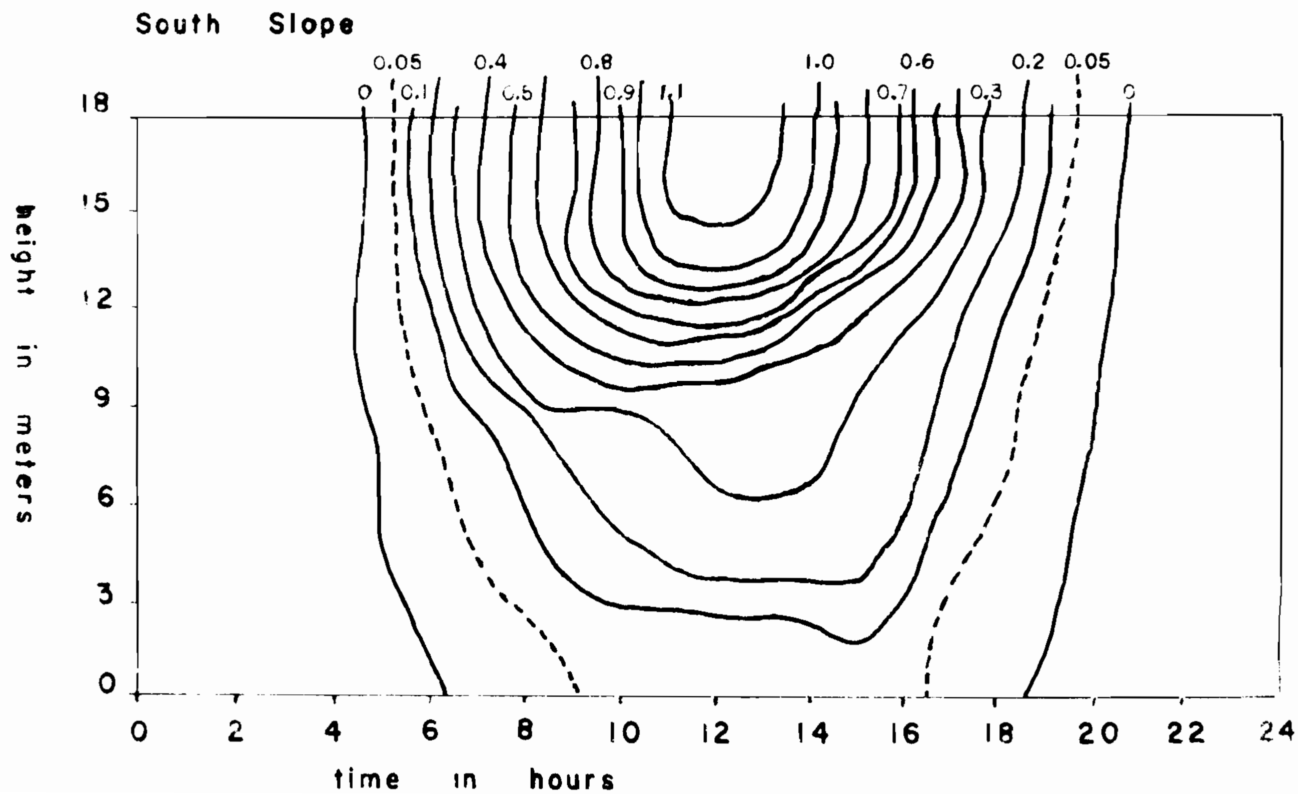


Figure 18

Morphological Structure of the Forest at the North Slope and the South Slope Sites

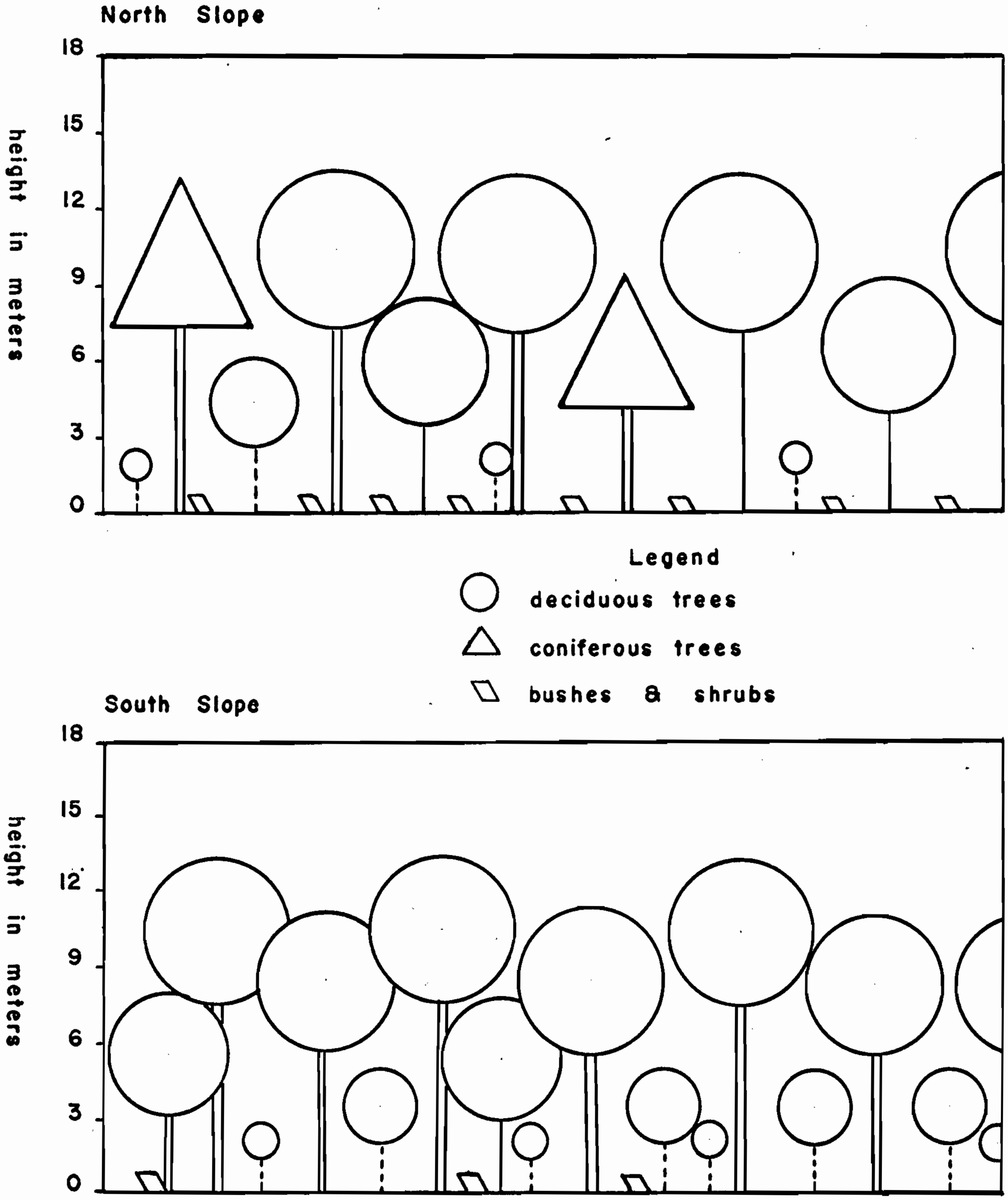


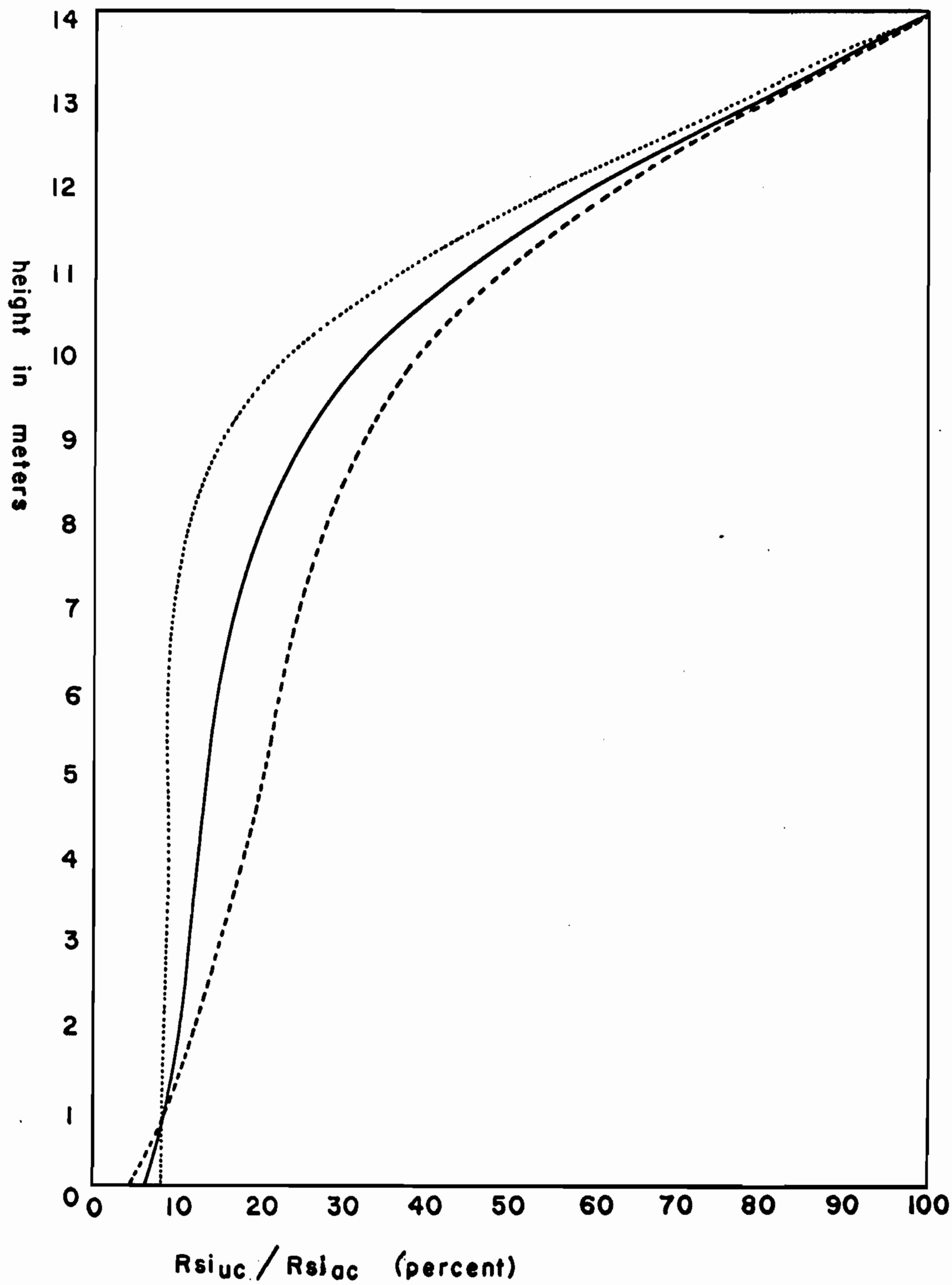
Figure 22

Rsi_{uc}/Rsi_{ac} at Different Heights Above the Forest Floor for North and South Slopes

·········· north slope

------ south slope

——— average trend for both slopes



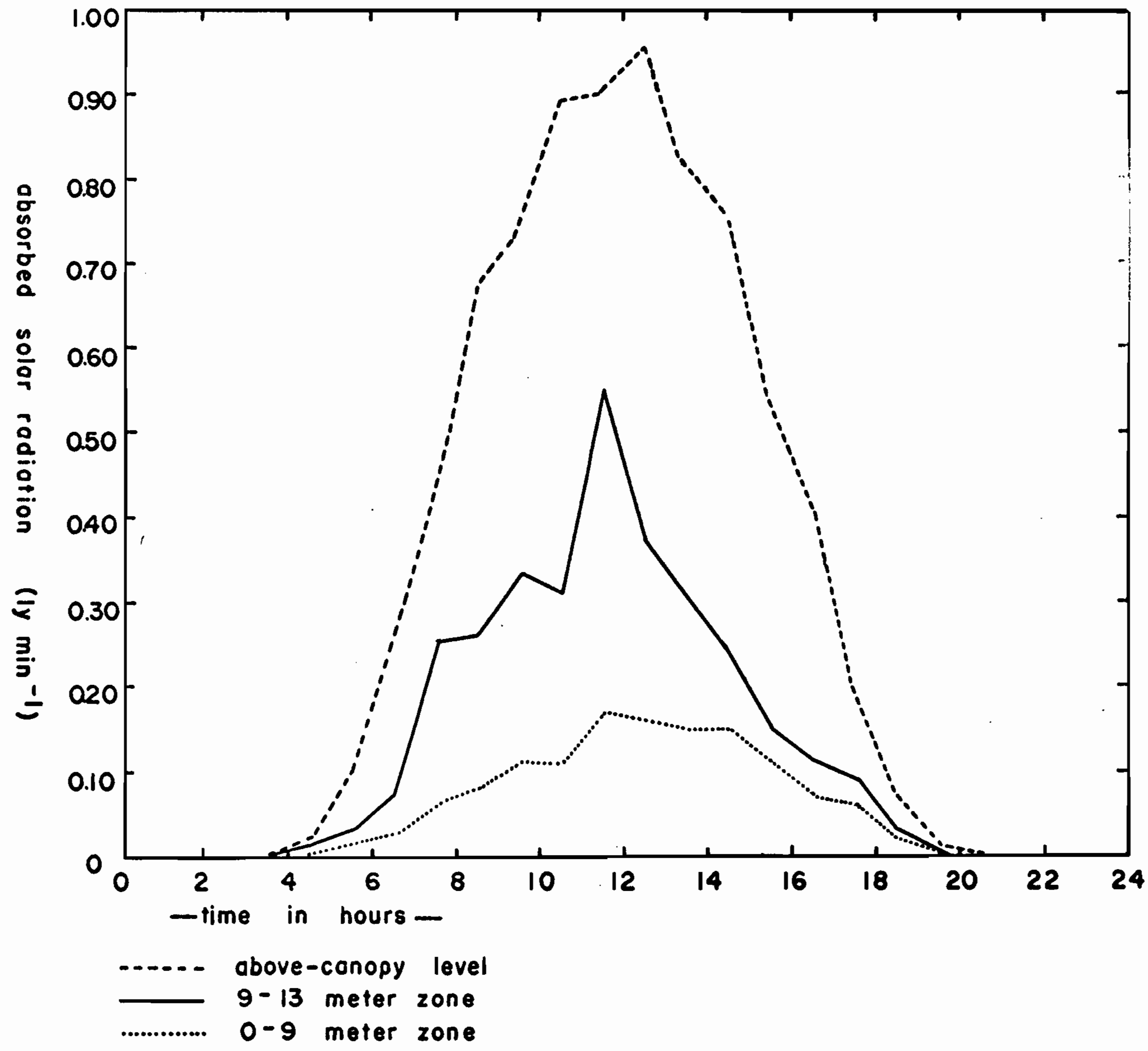
at the same locations. The radiation calculations are averaged for the North and South Slopes, and divided into heights from 0 to 9 M, from 9 to 14 M, and above the canopy. The results are tabulated in Appendix G, and are plotted in Fig. 23.

The absorbed solar radiation, plotted in Fig. 23, shows a regular decrease in amount with greater depth in the forest. For the long-wave balance, the curves have been smoothed by plotting the running means of hourly averages. This was necessary because of strong hourly fluctuations in the canopy zone. These fluctuations were caused by the shading of one of the net radiometers from incoming solar radiation, due to its particular position. This sufficiently decreased average net radiation to cause large hourly variations. An interesting period in the long-wave balance occurs between 1300 and 1800 hours, when the negative values above the canopy reach a peak, while those beneath achieve a minimum. This is the result of a temperature maximum at the canopy top, which creates a maximum black body radiation in both the upward and downward directions. Such a situation serves to increase the negative long-wave radiation above the canopy, and to decrease it within and beneath the crown. In fact, for two hours in the late afternoon (1600 to 1800 hours), the long-wave balance fell to zero. This is shown in Appendix G (i) but does not appear in Fig. 23 because of the smoothing. During these two hours the upward-directed infrared radiation was exactly balanced by that which was directed downward from the crown layer.

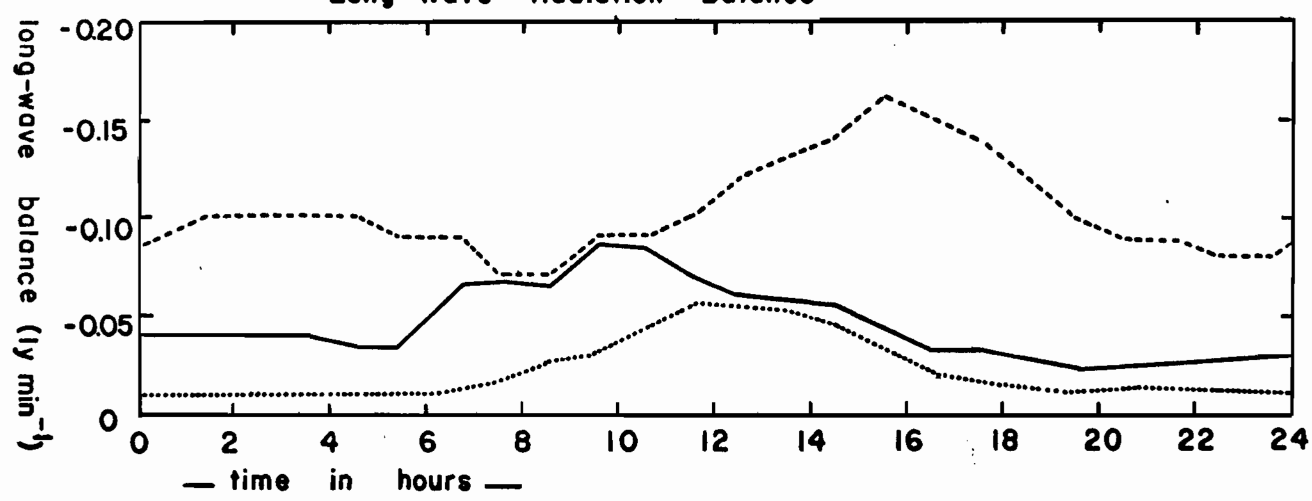
Figure 23

Radiant Energy Above, Within and Beneath the Forest Canopy

Absorbed Solar Radiation



Long-Wave Radiation Balance



D. Horizontal Variations in the Net Radiation Above the Forest.

It is important to know the horizontal variations in net radiation which occur above a forest cover in order to appreciate the characteristics of any single site. One would expect some variations due to albedo, though these should not be great where all the tree types are deciduous. One might further expect a variation due to shading, where a segment of forest is surrounding by higher trees.

For the investigations, two net radiometers were mounted about one meter above the tree canopy. One was fixed in position, while the other was free to rotate in a complete circle of 5 M radius. Measurements with the latter were made at every 45 degree interval through the circle, so that eight separate measurements were determined, each including a 24-hour period of relatively clear skies. The fixed radiometer was mounted above an oak tree. The mast for the variable radiometer rose above another oak and scanned oak, maple and ash trees during its complete rotation.

The results which were gathered between July 21 and August 6 are tabulated in Table 24. The ratio of Rn (variable) Rn (fixed) always fell within $\pm$ 7 percent of unity for both daytime and nightime readings and was apparently independant of tree species. As the figure of 7 percent includes the calibration and levelling errors of the instruments, it is concluded that for all practical purposes, horizontal variations in net radiation above the tree canopy are not important.

E. Variations in Above-canopy Net Radiation over North and South Slopes

Measurements of above-canopy net radiation over North and South forested slopes were taken for a period from May 6 to May 27.

Table 24. Horizontal variations in net radiation above the tree canopy. (all values in langleys per measuring period).

Main tree species	Position	Period	Fixed Rn	Variable Rn	Rn (variable) / Rn (fixed)
maple	south	day	859	886	1.03
		night	-102	-100	0.98
maple	south-west	day	671	685	1.02
		night	-95	-95	1.00
maple	west	day	471	475	1.01
		night	-86	-85	0.99
ash-maple	north-west	day	589	613	1.04
		night	-117	-121	1.04
maple-oak	north	day	456	433	0.95
		night	-28	-27	0.96
oak	north-east	day	535	497	0.93
		night	-47	-45	0.96
oak	east	day	633	625	0.99
		night	-86	-84	0.98
oak	south-east	day	597	625	1.05
		night	-63	-61	0.97

Because of a malfunctioning stepping switch and interrupted recording due to strong winds, only six full days of simultaneous readings on both slopes were suitable for analysis. The average values for each hour on each slope over the measuring period are tabulated in Appendix D.

Fig. 24 shows the daily progress of net radiation over North and South Slopes for three days with differing cloud amounts. On May 15 the period from midnight to noon was virtually cloudless, whereas the afternoon and evening hours showed a steadily increasing cloud cover. On May 7, it was the afternoon period during which skies were relatively clear. Finally, May 13 gives the net radiation for a primarily overcast day. Unfortunately, no data were recorded for a cloudless daily period. For the periods of heavy cloud cover the net radiation recorded on each slope was almost identical. Thus for May 13 the difference in daily net radiation between slopes was only 7 percent. Such a result is not unexpected since the clouds give diffuse solar radiation, and this shows no variation with topographic conditions. On May 15, the North Slope showed a higher net radiation during the clear morning hours, but on May 7 the South Slope gave the greatest net radiation during the clear afternoon period. In each case, the daily differences in net radiation between slopes were 11 percent and 13 percent.

More important than the small differences in net radiation, which appear in Fig. 25, are the similarities. The strong differences in the intensity of solar radiation which would be experienced on north and south-facing slopes with bare surfaces, even during the high sun

Figure 24

Diurnal Pattern of Net Radiation Above North and South Slopes

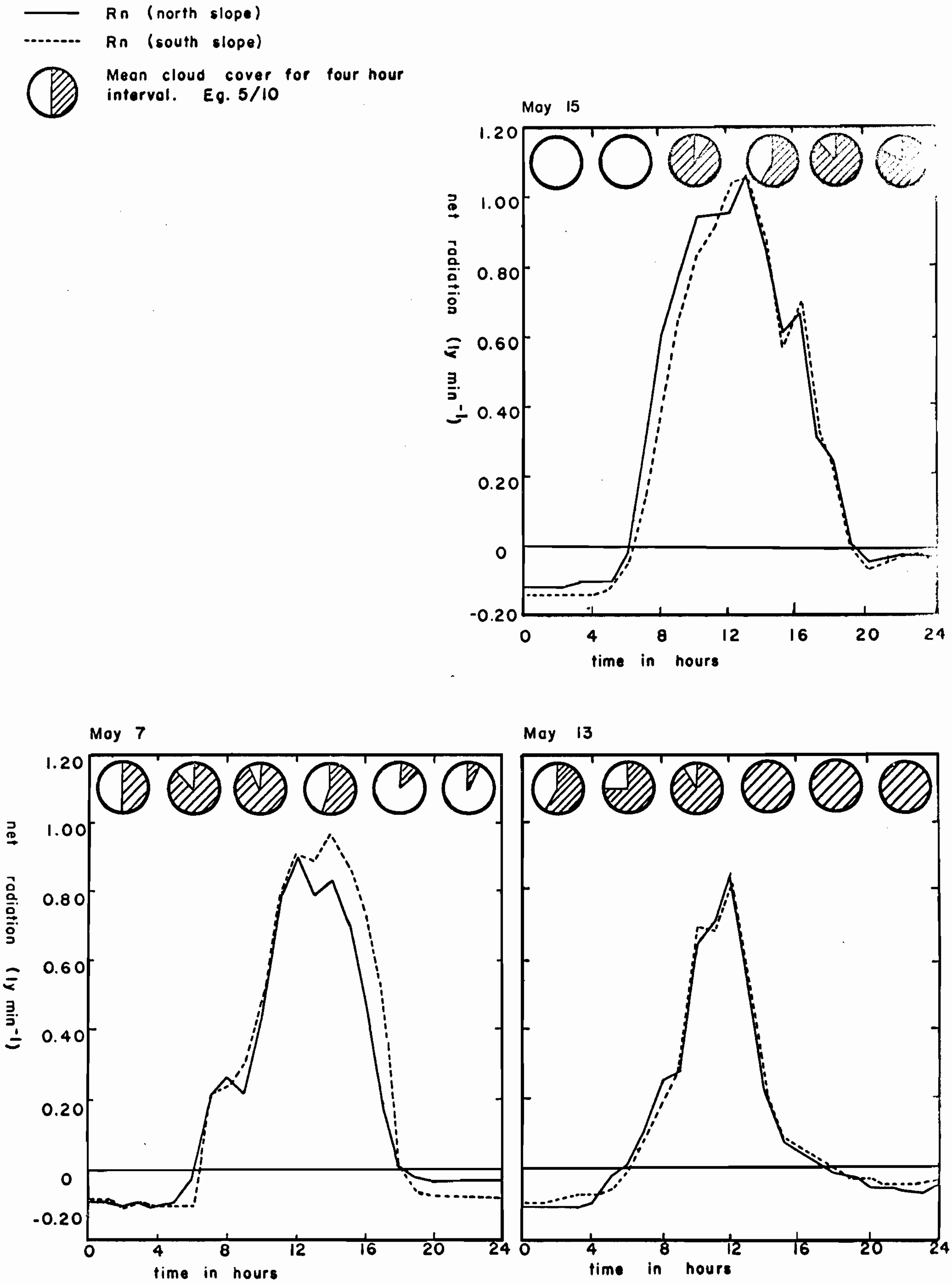
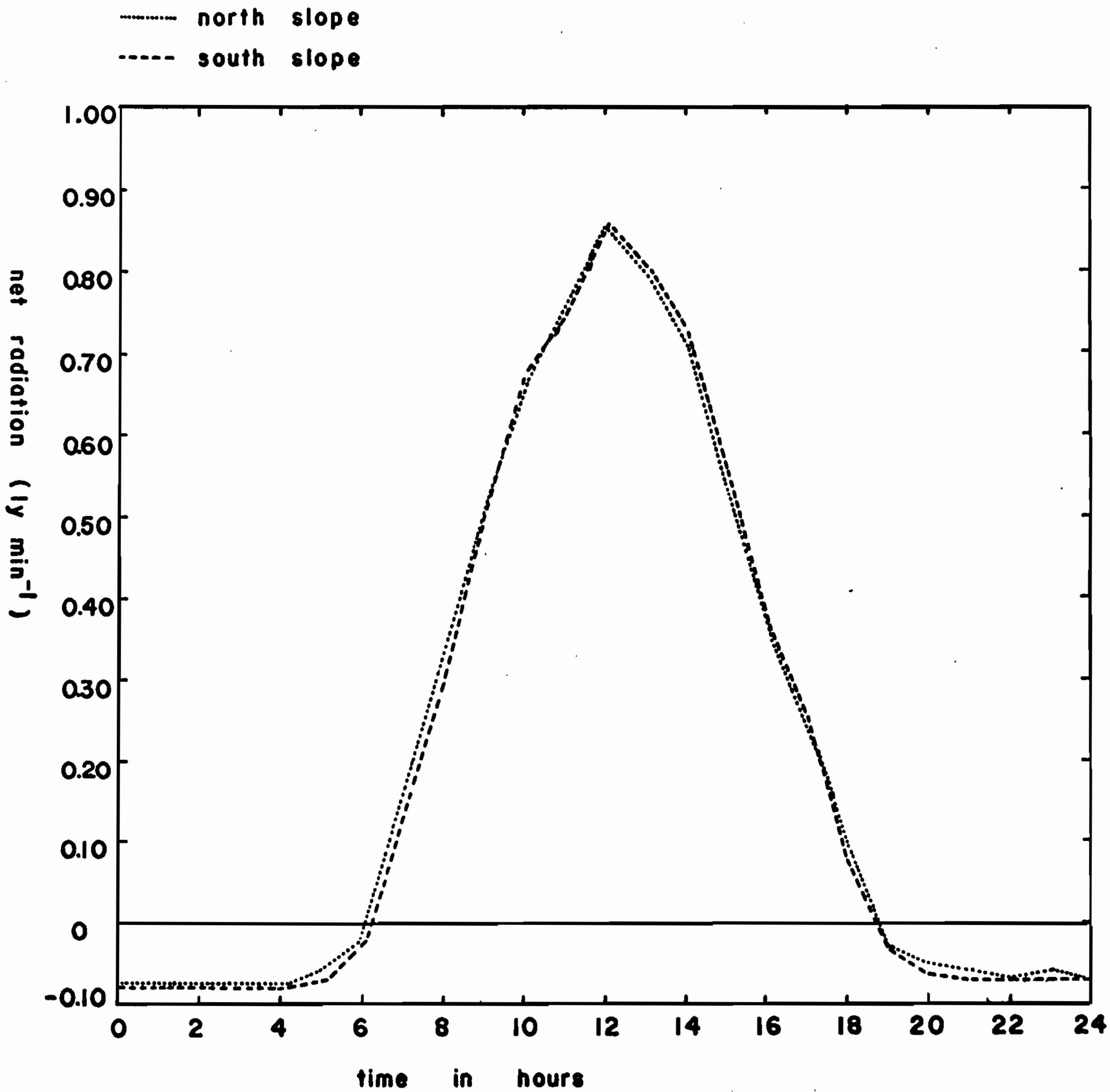


Figure 25

Average Above-Canopy Net Radiation Over North and South Slopes for a Six-Day Recording Period



period, are not evident in the net radiation over forest-covered hillsides. Indeed, for the six days of uninterrupted record, the total net radiation which was recorded over the North and South Slopes was 1907 and 1904 ly. respectively. Also as is shown in Fig. 25, the hourly measurements were closely coincident. These six days had a mean cloud cover of 7.0/10, a value which compares to a mean daily cloud cover of 6.3/10 for the 22 weeks of the growing season. It is evident that, for the growing season of 1964 at Mont St. Hilaire, major differences between the net radiation above forested North and South Slopes are not indicated. This is in contrast to differences with depth in the forest layer, where net radiation falls off more rapidly on the North Slope than it does on the South.

F. Net Radiation and Soil Heat Flow During the Changeover Period from Full-leaf to Leafless Forest Crown.

(a) Experimental Design

In the fall of 1963, a study of the influence of the full-leaf and leafless forest conditions on net radiation was carried out. The experimental area was a beech-maple forest which was moderately disturbed by man's activities. As such it displayed a somewhat more open character than the virgin forest sites which were investigated the following year. The results of the experiment are noteworthy to the extent that they show the forest still exerting a strong influence when in a leafless state. There was also an initial attempt to assess the influence of cloud cover on the depth to which radiation can penetrate into the forest layer. Finally, the vertical heat flow in the top layers of the forest floor was investigated under various conditions. The period of data collection involved a time span of a little over a month, between September 22 and October 26, 1963. This period covered the interval during which the foliage of the broadleaf forest changed to its death colour, and fell from the trees. Sky conditions showed a considerable period of little or no cloud. From the hourly records of cloud cover, based on the total analysed period, clear to scattered cloud conditions (0 to 5/10) were found for 65 percent of the time, with broken to overcast conditions (6/10 to 10/10) for the remaining 35 percent of the period.

(b) Results

(i) Net Radiation. During the period of full-leaf vegetation, the mean daytime ratio of net radiation at the one meter level to net radiation above the forest gave 0.09. This is slightly higher than was achieved in measurements in more natural forest stand. Under conditions of clear to scattered cloud, the ratio was little different from overcast conditions. Since the measurements beneath the canopy were taken at the one meter level, this result agrees in general with Fig. 20, which shows little difference between direct and diffuse conditions at the lowest levels in the forest.

For the leafless condition, the ratio rose to 0.41 as an average for all cloud conditions. This represents over a four-fold increase when compared to the full-leaf state. A higher ratio than this might be expected and it is evident that the tree trunks and branches are effective in reducing net radiation at ground level below half of that which is recorded above the living layer. The highest ratio, at 0.50, was achieved during the intervals of greatest solar altitude between 1100 hours and 1400 hours. Unlike the situation which existed during the full-leaf state, the ratio during cloudy weather reached 0.42 and fell to 0.36 for clearer sky periods.

(ii) Soil Heat Flow. The values of soil heat flow were low at all times, never exceeding 0.05 ly min^{-1} into the soil (positive) or -0.05 ly min^{-1} out of the soil. The measurements can be divided into two definite periods. The first period includes both the interval before the accumulation of a substantial leaf litter, and the interval after the layer of forest duff was removed artificially as part of the experiment. The second period covered the interval when the thick litter which had accumulated over the forest floor had a significant influence on the soil heat flow.

As is seen in Figure 26 (a), the soil heat flow during the first-mentioned stage showed a close dependance on net radiation under the forest canopy. A linear regression of soil heat flow on net radiation gave the regression formula $S = 0.250\ Rn - 0.015$, where S is soil heat flow and Rn is net radiation. The correlation coefficient was 0.90.

With the accumulation of leaf litter, the close association of soil heat movement with net radiation disappeared. A correlation coefficient of 0.66 was associated with the regression equation $S = 0.140\ Rn - 0.002$. The measurements are shown in Figure 26 (b). The newly-fallen leaf layer exhibits great insulating properties in preventing heat loss from the soil. At no time did the rate of energy loss exceed 0.02 ly min^{-1} when the forest floor was protected by a thick layer of leaf litter.

Figure 26 (a)

Linear Regression of Soil Heat Flow on Net Radiation for a Relatively Bare Forest Floor

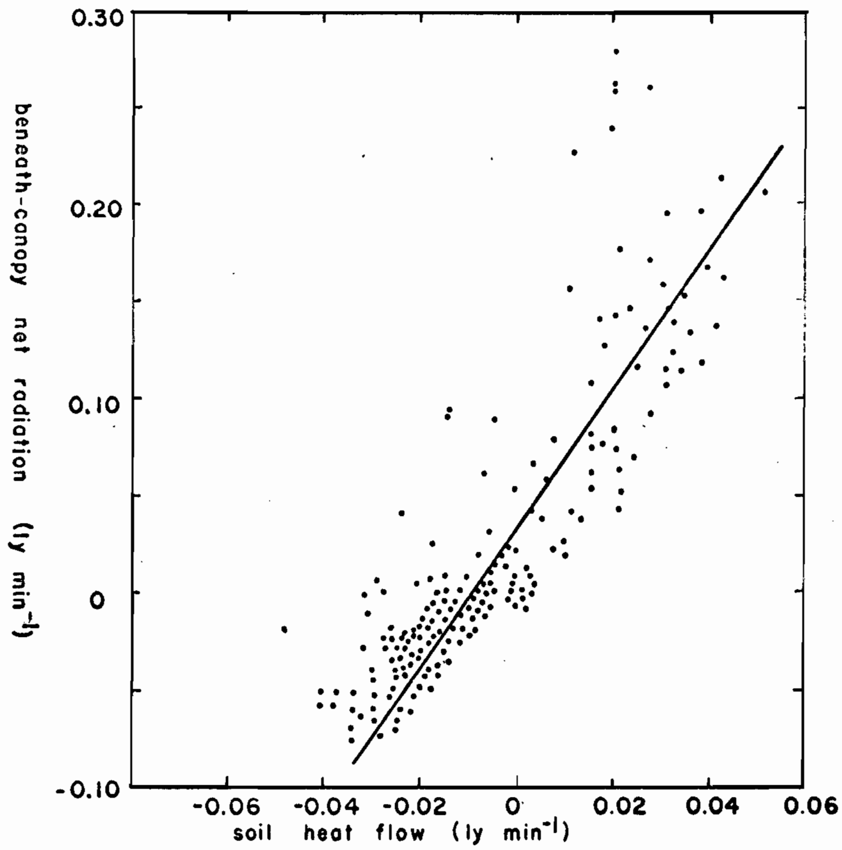
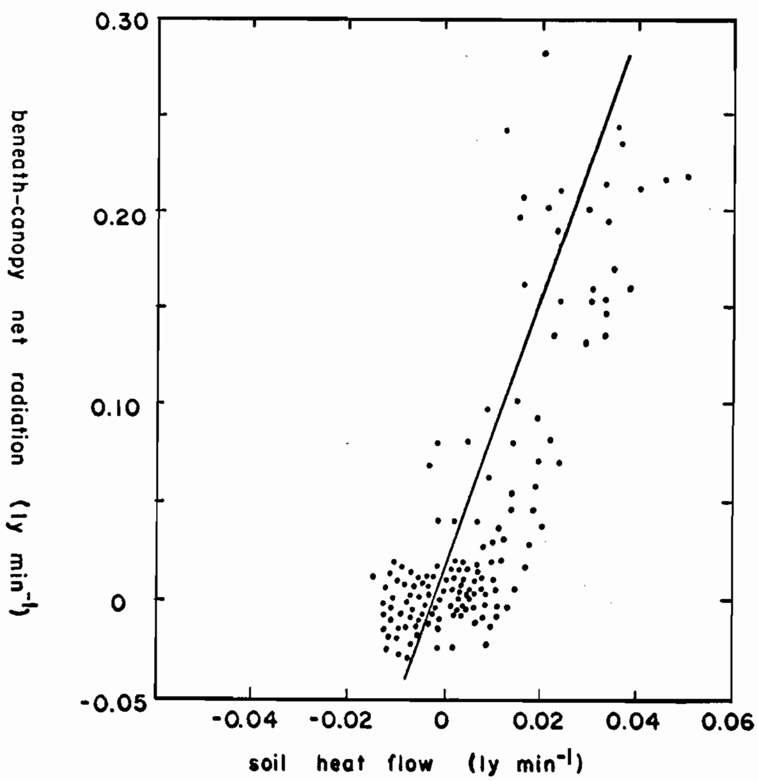


Figure 26 (b)

Linear Regression of Soil Heat Flow on Net Radiation for a Litter-Covered Forest Floor



G. Water Movements in the Forest Layer

(a) Water Interception by the Forest Canopy

It might be expected that the water which is intercepted by a forest crown would represent a complex function of total rainfall, rainfall intensity, morphology of the tree species, density of the trees and stage of leaf development. It would also be expected that there is a maximum amount of water which a given tree can intercept might also be anticipated.

The results for the experimental site on Mont St. Hilaire are shown in Fig. 27 (a). The weekly amount of rain which was intercepted by the canopy is plotted against the total weekly rainfall. A linear regression of rainfall interception on total rainfall is determined for all values except for those two which show the highest weekly rainfall. The two measurements are excluded because they are extremes and do not conform to an otherwise close-fitting line of regression. The regression curve reads $I = 0.73 + 0.288\ P$, where I is the amount of rain intercepted and P is total weekly precipitation, both measured in centimeters. The correlation coefficient of 0.93 indicates a good relationship. The maximum amount of rain which the canopy can intercept may or may not be given by the two points showing 1.20 cm of intercepted water, which correspond to weekly rainfall amounts of 4.15 and 5.57 cm. A firm statement cannot be based on two measurements. Moreover, for one period when rainfall was greater than 4.2 cm, there was an intercepted rainfall well below the possible maximum line. The conclusions which can be drawn from these data are that for low weekly rainfall amounts (less than 4.2 cm), the

Figure 27(a)

Linear Regression of Canopy Interception on Total Weekly Rainfall

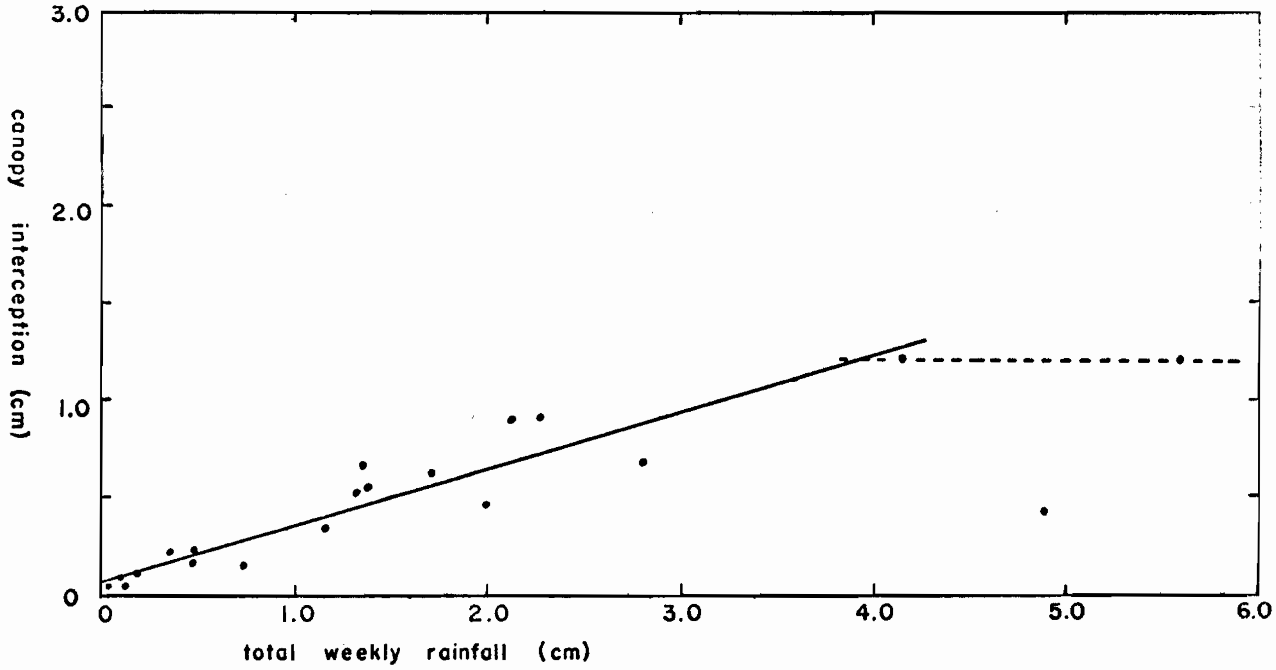
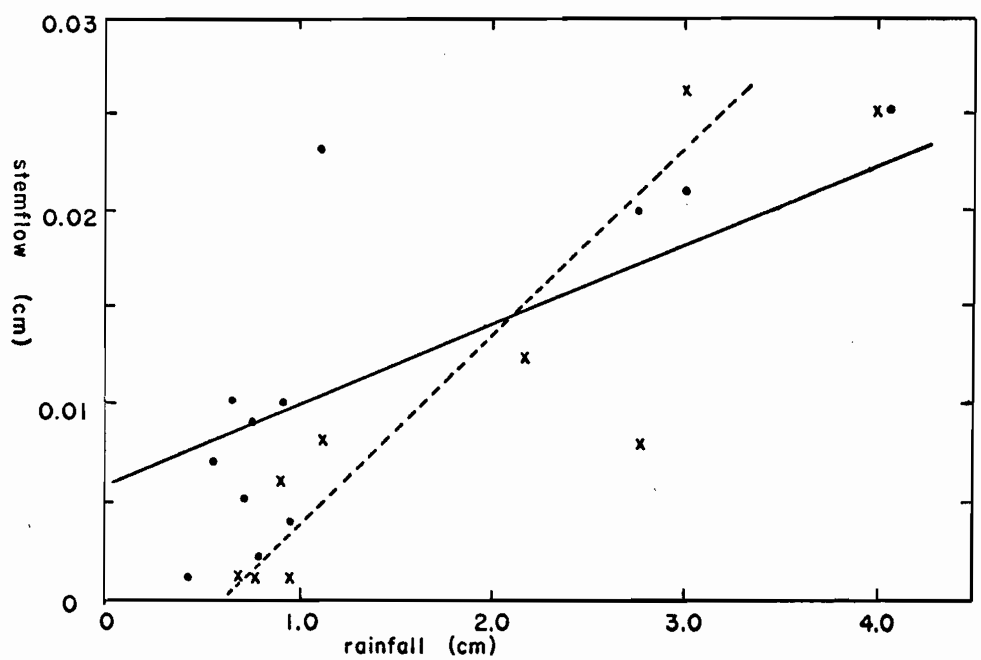


Figure 27(b)

Linear Regressions of Beech and Maple Stemflow on Total Weekly Rainfall

beech stemflow trend
maple stemflow trend



amount of water which is intercepted can be predicted from a linear relationship with total rainfall, at least for a natural forest cover of the type found at St. Hilaire. For higher rainfall amounts, no such relationship is evident. An accurate prediction of canopy interception during heavy rainfall will require further measurements.

Geiger (1965) provides a useful summary of data on canopy interception. The most widely-used and successful approach expresses interception as a linear function of total rainfall in the same way as has been presented above. The equation in the form $I = a + bP$ was first applied by Horton (1919). This equation was found to be generally satisfied, in a series of over 200 precipitation measurements of both snow and rain, for pine forest in California, by Rowe and Hendrix (1951). The average interception of 28 percent during the growing season at Mont St. Hilaire compares to 25 percent for mature beech forest measured by Buhler (1892) and quoted in Kittredge (1964), and 33 percent for various species of oak measured by Ovington (1954). Kittredge (1964) detailed a study by Millett (1944), which shows a rapidly decreasing percent interception with increasing monthly rainfall amounts for pine trees in Australia. Millett's data point to a maximum quantity of intercepted rain after a certain total fall is reached. This occured for monthly rainfalls greater than 9 cm.

(b) Stemflow

As was pointed out in Chapter IV, the quantitative significance of stemflow is minor. Some interesting conclusions into the nature of stemflow can be reached, however, for the mature beech-maple forest at

Mont St. Hilaire. For total single rainfalls of less than 0.40 cm., there is no measurable stemflow. For rainfall amounts between 0.40 and 0.70 cm., stemflow begins on beech trees. Only for rain amounts greater than 0.70 cm., can stemflow from maple trees be measured. Fig. 27 (b) shows the linear regression lines for stemflow plotted against total single rainfall amounts. For beech and maple trees, the regression equations become SF (beech) = 0.006 + 0.004 P, and SF (maple) = 0.007 P - 0.003, where SF is stemflow and P is precipitation, both measured in centimeters. Correlation coefficients are fairly low at 0.88 for beech and 0.87 for maple. The two regression lines should be treated as merely indicating trends. There are too few measurements to attach a quantitative significance to the correlation. For example, the linear regression line for beech stemflow intercepts the stemflow axis above zero, which would indicate a stemflow with very little rainfall. As can be seen from the statements above, this is quite unrealistic as stemflow was never measured for rainfalls of less than 0.40 cm. The trend line for maple has a steeper slope than that for beech, so that the two lines converge just beyond 2 cm. of rainfall. This is real, and is due to the morphology of the tree trunks and branches. The rough bark of the maple gives a greater surface for collecting water drops, and thus prevents coalescence of these drops during low rainfall amounts. The beech has the smoothest bark of any tree and presents an ideal surface for a rapid coalescing of rain drops, thus hastening the ascendancy of gravitational attraction over surface friction. It must not be inferred from Fig. 27 (b) that, for rainfall amounts greater than 2 cm., the maple trees yield a higher stemflow amount than do the beech. The present study yielded too little data to allow any statement of the relative stemflow amounts under conditions of heavy rainfall.

Ovington (1954) presented data for the stemflow of 12 tree species, both deciduous and coniferous. In no case did the stemflow exceed 0.32 percent of the annual rainfall. These data applied to conditions of small individual rainfalls (less than 1.7 cm.) in England. Delfs et al (1958), in a summary of their stemflow measurements, noted that the highest absolute flow takes place during light prolonged rain, but only after initial wetting of the stem is accomplished. The delay is greater on large trees with a large leaf area and rough bark. Penman (1963) outlines stemflow measurements by Eidmann (1959) for European beech. For the full-leaf state, Eidmann measured stemflow at 16 percent of the total rainfall. This compared to less than 1 percent for spruce forest. It is clear from the foregoing discussion that stemflow can vary greatly in importance, depending on the type of rainfall and the nature of the forest. For the type of rainfall experienced at Mont St. Hilaire for a beech-maple forest, stemflow is greater on beech trees than on maple, but its significance in terms of the total rainfall is small.

(c) <u>Runoff from the Forest Floor</u>

The general character of forest runoff was discussed in Chapter II. It was noted that a high infiltration rate, especially after dry

Figure 28 (a)

Seasonal Trends of Surface and Underground Runoff

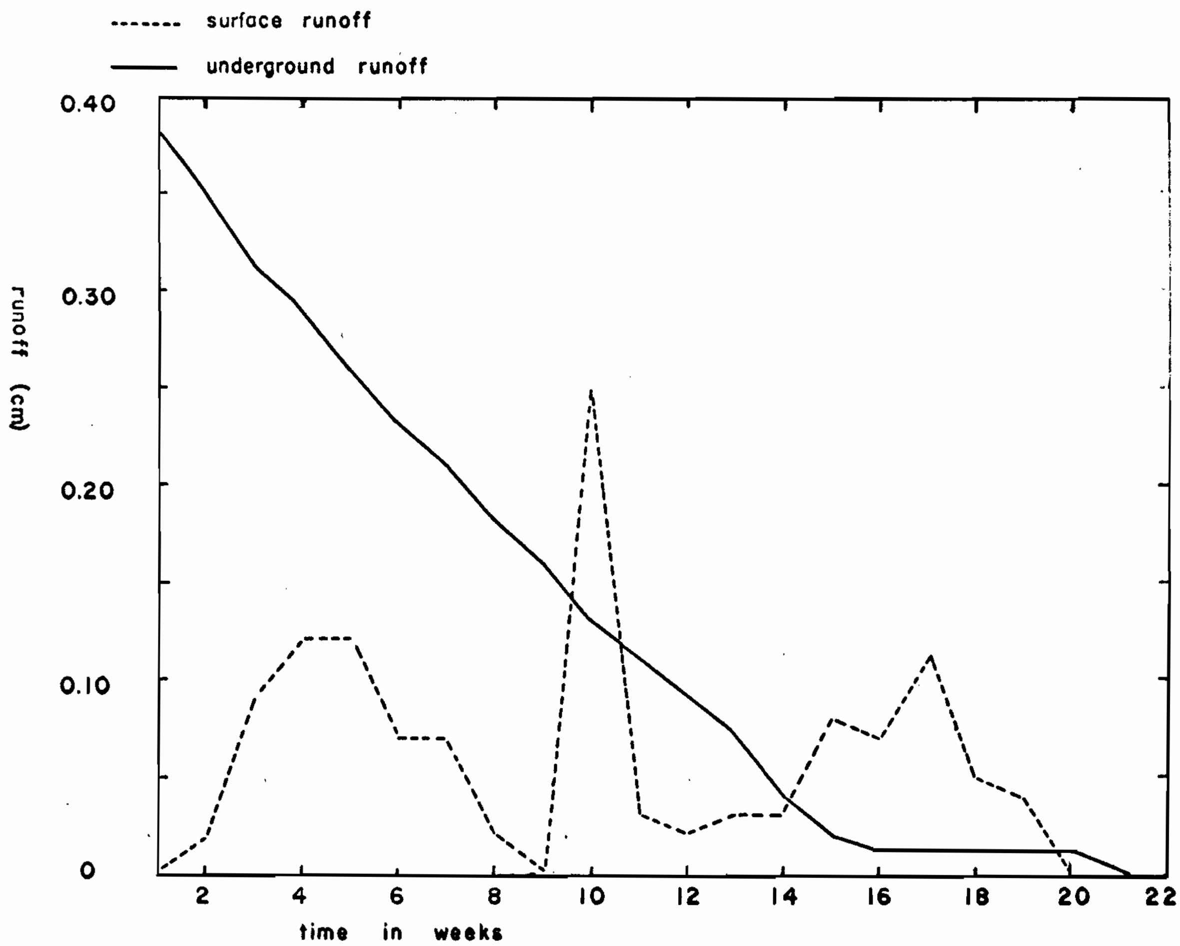
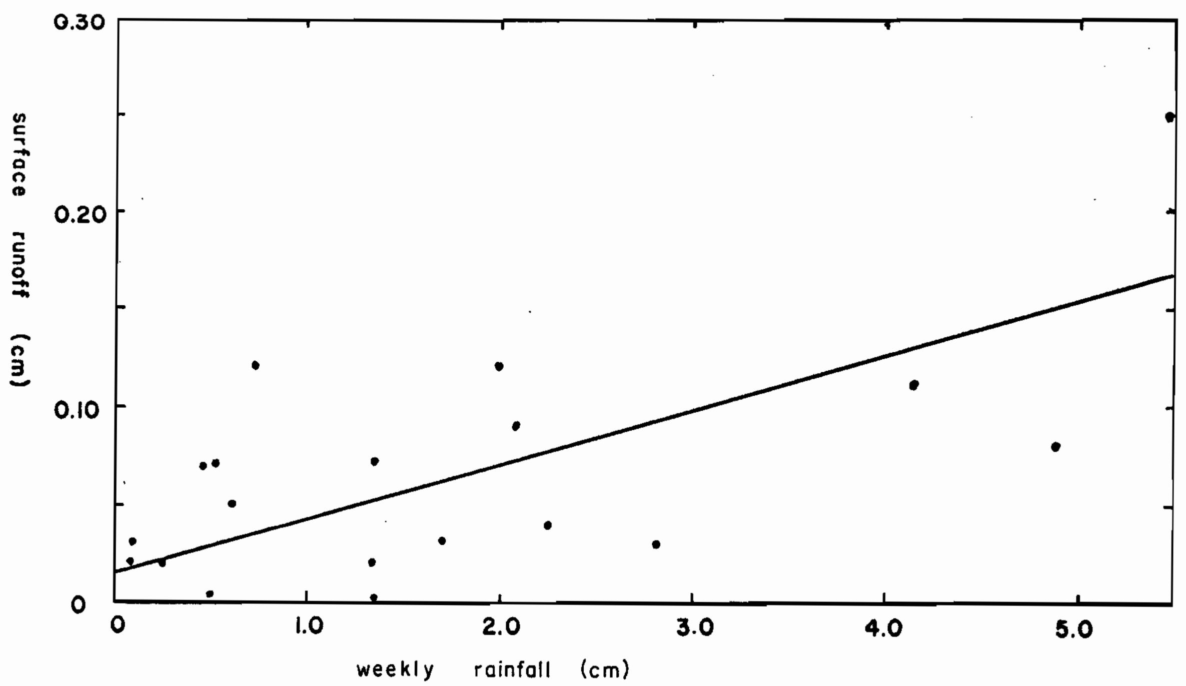


Figure 28 (b)

Linear Regression of Surface Runoff on Weekly Rainfall



periods, serves to reduce the surface runoff. In Chapter IV the total runoff was plotted for the twenty-two week growing season along with the total weekly rainfall. It was shown that runoff decreased throughout the growing season even as rainfall showed a general increase.

In Fig. 28 (a) the surface runoff and underground runoff (total runoff-surface flow) are plotted for each week of the growing season. The underground flow decreased in a linear fashion from Week 1 to Week 16, after which it maintained a constant low rate to Week 21, when runoff became so small that it no longer constituted a measurable figure. Surface runoff in contrast was variable from week to week.

Figure 28 (b) plots weekly surface runoff against the total weekly rainfall, in order to discover the extent to which the surface flow responded to rainfall. The regression curve of $Q = 0.015 + 0.028 P$, where Q is runoff and P is rainfall gives a correlation coefficient of 0.71. The low correlation is created by several factors. If a dry period is succeeded by a relatively wet one, much of the water reaching the soil during the latter period must be used in rewetting the surface layers of the soil before any surface runoff can proceed. Thus, as the ground becomes dry during the late stages of the period of active growth, much of the water penetrating to the soil surface will be absorbed immediately. Each case can be illustrated by consulting Figure 28 (a), and the weekly rainfall amounts given in Table 13. In Week 9, the surface runoff was nil in spite of a total rainfall of 1.34 cm. This contrasted to Week 16 where a rainfall of 1.35 cm. gave a surface runoff of 0.07 cm of water. For the preceding weeks in each case (i.e. Weeks 8 and 15), the rainfall was 0.08 and 4.90 respectively. Thus the drying of the surface soil layers during Week 8 was sufficient to prevent any surface runoff during the following period. Seasonal differences appear between Week 10 and Week 17. The rainfall in these periods was 5.90 and 4.14 cm respectively, and each week followed a preceding one of substantial rainfall. However, the surface runoff yielded 0.38 cm of water for Week 10, whereas in Week 17 the runoff was only 0.12 cm. The drying out of the ground layers during the late stages of the growing season served to reduce surface runoff substantially.

(d) Accuracy of Rainfall Measurements above the Forest.

The accuracy of rainfall measurements has long been a source of concern to earth scientists. In particular, the accuracy of determining rainfall above a forest from gauges in a clearing has been suspect. The problem is discussed in some detail by Geiger (1950, 1965) and Penman (1963).

For this study, a reasonably accurate estimate of rainfall over the West Creek watershed was needed. Two gauges were placed widely apart in the apple orchard in the positions noted in Chapter III. In case these measurements might prove unpresentative a control gauge was established above the forest canopy. There was good agreement between the measurements of each of the gauges. If the values from each gauge are added for the total period of measurement, the rainfall which was measured above the forest falls half way between the measurements in each orchard rain gauge. The average deviation from the mean was 1.5 percent. For single rainfalls, the highest deviation about the mean gave 8 percent, but again the above-canopy gauge gave readings between the two orchard values. For the site in question the rainfall which was measured in a forest clearing was representative of that which fell above the forest.

H. Summary

Differences in net radiation profiles for 20 degree north-facing and south-facing slopes are less than might be expected. Net radiation was somewhat higher at greater depths within the forest of the South Slope. Sixty percent of the net radiation was concentrated within the top three meters of the canopy on each slope. There was a more gradual decrease at lower levels. Net radiation was greater at lower heights on both slopes during periods of mainly diffuse solar radiation. The greatest ratio of net radiation at the one meter level to that above the canopy was found in the middle-afternoon, probably as a result of heating in the tree crown and an increased downward flux of infrared radiation. Differences of net radiation from place to place beneath the forest canopy were pronounced showing a mean deviation of 25 percent. This value applies to various topographic conditions and different amounts of cloud cover. The differences were more pronounced for direct than for diffuse solar radiation. The spatial variation of net radiation beneath the canopy increased to a noon hour maximum. Baumgartner's profile measurements of net radiation showed that in a spruce forest there was a 95 percent concentration of net radiation in the top three meters. This is a narrower zone of maximum energy exchange than was found for a deciduous forest in the present study.

Trapp, Gast, and Vezina all measured a greater penetration into the canopy by diffuse radiation than by direct. Tanner, Peterson and Love found that the net radiation under a corn crop was higher during periods of diffuse radiation, when the rows were oriented in an east-west direction.

Topographic variations in the penetration of solar radiation beneath the canopy were slightly greater than in the case of net radiation. Otherwise, the profiles of solar radiation within the forest resembled those of net radiation very closely. The absorbed solar radiation showed a regular decrease in amount with greater depth in the forest. The long-wave balance clearly illustrated the effect of canopy heating in the afternoon. During this period, the long-wave balance above the canopy gave higher negative values due to a greater upward infrared radiation. In contrast, the long-wave balance beneath the canopy eventually fell to zero due to the increased long-wave flux downward from the crown.

Net radiation showed little variation from place to place above deciduous forest under homogeneous topographic conditions. The available data showed that the differences in net radiation above the forest on North and South Slopes were also very small. For a six day recording period, which included both cloudy and sunny periods, the total net radiation above each slope was practically equal. For limited sunny periods differences did occur, but they exhibited no systematic topographic preferences.

In the fall, the ratio of the net radiation beneath the crown to that recorded above the canopy increased from 0.09 to 0.41 in the period between the full-leaf and leaf-bare states of the forest. Soil heat flow showed a close dependance on beneath-canopy net radiation when the leaf litter on the floor was thin or absent. With the development of a thick litter layer this dependance largely ceased due to the great insulating properties of the newly-fallen leaf cover.

The amount of water which was intercepted by the forest crown was a linear function of the total rainfall until a maximum weekly rainfall of over 4 cm was reached. Stemflow proceeded on the smooth-barked beech trees for all rain amounts greater than 0.40 cm, but for the maple trees it did not start until the total rainfall exceeded 0.70 cm. Underground runoff from the forest floor decreased regularly throughout the growing season. Surface flow, however, responded to a number of influences, and could not be predicted from indirect evidence. The rainfall which was measured in a forest clearing was representative of the rainfall above the forest for the area of the West Creek watershed.

Chapter VII Geographical Speculations and Implications.

A. Introduction

Forests constitute a considerable part of the living layer of the earth's surface, and as a result they assume geographical importance. Part of a forest's geographical distinctiveness results from the microclimatic features which are associated with it. Munn (1964) points out two important facts which have emerged from recent climatic studies. First, the forest is an active meteorological region. Exchange processes are vigorous in spite of the light winds beneath the crown. The canopy is not an impenetrable barrier for the transfer of heat and water vapour. As the second fact, Munn notes that forest climates are not all similar. Transpiration rates and soil moisture conditions differ, as do the physical characteristics of the foliage and the differences in slope and aspect. In short, he concludes that the forest contains strong sources of heat and moisture, which depend to a large extent on physiological factors. A further aspect of the distinctiveness of the forest's microclimate involves the differences which it displays compared to other geographical areas. Such differences result from the multiple surfaces which succeed one another downward from the canopy of the tallest trees to the earth's surface. Of the many surfaces, the two most important ones are found at the top of the canopy and at the forest floor. The discussion which follows is concerned with the distinctiveness of the microclimate of a middle-latitude deciduous forest during the growing season. It is necessarily confined to implications which result from observations at Mont St. Hilaire and is, therefore, limited in scope. Research into the nature of the microclimate of a deciduous forest is in too youthful a stage to allow many generalizations.

B. Global Radiation; its reflection and penetration.

The global radiation which reaches the active surface layer is largely independant of that surface, and is the same for all cover types which are experiencing the same meteorological conditions. Upon contact with the surface, however, significant differences become apparent. Miller (1955) writes: "Trees are the dark part of the landscape; this impression from the visible region of the solar spectrum is applicable to the entire spectrum." Such visible darkness results from the amount of reflection of insolation by the forest canopy. Table 24 shows mean albedo of various cover types as presented by Kondrat'ev (1954). It also shows the percent absorbed solar radiation. It is apparent that the solar radiation which is absorbed by a deciduous forest does not differ substantially from other green vegetation except alfalfa.

Microclimatic differences become pronounced with the penetration of the radiation into the active layer. Whereas surfaces which are barren of vegetation or which have only a thin cover absorb all the radiation in a very shallow layer, the forest displays a much

Table 25. Percent Absorption of solar radiation by different surfaces.

Surface Type	Mean Albedo (%)	Absorbed solar radiation (%)
Untilled field	8	92
Black earth	11	89
Coniferous forest	12	88
Ploughed field	14	86
Spring wheat	17	83
Deciduous forest	18	82
Winter wheat	20	80
Grey earth	20	80
Grass	21	79
Alfalfa	28	72

thicker zone of absorption. Thus, in the case of a close-crowned deciduous forest at Mont St. Hilaire, about 60 percent of solar radiation was absorbed in the top three meters of the canopy. However, the total absorption was spread through a depth of 14 meters. As was shown in Chapter VI, the solar radiation which reached the forest floor did not exceed 9 percent during the full-leaf period. The transfer of solar radiation through a deciduous forest also differs from other layers of green vegetation. Thus Tanner, Peterson and Love (1960) found that from 18 to 30 percent of the incoming solar radiation reached the ground beneath a corn crop, the actual amount depending on the density of the plants. Penman and Long (1960) found that up to one-half the solar radiation penetrated to the soil beneath a wheat field. As was noted previously, when discussing Baumgartner's work, even a spruce forest concentrates more solar radiation in a narrower zone than is the case with its deciduous counterpart. Thus, a deciduous forest presents a layer of maximum thickness for the absorption of solar radiation.

The ramifications of the radiation regime are several. The depth of the absorption layer and density of the wood mass result in a relatively large daily exchange of heat in the biomass in comparison to other vegetation types. The small penetration of radiation to the ground gives a small soil heat flow. The large absorption of radiation in the canopy results in strong daytime heating. The result, as stated by Munn (1964), is that by day the temperature wave originates at tree crown level and propagates both up and down with decreasing amplitude. This leads to inversion temperature profiles beneath the canopy and superadiabatic ones above. Beneath-canopy inversions during hot summer days have not, to the present writer's knowledge, been observed regularly in non-forest vegetation.

C. Net Radiation; horizontal and vertical variations.

The regime of net radiation within the forest layer depends on the pattern of solar radiation and on the long-wave balance. The latter will be influenced by differential heating in the forest, and the resulting temperature differences. For a bare or thinly-covered surface, the long-wave radiation upward is from the ground and the downward flux comes from the atmosphere. During the growing season the ground is always the warmer body. Therfore, the long-wave balance is always negative and usually large. A similar pattern occurs between the forest crown and the atmosphere above. Beneath a forest, however, the heating of the canopy can lead to a zero long-wave balance in late afternoon as was recorded in the present study, or even a positive balance such as that measured by Baumgartner in the evening hours. Within the canopy itself, the negative long-wave balance is still suppressed but to a lesser extent in comparison to the beneath-canopy regime. As a result of the small values of the long-wave balance beneath the forest, the magnitude of change in net radiation with depth closely parallels the changes in solar radiation. The radiation exchanges within the active layer are insulated to a considerable extent from the atmosphere above. The complexity of these exchanges is similar for all vegetation covers, but is more pronounced in a close-crowned forest because of the density of the canopy and the thickness of the active layer.

Place-to-place differences in hourly values of net radiation within the forest are large. They result from the variable penetration of solar radiation through the canopy. Such differences are greater than those which would be observed over an unvegetated surface. Above the forest, however, spatial differences should be smaller than for bare ground, for three reasons. First, the albedo of a deciduous forest will not be influenced by moisture differences which are found with the changing topography of bare soils. Second, variations due to changes in soil types which are inherent in a non-vegetated surface are not duplicated by similar differences in the forest crown. Finally, since heat energy can move more freely by convection in the air than by conduction in the ground, the surface temperature of the forest canopy will show less spatial variation than temperatures on a bare surface. It was shown in Chapter VI that areal differences in net radiation above the canopy are indeed small.

This writer has no evidence for greater differences over bare soil so the above argument, though soundly based, must remain speculative.

D. Topographical Influences on Microclimate

It has been shown in Chapter VI that net radiational difference which are due to topography are decreased by a fully-leafed forest layer during the period of high sun. Whereas for a barren hill with slopes oriented in a north-south direction, considerably more incoming radiation and hence higher net radiation is received per unit area on the south-facing than on the north-facing side, over a forest cover as was seen in Fig. 25, the varied orientation of tree leaves largely eliminates this difference for conditions of mean cloud cover. Because evapotranspiration is dependant on net radiation to a considerable degree, it would be expected that evapotranspiration differences for differently-oriented slopes would be lessened during the growing season.

Figs. 17 and 19, showed that the tree layer on the North Slope had its net radiation concentrated closer to the top of the forest crown than that on the South Slope. However, as was shown by Denmead (1964), this need not mean that more energy transformation on north slopes must occur in the top layers of the forest, since heat is readily diffused from its region of absorption.

While topographical differences between barren surfaces and forested surfaces may be considerable, the differences between forests and other surface types cannot be so clearly stated. It is likely that a field of grass or mature alfalfa would behave much the same as a forest in decreasing net radiational differences between slopes. Row crops if planted in a north-south direction, would display greater variations between differently oriented hillside.

Because of the radiational regime, temperature differences between slopes should be at a maximum at the soil surface and decrease steadily with height until at the tree-top level the temperature differences on each slope are small. Measurements by Cantlon (1950) tend to support this hypothesis.

The changeover from leafless to leafy condition in a deciduous woodland exerts the greatest microclimatic changes which occur in forested regions, changes which are not matched in rapidity for other areas except through changing snow cover. Over a period of several weeks in the spring the net radiation at the forest floor may be reduced from 41 to 9 percent or less. At the ground level, microclimatic changes occur most rapidly in the fall with the loss of leaves. The influences of the dead leaf layer in minimizing climatic variations within the soil is large, and is probably unequalled in most non-forested regions. The conservative influence is manifested both in the thermal and in the moisture characteristics of the forest soil.

E. Water and Heat Movements

The important influence of forest vegetation on runoff and ground moisture is documented by Kittredge (1948, 1962), Hoover (1962) and Penman (1963). The deep rooting system of trees allows them to tap a larger volume of soil water than is the case for most types of vegetation. As a result, the available soil moisture is generally large.

As discussed previously, forests will evapotranspire more water than vegetation with lesser rooting volume during dry periods. This, in turn, results in a soil which is able to absorb larger amounts of water in time of rain. Thus, runoff will be reduced. The crown layer lessens rainfall intensity by breaking the impact of the freely-falling raindrops. This allows the soil beneath to absorb moisture at a slower rate. Such a feature serves to diminish runoff during periods of high rainfall intensity. Evidence for reduced water yield from forest regions compared to other vegetated areas was presented in Chapter II.

If the net evapotranspiration from a forest is greater than for other surfaces, a lesser convective heat flux might result, because of less available sensible heat energy. Munn (1964) states that since air can "leak" down and up through the forest, large temperature gradients cannot be maintained above the canopy. In addition, strong transpiration reduces the daytime lapse rate, and in some situations may result in a temperature inversion. Munn further notes that pilots have recognized for many years the rarity of updrafts (free convection) over a forest.

The tree crown plays a further important role through its interception of rainwater. It was shown in Chapter IV that for the summer season of 1964 at Mont St. Hilaire more than 25 percent of the rainfall was intercepted. Such a high interception rate is greater than for non-forested regions, and since this is water which is not available for transpiration, it tends to compensate for the lesser runoff from the forest floor. It is not, however, completely wasted. The net radiational energy which is used to evaporate intercepted water is not used for transpiration, and the water demands on the tree are reduced accordingly.

The forest influences the rainfall reaching the soil surface in such a manner that it is less uniformly spread over area than in the case of more open sites. Such non-uniformity does not exert a strong geographical influence, though it may lead to short distance variations in soil structure, and in herb and bush growth.

F. The Forest; a conservative geographical area.

Because of its longevity and physical nature, the forest acts climatically as a conservative body. Due to its deep rooting system, it is less affected by drought conditions in comparison to other forms of vegetation. The deep layer which the forest provides for active energy exchange lessens temperature gradients. Evapotranspiration also proceeds through a relatively deep zone. The protection of the canopy lessens diurnal variations in temperature at the ground. The tree leaves break the impact of heavy rainfall and increase the length of intense showers. The forest strongly minimizes the topographical influences on microclimate, especially during the growing season. Even during the dormant period, its leaf litter dampens the movement of heat and water within the soil.

Chapter VIII Summary and Conclusions

A. Summary.

Observations into some aspects of the heat and water-balance of a middle-latitude deciduous forest are presented in this thesis. Measurements were started in the fall of 1963, and the bulk of the data was collected during the growing season of 1964. The site of Mont St. Hilaire is well-suited for microclimatic investigations. Much of the area is undisturbed and is covered with close-crowned, virgin, deciduous forest. Under Maycock's direction the distribution of tree species has been mapped, and quantitative studies have been carried out for select stands. The three most important deciduous trees are maple, beech and oak. The mountain possesses a topographic diversity which makes it ideal for investigations into the influence of slope orientation on microclimate. During the growing season of 1964, the temperature regime was normal for the region, but the rainfall was almost one-quarter less than the five-year average.

Previous investigations into the nature of a forest microclimate promote the following general observations. Few studies into the heat and water-balance of forested areas have been published. Measurements of net radiation and soil heat flow are obtained readily, but the latent and convective heat fluxes have proved difficult to ascertain. These latter terms of the heat-balance equation can be measured by employing aerodynamic methods, the Bowen Ratio within the energy-balance framework or covariance techniques. Aerodynamic calculations have been most favoured.

The relative importance of the latent and sensible heat fluxes is highly variable. Under moist conditions over land, the removal of heat by evapotranspiration is overwhelmingly larger than the convective movement of sensible heat. For periods of drought, the sensible heat flux assumes the greatest magnitude.

The forest creates its own unique microclimate because of its thick leaf canopy which intercepts radiation and rain. In addition, the canopy modifies the influence of topography on climate.

Instrumentation

Solar radiation was measured with a Kipp solarimeter and a bimetallic actinograph. The former was mounted premanently above the forest crown while the latter was used as a mobile field instrument. Four net radiometers were employed for the measurement of net radiation. They were mounted on various types of adjustable masts within and above the trees. Spatial sampling was achieved by moving them regularly. A number of thermopile discs were connected in series,and used to measure soil heat flow. Beneath-canopy rainfall and stemflow were measured in a select site, while above-canopy rainfall was recorded at three different places. The measurement of the stream height behind a weir with a 90 degree V-notch by an improvised stage-recorder, allowed a continuous calculation of stream flow. Soil-filled tanks to measure potential evapotranspiration were installed within a forest and in a forest clearing. They were irrigated daily except during periods of heavy rain.

The manufacturer's calibration for the Kipp was used throughout the study. The actinograph's response remained close to that of the solarimeter. A post-season check of the Kipp showed that its response was on the average within $\pm$ 5 percent of a recently-calibrated Eppley. Regular calibrations during the growing season showed large changes in the responses of two of the net radiometers. Fortunately, the responses from each side of the radiometers did not change significantly.

All radiation measurements were averaged for the hour. Stemflow and runoff were measured after each rain, and stream flow was evaluated both for individual storms and for each daily period. Potential evapotranspiration from the tanks was calculated on a daily basis.

Water-Balance

The period of active growth for the deciduous forest at Mont St. Hilaire was concentrated between May 1 and October 1. Of the total rainfall during the period, the tree crown intercepted 29 percent, 1 percent reached the forest floor as stemflow, and 70 percent penetrated to the ground as throughfall. Runoff from the forest amounted to 12 percent of the total rainfall. A soil moisture capacity of 15 cm was estimated for Mont St. Hilaire.

Calculations of evapotranspiration from the hydrologic-balance agree closely with those determined from the Thornthwaite mean temperature method. The potential evapotranspiration which was measured in the evapotranspirometer tanks gave large values which would require more latent heat energy than that available from net radiation. No single cause for the high potential evapotranspiration can be cited, and apparently multiple influences were in operation.

All water receipts and water expenditures are calculated for the growing season. The evapotranspiration process is divided into the evaporation from the tree canopy, evaporation from the forest floor and transpiration from the trees.

Heat-Balance

Measurements and estimations of the heat-balance over the forest are presented. Global radiation was measured and a mean albedo of 0.18 was assigned for deciduous forest, the difference giving the absorbed solar radiation. The long-wave balance is determined from the difference between the net radiation and the absorbed solar radiation. The various radiation measurements, as well as the energy used to evaporate the intercepted water, are plotted for each of the twenty-two weeks of the growing season.

A modified form of a calculation presented by Monteith and Szeicz gives a Ω value which shows the relative magnitude of the absorbed solar radiation and of the long-wave balance during various intervals of the growing season.

The latent heat flux used approximately two-thirds of the net radiational energy, and the sensible heat flux about one-third. Net heat storage in the biosphere and in photosynthesis is negligible over the growing season. Soil heat storage might represent $\pm$ 5 percent of the net radiational energy. The calculations of the heat-balance for Mont St. Hilaire agree closely with similar studies pursued in Germany and Russia.

The Forest as an Active Layer

Measurements of the profiles of solar and net radiation for north-facing and south-facing slopes during the high sun period show only small differences. An average of sixty percent of radiant energy is concentrated in the top three meters of the forest layer. Heating of the canopy in the afternoon is pronounced, and it exerts a strong influence on the net radiation regime. Hourly values of net radiation within the forest vary significantly from place to place. Diffuse radiation is generally more effective than the direct solar beam in penetrating the tree crown. Calculations show that the negative long-wave balance beneath the canopy is less than it is within or above the forest crown.

For homogeneous topography, simultaneous measurements of net radiation above the deciduous forest show only small differences from place to place. Above-canopy differences between north and south slopes are equally small for conditions of mean cloud cover.

In the fall, the ratio of net radiation beneath the canopy to the above-canopy measurement increased from 9 to 41 percent between the full-leaf and leaf-bare states of the forest. When the litter on the forest floor is thin, the soil heat flow shows a close dependance on the beneath-canopy net radiation. The insulating properties of a thick litter of newly-fallen leaves destroys this relationship.

The quantity of rain which is intercepted by the trees proves to be a linear function of total weekly rainfall for weekly rain amounts less than 4 cm. Stemflow occurs on beech trees for rain amounts greater than 0.40 cm, but does not start on the maple trees until 0.70 cm of rain has fallen. Runoff from the forest floor can be divided into ground flow and surface flow. The former decreases regularly throughout the growing season, but the latter varies with a number of interacting factors.

B. Conclusions

The energy-balance concept provides a theorectical framework within which the microclimatic features of a forest layer may be examined. The most important term of the energy-balance equation is net radiation, and it is this term which has been investigated most thoroughly in the present study.

It has proved possible to take measurements in the forest layer without erecting a permanent scaffold. This is desirable in some ways since there is little disturbance to the natural forest condition.

Exchange processes are active in all parts of the forest layer, but it is the tree canopy which exerts the strongest influences on microclimate. This study has shown that the canopy of a deciduous forest provides a deeper layer for active energy transformation than is found for most vegetated surfaces. The large interception of radiation in the crown reduces heat exchange at the forest floor. The soil heat flow and evaporation from the forest floor are quantitatively small. Strong daytime heating of the canopy results in an upward and downward heat movement from this zone.

It is an elementary observation that broad-leaved deciduous forests demand sites having abundant water. The corollary is two-fold.

First, forests evapotranspire a large amount of water. Their root systems tap large volumes of soil, and the removal of water from deep soil layer during dry periods increases the moisture capacity of the soil during rain. The result is a more sustained rate of evapotranspiration and lower yields of water from forested regions - the latter being an observation which is supported by an increasing body of evidence. The second part of the corollary involves a relatively small sensible heat flux due to a limited amount of available energy and the large evapotranspiration.

The radiation regime beneath the trees assumes a special character For most surfaces in summer, the long-wave balance is negative and large. It is shown in this study that within a forest it is either negative and small, or it is zero. There exists an almost closed system of infrared radiation exchange beneath the canopy.

During the growing season the top of the tree crown presents a distinctively homogeneous surface to the sun. Spatial variations in net radiation above the forest in flat areas are small. Differences in the absorbed solar radiation which are due to topography are substantially reduced. Major microclimatic differences due to slope orientation appear to exist only beneath the forest canopy during the period of high sun.

One of the strongest and most rapid changes in the forest microclimate is created when the trees lose their leaves in the fall. Radiation to the forest floor is increased more than four-fold, canopy interception of rain is reduced, wind speed increases at lower levels, and the forest floor is well-insulated by a thick leaf litter. The process is reversed during spring leafing but proceeds at a slower rate.

A forest acts climatically as a conservative body. It can withstand drought and promotes active energy exchange through a deep layer. It diminishes topographical influences on microclimate and protects the soil surface from intense rainfall. Runoff is strongly reduced. The forest fosters a distinctive microclimate which differs substantially from other vegetated and non-vegetated surfaces.

Appendix A

Incoming Solar Radiation (May 1 to October 1)

Values in ly day $^{-1}$

Date	May	June	July	August	September
1	659	197	620	535	455
2	659	512	279	325	508
3	660	200	334	622	352
4	413	446	303	383	076
5	654	678	203	117	289
6	588	612	323	593	276
7	494	387	352	584	431
8	387	670	467	355	250
9	335	571	629	271	450
10	122	194	584	379	299
11	482	694	474	415	172
12	715	684	556	183	514
13	289	513	214	268	494
14	207	392	243	325	484
15	684	370	431	287	495
16	128	597	581	275	309
17	648	662	501	153	231
18	639	657	495	491	262
19	506	600	565	431	493
20	744	513	670	282	418
21	655	456	377	588	309
22	656	671	466	075	407
23	641	516	656	158	304
24	567	316	482	409	353
25	207	694	414	525	084
26	717	322	202	462	345
27	462	684	487	510	092
28	362	697	555	457	405
29	367	384	311	151	407
30	426	597	353	507	177
31	592	---	643	337	399 (Oct.1)
Totals	15,869	15,645	14,220	11,620	10,972

Appendix B

Above-Canopy Net Radiation (May 1 to October 1)

Values in ly day $^{-1}$.

Symbols - (P) is Net Radiation Gain
(N) is Net Radiation Loss
(B) is Diurnal Net Radiation Balance

Date	May			June			July		
	(P)	(N)	(B)	(P)	(N)	(B)	(P)	(N)	(B)
1	460	-62	398	107	-16	091	421	-36	385
2	474	-58	416	384	-32	352	186	-22	164
3	468	-68	400	112	-16	096	271	-06	265
4	252	-33	219	321	-30	291	211	-18	193
5	463	-42	421	532	-68	464	137	-09	128
6	430	-25	405	503	-66	437	217	-06	211
7	463	-42	421	276	-25	251	229	-06	223
8	187	-41	146	513	-51	462	339	-36	303
9	202	-21	181	415	-34	381	440	-50	390
10	142	-19	123	139	-29	110	405	-61	344
11	333	-29	304	563	-54	509	299	-43	256
12	538	-53	485	496	-62	434	368	-53	315
13	228	-33	195	409	-43	366	128	-39	089
14	144	-49	095	110	-20	090	141	-36	105
15	500	-49	451	205	-35	170	242	-55	187
16	043	-37	006	415	-39	376	424	-56	368
17	497	-45	452	450	-62	388	327	-38	289
18	400	-46	354	460	-62	398	314	-38	276
19	480	-45	435	439	-32	407	376	-45	331
20	562	-52	510	374	-40	334	466	-52	414
21	500	-47	453	336	-33	303	304	-48	256
22	475	-68	407	492	-61	431	321	-40	281
23	484	-37	447	365	-32	333	456	-77	379
24	351	-11	340	244	-21	223	343	-67	276
25	091	-35	056	480	-74	406	276	-57	219
26	472	-40	432	234	-30	204	129	-34	095
27	330	-31	299	502	-17	485	316	-49	267
28	243	-26	217	473	-57	416	375	-20	355
29	258	-20	238	290	-25	265	219	-25	194
30	308	-22	286	424	-16	408	232	-37	195
31	454	-33	421	---	---	---	356	-44	312
Totals	11,232	-1,219	10,013	11,063	-1,182	9,881	9,268	-1,203	8,065

Appendix B (continued)

Date	August (P)	August (N)	August (B)	September (P)	September (N)	September (B)	
1	328	-17	311	257	-68	189	
2	209	-11	198	267	-80	187	
3	374	-34	340	179	-59	120	
4	238	-14	224	045	-30	015	
5	090	-04	086	209	-33	176	
6	356	-20	336	196	-72	124	
7	344	-23	321	337	-90	247	
8	234	-13	221	136	-40	096	
9	164	-20	144	336	-56	280	
10	259	-40	219	192	-62	130	
11	289	-44	245	102	-88	014	
12	174	-26	148	300	-125	175	
13	183	-53	130	292	-125	167	
14	240	-54	186	282	-123	159	
15	200	-28	172	294	-114	180	
16	232	-29	203	161	-65	096	
17	156	-35	121	097	-37	060	
18	377	-48	329	115	-85	030	
19	349	-43	306	263	-92	171	
20	215	-50	165	240	-92	148	
21	354	-71	283	197	-56	141	
22	023	-37	-14	260	-44	216	
23	108	-21	087	189	-63	126	
24	288	-63	225	232	-87	145	
25	361	-58	303	056	-63	-07	
26	318	-40	278	246	-32	214	
27	372	-64	308	045	-29	016	
28	282	-70	212	245	-77	168	
29	071	-43	028	347	-45	302	
30	369	-40	329	111	-47	064	
31	238	-56	182	271	-70	201	(Oct.1)
Totals	7,795	-1,169	6,626	6,499	-2,149	4,350	

Appendix C

Rainfall, Canopy Interception, Stemflow and Runoff
(May 1 to October 1)

All values in cm day $^{-1}$.

May

Date	Rainfall	Canopy Interception	Stem Flow	Runoff
1	0.00	0.00	0.00	0.07
2	0.00	0.00	0.00	0.06
3	0.00	0.00	0.00	0.06
4	0.00	0.00	0.00	0.05
5	0.00	0.00	0.00	0.05
6	0.00	0.00	0.00	0.05
7	0.00	0.00	0.00	0.04
8	0.49	0.21	0.00	0.09
9	0.34	0.11	0.00	0.08
10	0.46	0.19	0.00	0.10
11	0.00	0.00	0.00	0.03
12	0.00	0.00	0.00	0.02
13	0.05	0.03	0.00	0.03
14	0.00	0.00	0.00	0.02
15	0.05	0.03	0.00	0.03
16	1.31	0.61	0.01	0.14
17	0.13	0.04	0.00	0.06
18	0.08	0.03	0.00	0.04
19	0.51	0.19	0.00	0.06
20	0.00	0.00	0.00	0.04
21	0.00	0.00	0.00	0.03
22	0.17	0.05	0.00	0.06
23	0.00	0.00	0.00	0.03
24	0.48	0.13	0.00	0.07
25	0.03	0.02	0.00	0.03
26	1.05	0.19	0.01	0.15
27	0.25	0.07	0.00	0.04
28	0.00	0.00	0.00	0.03
29	0.00	0.00	0.00	0.03
30	0.00	0.00	0.00	0.03
31	0.00	0.00	0.00	0.04
Totals	5.40	1.90	0.02	1.64

Appendix C (continued)

June

Date	Rainfall	Canopy Interception	Stem Flow	Runoff
1	0.51	0.08	0.01	0.11
2	0.00	0.00	0.00	0.06
3	0.01	0.01	0.00	0.03
4	0.22	0.04	0.00	0.08
5	0.00	0.00	0.00	0.04
6	0.00	0.00	0.00	0.03
7	0.00	0.00	0.00	0.03
8	0.00	0.00	0.00	0.03
9	0.03	0.02	0.00	0.03
10	0.43	0.16	0.00	0.10
11	0.00	0.00	0.00	0.04
12	0.00	0.00	0.00	0.04
13	0.00	0.00	0.00	0.04
14	0.00	0.00	0.00	0.04
15	0.45	0.16	0.00	0.05
16	0.00	0.00	0.00	0.04
17	0.00	0.00	0.00	0.04
18	0.03	0.03	0.00	0.03
19	0.08	0.06	0.00	0.03
20	0.00	0.00	0.00	0.03
21	0.00	0.00	0.00	0.03
22	0.00	0.00	0.00	0.03
23	0.00	0.00	0.00	0.03
24	0.00	0.00	0.00	0.03
25	0.00	0.00	0.00	0.02
26	0.64	0.26	0.01	0.03
27	0.00	0.00	0.00	0.03
28	0.00	0.00	0.00	0.02
29	0.19	0.19	0.00	0.02
30	0.00	0.00	0.00	0.02
Totals	2.59	1.01	0.02	1.19

Appendix C (continued)

July

Date	Rainfall	Canopy Interception	Stem Flow	Runoff
1	0.00	0.00	0.00	0.02
2	0.49	0.21	0.03	0.02
3	1.23	0.34	0.00	0.04
4	0.00	0.00	0.07	0.05
5	3.79	0.54	0.02	0.10
6	0.51	0.28	0.00	0.07
7	0.03	0.02	0.00	0.05
8	0.00	0.00	0.00	0.04
9	0.00	0.00	0.00	0.03
10	0.00	0.00	0.00	0.02
11	0.47	0.15	0.00	0.02
12	0.00	0.00	0.00	0.02
13	0.99	0.36	0.01	0.02
14	0.24	0.11	0.00	0.02
15	0.00	0.00	0.00	0.02
16	0.00	0.00	0.00	0.02
17	0.00	0.00	0.00	0.02
18	0.00	0.00	0.00	0.02
19	0.00	0.00	0.00	0.02
20	0.00	0.00	0.00	0.01
21	0.23	0.08	0.00	0.01
22	0.00	0.00	0.00	0.01
23	0.00	0.00	0.00	0.01
24	0.00	0.00	0.00	0.01
25	0.00	0.00	0.00	0.01
26	0.00	0.00	0.00	0.01
27	0.00	0.00	0.00	0.01
28	0.00	0.00	0.00	0.01
29	2.34	0.59	0.04	0.04
30	0.46	0.09	0.01	0.01
31	0.00	0.00	0.00	0.01
Totals	10.78	2.77	0.18	0.78

Appendix C (continued)

August

Date	Rainfall	Canopy Interception	Stem Flow	Runoff
1	0.00	0.00	0.00	0.01
2	0.00	0.00	0.00	0.01
3	0.00	0.00	0.00	0.01
4	0.10	0.07	0.00	0.01
5	0.00	0.00	0.00	0.01
6	0.00	0.00	0.00	0.01
7	0.06	0.05	0.00	0.01
8	0.00	0.00	0.00	0.01
9	0.00	0.00	0.00	0.01
10	0.00	0.00	0.00	0.01
11	0.00	0.00	0.00	0.01
12	4.84	0.37	0.05	0.02
13	0.00	0.00	0.00	0.01
14	0.18	0.11	0.00	0.01
15	0.79	0.19	0.00	0.01
16	0.03	0.02	0.00	0.01
17	0.28	0.18	0.00	0.01
18	0.00	0.00	0.00	0.01
19	0.07	0.05	0.00	0.01
20	0.00	0.00	0.00	0.01
21	0.05	0.04	0.00	0.01
22	2.71	0.82	0.02	0.04
23	0.61	0.15	0.01	0.03
24	0.00	0.00	0.00	0.01
25	0.77	0.19	0.00	0.01
26	0.00	0.00	0.00	0.01
27	0.00	0.00	0.00	0.01
28	0.00	0.00	0.00	0.01
29	0.36	0.21	0.00	0.01
30	0.00	0.00	0.00	0.01
31	0.00	0.00	0.00	0.01
Totals	10.85	2.45	0.08	0.41

Appendix C (continued)

September

Date	Rainfall	Canopy Interception	Stem Flow	Runoff
1	0.00	0.00	0.00	0.01
2	0.00	0.00	0.00	0.01
3	0.00	0.00	0.00	0.00
4	0.72	0.28	0.00	0.01
5	0.36	0.15	0.00	0.01
6	0.00	0.00	0.00	0.00
7	0.10	0.07	0.00	0.00
8	0.89	0.31	0.00	0.02
9	0.00	0.00	0.00	0.01
10	0.18	0.08	0.00	0.01
11	0.03	0.03	0.00	0.01
12	0.00	0.00	0.00	0.00
13	0.00	0.00	0.00	0.00
14	0.00	0.00	0.00	0.00
15	0.00	0.00	0.00	0.00
16	0.00	0.00	0.00	0.00
17	0.02	0.02	0.00	0.00
18	0.00	0.00	0.00	0.00
19	0.00	0.00	0.00	0.00
20	0.00	0.00	0.00	0.00
21	0.00	0.00	0.00	0.00
22	0.00	0.00	0.00	0.00
23	0.00	0.00	0.00	0.00
24	0.00	0.00	0.00	0.00
25	0.05	0.05	0.00	0.00
26	1.09	0.31	0.01	0.00
27	0.00	0.00	0.00	0.00
28	0.00	0.00	0.00	0.00
29	0.00	0.00	0.00	0.00
30	0.00	0.00	0.00	0.00
1 (Oct)	0.00	0.00	0.00	0.00
Totals	3.44	1.30	0.01	0.09

Appendix D

Above-Canopy Net Radiation on North and South Slopes for Select Days between May 6 and May 27.

Values in ly min. $^{-1}$ averaged for the hour.

Symbols: N.S. - North Slope
S.S. - South Slope

Hour Ending	May 7		May 13		May 14	
	N.S.	S.S.	N.S.	S.S.	N.S.	S.S.
1	-0.09	-0.08	-0.10	-0.09	-0.05	-0.04
2	-0.10	-0.11	-0.10	-0.08	-0.05	-0.04
3	-0.09	-0.09	-0.10	-0.07	-0.03	-0.04
4	-0.10	-0.10	-0.09	-0.07	-0.01	0.00
5	-0.09	-0.10	-0.02	-0.05	-0.03	0.00
6	-0.02	-0.10	0.02	0.00	-0.01	0.00
7	0.20	0.21	0.12	0.11	0.06	0.05
8	0.26	0.24	0.25	0.19	0.10	0.08
9	0.21	0.30	0.28	0.29	0.22	0.19
10	0.42	0.49	0.64	0.73	0.21	0.20
11	0.76	0.69	0.71	0.72	0.21	0.19
12	0.90	0.92	0.86	0.81	0.29	0.27
13	0.79	0.89	0.51	0.54	0.32	0.31
14	0.83	0.96	0.22	0.23	0.30	0.33
15	0.70	0.87	0.08	0.09	0.21	0.22
16	0.46	0.72	0.05	0.04	0.21	0.21
17	0.18	0.46	0.01	0.04	0.16	0.17
18	0.01	0.01	-0.01	0.01	0.12	0.08
19	-0.02	-0.06	-0.02	-0.02	-0.05	-0.03
20	-0.03	-0.07	-0.05	-0.02	-0.09	-0.08
21	-0.03	-0.07	-0.05	-0.04	-0.11	-0.12
22	-0.03	-0.07	-0.06	-0.04	-0.13	-0.14
23	-0.03	-0.07	-0.06	-0.04	-0.12	-0.14
24	-0.03	-0.08	-0.05	-0.03	-0.12	-0.14

Appendix D (continued)

Hour Ending	May 15 N.S.	May 15 S.S.	May 19 N.S.	May 19 S.S.	May 26 N.S.	May 26 S.S.
1	-0.12	-0.14	0.00	0.00	-0.14	-0.14
2	-0.12	-0.14	0.00	0.00	-0.14	-0.14
3	-0.11	-0.14	0.00	0.00	-0.13	-0.14
4	-0.11	-0.14	0.00	0.00	-0.13	-0.14
5	-0.11	-0.13	0.00	0.01	-0.13	-0.11
6	-0.02	-0.05	0.01	0.02	0.03	0.05
7	0.33	0.15	0.22	0.21	0.35	0.31
8	0.59	0.40	0.39	0.43	0.61	0.59
9	0.78	0.64	0.76	0.78	0.84	0.81
10	0.94	0.83	0.86	0.84	0.98	0.94
11	0.95	0.91	0.87	0.83	1.05	1.00
12	0.99	1.03	0.90	0.94	0.74	0.76
13	1.07	1.05	0.96	0.98	1.00	1.06
14	0.86	0.88	1.00	0.96	0.99	0.96
15	0.60	0.56	0.73	0.71	0.71	0.78
16	0.66	0.70	0.79	0.77	0.44	0.45
17	0.31	0.33	0.38	0.30	0.15	0.12
18	0.24	0.22	0.16	0.16	0.08	0.05
19	-0.01	-0.01	-0.10	-0.08	-0.01	0.04
20	-0.05	-0.06	-0.10	-0.12	0.00	0.00
21	-0.05	-0.05	-0.12	-0.13	0.00	0.00
22	-0.03	-0.03	-0.15	-0.14	0.00	0.00
23	-0.04	-0.03	-0.15	-0.14	0.00	0.00
24	-0.04	-0.05	-0.15	-0.14	0.00	0.00

Appendix E (i)

Average Values of Net Radiation at Various Levels within Forest on South Slope Location

Values in ly min. $^{-1}$ averaged for the hour.

Symbols: UC - Under Canopy
AC - Above Canopy
UC/AC - Ratio of under canopy to above canopy net radiation at each level.
1M, 2M, etc. - Height in meters of under canopy net radiometers.
ND - No Data

Hour	1M		2M		4M		8M		10M	
Ending	UC	AC	UC	AC	UC	AC	UC	AC	UC	AC
1	-0.02	-0.08	-0.01	-0.07	-0.02	-0.10	-0.03	-0.08	-0.04	-0.07
2	-0.02	-0.08	-0.02	-0.07	-0.02	-0.10	-0.04	-0.08	-0.05	-0.08
3	-0.02	-0.07	-0.01	-0.07	-0.02	-0.10	-0.04	-0.09	-0.03	-0.06
4	-0.01	-0.05	-0.02	-0.07	-0.02	-0.10	-0.04	-0.09	-0.04	-0.06
5	-0.02	-0.04	-0.01	-0.06	-0.01	-0.07	-0.03	-0.04	-0.03	-0.06
6	-0.01	-0.03	0.00	-0.02	-0.01	0.05	0.04	0.16	0.01	0.06
7	0.01	0.27	0.00	0.04	0.02	0.24	0.04	0.14	0.04	0.19
8	0.03	0.52	0.02	0.23	0.05	0.45	0.08	0.24	0.07	0.34
9	0.03	0.55	0.02	0.13	0.06	0.68	0.11	0.40	0.25	0.61
10	0.05	0.46	0.03	0.17	0.08	0.64	0.10	0.38	0.15	0.74
11	0.05	0.50	0.03	0.46	0.07	0.61	0.16	0.52	0.30	0.78
12	0.04	0.53	0.04	0.51	0.09	0.80	0.14	0.62	0.20	0.87
13	0.04	0.53	0.04	0.59	0.08	0.89	0.16	0.58	0.16	0.86
14	0.04	0.28	0.03	0.50	0.10	0.69	0.17	0.50	0.31	0.76
15	0.05	0.33	0.03	0.42	0.09	0.60	0.18	0.40	0.31	0.67
16	0.04	0.23	0.02	0.23	0.05	0.44	0.12	0.26	0.29	0.50
17	0.02	0.14	0.01	0.15	0.06	0.22	0.07	0.17	0.30	0.31
18	0.01	0.00	0.01	0.05	0.03	0.07	0.05	0.11	0.08	0.10
19	0.00	-0.04	0.00	-0.02	0.00	-0.04	-0.01	-0.03	0.00	0.00
20	-0.01	-0.09	-0.01	-0.05	-0.01	-0.08	-0.02	-0.08	-0.02	-0.04
21	-0.01	-0.10	-0.01	-0.06	-0.01	-0.08	-0.02	-0.09	-0.03	-0.06
22	-0.01	-0.10	-0.02	-0.05	-0.01	-0.08	-0.02	-0.08	-0.03	-0.07
23	-0.01	-0.10	-0.02	-0.06	-0.01	-0.09	-0.02	-0.09	-0.03	-0.08
24	-0.02	-0.09	-0.02	-0.07	-0.02	-0.10	-0.03	-0.08	-0.03	-0.06
Mean Daytime UC/AC		0.08		0.07		0.12		0.32		0.37
Mean Nightime UC/AC		0.13		0.17		0.11		0.43		0.50

Appendix E (i) (continued)

Hour Ending	11M UC	11M AC	12M UC	12M AC	13M UC	13M AC	Average AC
1	-0.05	-0.11	ND	ND	-0.12	-0.10	-0.09
2	-0.05	-0.11	ND	ND	-0.12	-0.09	-0.09
3	-0.04	-0.08	ND	ND	-0.10	-0.09	-0.08
4	-0.04	-0.07	ND	ND	-0.11	-0.09	-0.08
5	-0.04	-0.05	ND	ND	-0.08	-0.04	-0.05
6	-0.01	0.09	0.05	0.11	0.05	0.10	0.09
7	0.05	0.32	0.08	0.29	0.10	0.16	0.21
8	0.13	0.47	0.15	0.48	0.26	0.45	0.40
9	0.38	0.68	0.15	0.77	0.58	0.70	0.58
10	0.39	0.89	0.58	0.93	0.65	0.77	0.65
11	0.66	0.95	0.29	0.69	0.66	0.76	0.66
12	0.46	1.01	0.15	0.70	0.63	0.77	0.73
13	0.21	0.99	0.24	0.56	0.59	0.72	0.72
14	0.18	0.85	0.46	0.66	0.51	0.66	0.61
15	0.38	0.72	0.38	0.50	0.43	0.50	0.52
16	0.24	0.44	0.16	0.33	0.30	0.36	0.37
17	0.17	0.25	0.08	0.16	0.20	0.22	0.20
18	0.07	0.13	0.07	0.15	0.12	0.12	0.12
19	-0.01	-0.07	0.00	0.01	-0.06	-0.04	-0.03
20	-0.02	-0.08	0.00	-0.01	-0.10	-0.11	-0.06
21	-0.03	-0.09	0.00	-0.02	-0.12	-0.12	-0.08
22	-0.03	-0.09	ND	ND	-0.12	-0.12	-0.09
23	-0.04	-0.09	ND	ND	-0.12	-0.11	-0.09
24	-0.04	-0.09	ND	ND	-0.11	-0.10	-0.09

	11M	12M	13M
Mean Daytime UC/AC	0.42	0.45	0.81
Mean Nightime UC/AC	0.50	ND	1.10

Appendix E(ii)

Average Values of Net Radiation at Various Levels within Forest on North Slope Location

Values in ly min. $^{-1}$ averaged for the hour.

Symbols: UC - Under Canopy
AC - Above Canopy
UC/AC - Ratio of under canopy to above canopy net radiation at each level.
1M, 2M, etc. - Height in meters of under canopy net radiometers.
ND - No Data.

Hour	1M		2M		4M		8M		10M	
Ending	UC	AC	UC	AC	UC	AC	UC	AC	UC	AC
1	0.00	-0.10	0.00	-0.06	0.00	-0.11	-0.01	-0.08	-0.03	-0.10
2	0.00	-0.09	0.00	-0.09	0.00	-0.11	-0.01	-0.08	-0.03	-0.10
3	0.00	-0.06	0.00	-0.08	0.00	-0.11	-0.01	-0.08	-0.03	-0.09
4	0.00	-0.01	0.00	-0.08	0.00	-0.11	-0.01	-0.09	-0.03	-0.09
5	0.00	0.00	0.00	-0.05	0.00	-0.08	-0.01	-0.07	-0.03	-0.06
6	0.00	0.03	0.00	0.05	0.00	-0.02	0.00	-0.02	-0.02	-0.01
7	0.02	0.25	0.02	0.34	ND	ND	0.02	0.17	0.00	0.19
8	0.03	0.51	0.03	0.57	ND	ND	0.04	0.48	0.03	0.43
9	ND	ND	0.04	0.81	ND	ND	0.06	0.65	0.15	0.57
10	0.03	0.95	0.04	0.67	0.03	1.02	0.08	0.71	0.11	0.86
11	0.04	1.03	0.05	0.75	0.09	1.10	0.10	0.86	0.11	0.98
12	0.06	1.05	0.07	0.73	0.08	1.12	0.12	0.89	0.31	1.03
13	0.04	1.02	0.07	0.67	0.08	1.09	0.12	0.96	0.22	1.00
14	0.05	0.92	0.07	0.77	0.11	1.00	0.12	0.79	0.17	0.88
15	0.06	0.73	0.06	0.58	0.13	0.54	0.19	0.63	0.36	0.71
16	0.06	0.54	0.05	0.44	0.11	0.37	0.13	0.47	0.32	0.48
17	0.05	0.32	0.07	0.34	0.08	0.27	0.10	0.30	0.13	0.25
18	0.04	0.11	0.03	0.10	0.09	0.15	0.05	0.12	0.03	0.09
19	0.03	-0.09	0.02	-0.03	0.01	0.02	0.01	-0.03	0.00	-0.04
20	0.00	-0.11	0.02	-0.07	0.00	-0.07	-0.01	-0.07	-0.03	-0.08
21	0.00	-0.11	0.00	-0.08	0.00	-0.08	-0.01	-0.07	-0.03	-0.09
22	0.00	-0.11	0.00	-0.08	0.00	-0.08	-0.01	-0.08	-0.03	-0.09
23	0.00	-0.11	0.00	-0.07	0.00	-0.09	-0.01	-0.08	-0.03	-0.10
24	0.00	-0.10	0.00	-0.06	0.00	-0.10	-0.01	-0.08	-0.03	-0.10
Mean Daytime UC/AC	0.06		0.09		0.12		0.16		0.26	
Mean Nighttime UC/AC	0.00		0.00		0.00		0.14		0.33	

Appendix E (ii) (continued)

Hour Ending	11M UC	11M AC	13M UC	13M AC	14M UC	14M AC	Average AC
1	-0.03	-0.08	-0.05	-0.08	-0.10	-0.11	-0.09
2	-0.03	-0.08	-0.05	-0.08	-0.08	-0.08	-0.09
3	-0.03	-0.08	-0.04	-0.07	-0.09	-0.10	-0.09
4	-0.03	-0.07	-0.03	-0.07	-0.09	-0.10	-0.07
5	-0.03	-0.05	-0.02	-0.05	-0.05	-0.06	-0.05
6	0.02	0.00	-0.01	-0.01	0.00	-0.02	0.00
7	0.03	0.19	0.10	0.22	0.25	0.22	0.26
8	0.05	0.47	0.13	0.56	0.47	0.54	0.54
9	0.17	0.58	ND	ND	0.74	0.78	0.70
10	0.15	0.77	ND	ND	0.64	0.69	0.83
11	0.28	0.89	1.05	0.95	0.89	0.92	0.94
12	0.56	0.96	1.09	1.05	0.80	0.84	0.95
13	0.44	0.94	1.07	1.05	0.83	0.73	0.91
14	0.34	0.80	0.73	0.86	0.73	0.81	0.81
15	0.22	0.58	0.14	0.64	0.54	0.52	0.60
16	0.15	0.38	0.12	0.43	0.41	0.37	0.44
17	0.09	0.22	0.07	0.32	0.23	0.21	0.27
18	0.03	0.09	0.00	0.13	0.13	0.07	0.10
19	0.00	-0.05	-0.05	-0.03	-0.01	-0.04	-0.03
20	-0.02	-0.06	-0.06	-0.08	-0.07	-0.08	-0.07
21	-0.03	-0.07	-0.05	-0.08	-0.09	-0.09	-0.08
22	-0.03	-0.08	-0.05	-0.08	-0.10	-0.11	-0.09
23	-0.03	-0.08	-0.05	-0.08	-0.10	-0.11	-0.09
24	-0.03	-0.08	-0.05	-0.08	-0.10	-0.11	-0.09

	11M	13M	14M
Mean Daytime UC/AC	0.37	0.73	1.00
Mean Nightime UC/AC	0.43	0.63	1.00

Appendix F (i)

Average Values of Incoming Solar Radiation at Various Levels within Forest on South Slope Location

Values in ly min. $^{-1}$ averaged for the hour.

Symbols: UC - Under Canopy
AC - Above Canopy
UC/AC - Ratio of under canopy to above canopy solar radiation at each level.
1M,2M, etc. - Height in meters of under canopy solarimeter.
ND - No Data

Hour Ending	0.3M		1M		2M		4M	
	UC	AC	UC	AC	UC	AC	UC	AC
1	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
2	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
3	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
4	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
5	0.00	0.05	0.00	0.02	0.00	0.00	0.00	0.02
6	0.02	0.19	0.00	0.13	0.01	0.02	0.02	0.05
7	0.05	0.44	0.00	0.37	0.02	0.08	0.04	0.34
8	0.12	0.72	0.03	0.64	0.03	0.15	0.08	0.56
9	ND	1.03	0.05	0.63	0.02	0.23	ND	0.82
10	0.05	1.17	0.05	0.66	0.06	0.46	0.13	0.71
11	0.06	1.25	0.07	0.71	0.09	0.67	0.15	0.84
12	0.05	1.33	0.07	0.56	0.10	0.71	0.22	0.98
13	0.14	1.30	0.06	0.73	0.11	0.82	0.20	1.11
14	0.02	1.20	0.06	0.53	0.13	0.74	0.16	0.81
15	0.10	1.03	0.19	0.83	0.07	0.63	0.24	0.96
16	0.06	0.82	0.07	0.63	0.04	0.42	0.15	0.71
17	0.04	0.60	0.02	0.33	0.03	0.32	0.04	0.37
18	0.03	0.37	0.00	0.21	0.01	0.17	0.03	0.20
19	0.01	0.14	0.00	0.12	0.01	0.05	0.01	0.07
20	0.00	0.01	0.00	0.01	0.00	0.00	0.00	0.01
21	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
22	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
23	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
24	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
UC/AC	0.06		0.09		0.13		0.18	

Appendix F(i) (continued)

Hour Ending	8M UC	8M AC	10M UC	10M AC	Average AC
1	0.00	0.00	0.00	0.00	0.00
2	0.00	0.00	0.00	0.00	0.00
3	0.00	0.00	0.00	0.00	0.00
4	0.00	0.00	0.00	0.00	0.00
5	0.00	0.02	0.01	0.04	0.03
6	0.02	0.11	0.06	0.20	0.13
7	0.06	0.22	0.13	0.36	0.30
8	0.11	0.34	0.52	0.62	0.51
9	0.22	0.40	0.40	0.83	0.66
10	0.14	0.49	0.52	0.98	0.74
11	0.20	0.71	0.29	0.88	0.84
12	0.30	0.84	0.42	1.02	0.91
13	0.40	0.81	0.23	0.98	0.96
14	0.36	0.74	0.30	1.00	0.84
15	0.27	0.60	0.36	0.82	0.81
16	0.32	0.43	0.19	0.63	0.61
17	0.18	0.37	0.13	0.38	0.39
18	0.04	0.23	0.08	0.28	0.24
19	0.01	0.09	0.03	0.11	0.10
20	0.00	0.01	0.01	0.01	0.01
21	0.00	0.00	0.00	0.00	0.00
22	0.00	0.00	0.00	0.00	0.00
23	0.00	0.00	0.00	0.00	0.00
24	0.00	0.00	0.00	0.00	0.00
UC/AC	0.41		0.40		

Appendix F(ii)

Average Values of Incoming Solar Radiation at Various Levels within Forest on North Slope Location

Values in ly min. $^{-1}$ averaged for the hour.

Symbols: UC - Under Canopy
AC - Above Canopy
UC/AC - Ratio of under canopy to above canopy solar radiation at each level.
1M, 2M, etc. - Height in meters of under canopy solarimeter.
ND - No Data.

Hour Ending	0.3M		1M		2M		6M	
	UC	AC	UC	AC	UC	AC	UC	AC
1	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
2	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
3	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
4	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
5	0.00	0.04	0.00	0.04	0.00	0.03	0.00	0.04
6	0.00	0.21	0.00	0.26	0.00	0.04	0.00	0.25
7	0.02	0.44	0.01	0.45	0.00	0.11	0.02	0.50
8	0.03	0.50	0.03	0.59	0.00	0.16	0.05	0.74
9	0.04	0.88	ND	ND	0.03	0.05	0.06	0.96
10	0.06	0.78	ND	ND	0.05	0.77	0.13	1.13
11	0.07	0.89	ND	ND	0.05	1.05	0.10	1.25
12	0.05	0.50	0.14	1.30	0.06	1.20	0.18	1.29
13	0.04	0.41	0.08	1.28	0.07	1.17	0.18	1.28
14	0.08	0.88	0.07	1.19	0.08	0.86	0.17	1.21
15	0.06	0.58	0.11	1.05	0.11	0.90	0.09	0.67
16	0.06	0.56	0.08	0.86	0.10	0.77	0.11	0.52
17	0.07	0.53	0.06	0.63	0.05	0.52	0.06	0.40
18	0.02	0.23	0.04	0.37	0.01	0.20	0.04	0.28
19	0.00	0.08	0.01	0.14	0.00	0.08	0.02	0.12
20	0.00	0.01	0.00	0.01	0.00	0.01	0.00	0.01
21	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
22	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
23	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
24	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
UC/AC	0.08		0.08		0.08		0.11	

Appendix F (ii) (continued)

Hour Ending	8M UC	8M AC	11M UC	11M AC	Average AC
1	0.00	0.00	0.00	0.00	0.00
2	0.00	0.00	0.00	0.00	0.00
3	0.00	0.00	0.00	0.00	0.00
4	0.00	0.00	0.00	0.00	0.00
5	0.00	0.02	0.01	0.20	0.03
6	0.00	0.13	0.02	0.16	0.16
7	0.01	0.27	0.05	0.38	0.36
8	0.02	0.46	0.08	0.58	0.51
9	0.04	0.63	0.24	0.78	0.66
10	0.06	0.75	0.27	0.95	0.87
11	0.09	0.91	0.47	1.18	1.05
12	0.13	0.97	0.91	1.12	1.06
13	0.12	1.08	0.66	1.13	1.06
14	0.15	0.99	0.41	1.01	1.02
15	0.09	0.94	0.21	0.84	0.83
16	0.06	0.82	0.16	0.65	0.70
17	0.06	0.54	0.14	0.46	0.51
18	0.05	0.30	0.14	0.29	0.28
19	0.01	0.08	0.04	0.11	0.10
20	0.00	0.01	0.00	0.02	0.01
21	0.00	0.00	0.00	0.00	0.00
22	0.00	0.00	0.00	0.00	0.00
23	0.00	0.00	0.00	0.00	0.00
24	0.00	0.00	0.00	0.00	0.00
UC/AC	0.10		0.39		

Appendix G (i)

Diurnal Radiation Regime within the 0-9M Level of the Forest

All values in ly min. $^{-1}$ averaged for the hour

Hour Ending	Solar Radiation	Albedo	Absorbed Solar Radiation	Net Radiation	Long Wave Balance
1	0.00	0.00	0.00	-0.01	-0.01
2	0.00	0.00	0.00	-0.01	-0.01
3	0.00	0.00	0.00	-0.01	-0.01
4	0.00	0.00	0.00	-0.01	-0.01
5	0.00	0.00	0.00	-0.01	-0.01
6	0.01	0.00	0.01	-0.01	-0.01
7	0.03	0.00	0.03	0.01	-0.02
8	0.07	0.01	0.06	0.04	-0.02
9	0.09	0.01	0.08	0.05	-0.03
10	0.13	0.02	0.11	0.06	-0.05
11	0.13	0.02	0.11	0.08	-0.03
12	0.20	0.03	0.17	0.09	-0.08
13	0.19	0.03	0.16	0.08	-0.08
14	0.17	0.02	0.15	0.11	-0.04
15	0.17	0.02	0.15	0.11	-0.04
16	0.13	0.02	0.11	0.08	-0.03
17	0.08	0.01	0.07	0.07	0.00
18	0.07	0.01	0.06	0.06	0.00
19	0.02	0.00	0.02	0.01	-0.01
20	0.00	0.00	0.00	-0.01	-0.01
21	0.00	0.00	0.00	-0.01	-0.01
22	0.00	0.00	0.00	-0.01	-0.01
23	0.00	0.00	0.00	-0.01	-0.01
24	0.00	0.00	0.00	-0.01	-0.01

Appendix G(ii)

Diurnal Radiation Regime within the 9-14M Level of the Forest

All values in ly min.$^{-1}$ averaged for the hour

Hour Ending	Solar Radiation	Albedo	Absorbed Solar Radiation	Net Radiation	Long Wave Balance
1	0.00	0.00	0.00	-0.04	-0.04
2	0.00	0.00	0.00	-0.04	-0.04
3	0.00	0.00	0.00	-0.04	-0.04
4	0.00	0.00	0.00	-0.04	-0.04
5	0.01	0.00	0.01	-0.04	-0.04
6	0.04	0.01	0.03	0.01	-0.02
7	0.09	0.02	0.07	0.04	-0.03
8	0.30	0.05	0.25	0.09	-0.16
9	0.32	0.06	0.26	0.28	-0.02
10	0.40	0.07	0.33	0.27	-0.06
11	0.38	0.07	0.31	0.47	-0.16
12	0.67	0.12	0.55	0.51	-0.04
13	0.45	0.08	0.37	0.33	-0.04
14	0.36	0.06	0.30	0.26	-0.04
15	0.29	0.05	0.24	0.30	-0.06
16	0.18	0.03	0.15	0.20	-0.05
17	0.14	0.03	0.11	0.13	-0.02
18	0.11	0.02	0.09	0.05	-0.04
19	0.04	0.01	0.03	-0.01	-0.04
20	0.01	0.00	0.01	-0.02	-0.03
21	0.00	0.00	0.00	-0.03	-0.03
22	0.00	0.00	0.00	-0.03	-0.03
23	0.00	0.00	0.00	-0.04	-0.04
24	0.00	0.00	0.00	-0.04	-0.04

Appendix G(iii)

Diurnal Radiation Regime Above the Forest Canopy

All values in ly min. $^{-1}$ averaged for the hour

Hour Ending	Solar Radiation	Albedo	Absorbed Solar Radiation	Net Radiation	Long Wave Balance
1	0.00	0.00	0.00	-0.10	-0.10
2	0.00	0.00	0.00	-0.10	-0.10
3	0.00	0.00	0.00	-0.10	-0.10
4	0.00	0.00	0.00	-0.10	-0.10
5	0.02	0.00	0.02	-0.07	-0.09
6	0.12	0.02	0.10	0,01	-0.09
7	0.30	0.05	0.25	0.17	-0.08
8	0.52	0.09	0.43	0.36	-0.07
9	0.82	0.15	0.67	0.62	-0.05
10	0.89	0.16	0.73	0.65	-0.08
11	1.09	0.20	0.89	0.77	-0.12
12	1.10	0.20	0.o0	0.79	-0.11
13	1.16	0.21	0.95	0.84	-0.11
14	1.00	0.18	0.82	0.67	-0.15
15	0.92	0.17	0.75	0.59	-0.16
16	0.66	0.12	0.54	0.36	-0.18
17	0.50	0.09	0.41	0.24	-0.17
18	0.25	0.05	0.20	0.07	-0.13
19	0.08	0.01	0.07	-0.02	-0.09
20	0.01	0.00	0.01	-0.07	-0.08
21	0.00	0.00	0.00	-0.09	-0.09
22	0.00	0.00	0.00	-0.06	-0.06
23	0.00	0.00	0.00	-0.08	-0.08
24	0.00	0.00	0.00	-0.09	-0.09

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