WATER, HEAT and CROP GROWTH

FEDDES

Dit proefschrift met stellingen van

REINDER AUKE FEDDES,

landbouwkundig ingenieur, geboren te De Krim op 2 november 1939, is goedgekeurd door de promotoren, Dr. Ir. W. H. VAN DER MOLEN, hoogleraar in de agrohydrologie en Dr. Ir. J. SCHENK, hoogleraar in de natuur- en weerkunde.

De Rector Magnificus van de Landbouwhogeschool,
J. M. POLAK

Wageningen, 10 mei 1971

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WATER, HEAT AND CROP GROWTH

(met een samenvatting in het Nederlands)

PROEFSCHRIFT

TER VERKRIJGING VAN DE GRAAD
VAN DOCTOR IN DE LANDBOUWWETENSCHAPPEN
OP GEZAG VAN DE RECTOR MAGNIFICUS, MR. J. M. POLAK,
HOOGLERAAR IN DE RECHTS- EN STAATSWETENSCHAPPEN
VAN DE WESTERSE GEBIEDEN,
TE VERDEDIGEN TEGEN DE BEDENKINGEN VAN EEN
COMMISSIE UIT DE SENAAT
VAN DE LANDBOUWHOGESCHOOL TE WAGENINGEN
OP VRIJDAG 25 JUNI 1971 TE 16.00 UUR

DOOR

R. A. FEDDES

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STELLINGEN

I

Voor een juiste voorspelling van kieming en opkomst van gewassen kan zeer wel gebruik worden gemaakt van temperatuursommen.

GROBBELAAR, W. P. 1963. Responses of young maize plants to root temperatures. Proefschrift Landbouwhogeschool, Wageningen, pp. 71.

Dit proefschrift

II

Bij berekeningen die betrekking hebben op de warmtehuishouding van diep dan wel ondiep gedraineerde gronden, is de aanname dat hun minimum temperaturen gelijk zijn, niet geoorloofd.

DUIN, R. H. A. VAN. 1963. The influence of soil management on the temperature wave near the soil surface. Techn. Bull. ICW 29: pp. 21.

Dit proefschrift

III

Het draineren met buizen van de in Nederland gebruikelijke kleine diameters valt, zowel uit het oogpunt van werking als van onderhoud, te ontraden.

IV

Bij de berekening van drainafstanden met behulp van drainageformules dient rekening te worden gehouden met de intreeweerstand van het drainagesysteem.

WIDMOSER, P. 1968. Der Einfluss van Zonen geänderter Durchlässigkeit im Bereich von Drain- und Brunnenfilterrohren. Schweizerische Bauzeitung 86, 9:1–11.

V

Inpoldering van de Markerwaard zal zeer waarschijnlijk een invloed op het klimaat hebben die ongunstig is voor de bolopbrengst van tulpen in de Noordoostpolder.

VI

Het hogere produktieniveau dat gewassen in droge jaren bij voldoende watertoevoer bereiken, wordt meer veroorzaakt door de hogere temperaturen in het voorseizoen dan door de grotere hoeveelheden straling.

RIJTEMA, P. E. and G. ENDRÖDI. 1970. Calculation of production of potatoes. Neth. J. Agric. Sci. 18, 1:26–36.

VII

Uit de onderzoekingen van Hadas mag niet de conclusie worden getrokken dat de wet van Darcy niet geldt voor stroming in onverzadigde grond.

HADAS, A. 1964. Deviations from Darcy's law for the flow of water in unsaturated soils. Israel J. Agric. Res. 14:159–168.

VIII

Het gebruik van numerieke methoden ter bepaling van de temperatuurvereffeningscoëfficiënt is slechts in een beperkt aantal gevallen mogelijk.

Dit proefschrift

IX

In verband met de te verwachten produktieverhoging van tulpen bij de overgang van ruggenteelt naar beddenteelt is de ontwikkeling van een beddenrooier voor kleigronden dringend gewenst.

X

Voor de beschrijving van de energietoestand van een vloeistof dient de term potentiaal uitsluitend gereserveerd te worden voor de potentiële energie per eenheid van massa.

XI

Bestemmingsverandering van landbouwgronden tot natuur- of recreatiegebieden zal primair afhankelijk zijn van de ontwikkeling van nieuwe beheersvormen en van een regeling van de kosten van het toekomstig onderhoud.

STUDIEGROEP VOLTHE-DE LUTTE. 1971. De landinrichting van het gebied Volthe-De Lutte, verkenning, analyse en modellen. Wageningen.

XII

Het is onjuist dat door een makelaar bij een transactie in onroerend goed anderhalf procent provisie aan beide partijen wordt berekend.

Proefschrift van R. A. FEDDES
Wageningen, 25 juni 1971

Aan mijn ouders

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CONTENTS

I. INTRODUCTION

1. GENERAL

To a large extent the result of a farmer's efforts to get higher crop yields will be determined by the prevailing environmental conditions, i.e. by the existing complex of physical, chemical and biological factors. The possibilities of an efficient use of these factors are enlarged by our knowledge of their effects on agricultural production. Such knowledge is in first instance mostly gained from practical experience or field experiments often giving, however, only empirical information of local value.

To apply local results with success to analogous problems in other regions a fundamental treatment of research data is necessary, by which the occurring phenomena and processes can be analyzed more objectively. Nowadays such approaches are feasible, the more so because of the recently rapid increasing knowledge on the processes taking place in the soil-plant-atmosphere system. In this context the progress in knowledge during the past ten years on the movement of water in the unsaturated zone must be mentioned in particular.

In this paper attention will mainly be paid to the effects of the physical environmental conditions on plant development. As local research object, part of the economically important horticultural region of Geestmerambacht was taken (5300 ha; 1 ha = 2.47 acre), situated in the Netherlands in the northern part of the province of North-Holland. In this area 70% of the land lies in open horticulture, with as main crops early and store varieties of cabbage (covering 80% of the vegetable-growing area), early potatoes and onions. This one sided cropping pattern, of which the crops are mostly intended for export, offers only low economic possibilities with the result that for example in 1968 the total net return of the holdings in this area was negative.

There are a number of reasons for this situation. A large part of the area is covered with heavy sea clay soils, the so-called 'pikklei' or sticky clay soils (DU BURCK, 1957). These soils consist of heavy, very compacted, non-calcareous clay with a prismatic structure and are very susceptible to drought (table 1).

Usually 0.20 to 0.30 m heavy clay is underlain by a 0.60 m thick sticky clay. The granular composition of these layers is almost the same. Under the sticky clay layer a peaty horizon of 0.20 m deep is often found, this overlying a calcareous light sandy loam.

Cultivation, sowing and planting can be started only relatively late, the development of the crops is slow and often only one crop can be grown per

TABLE 1. Composition (weight per cent) of a sticky clay soil (partly after DU BURCK, 1957)

Depth m	Mineral particles < 2µm	2–16µm	16–58µm	58–150µm	150–300µm	> 300µm	Humus	$CaCO_3$	pH
0.00–0.30	46	25	22	7.5	0.6	0.2	3.7	0.1	6.7
0.30–0.60	47	23	22	6.4	0.2	0.1	2.4	0.2	7.1
0.60–0.80	51	26	18	4	0.3	0.5	3.1	1.1	7.4
0.80–1.00	41	19.5	21	16.5	1	2	11.5	0.6	7.3
1.00–1.20	12	8.5	47	31.5	0.2	–	2	13.7	7.6
1.20–1.80	10.2 <2µm	20.9 2-50µm	48.7 50–105µm	13.3 105–150µm	7.0 >150µm				

growing season. Furthermore, the accessibility of the fields in this region with its many water ways, fragmented parcelling and insufficient water management is poor and makes land exploitation difficult; in some cases the fields can only be reached by boat.

In this situation there is limited scope for the cultivation of more profitable intensively managed crops. Therefore, market gardeners asked for a consolidation scheme, including new roads, improved drainage and better workable soils. It was decided to carry out an integrated land consolidation program for the region and in connection with it to design and construct in 1958 a groundwater level experimental field at Oudkarspel with the purpose to investigate the possibilities of improving the heavy clay soils and the water management of the region (details on the layout of the experimental field are given in Chapter II). On this experimental field the original clay soil was improved by changing the succession of soil layers, after which the reactions of a number of horticultural crops were investigated during the years 1959 to 1966 at different constant depths of the groundwater table.

2. PREVIOUS INVESTIGATIONS

The previous investigations on the groundwater level experimental field (VAN DER VALK and SCHONEVELD, 1963, 1964; VAN DER VALK and NICOLAI, 1969; ANNUAL REPORTS, 1958 through 1966) had the purpose of finding relationships between yield and mean optimum groundwater depth during the growing season on the different profiles. Due to change in weather conditions the results differed of course from year to year. In general the optimum water level on the clay soil proved in dry years to be between 0.60–0.80 m minus surface for horticultural crops and between 0.40–0.60 m for bulbous crops, and in wet years for both categories approximately 1.00 m. For the improved clay soil, i.e. the sandy loam, a water level between 1.00–1.20 m minus surface appeared to be the most favourable, the occasional depressions were sharper, however. As a general rule it could also be established that late crops need a deeper water table than earlier crops.

The clay soil showed a number of disadvantages as compared with the sandy loam. Tillage depended much more on weather conditions and often had to start later. Germination, emergence and establishment of the crop met with great difficulties, because the top layer is very susceptible to drought. In autumn, mechanical harvesting was only possible under dry soil conditions. The yield was not always highest on the sandy loam, however, especially in the case of late crops. The growth on this profile is faster early in the season, but generally stops in autumn, while on the clay soil growth is still going on.

Based on these earlier experiments, advice was given on the most desirable water level in the land consolidation area of Geestmerambacht, resulting in a freeboard of 1.25 m below the future land surface of the open water conduits.

3. SCOPE OF PRESENT INVESTIGATIONS

The main aim of the present study is to give a more fundamental treatment of some environmental conditions influencing plant growth, based on data from the Geestmerambacht area. This treatment could be realized by taking into account the laws of conservation of matter and energy of the plant environment, i.e. by considering by means of balance approaches what happens to mass fluxes and energy fluxes reaching the surface. With a balance approach the unknown flux can be derived from detailed measurements and/or theoretical analyses of the remaining component fluxes. With lysimeters, for example, the evaporation flux from a surface can be obtained from the water balance equation when the fluxes of precipitation, capillary rise, subsurface outflow and change in soil moisture content are measured. Such comprehensive balance studies make it possible to determine the total influence the physical conditions of the environment have upon plant growth by means of detailed quantitative analyses of the transport processes occurring in the soil-plant-atmosphere system. Necessary in this context is an evaluation of the physical properties of the system by which the transport processes are determined.

In Chapter II a description of the layout and construction of the groundwater level experimental field is given. Because this field initially did not lend itself to water balance studies, a special type of lysimeter was developed and installed on the experimental field, the details of which are also presented.

In Chapter III the principles of the transport of water through the soil via the plant into the atmosphere are treated. Part III A deals with the flow of water through the unsaturated zone of the soil. The results of the measurements of the various physical parameters involved are presented. The amount of water available for evaporation during a number of days and at different depths of groundwater table is calculated for different soil profiles. In part III B the estimation of evaporation from a combined energy balance – vapour transport approach which is based on rather easily measurable meteorological as well soil and crop quantities is discussed. The in this way derived evaporation data of different crops are compared with evaporation data obtained by means of water balance approaches. Special attention is given to the problem of estimation of net radiation and reflection. Finally a mathematical description of water uptake by roots is derived and evaluated.

Another important physical condition of the soil is its temperature regime. A detailed treatment of heat transfer in soil, including the determination of the thermal properties involved, is presented in Chapter IV. Temperature effects in relation to soil profile and groundwater table depth are quantified and discussed.

Chapter V is devoted to a synthesis of the combined effects of the mentioned physical processes on plant growth. To that end two stages of plant development are distinguished: the stage from germination to seedling emergence and the stage from seedling emergence to maturity. For the first stage the effects of soil temperature and soil moisture content are evaluated and optimum conditions are formulated. For the second stage photosynthesis rates are compared with theoretically calculated rates of a standard crop. Both photosynthetic as well as water use efficiencies of a number of crops are discussed. Based on the concept evolved for moisture available for evaporation, some earlier results of the experimental field are evaluated and optimum amounts of water for maximum fresh yield production of some crops are established. From these latter data and from given frequency distributions of rainfall, lowest and highest admissable as well as optimum groundwater table depths are calculated.

In the manner described above it has been tried to develop some theoretical concepts and calculation methods to quantify the effects of some environmental conditions on seedling emergence and crop production, which can possibly be used in evaluating existing as well as planned water, soil and crop management practices.

II. EXPERIMENTAL FIELD AND LYSIMETERS

1. GROUNDWATER LEVEL EXPERIMENTAL FIELD

As already mentioned in the Introduction, the soil improvement of part of the experimental field Geestmerambacht was carried out in 1958 and had the purpose to improve the arable top layer as well as eliminating the negative influences of the sticky clay layer. Three objects were laid out in replicate (see fig. 1 and 3) of which the first is an untreated reference object (A), also used to evaluate the influence of groundwater depth on the quality of the original sticky clay. In the second object a re-assortment of layers was realized by transferring the top layer and the underlying sticky clay layer to 0.90 m minus original soil surface and bringing up the very fine sandy loam from beneath this depth (C)*. The third type of profile (D) was obtained in the same way as profile C, with the exception that the original clay top layer was removed and brought back to prevent puddling, as was expected to occur on profile C. In future the original sticky clay profile (A) will be denoted as clay, the very fine sandy loam (C) as sandy loam and the third profile (D) as clay on sandy loam.

As to the water management, first a plan with inclining groundwater levels was developed. After testing, this plan was rejected because the permeability factors proved to be too low and the differences in soil fertility could not sufficiently be eliminated in this way. From a number of other alternatives, a system with stepped groundwater levels as developed by Ir. W. C. VISSER of the Institute for Land and Water Management Research at Wageningen, was finally chosen. A general view of the set-up is given in fig. 1 (VAN DER VALK,

* An intended B-profile, breaking the sticky clay layer in situ, was never realized.

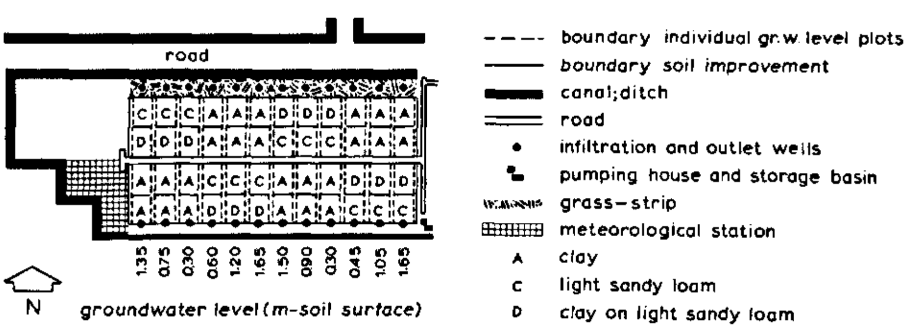


FIG. 1. *Layout of the experimental field.*

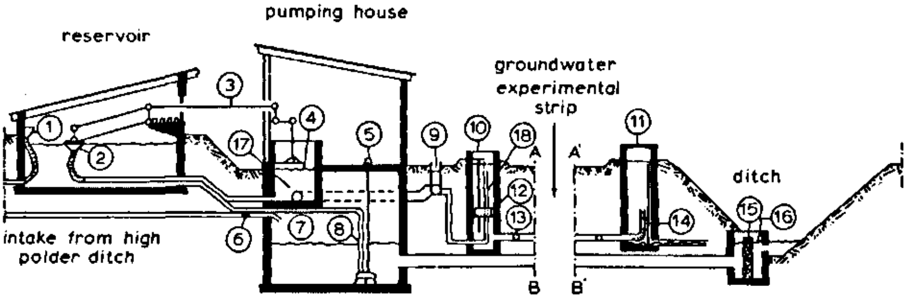
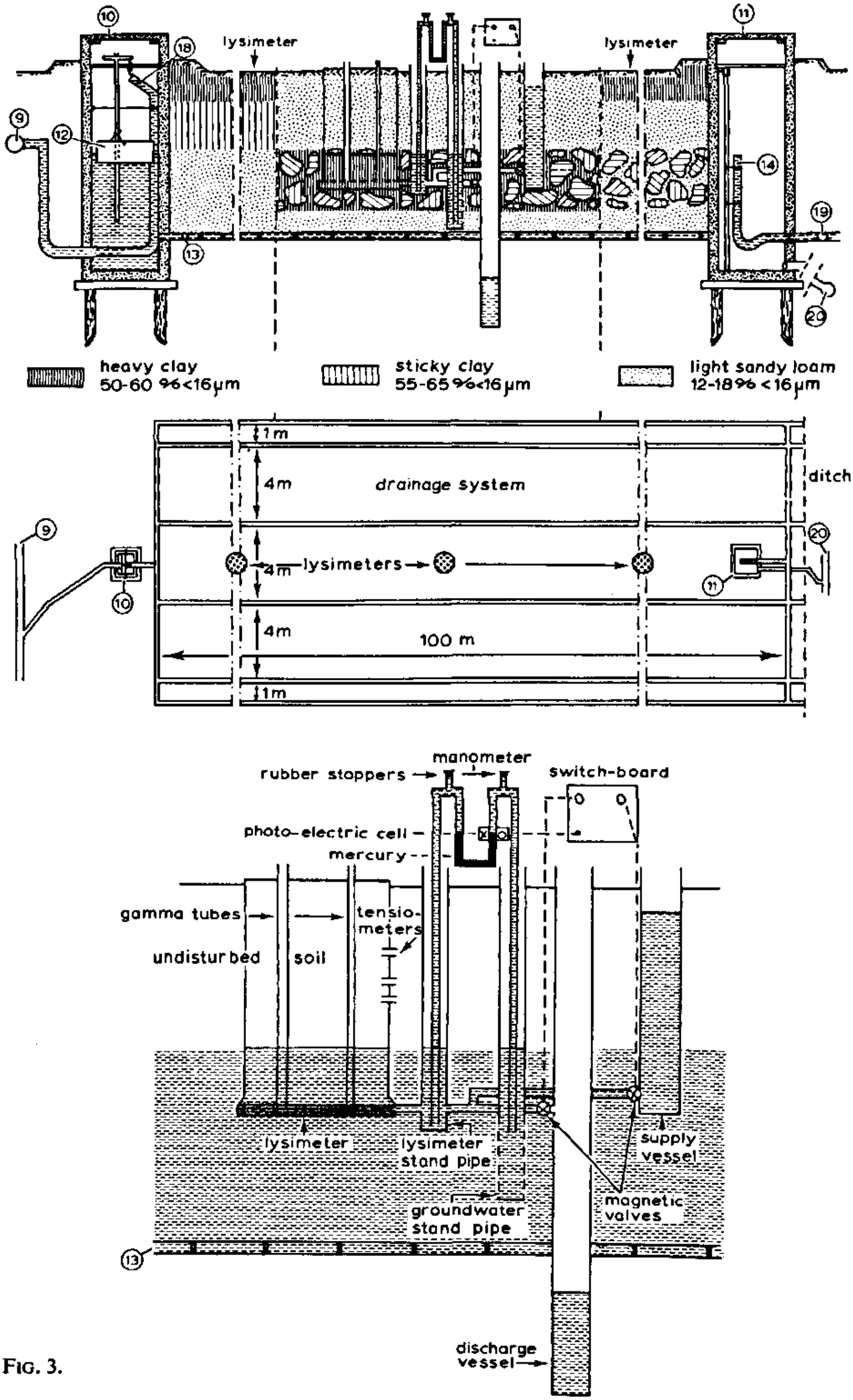


FIG. 2. *Cross-section of the water level regulating system (along the length of an experimental strip) at the experimental field. 1, overflow; 2, floating overflow; 3, master rod; 4, master float; 5, pump; 6, valve; 7, delivery pipe; 8, nonreturn valve; 9, infiltration main; 10, infiltration well; 11, outlet well; 12, floating reduction valve; 13, groundwater level drain; 14, overflow; 15, coke filter; 16, servo-switch; 17, concrete tank; 18, supply pipe.*

1961). Groundwater levels are stepped with 0.15 m steps from 1.65 m to 0.30 m minus surface. From fig. 1 it can be seen that the largest difference in phreatic level between two groundwater plots is 0.60 m. The groundwater levels are maintained independently and automatically in strips of 14 × 100 m². Each groundwater section (each including the three profiles) contains a subirrigation system of 6 tile lines laid at 1.80 m depth with spacings of 4 m in the centre and of 1 m at the borders (fig. 3). The strips are separated from each other by bufferstrips of 8 m wide. A cross-section of the water level regulating system is given in fig. 2. Via a floating overflow (2) in the reservoir and a concrete tank (17) in the pumping house, water is let in from the reservoir to the infiltration main (9) to which all infiltration wells are connected. The supply is regulated by a master float (4) in the concrete tank in such a way that the water level in the infiltration main and therefore also in the supply pipes (18) of the infiltration wells is almost constant. A decline of the groundwater level in the strip by evaporation causes a decline of the water level in the infiltration well (10) and by that of a floating reduction valve (12) which is connected to the supply pipe (18). This pipe is turnable at the upper end, so the more the reduction valve is going down the more water is supplied to the infiltration well (see also fig. 3).

A rise of the groundwater level by rainfall or subsurface inflow will give a discharge via an overflow (14) to the lower ditch and this water is returned via a coke filter (15) to the pumping house, where an electric pump (5) governed by

FIG. 3. *Some appliances at the experimental field, from top to bottom: cross-section of the 0.90 m groundwater plot with built in lysimeter; view of the tile drainage system at 1.80 m depth of the 0.90 m groundwater plot and the site of its three lysimeters; cross-section of one of the lysimeters. For the meaning of the circled numbers from 9 through 18, see fig. 2; 19, outlet well drain; 20, discharge main.*



lysimeter
lysimeter
heavy clay
50-60 %<16µm
sticky clay
55-65 %<16µm
light sandy loam
12-18 % < 16µm
1m
4m
drainage system
ditch
4m
lysimeters
4m
100 m
1m
manometer
rubber stoppers
switch-board
photo-electric cell
mercury
gamma tubes
tensiometers
undisturbed
soil
lysimeter
lysimeter stand pipe
supply vessel
groundwater stand pipe
magnetic valves
discharge vessel

FIG. 3.

a floating servo-switch (16) pumps the water back to the reservoir. To prevent leakages from the high groundwater level strips through the grass covered border strip to the lower ditch, a discharge main (20) which is connected to all outlet wells was in a later stage laid on the bottom of the lower ditch to take care of the discharge. Now the water level in the lower ditch can be raised to prevent the mentioned leakages. In periods of water surplus the overflow in the reservoir (1) is working, in periods of water shortage water is let in from the high polder ditch (6) via the direct inlet.

2. LYSIMETERS

For two main reasons the construction of the experimental field makes it impossible to set up a closed water balance for each individual strip. In the first place it is not possible to determine how much water is taken by means of capillary rise from the groundwater by each of the three different profiles separately. Secondly, in spite of the buffer strips there still is seepage from neighbouring sections when they have a higher groundwater level; this quantity can hardly be measured or calculated.

To rectify these disadvantages, the present author in 1966 designed a non-weighable lysimeter in which the same groundwater level, including the minor fluctuations is maintained as exists in the surrounding field (fig. 3). Plastic containers of 0.80 m diameter and 1.30 m height, containing the specific undisturbed soil profile, have been installed by means of hydraulic pressure in each of the three profiles of the 0.90 m as well as in those of the 1.20 m groundwater level strip. The groundwater level in the field and the water level in the container are registrated by groundwater standpipes. The lysimeter standpipe is erected on the tube connecting the container with a supply and discharge vessel. On the two standpipes a mercury manometer system* was built, provided with a photo-electric cell (23). Differences in water level between lysimeter and field are transferred to the manometer system. Then the photo-electric cell via the switch-board gives a command to open one of the magnetic valves till the water levels equal each other. The system is working with an accuracy within 1 cm difference in head. By daily measuring of the decline in water level in the supply vessel and the rise in the discharge vessel, in dry periods the amount of capillary rise from the groundwater table is determined, and in wet periods the subsurface outflow.

* The manometer system was developed by Ir. A. J. W. Borghorst of the Department of Physics and Meteorology of the Agricultural University at Wageningen.

III. WATER

A. TRANSPORT IN THE UNSATURATED ZONE

1. INTRODUCTION

To describe the condition of water in soil, mechanical and thermodynamical (or energy) concepts are used. In the mechanical concept only the mechanical forces moving water through the soil, are considered. It is based on the idea that at a specific point, water in unsaturated soil is under a pressure deficit as compared with free water so the potential is negative. Often the negative sign is avoided by using the term suction, which is negative pressure.

However, water movement in soil is not only due to differences in suctions, but water may also move through unsaturated soil by other driving forces such as thermal, electrical or solute concentration gradients. Therefore a thermodynamic or energy concept has been developed over the years by various authors, which concept combines all the various components involved. BUCKINGHAM (1907) introduced the concept for soil systems, the theory was then further developed for biological systems and a thermodynamic approach was evolved of which SLATYER (1967) and TAYLOR (1968) give clear reviews.

For a given temperature the state of water can be described by the partial specific Gibbs' function of water* often simply called the water potential. It is an expression for the capacity of a unit mass of water to do work as compared to the work capacity of the same mass of pure free water. Pure free water at the same temperature is defined as having a potential of zero. Since water in the unsaturated zone is retained in the soil by binding forces at the surfaces of the soil particles, soil moisture is not capable of doing as much work as pure free water, hence the potential is negative. Apart from forces resulting from the attraction of soil matrix and water, forces associated with the osmotic characteristics of the soil solution and forces which effect the pressure on the soil water are contributing to the water potential. Hence the water or moisture potential Ψ can be written as:

* The term Gibbs' function or Gibbs' free energy of water per unit mass (G), in Europe mostly called free enthalpy, is used here for the function $G = U - TS + pV$, where T is temperature, p is pressure, S is entropy, U is internal energy and V is volume. In this paper the symbols, S, U and V represent the differences with entropy, energy and volume relative to a unit mass of pure free water of the same temperature.

$$\Psi = \Psi_m + \Psi_s + \Psi_p \tag{1}$$

where Ψ_m is the matric potential, Ψ_s the solute or osmotic potential and Ψ_p the pressure potential. When the water is located at an elevation different from that of the reference level, gravitational potential due to the force of gravity, Ψ_g, has to be added. Hence the total water potential Ψ_{total} is given by

$$\Psi_{total} = \Psi + \Psi_g = \Psi_m + \Psi_s + \Psi_p + \Psi_g \tag{2}$$

Potentials are expressed on a unit mass basis ($J.kg^{-1}$)*, but may also be expressed as energy per unit volume or per unit weight, as is illustrated below (see also ROSE, 1966). For conversion of one system into another it is necessary to keep in mind that the density ρ is the mass m per volume V. From the definition of potential it follows that the capacity to do work (W) of a mass of water ($m = \rho V$), which is raised through a vertical distance z from the reference level, is equal to mgz or ρVgz, if g is the gravitational acceleration. Thus the (gravitational) potential expressed per unit mass, volume or weight will be:

$$\Psi_{mass} = \rho Vgz/\rho V = gz \qquad (J.kg^{-1}) \tag{3}$$

$$\Psi_{volume} = \rho Vgz/V = \rho gz \qquad (J.m^{-3} \text{ or } N.m^{-2}) \tag{4}$$

$$\Psi_{weight} = \rho Vgz/\rho Vg = z \qquad (m) \tag{5}$$

From eqs. (3), (4) and (5) it follows that:

$$\Psi_{mass} : \Psi_{volume} : \Psi_{weight} = g : \rho g : 1 \tag{6}$$

As can be seen from eq. (4), potential expressed as energy per unit volume is in agreement with the mechanical or hydraulic terminology, since the energy per unit volume ($J.m^{-3}$) is equivalent with a pressure unit ($N.m^{-2}$). Therefore in this paper, potentials expressed on a volume basis (Ψ_{volume}), will be called pressure equivalents and denoted by ψ, where $\psi = \rho\Psi_{mass}$, in bar (10^5 $N.m^{-2}$).

When dealing with water flow, the energy per unit weight is often used. In that case energy has the dimension of length and is usually expressed in m head of water (eq. 5). The gravitational potential in this system is a height which is easy to measure as a hydraulic head. Potentials expressed on a unit weight basis therefore will be referred to as a hydraulic head with symbol h, where $h = \Psi/g$, in m. To facilitate conversion of one system into another, table 21 modified after TAYLOR (1968), has been inserted.

* In this study the units principly used are those of the so-called International System of Units (SI) as recommended by the International Organization of Standardization. These units are presented in table 20. As other texts often use different symbols and units, this table also gives some conversion factors. Furthermore, for conveniency of differently oriented readers, almost all graphs contain secondary axes in colloquially generally used units.

2. FIELD AND LABORATORY EXPERIMENTS

The soil moisture characteristics or moisture retention curves (fig. 4) of the three soil profiles which give the relationship between moisture content θ and matric pressure ψ_m have been determined according to methods described by STAKMAN, VALK and VAN DER HARST (1969).

In the pressure range of zero to -0.49 bar (pF 2.7), undisturbed soil cores of 100 cm³ (Kopecky) are used. The samples which are pre-saturated with water, are placed on a tension plate or on a water saturated porous medium, to which the required negative pressure is applied. To determine the moisture percentages at -2.5 bar (pF 3.4) and -15.6 bar (pF 4.2) pre-wetted disturbed soil samples are placed in a pressure membrane apparatus. Here the water content is obtained on a dry weight basis and it is converted to a volume percentage based on a mean value of dry bulk density. For pressures below -15.6 bar (pF 4.2), disturbed soil samples are brought into equilibrium with air of known and constant humidity. After measuring the relative vapour pressure of the soil water ψ_m can be determined (e.g. STAKMAN, 1968). Volume moisture percentages have to be derived as mentioned above. The relative vapour pressure is also affected by the component ψ_s, but for the soils under investigation this influence could be neglected.

In order to follow the change in soil moisture content with depth and time in the field, undisturbed soil cores can be taken at several sites. However, serious difficulties occur when the erratic variation in dry bulk density of the soil is

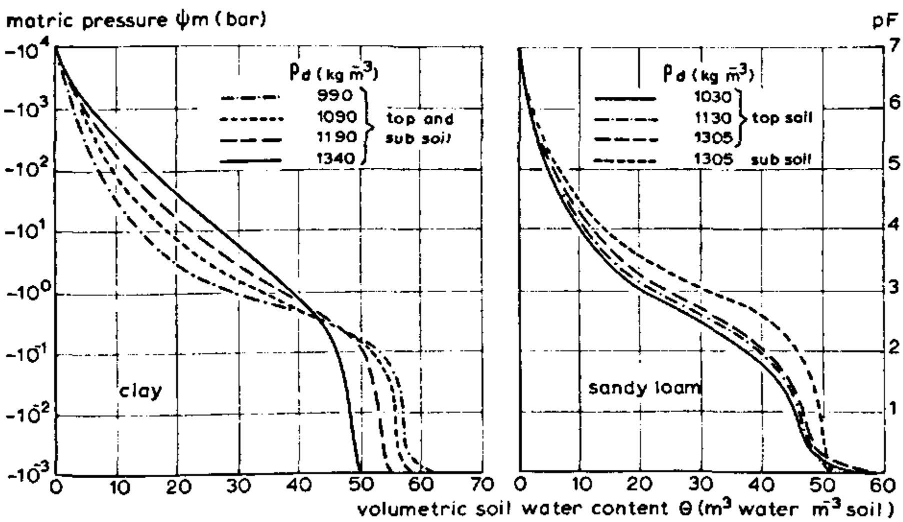


FIG. 4. *Soil moisture retention curves of clay and sandy loam at a few dry bulk densities.*

considerable as is the case with the soils of the experimental field. Therefore use was made of the gamma radiation transmission method, which measures changes in soil water content by the attenuation of gamma radiation. This method has the advantage that measurements of soil density are always taken at the same site. If the dry bulk density of the soil remains constant with time, changes in wet bulk density are only due to changes in water content. However, to convert the readings into the absolute values of the soil moisture content, the dry bulk density has to be determined too.

The gamma transmission apparatus used was from Frieseke and Hoepfner. It had a 20 mc ^{137}Cs sealed radiation source with a peak gamma energy of 0.662 MeV. Collimation at the source and at the detector was not applied. For measurements in the field this is neither practical nor essential (see J. DE VRIES, 1969). The copper or iron access tubes were placed at a mutual distance of 0.40 m. Measurements were performed at 0.10 m depth intervals with the deepest measurement under the groundwater table. The first two years of the experiments, measurements in duplicate were made outside the lysimeters, the last year one of each pair was taken inside the lysimeter.

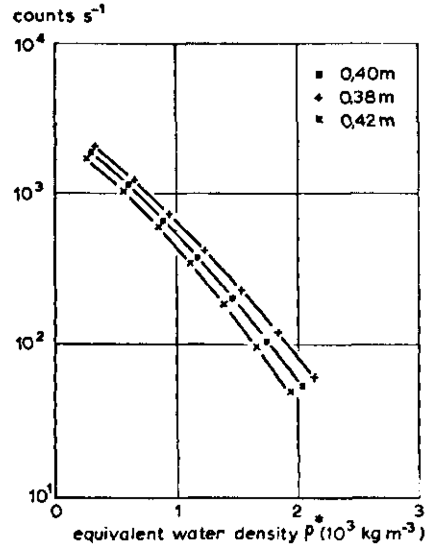
Many authors found a linear relationship between log count rate and dry bulk density from their calibration experiments (e.g. JENSEN, 1966; SMITH et al., 1967). In order to know the relationship between log count rate and wet bulk density, however, in principle a new calibration curve has to be determined for each separate value of dry bulk density (RYHINER and PANKOW, 1969). The latter authors propose to use as density scale, not the mass densities of the various components of the soil but the densities of these components expressed as an equivalent 'electron' density of water (ρ^*). This resulted in an equivalent water density:

$$\rho^* = 0.9\,\rho_d + \rho_w + 1.01\,\rho_o \qquad (\mathrm{kg.m^{-3}}) \tag{7}$$

where ρ_d is the dry bulk density of the mineral soil, ρ_w the volume fraction of water and ρ_o the dry bulk density of organic matter. Following this method, they obtained a single calibration curve for different soils for each type of access tube.

The gamma access tubes must be inserted precisely parallel and vertical at an exact constant distance, because the gamma transmission method is highly susceptible for deviations in distance as is clearly shown in fig. 5. In this graph, a measurement of 100 impulses per second is equivalent with a ρ^* value of 1877 kg.m^{-3}, if the distance between the access tubes is 0.40 m. If the distance had been for example 0.42 m one obtains a ρ^* value of 1837 kg.m^{-3}. As the accuracy of the gamma radiation measurements was in the range of 5 kg.m^{-3}, this will, when using the calibration curve of 0.40 m distance, introduce a systematic error in the calculated moisture content that is 8 times as large as the

FIG. 5. *Calibration curves of count rate of the gamma transmission apparatus and equivalent water density ρ* for three different distances between source and detector (after* RYHINER *and* PANKOW, *1969).*



accuracy of the gamma measurement itself. The corresponding error in the soil moisture content is in the range of 4 vol. %. Because a truly parallel installation of the tubes at a certain desired distance is very difficult, a theoretical method was developed to apply a correction for distance deviations and non-parallelism of the tubes at each measuring depth (RIJTEMA, 1969).

Before the installation of the access tubes, soil samples were taken while making the holes. Of these samples dry weight moisture percentage and organic matter content were determined. Immediately after the installation of the tubes, gamma radiation measurements were performed. At the end of the measuring period, volumetric soil samples (100 cm^3 Kopecky cores) were taken between the access tubes to obtain an ultimate check.

An exact measurement of the volume moisture percentage of the surface layer by means of the gamma transmission or volumetric sample method is impossible, because of changes in dry bulk density. Only estimates can be made and therefore another method was followed. Many samples of about 5–15 mm height were taken of the upper top layer. The volumes of these samples were determined by measuring approximately the average thickness and the circumference of the clods. After weighing and drying at 105 °C, the clods were reweighed and dry bulk density and moisture content on volume basis could be obtained.

Values of hydraulic conductivity at different saturation conditions were obtained from field and laboratory experiments, which were based on a steady state solution of the flow problem. This requires a measurement of the total water head at different depths at known flux.

In the field mean hydraulic conductivity values were obtained in dry periods of a week duration, when it could be expected that a continuous flow of water existed upwards from the groundwater table to the surface. The flux through the phreatic surface was determined from lysimeter experiments, while the amount of moisture extracted from the soil was measured with the gamma transmission method. From the soil moisture profile the water head at various depths could be derived.

In the laboratory the infiltration method as developed by WESSELING and WIT (1966) was used (fig. 6). This method is based on the infiltration of constant small amounts of water into vertical undisturbed soil columns. The matric head (tension) gradients were measured along the columns with small tensiometers which had different types of fillings, namely fine sand, very fine sand or a mixture of this and kaolinite and could therefore react rather fast. Tensions up to some 5 m head of water could be measured.

Saturated hydraulic conductivity values were determined using the constant and falling-head method of WIT (1967), the auger-hole method (ERNST, 1950, see also VAN BEERS, 1963) and pumping tests.

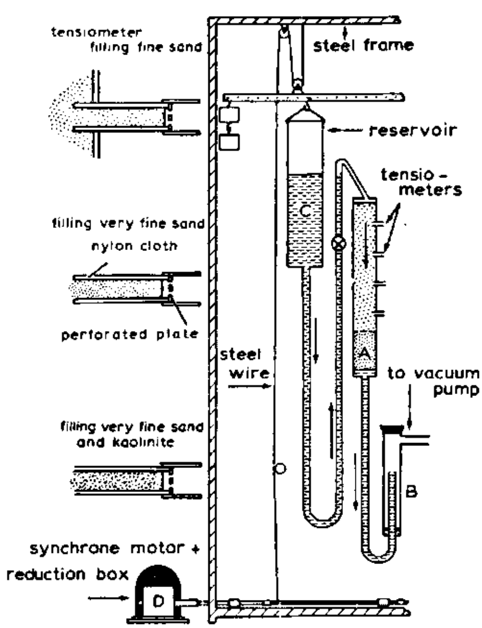


FIG. 6. *Experimental arrangement for the determination of hydraulic conductivity. The three types of tensiometers are, from top to bottom, for h_m values down to -1, -2 and -5 m respectively (after* WESSELING *and* WIT, *1966).*

3. MOISTURE CONTENT, MATRIC PRESSURE AND BULK DENSITY

From soil sampling with volume cores it appeared that the variation in dry bulk density of each separate layer in the field was quite large ($1000 < \rho_d < 1500$ kg.m^{-3}). Especially when determining the moisture retention curves the variability in dry bulk density causes serious trouble as can be seen from fig. 7A. At a moisture content of for example 47 vol. %, the matric pressure varies from -1.0×10^{-3} to -0.7 bar while the dry bulk density ranges from 1444 to 1194 kg.m^{-3}. For this reason replicate sets of samples are required to obtain more valid mean estimates of bulk soil water content or water pressure. SLATYER (1967), referring to measurements of Aitchison et al. and of Staple and Lehane, indicates that in a specific loam soil more than 10 replicate samples were required to keep the differences in dry weight soil water content below 1% with a probability of 95%. PEERLKAMP (1954) found that for water balance purposes at least 12 replicate samples were required at an experimental field in the Rottegatspolder in the Netherlands. For a discussion on errors in hydraulic conductivity and dry bulk density determinations in relation to the number of samples taken, one is referred to MASON et al. (1957).

Because of these problems, RIJTEMA and FEDDES (1971) analyzed a number of retention data obtained from sticky clay samples which were taken in the spring of various years. The samples taken in the autumn were not used for the determination of the retention curves because air was trapped in the samples when

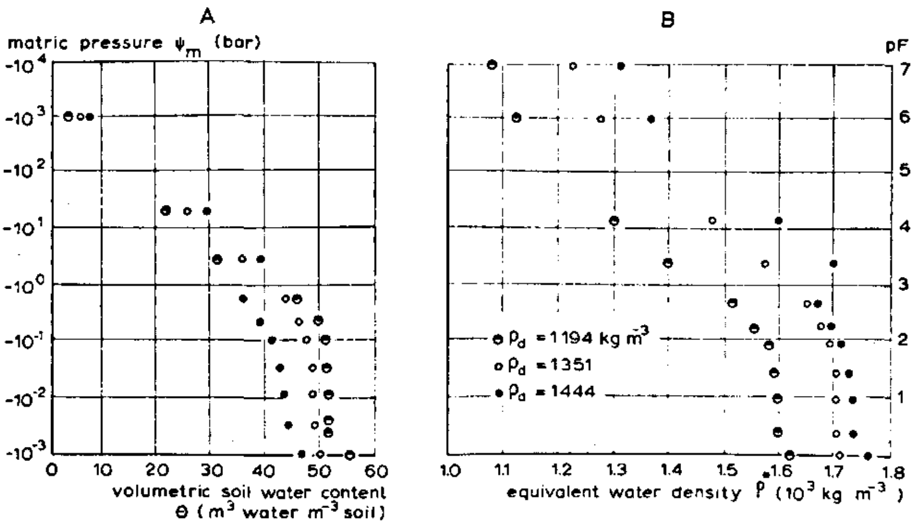


FIG. 7. *A, relation between matric pressure and volumetric soil water content; B, equivalent water density. Both for clay at three dry bulk densities and a sampling depth of 0.45 to 0.50 m.*

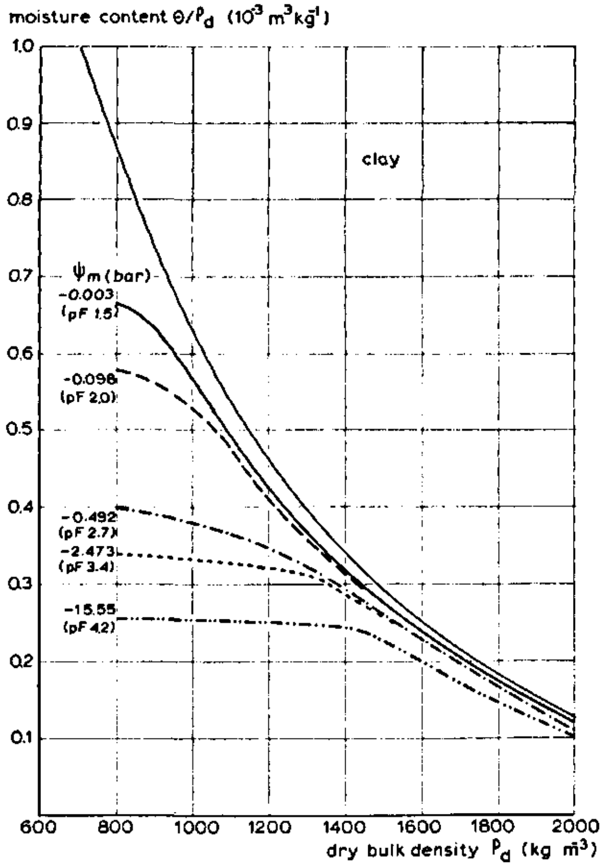


FIG. 8. *Relation between moisture content on dry weight basis and dry bulk density at various matric pressures for clay.*

saturating them. This phenomenon was already reported for heavy clay soils by MAKKINK and VAN HEEMST (1965). From the spring data relationships between moisture content on dry weight basis θ/ρ_d, matric pressure ψ_m, and dry bulk density ρ_d were found, the results of which are shown in fig. 8. From this graph moisture retention curves at various dry bulk densities can be derived. This method offers some advantages especially when using it with the gravimetric sampling procedure or the gamma transmission method.

With the gravimetric method changes in soil moisture content can be followed if the ρ_d values of the individual samples are equal. If this is not the case, it can be recommended to use the mean ρ_d value and to obtain with aid of fig. 8 the average moisture retention curve.

The retention curves for the ρ_d values of the individual samples can be reconstructed. The procedure to be followed then is to derive the actual ψ_m value of each sample using the measured θ value and ρ_d value of the actual sample in combination with the pertinent soil moisture retention curve. Having obtained the actual matric pressure, the moisture content belonging to this pressure at $\bar{\rho}_d$ can be read from the average retention curve. In this way one is less dependent on the differences in dry bulk densities between the individual samples.

With the gamma transmission method the dry bulk density is incorporated in the obtained data and when ρ_d is known one can determine ψ_m directly from the measurements with the aid of fig. 7B. This graph represents in fact the same relationship as fig. 7A, except that the variable in fig. 7B is the equivalent water density ρ^*, instead of the moisture content θ.

An important conclusion that can be drawn from fig. 8 is, that the procedure of converting dry weight moisture percentage to volume moisture percentage based on a mean value of dry bulk density is for values of ρ_d exceeding 1100 kg.m^{-3} not valid at matric pressures equal to -2.5 bar. The same holds when $\rho_d > 1300$ kg.m^{-3} at matric pressures of -15.6 bar and lower. As was mentioned earlier, this procedure is normally used for the determination of volume moisture percentages at low pressures, where the weight percentage is determined from disturbed samples. As these samples have generally dry bulk densities below 1000 kg.m^{-3}, this conversion procedure may involve some errors.

4. CAPILLARY RISE

In general, flow of water through soil can be described as the sum of the products of all acting forces and corresponding conductivities. The driving force may be a water pressure gradient, or a thermal or electrical gradient (TAYLOR, 1968). In this paper only the total water pressure gradient will be considered. In that case the general flow equation reduces to Darcy's law. Considering the flow of water in response to a force in vertical direction only, the equation can be written as:

$$q = -k^*(\nabla\psi_{total}) = -k^*(\nabla\psi + \nabla\psi_g) =$$
$$= -k^*(\nabla\psi_m + \nabla\psi_p + \nabla\psi_s + \nabla\psi_g) \qquad (8)$$

where:

q = volume flux of water passing per unit time through a unit horizontal area (m^3.m^{-2}.s^{-1} = m.s^{-1})
k^* = hydraulic conductivity (m^3.s.kg^{-1})
ψ_{total} = total water pressure equivalent (bar)
z = vertical co-ordinate (m), positive in upward direction

A negative sign has been introduced in order to make q positive in upward direction, i.e. positive in the direction of decreasing pressure. It has become customary to refer to k as hydraulic conductivity for water flow in the saturated as well as the unsaturated zone, since in both cases the flux of water is expressed per unit of hydraulic (water) pressure. In saturated soil, the pores filled with water, $\nabla\psi$ is approaching zero, the flow is largely controlled by the gravitational component ψ_g. In unsaturated soil, the contribution of ψ_g diminishes at increasing desiccation of the soil and ψ begins to dominate flow. Because of the coherent nature of air spaces we can neglect ψ_p. In general the influence of ψ_s is low, because both solutes and water are moving together. Hence, eq. (8) can be written as:

$$q = -k^* (\nabla\psi_m + \nabla\psi_g) \qquad (\mathrm{m.s^{-1}}) \tag{8a}$$

As was stated earlier, when dealing with flow of water, it is convenient to express energy on a unit weight basis. Using the appropriate symbols, eq. (8a) turns to:

$$q = -k (\nabla h_m + 1) \qquad (\mathrm{m.s^{-1}}) \tag{8b}$$

where k is in $\mathrm{m.s^{-1}}$, h_m is matric head and h_g is gravitational head in m water column [$\nabla h_g = 1$, see eq. (5)].

There are three variables in eq. (8b), namely q, k and h_m, and solving it becomes complicated because k is a function of the matric head h_m. If this function is known eq. (8b) may be integrated, when considering steady state flow.

Several analytical approximations for the relation between k and h_m have been proposed (e.g. WIND, 1955; WESSELING, 1957; GARDNER, 1958). RIJTEMA (1965) used in his approximation the functions:

$$k = k_0 \quad \text{for} \quad h_m \geqslant h_a \tag{9}$$

$$k = k_0 \, e^{\eta (h_m - h_a)} \quad \text{for} \quad h_a > h_m \geqslant h_{limit} \tag{10}$$

$$k = a (-h_m)^{-n} \quad \text{for} \quad h_m < h_{limit} \tag{11}$$

where k_0 is the saturated hydraulic conductivity, h_a the matric head at air entry point, i.e. the head at which a water-saturated porous medium starts to let air go through (e.g. STAKMAN, 1966), h_{limit} is the head below which eq. (10) does not hold (see fig. 9) and a and n are constants.

From the data of the experimental field it is possible to compute the hydraulic conductivity with the aid of eqs. (9), (10) and (11).

The flux $q(z, t)$ is not the same for all depths, but is the sum of the volume flux through a constant phreatic surface $q(0, t)$ and the amount of moisture M extracted per unit of time from the soil above this surface. The flux $q(0, t)$ was measured with the lysimeter. M was determined with the gamma transmission method and is defined by:

$$M = -\int_{t_1}^{t_2}\int_0^z \frac{\partial\theta}{\partial t}\,\mathrm{d}z\,\mathrm{d}t \qquad \text{(m)} \tag{12}$$

Hence the flux $q(z, t)$ is:

$$q\,(z, t) = q\,(0,t) + \frac{\partial M}{\partial t} = -\,k\,(\nabla h_m + 1) \qquad (\text{m.s}^{-1}) \tag{13}$$

The integral of eq. (13) over the time interval $(t_2\text{-}t_1)$

$$\int_{t_1}^{t_2} q\,(z, t)\,\mathrm{d}t = \int_{t_1}^{t_2} q\,(0, t) + M = \int_{t_1}^{t_2} -\,k\,(\nabla h_m + 1)\,\mathrm{d}t \qquad \text{(m)} \tag{14}$$

has been used to determine the hydraulic conductivity. To simplify the calculations, time averaged values at depths z were taken:

$$\bar{q}\,(z) = \frac{1}{(t_2 - t_1)}\int_{t_1}^{t_2} q\,(z)\,\mathrm{d}t \qquad (\text{m.s}^{-1}) \tag{15}$$

An example of the calculation of the hydraulic conductivity k of clay in the field is given in table 2 for the period July 11 through July 18, 1967.

The results of such calculations are shown in fig. 9. As can be seen, the laboratory data agree reasonably well with the values obtained in the field. It is important for this type of calculation not to use the h_m values of the root zone, as the movement of the water through the roots is independent of the hydraulic gradient in the soil. Otherwise, too high values for k at certain h_m are obtained. For the sandy loam sufficient measurements without any root influence were available. For clay, however, h_m values of the root zone had to be used. Therefore, according to data of WIND (1955) and RIJTEMA (1965), the function $k = 0.993 \times 10^{-9}(-h_m)^{-1.35}$ was taken for h_m values lower than –0.50 m (fig. 9).

The factors k_0, η and a for the three profiles, derived from fig. 9 and the corresponding values of h_a and h_{limit} are shown in table 3.

Substitution of eqs. (9) and (10) into eq. (8b) gives after integration respectively:

$$z = -\,k_0 h_m/(q + k_0) \tag{9a}$$

$$z = 1/\eta \ln \frac{q + k_0}{q + k_0\,\mathrm{e}^{\eta\,(h_m - h_a)}} - \frac{k_0 h_a}{q + k_0} \tag{10a}$$

TABLE 2. Calculation of hydraulic conductivity of clay for $z = 0$ at 0.98 m below surface and $q(0,t) = 1.7$ mm.week^{-1}, t_1 = July 11 and t_2 = July 18, 1967

Depth z m below surface	$\rho^*(t_1)$	$\rho^*(t_2)$	$\bar{\rho}^*$	h^*_m	$\rho^*(t_1) + -\rho^*(t_2) = \Delta M_i$	$\bar{q}(z,t) = \bar{q}(0,t) + \dfrac{\Sigma \Delta M_i}{\Delta t}$	$\bar{q}(z,t)$	$\bar{q}(z,t)$	$\dfrac{\Delta \bar{h}_m}{\Delta z} + 1$	$\bar{k}$	$\bar{h}_m$
	kg.m^{-3}	kg.m^{-3}	kg.m^{-3}	m	mm	mm. week^{-1}	mm. day^{-1}	mm. day^{-1}	mm. day^{-1}	mm. day^{-1}	m
0.10	1551	1511	1531	−12.90	4.0	14.7	2.100	1.814	−70.4	0.026	−9.33
0.20	1572	1546	1559	− 5.76	2.6	10.7	1.529	1.343	−27.8	0.048	−4.32
0.30	1552	1530	1541	− 2.88	2.2	8.1	1.157	1.000	−14.0	0.071	−2.13
0.40	1530	1504	1517	− 1.38	2.6	5.9	0.843	0.657	− 4.9	0.135	−1.08
0.50	1580	1568	1574	− 0.79	1.2	3.3	0.471	0.386	− 2.3	0.166	−0.62
0.60	1659	1656	1657	− 0.46	0.3	2.1	0.300	0.279	− 0.2	1.396	−0.40
0.70	1675	1675	1675	− 0.33	0.0	1.8	0.257	0.257	− 0.4	0.659	−0.27
0.80	1652	1651	1652	− 0.20	0.1	1.8	0.257	0.250	—	—	—
0.90	1637	1637	1637	− 0.11	0.0	1.7	0.243	0.243	—	—	—
1.00	1635	1635	1635	− 0.01	0.0	1.7	0.243				

* $\bar{h}_m$ for each layer from graphs similar to fig. 8

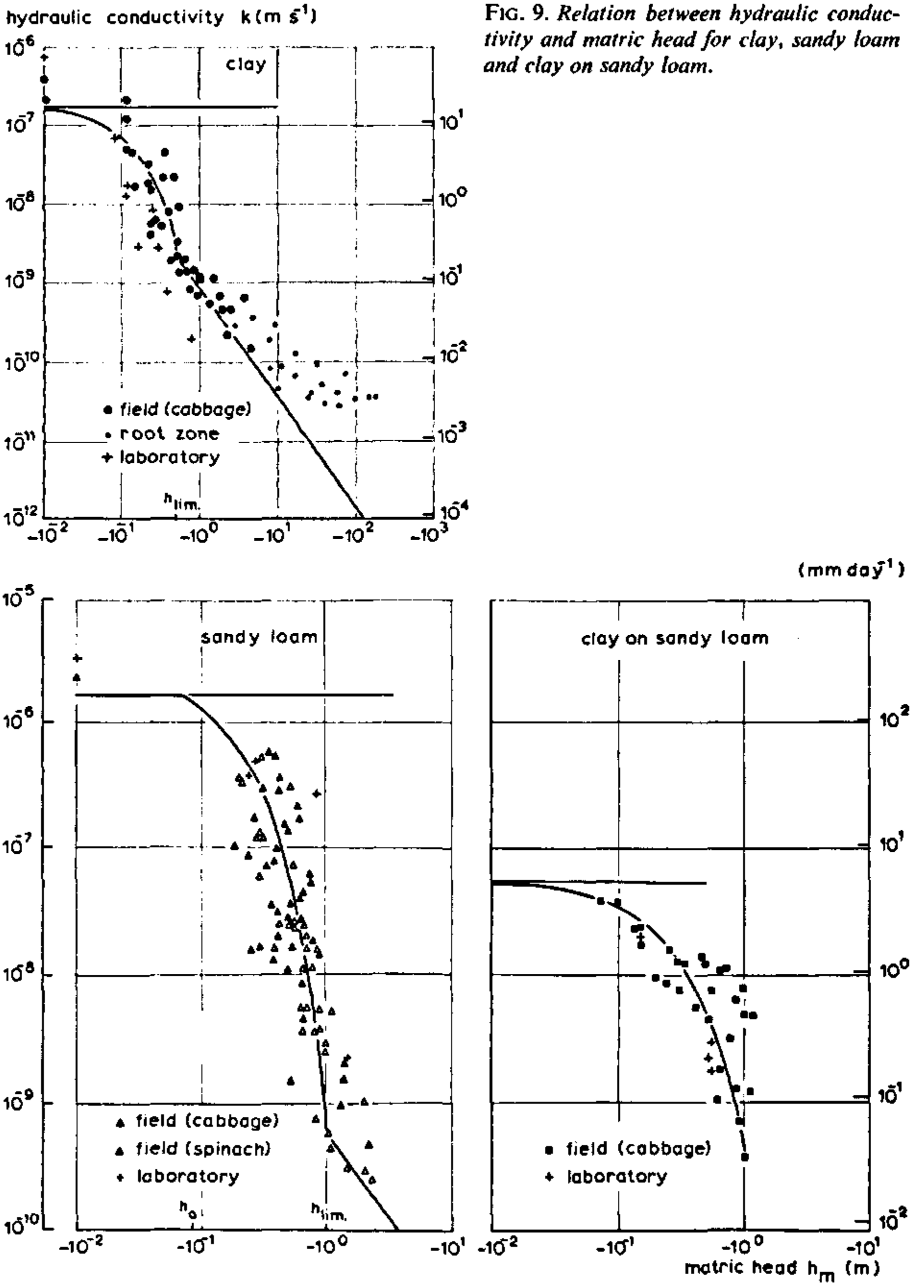
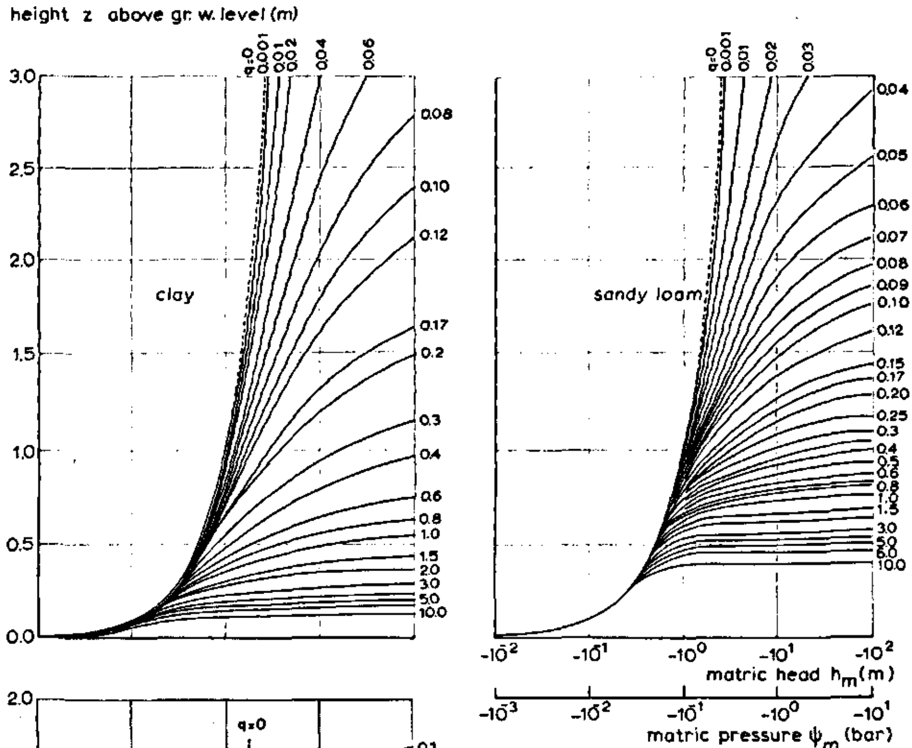


FIG. 9. *Relation between hydraulic conductivity and matric head for clay, sandy loam and clay on sandy loam.*



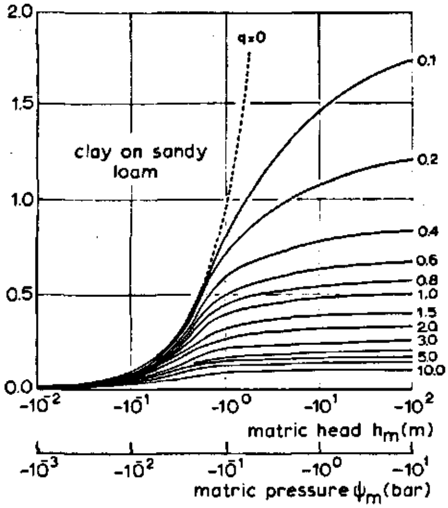


FIG. 10. *Relation between height of capillary rise above the groundwater table z and matric head h_m at different fluxes of upward flow q (mm.day^{-1}) for clay, sandy loam and clay on sandy loam.*

Table 3. Values of k_0, η, a, h_a and h_{limit} for the three soil profiles

Profile	k_0 m.s^{-1}	k_0 mm.day^{-1}	η m^{-1}	n	a	a	h_a m	h_{limit} m
Clay	1.6×10^{-9}	(14)	8.61	1.35	0.993×10^{-9}	(0.086)	0	-0.50
Sandy loam	17.4×10^{-9}	(150)	8.24	1.40	0.570×10^{-9}	(0.049)	-0.09	-1.00
Clay on sandy loam	0.6×10^{-9}	(5)	4.73	1.40	0.570×10^{-9}	(0.049)	0	-1.00

For values below h_{limit}, analytical expressions are available, for special values of n (see GARDNER, 1958). In non-specific cases numerical integration must be applied, according to:

$$z_1 - z_2 = - \int_{h_m(z_2)}^{h_m(z_1)} \left(\frac{\mathrm{d}h_m}{1 + q/a\,(-h_m)^{-n}} \right) \tag{11a}$$

In fig. 10, the relationships between q, h_m and z for the three profiles are given. For regions close to the water table, a precise knowledge of these relationships is not very important, because flow resistance is low and therefore loss of matric head is very small. Data at various fluxes nearly coincide with the equilibrium curve where $q = 0$ mm.day^{-1}. For low matric heads the various fluxes remain almost constant and therefore an exact knowledge of the relationship between q, h_m and z is in these cases also of lesser importance.

5. DRAWDOWN OF THE GROUNDWATER TABLE

In soils where a groundwater table is present, often the problem is encountered to quantify the influence of surface evaporation on depth and drawdown of the water table. Knowing the water properties of the soil, such a problem can be solved for certain boundary conditions.

One approach is to impose independently of the soil conditions a uniform specified evaporation flux at the surface dependent on meteorological conditions only. As will be pointed out in Chapter IIIB-3c, a constant evaporation rate will, due to the development of a dry surface layer, soon be followed by a falling rate phase, and the existence of a moisture profile satisfying the requirements becomes quite impossible (see CHILDS, 1969). Another, more realistic, approach is the imposition of a fixed matric head at the soil surface. In Chapter IIIB-3c it will be shown that even in rather wet periods such a situation is likely to occur because of rapid desiccation of the top layer resulting in rather low matric heads at the surface.

For the clay and sandy loam a theoretical evaluation of the influence of evaporation on the drawdown of the groundwater table was carried out assuming the following boundary conditions. The initial condition is the equilibrium situation ($q = 0$), with the water table depth z put at 0.39 m below surface. The latter value taken arbitrarily for conveniency of the calculations only. Next a sudden fall of the matric head from $h_m = -0.39$ m to -158.5 m (pF 4.2) is imposed at the surface. The imposed head value also arbitrarily taken. However, it has been shown in section 4 of this Chapter that at least for the shallow groundwater levels, the fluxes q remain rather constant at matric heads between

-10^2 and -10^4 m. This fall in matric head causes an evaporation flux at the surface, soil moisture is released from the profile and the result is a drawdown of the groundwater table. The procedure applied can be shortly summarized as follows:

- The relationships between q, h_m and water table depth z are known from fig. 10.
- With the aid of fig. 10, and using the retention curves, the courses of the moisture content with depth for $q = 0$ and $q =$ q (with h_m is -158.5 m at the surface) are obtained for the various water table depths z being read from fig. 10 at $h_m = -158.5$ m for various q values.
- The differences in moisture content with depth at $q = 0$ and $q =$ q are plotted on graphs.
- The areas between the successive q values are then determined from the graphs by means of an integrimeter (or planimeter). These areas represent the amounts of moisture released from the profile at certain drawdowns of the water table. For example if the water table in sandy loam is lowered from 1.29 m to 1.37 m below surface, i.e. the moisture profile is changing from the situation $q = 0.2$ mm.day^{-1} to the situation $q = 0.17$ mm.day^{-1}, 6.0 mm water is released.
- Starting with the water table depth z at 0.39 m below surface, now the cumulative amounts of water released at increasing depths of the groundwater table can be obtained (fig. 11A).
- For interpolation purposes the groundwater table depth z at $h_m = -158.5$ m is plotted against q on half-logarithmic paper.

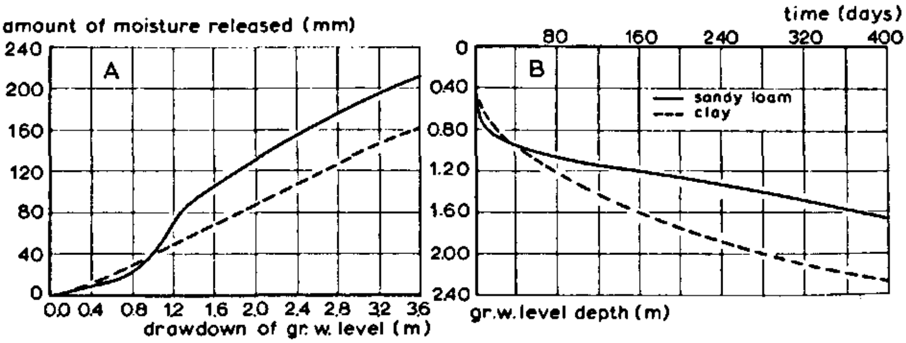


FIG. 11. *A, cumulative amount of moisture released from the profile at certain drawdowns of the water table; B, lowering of the groundwater table with time. Both under influence of evaporation from the surface while maintaining a constant equivalent matric head h_m of -158.5 m at the surface.*

Taking sandy loam as an example, the numerical analysis is carried out in the following way.

In the initial situation $q = 0$ and $z = 0.39$ m. After the sudden drawdown of the matric head at the surface, it is seen from fig. 10 for $z = 0.39$ m and $h_m = -158.5$ m that $q = 10.0$ mm.day^{-1}. Assuming that the drawdown of the water table is 0.01 m then $\bar{z} = 0.395$ m and $q = 9.7$ mm.day^{-1} (as read from the half-logarithmic graph). It follows from fig. 11A that the released amount of moisture is 0.2 mm. Hence the time taken for a drawdown of 0.01 m of the groundwater table, is as a mean 0.2 mm/9.7 mm.day^{-1} = 0.0203 day = 27 min. For a lowering of the water table from 0.40 to 0.41 m a time of 32 min. is calculated, etc. The final results for clay and sandy loam are presented in fig. 11B showing that in the beginning the water table in the sandy loam is falling at a larger mean speed than that of the clay soil. Fig. 11B also shows, however, that the final water table for clay lies deeper than for sandy loam. After 360 days the calculated depths are 2.23 m and 1.60 m respectively.

The calculated depth for the sandy loam seems to be somewhat small, as groundwater table depths between 2–2.5 m in similar soils are often found in arid zones (VAN DER MOLEN, personal communication). This discrepancy may be due to the following reasons. At the low relative humidities in arid climates, the matric head of the surface layer will be lower than assumed in the calculations. Then values of -10^4 m may not be unusual. Especially for the later stages of desiccation when deeper groundwater tables occur, this difference in matric head may be important (see fig. 10). If the soil is covered by a vegetation, the h_m value of -158.5 will perhaps be reached at the underside of the root zone. Then the depth of the root zone, say 0.50 m, has to be added to the calculated water table depth.

The evaporation flux from a deep uniform soil initially unsaturated at a uniform moisture content ($q = 0$) decreases with the square root of time (GARDNER, 1959). So when plotting cumulative evaporation against $t^{\frac{1}{2}}$ a straight line should be obtained, with the slope dependent on the type of soil. From fig. 11A and B the variation of the cumulative amounts of water released with time (i.e. the variation of the cumulative evaporation with time) was derived. By plotting cumulative evaporation against $t^{\frac{1}{2}}$ (days$^{\frac{1}{2}}$), a slight departure from the $t^{\frac{1}{2}}$ relationship was obtained as different boundary conditions were applied. GARDNER and GARDNER (1969) in laboratory studies and BLACK et al. (1969) in field experiments also found in cases of finite depth of wetting departures from the $t^{\frac{1}{2}}$ relationship, so it seems that Gardners hypothesis also holds reasonably well for soils under the influence of a groundwater table.

6. AVAILABLE WATER

To describe the amount of water available for plants, often the terms field capacity and permanent wilting percentage have been used. Field capacity is taken to be the upper limit of stored soil water available for plant growth and permanent wilting percentage as the lower limit.

SLATYER (1967) and KRAMER (1969) discussing these terms refer for field capacity to measurements of Marshall on undisturbed samples, who proposed to use a pressure value of -0.1 bar, and to Richards and Weaver, who recommended a pressure of -0.3 bar on samples which are dried, ground and sieved. These values may be useful for soils without influence of groundwater. However, in the presence of a groundwater table the term field capacity must be defined as the moisture content of the unsaturated soil in the situation where the flux q is zero. In this situation the matric pressure gradient is in equilibrium with the gravitational pressure gradient. If the gravitational energy is expressed on a unit weight basis, field capacity is the moisture content belonging to a gravitational head of $-\Delta h$ m, if Δh is the elevation above the phreatic surface.

According to SLATYER (1967) the permanent wilting percentage is neither a constant as it varies with the variation in osmotic pressure of plant-tissues, the latter ranging from -5 to -200 bar. Both SLATYER (1967) and KRAMER (1969) pointed out, however, while referring to measurements of Richards and Wadleigh who found that the variation in soil water pressure at wilting point approximates -10 to -20 bar, that the mean value of about -15 bar can still for many practical purposes be regarded as an important soil value. In general one can say that a plant starts wilting when the water pressure in the plant (ψ_{plant}) is equal to the total water pressure in the soil (ψ_{total}), i.e. $\psi_{plant} = \psi_{total}$.

Under certain restrictions the maximum amount of water that will be available for the evaporation of the crop during a number of days, can be calculated if one is informed about precipitation, depth of the root zone, depth of the groundwater table and the hydrological properties of the soil (WESSELING, 1957; RIJTEMA, 1965; FEDDES, 1969; FEITSMA, 1969). The amount of moisture available in the soil is the sum of water available in the root zone plus the amount of water delivered via upward movement from below the root zone as well as from the groundwater table. The water available in the root zone is often defined as the water content between the equilibrium situation ($q = 0$) and $\psi = -15$ bar. It is known, however, that at increasing desiccation of the soil the availability of water for the plant decreases progressively. Therefore in order to get optimum production, it is better not to allow the soil to dry out that far. This holds especially for soils under leaf crops like many horticultural crops. FEDDES (1969b) concludes in a review dealing with moisture requirements and effects of water pressure on yield and quality of various vegetable crops, that in general for these crops the admissable pressure at which soil water begins to limit plant growth is about −0.4 bar (pF = 2.6). In the present calculations this value is adopted as the upper drought limit for the root zone. To evaluate the upward flow from the groundwater table an average pressure of 0.224 bar (pF = 2.35) was adopted for the whole growing period of the crop. This is an arbitrary value, but as can be seen from fig. 9 only the magnitude and not the exact value is of importance. For the shallower groundwater table depths of the experimental field the values of q do not alter much around this pressure value, but it is to be noted that this is not true for rather deep water tables. The effective root zone which is the depth above which most roots are found, depends on the type of the crop, the profile and the meteorological conditions. In the clay on sandy loam there is a sharp boundary of the rooting depth at 0.40 m, while in the sandy loam the boundary is less distinctive. On the clay profile the effective rooting depth was larger: about 0.60 m.

The amounts of water coming available via capillary rise from the groundwater table can be derived from fig. 9 by reading at $\psi_m = -0.224$ bar the various fluxes for the respective depths z. The result is presented in table 4.

Tabel 4. Capillary rise ($mm.day^{-1}$) in three profiles at various groundwater levels and an average pressure of −0.224 bar (pF = 2.35) at the lower side of the effective root zone

Profile	Depth effective root zone m	Groundwater level m below surface							
		0.60	0.75	0.90	1.05	1.20	1.35	1.50	1.65
Clay	0.60	–	6.5	1.9	0.8	0.4	0.3	0.2	0.1
Sandy loam	0.40	–	11.0	3.8	1.2	0.4	0.2	0.1	0.07
Clay on sandy loam	0.40	3.5	1.5	0.8	0.4	0.3	0.2	0.1	0.07

The computation of the amount to be extracted in and below the root zone is elucidated with fig. 12A for the clay soil with a constant groundwater table at 1.05 m below the surface and an effective rooting depth of 0.60 m. The lowest admissable pressure in the root zone is, as was already mentioned, −0.4 bar and the effective rooting depth of 0.60 m reaches to 0.45 m above the groundwater table. Under these conditions the maximum flux can be read from fig. 10 ($1\ mm.day^{-1}$) and the pressure profile can be found from interpolation. Starting from the equilibrium situation OAB, the total amount of moisture available represented by the area OABCDO, can be determined with the aid of the retention curves of the various layers. With these curves the area can be translated in moisture content and then be measured with an integrimeter or a planimeter. In fig. 12B once more the extraction from the layers below the root zone (which in fig. 12A is represented by the area OADO for a 1.05 m water table depth) is presented for various water table depths.

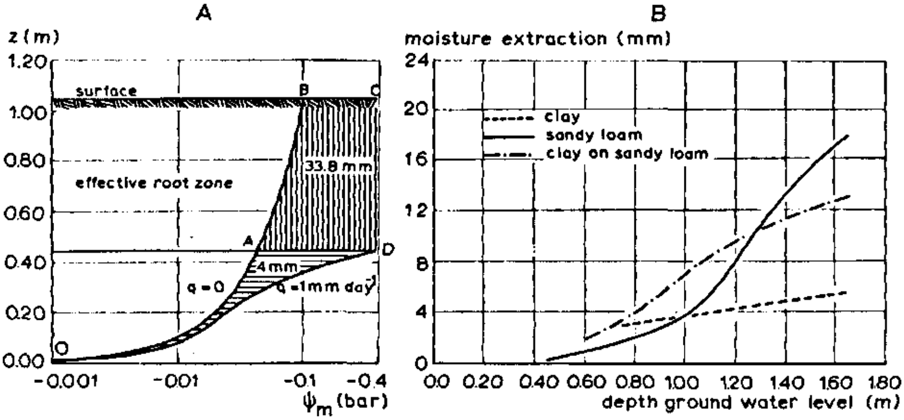


FIG. 12. *A, example of the amount of water available for the crop in clay when the minimum value of matric pressure ψ_m in the root zone is -0.4 bar (pF 2.6) and a groundwater table at 1.05 m below surface; B, moisture extraction from the layers below the effective root zone for clay, sandy loam and clay on sandy loam at various groundwater table depths when ψ_m of the root zone is -0.4 bar.*

After summarizing the various quantities of moisture available in and coming available from below the root zone and from capillary rise from the groundwater table over a certain growing period of the crop, one obtains the amount of moisture available for evaporation, which is given in table 5.

TABLE 5. Amount of moisture (mm) available for evaporation in periods of 30, 60, 90 and 120 days in relation to depth of groundwater table, when the lower limit of the matric pressure in the root zone is allowed to be −0.4 bar (pF = 2.6)

Profile	Period	Groundwater depth m below surface							
		0.60	0.75	0.90	1.05	1.20	1.35	1.50	1.65
Clay									
	30 days	—	235	96	63	50	44	38	35
	60	—	430	153	88	64	52	44	38
	90	—	625	210	112	77	60	49	42
	120	—	820	267	137	91	68	54	46
Sandy loam									
	30 days	—	—	168	87	62	53	48	44
	60	—	—	282	125	77	61	53	49
	90	—	—	396	162	90	67	57	53
	120	—	—	510	200	104	73	60	55
Clay on sandy loam									
	30 days	138	75	54	45	39	36	32	30
	60	242	120	78	58	48	42	37	34
	90	348	164	101	72	57	47	40	36
	120	452	208	124	85	65	52	44	38

B. TRANSPORT TO THE ATMOSPHERE

1. INTRODUCTION

There is little need to emphasize the importance of a reliable method for the determination of both actual as well as potential evaporation from different surfaces, under varying conditions of climate, soil profile and plant cover.

For example, agriculturists who are concerned with optimum crop production are likely to wish information on the effects of water supply on production, i.e. the specific water requirements of a crop, and whether these requirements under the prevailing environmental conditions are met or not. When from regularly taken evaporation estimates it appears that water shortages frequently occur, technical measures can be taken to establish better irrigation treatments.

Another group dealing with evaporation that can be mentioned is that of hydrologists concerned with the water balance of catchment areas, who when for example dealing with stream flow and groundwater recharge want to know how much water is lost by means of evaporation.

An estimation method which is of value in both agricultural and hydrological problems and which is based on rather easily measurable meteorological as well as soil and crop factors, is the combination method treated in section 3 of this Chapter. This method has been used on agricultural crops, but little is known about its applicability to vegetable crop production. It is the purpose of this part of the present paper to assess this. In this context attention is also given to the estimation of net radiation and reflection. The calculations by means of the various equations involved will be verified experimentally with data obtained from four vegetable crops, namely spinach, red cabbage, dwarf French beans and celery.

2. FIELD EXPERIMENTS

During the years 1967 to 1970 the following main crops were grown on the experimental field:

Spinach (*Spinacia oleracea* L., cv. *Vital R.*)	April-May 1967
Red cabbage (*Brassica oleracea* L., cv. *Rode Herfst*)	June-Nov. 1967
Dwarf French beans (*Phaseolus vulgaris* L., cv. *Centrum*)	May-Sept. 1968
Spinach (*Spinacia oleracea* L., cv. *Vital R.*)	Sept.-Nov. 1968
Celery (*Apium graveolens* L., cv. *Marmer Kogel*)	April-Oct. 1969

Growth and development of these crops were analyzed at regular times. In the main, measurements of crop height, soil cover, and leaf area were made twice a week. Dependent on their development stage the crops were periodically harvested. The fraction of soil covered was estimated with aid of a frame of 1 m^2. The leaf area was determined by measuring length and width of all leaves of some individual plants.

Dependent on crop type, fresh and dry weight were determined either per plant or per unit surface (1 m^2). The groundwater table depths at the different plots were measured in plastic pipes (inner diameter 16 mm) placed in regularly spaced auger holes. The pipes were perforated and wrapped in nylon filter cloth at the bottom ends. The measurements were taken with a thin metallic tape fitted with a bell sounding device. In the 0.90 and 1.20 m groundwater plots (having lysimeters) the water levels were measured every day, in the other plots biweekly.

The necessary meteorological data were obtained from a weather station located at the open western edge of the experimental field (fig. 1). In a Stevenson screen continuous records of air temperature (T_z) and relative humidity ($100e_z/e_z{}^*$) were obtained using two bimetallic thermographs* and two hair-hygrographs respectively. Daily (24 hours) mean values were estimated by using an Ott-integrimeter. The recording devices were calibrated each day with an unaspirated mercury-in-glass psychrometer consisting of a dry bulb and a wet bulb ordinary thermometer. Maximum and minimum air temperatures at 2 m and 0.10 m above soil surface were measured with Six's thermometers. Details of the measurements of soil temperatures are described in Chapter IV3.

Wind velocities (u) were measured at a height of 2 m above soil surface with a totalizing cup anemometer, which was provided with three conical beaded-edge cups and a mechanical counter mechanism. In a few cases single values of (u) were derived from data of the meteorological station in Den Helder.

Radiation was measured with the following instruments. The duration of bright sunshine (n) was measured with a heliograph, i.e. a Campbell-Stokes sunshine recorder. Incident direct and diffuse short wave (0.3–ca. 2 μm) radiation flux (R_s) was measured with a pyranometer known as the Moll-Gorczynski solarimeter, manufactured by Kipp. It consists of a Moll thermopile of 14 elements of constantan-manganin (resistance: 9.6Ω; sensitivity: 1.3×10^{-5} V/W.m^{-2}) mounted under two concentric hemispherical glass domes of sufficient diameter (> 50 mm) to satisfy the cosine law. Two domes are used to prevent convection caused by atmospheric influences. A white lacquered screen (diameter 0.30 m) parallel to and slightly above the plane of the sensing surface prevented the housing to be heated by radiation.

The solar radiation reflected by the surface (R_s^{re}) was measured by a second

* Unless stated otherwise, instruments were obtained from Casella, London.

pyranometer of the same type (resistance: 11.5Ω; sensitivity: 1.2×10^{-5} $V/W.m^{-2}$) mounted in an inverted position at a height of 2.25 m above the ground surface on a movable 'sulky' type installation. To obtain information on the reflection of different soils and different vegetable crops at different percentages of covered soil, measurements were carried out above 36 rectangular clay and sandy loam plots (each 6 × 6 m) under the following crops: broad beans, dwarf French beans, garden beet, red cabbage, carrots (summer and winter), cauliflower, celeriac, celery, endive, lettuce, spinach, brussels sprouts and onion.

Net radiation (R_n) was measured at 1.10 m above each of the 4 main crops mentioned at the beginning of this Chapter. Measurements were performed with a miniaturized unaspirated net radiometer developed by FUNK (1962) and manufactured by CSIRO (model CN 6). The ribbon thermopile consists of 250 elements of cropper-constantan (125 facing upwards ,125 downwards) and has an internal resistance of 87Ω and a sensitivity of 0.83×10^{-5} $V/W.m^{-2}$. The thermopiles were protected by thin polyethylene hemispheres which were continuously kept inflated and free of internal condensation by supplying dry nitrogen from a gas cylinder. The outputs of the two pyranometers and the net radiometer were recorded on three points of a 12-point Honeywell (Brown) stripchart recorder, which was calibrated in °C by a built-in cold-junction compensation. To make these three points suitable for radiation measurement, a constant reference temperature is necessary for which the temperature of the groundwater at a depth of 1.20 m below the surface was taken. A system of resistances was wired between the radiation instruments and the recorder. This made 1 cm on the recorder represent 94.6 $W.m^{-2}$ for the upward facing pyranometer, 45.3 $W.m^{-2}$ for the downward facing pyranometer and 61.6 $W.m^{-2}$ for the net radiometer. The two pyranometers were calibrated (accuracy ± 2%) against a pyrheliometer at the Department of Physics and Meteorology of the Agricultural University at Wageningen, for the net radiometer the manufacturer's calibration (accuracy ± 5%) was accepted. The accuracy of determining the daily totals of R_s, R_s^{re} and R_n by means of the Ott-integrimeter from the charts, depends on the shape of the recorded curves. It was estimated to be about 3 to 5% on the average. Hence the degree of accuracy of individual values of R_s and R_s^{re} was, as far as determined by random errors, about 5% and of R_n about 8%.

Precipitation was measured with two collecting rain gauges of the type used by the Royal Netherlands Meteorological Institute (cross-sectional area: 0.04 m^2) with height of the rims at 0.40 m above soil surface, as well as with a recording rain gauge (Van Doorne, De Bilt) of the same cross-sectional area placed with the rim at soil surface.

The amount of water intercepted by the cabbage crop was measured at 6 different sites by covering the ground around a number of individual plants with plastic sheets (areas about 3 m^2). The amounts of water reaching these covers, i.e. the throughfall, were collected in cylinders (diameter 0.395 m, length 1.20 m). The thus collected data were compared with the rainfall measurements.

To relate the evaporation of bare soil and cropped surfaces to a standard surface like water, two circular shaped tanks (diameter 0.50 m, depth 0.25 m) were installed in the ground keeping the rim at the surface and the water level at 25 mm below the rim. Each day the changes in water levels were measured by means of hook gauges, after which the levels were brought back to their original position. Heat exchange with the surrounding soil was avoided as much as possible by insulating the outside of the pan with a layer of gravel.

3. EVAPORATION

a. THEORY

In this paper the term evaporation will be used to indicate the change of liquid water into vapour, whether it is evaporation from water, soil or plant surfaces.

The process of evaporation has three basic physical requirements. There must be: A. a continuous supply of water; B. energy available to change liquid water into vapour and C. a vapour gradient to maintain a flux from the evaporating surface to the atmosphere. The various methods to determine evaporation are based on one or more of these requirements. The water balance approach is based on A, the energy balance approach on B, the aerodynamic or profile method on C and the combination method on parts of B and C. In the present context a full description of all these methods cannot be given. Reference can be made, however, to for example Rose (1966), Penman et al. (1967), Slatyer (1967), Tanner (1967, 1968), Rosenberg et al. (1968). Here, only a short outline of the two methods used in the investigations will be given. The first method is the water balance approach, the second the combination method based on the use of the energy balance and a Dalton-type transport formula.

When expressing for the soil surface the various terms of the water balance or conservation equation as depth rate equivalents, the equation may be written as:

$$E = \chi - \chi_i + q_u(0, t) - q_d(0, t) - \int_0^{z_s} \frac{\partial \theta}{\partial t} \mathrm{d}z \qquad (14a)$$

where E, χ and χ_i are the fluxes ($m.s^{-1}$) of evaporation, precipitation and intercepted precipitation at the soil surface (z_s) respectively, and $q_u\,(0,t)$ and $q_d(0,t)$ the upward and downward fluxes through the phreatic surface at a constant water depth ($z = 0$), θ is the volumetric moisture percentage and t the period considered (s). The integral of eq. (14a) over the time interval (t_2–t_1)

$$\int_{t_1}^{t_2} E\,\mathrm{d}t = \int_{t_1}^{t_2} \left[\chi - \chi_i + q_u(0,t) - q_d(0,t)\right] \mathrm{d}t + \\ - \int_{t_1}^{t_2} \int_{0}^{z_s} \frac{\partial \theta}{\partial t}\,\mathrm{d}z\,\mathrm{d}t \qquad \text{(m)} \tag{14b}$$

has been used to check and to calibrate the combination method.

The vertical energy balance equates all the incoming and outgoing energy fluxes at the earth's surface. Neglecting the part of energy involved in metabolic processes, one can write this equation as:

$$R_n = LE + H + G \qquad (\mathrm{W.m^{-2}}) \tag{16}$$

where R_n represents the energy flux of net incoming radiation, LE the flux of latent heat into the air, H the flux of sensible heat into the air and G the flux of heat into the soil. Eq. (16) can also be written as a transport equation:

$$R_n - G = -(L\rho_a\varepsilon/p_a)\,K_v\nabla\bar{e} - \rho_a c_p K_h \nabla\bar{T} \tag{16a}$$

where ρ_a is density of the air ($kg.m^{-3}$), c_p is specific heat of air at constant pressure ($J.kg^{-1}.K^{-1}$), K_h and K_v are eddy transfer coefficients for heat and vapour ($m^2.s^{-1}$), $\bar{e}$ is time averaged vapour pressure (bar), ε is ratio of molecular weight of water vapour to dry air (= 0.622), p_a is atmospheric pressure (bar) and $\bar{T}$ is time averaged temperature (K) and z is elevation above ground surface. By using the so-called Bowen ratio H/LE, the measured net energy (R_n–G) can be partitioned in LE and H. Assuming equality of K_v and K_h and expressing the Bowen ratio in finite form one gets:

$$\beta = H/LE = \gamma\Delta\bar{T}/\Delta\bar{e} \tag{17}$$

where $\Delta T = T_2 - T_1$, $\Delta\bar{e} = e_2 - e_1$ (both measured over $\Delta z = z_2 - z_1$) and $\gamma = c_p p_a/L\varepsilon$ ($bar.K^{-1}$) is the psychrometer constant. Combination of eq. (16) and (17) yields:

$$LE = (R_n - G)\,/\,(1 + \beta) \qquad (\mathrm{W.m^{-2}}) \tag{18}$$

In order to compute evaporation from water surfaces PENMAN (1948) combined eq. (18) with the empirical formula of Dalton:

$$LE = \mathrm{f}(\bar{u})\,(e_0^* - e_z) \tag{19}$$

where $\mathrm{f}(\bar{u})$ is an empirically determined function of mean horizontal wind speed, e_0^* is the saturated vapour pressure at the water surface and e_z the vapour pressure in the air at height z. In order to know e_0^* the surface temperature must be determined. To avoid the difficulties involved in measuring surface temperature Penman introduced an auxiliary value E_a so that e_0^* in eq. (19) is replaced by e_z^*, the saturated vapour pressure at air temperature T_z, hence:

$$LE_a = \mathrm{f}(\bar{u})\,(e_z^* - e_z) \tag{19a}$$

The variables e_0^* and T_0 can be eliminated by introducing the slope of the saturation vapour pressure curve $\mathrm{d}e^*/\mathrm{d}T$ or δ (bar.K^{-1}) at the mean temperature $(T_0 + T_z)/2$ or at T_z, provided $(T_0 - T_z)$ is small. Combination of eqs. (18), (19) and (19a) gives the formula of PENMAN (1948):

$$LE = \frac{\delta\,(R_n - G) + \gamma L E_a}{\delta + \gamma} \qquad (\mathrm{W.m^{-2}}) \tag{20}$$

Nomographs and tables to compute LE can be found in RIJKOORT (1954), WESSELING (1960) and KOOPMANS (1969). CHIDLEY and PIKE (1970) developed a computer programme.

In eq. (20) the term E_a is empirically expressed as:

$$E_a = 4.67 \times 10^{-4}\,(0.5 + 0.54\,\bar{u}_z)\,(\bar{e}_z^* - \bar{e}_z) \qquad (\mathrm{mm.day^{-1}}) \tag{19b}$$

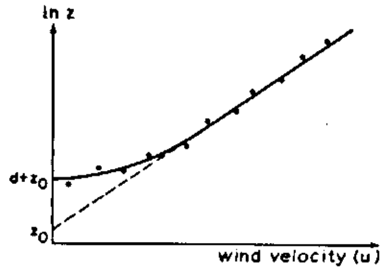
where $\bar{u}_z$ is average wind speed (m.s^{-1}).

The Dalton formula can be improved by using turbulent transport theories, from which follows that under neutral conditions (temperature lapse rates between isothermal and dry adiabatic) when all the turbulence is caused by frictional effects, the wind profile above a surface can be represented by:

$$\bar{u}/u_* = (1/K)\ln\left[(z - d)/z_0\right] \tag{21}$$

where u_* is the friction velocity (m.s^{-1}), K is Von Karman's constant, z is elevation above the ground surface, z_0 is roughness length (m) and d the displacement of the zero plane of wind speed. Due to the logarithmic form of eq. (21), z_0 is the extrapolated height at $\bar{u} = 0$ when plotting $\bar{u}$ versus ln z. To find the height at which turbulent exchange commences, one can extrapolate the actual wind profile (fig. 13), yielding at $\bar{u} = 0$ an intercept equal to $(z_0 + d)$. MONTEITH (1963) refers to this height as the effective crop surface. Both factors

FIG. 13. *Example of the relation between wind velocity profile above a rough surface and the natural logarithm of height z above the soil surface (after* GATES, 1962*)*.



z_0 and d depend on $\bar{u}$ (e.g. TANI, 1960). Assuming under neutral conditions similarity of the transfer coefficients for momentum, vapour and heat (RIDER, 1954), the profiles of $\bar{u}$, $\bar{e}$ and $\bar{T}$ in the air layer above the evaporating surface are geometrically similar. Hence plotting of measurements of $\bar{T}$ and $\bar{e}$ taken at several heights above the surface on a logarithmic scale against $\bar{u}$, will give straight lines from which $\bar{T} = T_0$ and $\bar{e} = e_0$ can be read at $\bar{u} = 0$. The transfer coefficients for momentum found from the wind profile can then be used to calculate the fluxes of sensible and latent heat.

Under non-neutral conditions, part of the turbulence is caused by thermal buoyancy, and the logarithmic wind profile does not exist: during stable conditions it becomes concave downward and during unstable conditions concave upward. However, the assumption of similarity of transfer coefficients may still be used by introducing stability corrections, such as for example proposed by MONIN and OBUKHOV (1954) and SELLERS (1965). For a survey of wind profile formulae see RIJKOORT (1968). Under neutral conditions it was shown (e.g. LETTAU, 1952) that the turbulent transfer coefficients (K_h and K_v) must increase linearly with height. STERK (1956) found from experiments some 100 km from the coast during the years 1948–1954 that this was valid both for periods of a day and a year. Hence with the stronger winds prevailing in the coastal regions it was assumed that for the circumstances at the experimental field the logarithmic wind profile could be used.

Assuming now constancy of vertical fluxes with height, PENMAN and LONG (1960) expressed the evaporation flux for a crop in an improved Dalton type formula as:

$$E = \bar{u}K^2 (C_0 - C_z) / \{\ln [(z - d)/z_0]\}^2 \qquad (\mathrm{kg.m^{-2}.s^{-1}}) \qquad (22)$$

where C_0 and C_z are the vapour concentrations ($\mathrm{kg.m^{-3}}$) at height ($z_0 + d$) and z respectively.

When evaporation is limited by water supply it is often written in the form of a diffusion resistance model in analogy with Ohm's law (PENMAN and SCHOFIELD, 1951; MONTEITH, 1963):

$$E = \frac{C_0 - C_z}{r_a} = \frac{C_0^* - C_0}{r_s} = \frac{C_0^* - C_z}{r_a + r_s} \qquad (\mathrm{kg.m^{-2}.s^{-1}}) \tag{23}$$

where C_0^* is the saturated vapour concentration (kg.m^{-3}) and r_a and r_s (s.m^{-1}) are the external atmospheric or aerodynamic resistance and the internal canopy resistance respectively. (For conversion of concentrations C into pressures e, a conversion factor of $\varepsilon\rho_a/p_a = 7.40 \times 10^{-6}$ m^{-2}.s^2 can be used). Comparing eqs. (22) and (23) MONTEITH (1963) arrived at the expression:

$$r_a = \{\ln [(z - d)/z_0]\}^2/\bar{u}K^2 \qquad (\mathrm{s.m^{-1}}) \tag{24}$$

from which functions of r_a at various z_0 and $\bar{u}$ were derived. RIJTEMA (1965), following the same procedure, expressed eq. (24) as:

$$r_a = \frac{\varepsilon\rho_a}{p_a}\left[\mathrm{f}(z_0, d)\ \bar{u}\right]^{-1} \qquad (\mathrm{s.m^{-1}}) \tag{24a}$$

where $\mathrm{f}(z_0,d)$ is a roughness function (m^{-2}.s^2), which is assumed to be the product of a function of crop height $\mathrm{f}_1(l)$ (m^{-2}.s^2) and a dimensionless function of wind speed $\bar{u}$, $\mathrm{f}_2(\bar{u})$:

$$r_a = \frac{\varepsilon\rho_a}{p_a}\left[\mathrm{f}_1(l)\,\mathrm{f}_2(\bar{u})\bar{u}\right]^{-1} \tag{24b}$$

The function $\mathrm{f}_1(l)$ originally derived for grass but recommended and used for other crops, has been tested under conditions of optimum water supply for arable crops by SLABBERS (1969), whose results are presented in fig. 14. As can be seen the data agree fairly well with those for grass. The deviating points for sunflower have been explained as resulting from maturation of the crop at the end of the growing season, giving a reduction of evaporation caused by a lesser covering of the soil and a higher transport resistance in the crop. The points for

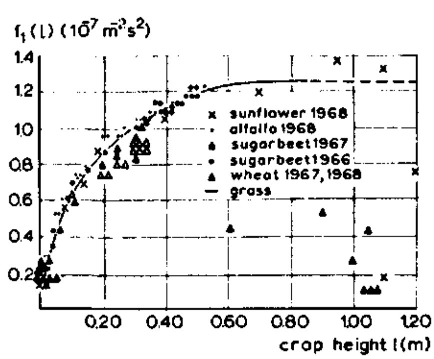


FIG. 14. *Dependence of the function $f_1(l)$, see eq. (24b), on crop height l (after* SLABBERS, *1969; the curve for grass after* RIJTEMA, *1965).*

wheat under winter conditions fit the line, whereas under summer conditions both high potential evaporation fluxes and high transport resistances are reducing evaporation (see also RIJTEMA and RYHINER, 1966).

BRUTSAERT (1965) using a model based on evaporation as a molecular diffusion process into a turbulent atmosphere, found that evaporation (E) is proportional to $\bar{u}_*^{0.75}$. Using his data, an alternative expression for r_a can be derived by substitution of the relationship $f_2(\bar{u})\bar{u} = 1.15\, \bar{u}^{0.75}$ in eq. 24b (see also RIJTEMA, 1966) yielding:

$$r_a = \frac{\varepsilon\rho_a}{p_a}\left[f_1(l)\ 1.15\ \bar{u}^{0.75}\right]^{-1} \qquad (\mathrm{s.m^{-1}}) \tag{24c}$$

Using eq. (24c), values of r_a for various crop heights (l) and wind velocities (u) were calculated (table 6).

The factor r_s (eq. 23) is regarded by MONTEITH (1963; 1965) to be a factor mainly reflecting changes in stomatal resistance. He assumes from a number of experiments that the fraction of net heat available for evaporation is closely related to r_s as $LE/(R_n - G) = \log\,(25/r_s)^{0.5}$ (see also COWAN and MILTHORPE, 1968; SZEICZ et al., 1969). Rijtema, taking into account the geometry of the evaporating surfaces, calls r_s a diffusion resistance built up by a stomatal resistance (r_l), a resistance dependent on fraction of soil covered (r_c) and one dependent on availability of soil moisture and on liquid flow in the plant (r_h). Hence

$$r_s = r_l + r_c + r_h \qquad (\mathrm{s.m^{-1}}) \tag{25}$$

To describe evaporation from a canopy, from a leaf or from another surface as a diffusion process has been strongly criticized by TANNER (1963) and FUCHS and TANNER (1967), because a diffusion model does not include the spatial distribution of sensible and latent heat sources and sinks, and neither the difference in transfer coefficients. However, it has been pointed out by SLATYER (1967) that for practical purposes this approach has a number of advantages above such laborious and complicated methods as for example the profile method.

With the substitution of eq. (23) into eq. (20) the evaporation of surfaces with either optimum or limited water supply can be computed (MONTEITH, 1965; RIJTEMA, 1965). For a wet surface with ample water supply the equation of Penman (20) can be modified into:

$$LE^* = \frac{\delta\,(R_n - G) + \rho_a c_p\,(e_z^* - e_z)/r_a}{\delta + \gamma} \qquad (\mathrm{W.m^{-2}}) \tag{26}$$

TABLE 6. Values of r_a (s.m^{-1}) calculated with eq. (24c) for various crop heights l (m) and wind velocities u (m.s^{-1})

l m	u m.s^{-1}												
	0.3	0.5	0.7	1.0	1.5	2.0	2.5	3.0	3.5	4.0	5.0	6.0	7.0
0.00	1020	693	539	412	304	245	207	181	161	146	123	108	95.8
0.01	962	656	509	390	288	232	196	171	152	138	117	102	90.6
0.02	793	541	420	322	237	191	162	141	126	114	96.2	83.9	74.7
0.03	590	402	312	239	176	142	120	105	93.4	84.5	71.5	62.4	55.6
0.04	468	319	248	190	140	113	95.4	83.2	74.1	67.1	56.7	49.5	44.1
0.05	389	265	206	158	116	93.7	79.3	69.2	61.6	55.7	47.1	41.1	36.6
0.06	338	231	179	137	101	81.5	69.0	60.2	53.6	48.5	41.0	35.8	31.9
0.07	305	208	161	123	91.1	73.4	62.1	54.2	48.2	43.6	36.9	32.2	28.7
0.08	281	192	149	114	84.1	67.8	57.4	50.0	44.6	40.3	34.1	29.7	26.5
0.09	261	178	138	106	78.1	62.9	53.2	46.4	41.3	37.4	31.6	27.6	24.6
0.10	247	168	131	100	73.8	59.5	50.3	43.9	39.1	35.4	29.9	26.1	23.2
0.12	226	154	120	91.5	67.5	54.4	46.0	40.1	35.8	32.3	27.4	23.9	21.3
0.14	210	143	111	85.2	62.9	50.7	42.9	37.4	33.3	30.1	25.5	22.2	19.8
0.16	199	135	105	80.5	59.4	47.9	40.5	35.3	31.5	28.5	24.1	21.0	18.7
0.18	190	130	101	77.1	56.9	45.9	38.8	33.8	30.1	27.3	23.1	20.1	17.9
0.20	183	125	96.8	74.1	54.7	44.1	37.3	32.5	29.0	26.2	22.2	19.3	17.2
0.25	171	116	90.5	69.2	51.1	41.2	34.8	30.4	27.1	24.5	20.7	18.1	16.1
0.30	163	111	86.5	66.2	48.8	39.3	33.3	29.0	25.9	23.4	19.8	17.3	15.4
0.35	156	106	82.7	63.3	46.7	37.6	31.8	27.8	24.7	22.4	18.9	16.5	14.7
0.40	150	102	79.4	60.7	44.8	36.1	30.5	26.6	23.7	21.5	18.2	15.8	14.1
0.45	144	98.1	76.2	58.3	43.0	34.7	29.3	25.6	22.8	20.6	17.4	15.2	13.6
0.50	138	94.4	73.3	56.1	41.4	33.4	28.2	24.6	21.9	19.8	16.8	14.6	13.0
0.55	135	92.3	71.7	54.9	40.5	32.6	27.6	24.1	21.4	19.4	16.4	14.3	12.8
0.60	133	90.9	70.6	54.1	39.9	32.1	27.2	23.7	21.1	19.1	16.2	14.1	12.6
0.65	130	89.0	69.1	52.9	39.0	31.5	26.6	23.2	20.7	18.7	15.8	13.8	12.3
0.70	129	87.7	68.2	52.2	38.5	31.0	26.2	22.9	20.4	18.4	15.6	13.6	12.1
0.80	124	84.8	65.9	50.4	37.2	30.0	25.4	22.1	19.7	17.8	15.1	13.1	11.7
0.90	122	83.1	64.6	49.4	36.4	29.4	24.8	21.7	19.3	17.5	14.8	12.9	11.5

where $E_a' = \varepsilon\rho_a\,(e_z^* - e_z)/p_a r_a$, and E^* the evaporation of a wet surface.

For surfaces with limited water supply r_s has to be included. Thus by combining eq. (26) with eq. (23), one gets:

$$LE = \frac{\delta\,(R_n - G) + \rho_a c_p\,(e_z^* - e_z)/r_a}{\delta + \gamma\,(1 + r_s/r_a)} \qquad (\mathrm{W.m^{-2}}) \qquad (27)$$

It is clear that if r_s is zero, E becomes equal to E^*. If r_s consists only of the sum of r_l and/or r_c (see eq. 25), eq. (27) yields the potential evaporation flux

(E_p) under the prevailing meteorological conditions and surface structure.

The reduction in evaporation can be found as the ratio of eq. (27) to eq. (26).

$$E/E^* = (\delta + \gamma)/[\delta + \gamma (1 + r_s/r_a)] \qquad (28)$$

During periods with precipitation, evaporation will increase due to evaporation of intercepted water. Evaporation may then be approximated as (see RIJTEMA, 1965):

$$E = \frac{\delta + \gamma}{\delta + \gamma (1 + r_s/r_a)} (E^* - E_i) + E_i \qquad (\mathrm{kg.m^{-2}.s^{-1}}) \qquad (29)$$

where E_i is denoting the evaporation flux of the intercepted water. RUTTER (1968) developed another formula being an alternative expression of the same concepts. Because E^* is the maximum possible evaporation of a cropped surface, $E^* \geqslant E_i$ holds.

Eqs. (26), (27) and (29), together with table 6 were used to calculate and analyze evaporation fluxes for various vegetable crops grown under different meteorological conditions, on different soil profiles with different depths of groundwater tables. The calculated data were compared with those obtained from water balance studies, and the significance of various relationships were derived and tested (see IIIB-3d and IIIB-4).

b. RADIATION

As can be seen from eqs. (26) and (27), one of the most important terms determining evaporation flux from a surface is the net radiation flux (R_n) absorbed by that surface. In case no direct measurements of R_n are available, which up to now is the rule rather than the exception, R_n has to be derived from empirical formulae. These are based on physical considerations and need other meteorological data and locally adjusted constants. Some of these formulae were used and the calculation results were compared, by means of linear regression techniques, with measured values of R_n and with data known from literature.

From a practical point of view it is important that R_n can be determined from relationships which are not locally determined, but more universally applicable and more easier in use. Therefore a few empirical expressions based on the often noted high correlationship between net and shortwave radiation, were derived and analyzed by linear regression.

Because as in the present paper evaporation is usually calculated on a weekly basis, weekly mean data were mainly used. The weekly means were calculated from data covering 24 hours (0800–0800 local standard time) during the period of interest for plant growth which ranges from about mid March till end October. For radiation totals data both on a 24-hour basis as well as on a daylight basis were used. The main results are shown in table 7.

Net radiation flux (R_n) can be written as:

$$R_n = (1 - \alpha) R_s - R_t \qquad (\mathrm{W.m^{-2}}) \tag{30}$$

where R_s is the flux of incident shortwave (about 0.3 to 2 μm) radiation, R_t is the flux of net outgoing thermal or longwave (about 4 to 100 μm) radiation and α is the surface reflection coefficient of shortwave radiation. According to this sequence the various terms will be discussed.

TABLE 7. Results of linear regression analysis of radiation data (R_s, R_n in $\mathrm{W.m^{-2}}$; n in min.) for from stat. on exp. field

Line	Object	x measured	y	Derivation of y with	Period of measurement
	WEEKLY MEAN DATA (24 HOURS)				
1	–	R_s	R_s	31a	April 13–October 31, 1967
2	–	n**	n	meas.	March 13–December 26, 1967
3	–	R_s	R_s	32	April 20–October 31, 1967
4	–	R_s*	R_s	meas.	April 13–October 31, 1967
5	Red cabbage on sandy loam	R_s	R_t	31a	June 21–October 31, 1967
6	Red cabbage on sandy loam	R_n	R_n	31a, 33a	June 21–October 31, 1967
7	Celery on clay	R_n	R_n	31a, 33a	June 24–October 31, 1969
8	line (6) and (7)	R_n	R_n	31a, 33a	
	DAILY DATA (24 HOURS: 0800–0800 LOCAL STANDARD TIME)				
9	Red cabbage on sandy loam	R_s	R_n	meas.	June 21–October 31, 1967
10	Bare sandy loam	R_s	R_n	meas.	April 4–May 25, 1969
11	Celery on clay	R_s	R_n	meas.	June 26–October 21, 1969
12	line (9), (10) and (11)	R_s	R_n	meas.	
	DAYLIGHT DATA (SUNRISE – SUNSET)				
10′	Bare sandy loam	R_s	R_n	meas.	March 27–June 11, 1969
11′	Celery on clay	R_s	R_n	meas.	June 26–October 22, 1969
12′	line (10′) and (11′)	R_s	R_n	meas.	
	WEEKLY MEAN DATA (24 HOURS)				
13	Red cabbage on sandy loam	R_s	R_n	meas.	June 21–October 31, 1967
14	Red cabbage on sandy loam	R_s*	R_n	meas.	June 21–October 31, 1967
15	Celery on clay	R_s*	R_n	meas.	June 24–October 21, 1969
16	line (14) and (15)	R_s*	R_n	meas.	
16′	Transformation of line (16) with aid of line (4)	R_s	R_n	meas.	

1. Shortwave radiation

An empirical expression frequently used for the calculation of shortwave radiation flux (R_s) is the one proposed by ANGSTRÖM (1924):

$$\bar{R}_s = a + b\bar{n}/\bar{N} = [c + (1 - c)\,\bar{n}/N]\,\bar{R}_s^c \qquad (\mathrm{W.m^{-2}}) \tag{31}$$

where c is a local constant, $\bar{n}$ is the mean number of minutes of bright sunshine in a mean day-length of $\bar{N}$ minutes (the maximum $\bar{n}$ can reach on clear days) and $\bar{R}_s^c$ is the mean value of R_s under clear sky conditions. The values of N and R_s^c are depending on the latitude and the time of the year. N can be found in the SMITHSONIAN METEOROLOGICAL TABLES (1951) table 171. From a linear regression of mean monthly day-values during the years 1938 to 1953 of $\bar{R}_s$ on $\bar{n}/\bar{N}$, DE VRIES (1955) estimated a and b for Wageningen, after which c was

$y = ax + b$. Data with * from meteor. stat. Den Helder, with ** from local stat. Castricum, other data

Number of data	$\bar{x}$ W.m^{-2}	$\bar{y}$ W.m^{-2}	Slope a	Intercept b W.m^{-2}	Correlation coefficient r	Standard deviation from regression		
						$S_{y,x}$	S_a	S_b
21	150	150	0.976	3	0.99	10	0.036	6
33	274	316	1.110	12.6	0.99	25.6	0.027	8.6
21	148	142	0.932	5	0.98	11	0.041	6
21	165	150	0.917	− 2	0.99	7	0.022	4
15	130	44	0.142	26	0.70	10	0.041	6
15	60	55	0.827	5	0.99	7	0.038	3
8	97	87	1.034	−14	0.96	12	0.121	12
23	73	66	0.875	2	0.98	9	0.042	4
115	134	64	0.692	−29	0.96	14	0.018	3
11	139	75	0.489	7	0.95	13	0.053	8
19	177	87	0.552	−11	0.96	15	0.038	8
145	140	68	0.649	−23	0.96	15	0.016	3
21	191	110	0.476	19	0.97	12	0.028	6
40	157	87	0.591	− 5	0.97	12	0.024	4
61	168	95	0.547	3	0.96	13	0.020	4
15	130	60	0.753	−38	0.99	7	0.029	4
15	144	60	0.694	−40	0.99	7	0.028	4
8	189	97	0.634	−22	0.97	10	0.066	13
23	159	73	0.693	−37	0.98	9	0.027	5
23	144	73	0.755	−36	0.98	9	0.030	5

found to be $a/(a + b)$. The factor c varied during the year, but an average value of 0.29 was found for Wageningen, which was adopted for the present author's experimental field. So eq. (31) becomes:

$$\bar{R}_s = (0.29 + 0.71\ \bar{n}/N)\,\bar{R}_s^c \qquad (31a)$$

From lines drawn through the points ($\bar{R}_s$, $\bar{n}/\bar{N}$) for each mean monthly day-value, R_s^c values are found, which can be plotted against time, from which finally smoothed daily R_s^c values are obtained. Application of eq. (31a) and R_s^c data given by De Vries gave rather good results, as can be seen from a comparison between measured and calculated mean weekly data of R_s at the experimental field (see line 1, table 7).

To avoid tedious determination of n, a very good correlation with values of n from a neighbouring meteorological station can sometimes be obtained, as in the present case from the relevant 33 mean weekly day-values from Castricum (see line 2, table 7). Another empirical equation easier to use, is that first proposed by KIMBALL (1927):

$$\bar{R}_s = (p + q\ \bar{n}/\bar{N})\,\bar{R}_s^{top} \qquad (\mathrm{W.m^{-2}}) \qquad (32)$$

where $\bar{R}_s^{top}$ is the extra-terrestrial radiation flux at the top of the atmosphere which can be found in the SMITHS. METEOROL. TABL. (1951), table 132.

The value of p for a latitude of $\varphi < 60°$, is found as $p = 0.29 \cos \varphi = 0.18$ and for q the value of 0.54 was taken (SCHOLTE UBING, 1959). A comparison of measured data and data calculated with Kimball's equation also shows a rather good correlation (see line 3, table 7). HELDAL (1970) obtained for Ås in Norway even better results with eq. (32) than with eq. (31). LINACRE (1967) presented values for p and q from 39 different stations, mostly near the means 0.25 and 0.50 respectively. Generally it is stated that the eqs. (31) and (32) are superior to those empirical formulae which contain values of fraction of sky covered by clouds instead of values of n/N (e.g. SCHOLTE UBING, 1959; VAN WIJK, 1966; HELDAL, 1970).

Comparison of measured R_s values at the experimental field and at Den Helder yielded a high correlation coefficient ($r = 0.99$, see line 4, table 7), emphasizing the fact that in this situation it is allowable to derive R_s from a relatively distant (about 30 km) meteorological station.

2. Thermal radiation

A review on a number of empirical formulae approximating thermal radiation flux (R_t) has been given by SCHOLTE UBING (1959). Probably the most applied equation is a Brunt-type formula as used by PENMAN (1948):

$$R_t = \varepsilon_{ea}\sigma \bar{T}_z^4 [1 - (a + b\sqrt{\bar{e}_z})] [1 - d(1 - \bar{n}/\bar{N})] \qquad (\mathrm{W.m^{-2}}) \qquad (33)$$

where ε_{ea} is the longwave emissivity of the earth (about 1), σ is the Stefan-Boltzmann constant (5.67×10^{-8} W.m^{-2}.K^{-4}), $\bar{e}_z$ is mean vapour pressure (mbar), $(1 - \bar{n}/\bar{N})$ is the mean fractional cloudiness, $\bar{T}_z$ is the mean air temperature (K) at screen height and a, b and d are empirically determined constants. Following PENMAN (1948) the values $a = 0.44$ and $b = 0.080$ were used. For d, which depends on the type and the height of the clouds, the maximum value of 0.9 was taken (SCHOLTE UBING, 1959; GEIGER, 1961), so:

$$R_t = 5.67 \times 10^{-8} \bar{T}_z^4 (0.56 - 0.080\sqrt{\bar{e}_z})(0.10 + 0.90\,\bar{n}/\bar{N}) \qquad (33a)$$

Because R_t was not measured directly, no comparison of measured and calculated data could be made. However, the data of R_s and R_t can be correlated because both increase with decreasing cloudiness, and vice versa. The correlation is rather poor (see line 5, table 7, r is about 0.70), chiefly due to the influence of the term $(0.10 + 0.90\,\bar{n}/\bar{N})$, which changes too much in periods with alternating cloud cover and does not take into account the differences in cloud cover between day and night.

It is obvious that eq. (33a) only holds for periods with constant cloud conditions, be it overcast or clear sky. Besides that, eq. (33a) is subject to the assumption that $T_z \approx T_0$, which is only a good approximation in the case of well-watered crops. If T_0 exceeds T_z, R_t increases roughly with 5 W.m^{-2}.K^{-1} (SCHOLTE UBING, 1959; LINACRE, 1968).

Because over short periods the results of eq. (33a) are liable to strong deviations other approaches have come in use. These will be discussed now.

3. Net radiation

If the reflection coefficient α is known (for data see fig. 21) estimation of net radiation flux (R_n) using semi-empirical eqs. as (31a), (33a) seems to yield rather good results (see lines 6, 7 and 8 of table 7). Despite the low accuracy of the estimation of R_t, the sample standard deviation from regression ($S_{y,x}$) for both red cabbage on sandy loam, and celery on clay as well as for the pooled objects, is about 10 W.m^{-2}. This is approximately 10% of the net radiation flux received on an average day. The random errors in measuring R_n, as quoted by LINACRE (1968), are in the same order of magnitude. Even measurements of two net radiometers exposed side by side above pasture land, as reported by HOLMES and WATSON (1967), showed differences of 10%. From own measurements it was found that on days with alternating cloud cover, errors in the area determination of the radiation curves with an integrimeter also amounted 5 to 10%. The latter deviation can be reduced by using an electronic integrator. For well-watered surfaces ($T_z \approx T_0$), LINACRE (1968) developed a number of approximate expressions for R_n with decreasing accuracy but increasing simplicity of estimation, involving only the three terms n, N and R_s^{top}. Still the incorporation of locally determined values of p and q in the Kimball equation (32) seems to be desirable, however.

Instead of using empirical formulae such as (31a) and (33a), R_n is often derived from R_s data only, by considering the correlation:

$$R_n = a\,R_s - b \tag{34}$$

Comparison of the eqs. (30) and (34) shows that the slope a of R_n mainly depends on the reflection coefficient α, and the intercept b will be a function of the other terms in eq. (33), i.e. of cloud cover and air temperature.

As mentioned earlier, b will be a constant for days with clear or overcast skies. SCHOLTE UBING (1959) for example reports for clear 24-hour days ($n/N \geqslant 0.65$) $b \approx 80$ W.m^{-2} and for overcast 24-hour days ($n/N \leqslant 0.10$) $b \approx 20$ W.m^{-2}. For a the values found by the latter author are 0.86 and 0.80 respectively, which means that data for clear days show a larger slope than those for overcast days.

From a practical point of view a separation in clear and overcast conditions seems to be less attractive because all kind of sky conditions occur in the period of plant growth. For evaporation calculations one uses mean values of n/N over a balance period. For weekly periods during 1967 (April 13–October 31) the gross of $\bar{n}/\bar{N}$ values was in the range of 0.4 to 0.6 with extremes ranging from 0.14 to 0.77, indicating that the extreme conditions reported by Scholte Ubing are seldom reached.

In table 7 the results of regression analyses of R_n on R_s of measured 24-hour data are listed for red cabbage on sandy loam in line 9, for bare sandy loam in line 10, for celery on clay in line 11 and for the pooled objects in line 12. The correlation coefficients r are exceeding 0.95 and the standard deviations from regression $S_{y,x}$ are ranging from 13 to 15 W.m^{-2}, which is about 10% of the shortwave flux received on an average day. In fig. 15A the regression line of the pooled objects, with equation

$$R_n = 0.649\,R_s - 23 \qquad (\mathrm{W.m^{-2}}) \tag{34a}$$

together with the 145 individual data are shown. In this graph also the regression line ($R_n = 0.66\;R_s - 25$) as found by FITZPATRICK and STERN (1970) from 309 radiation data above irrigated cotton in Kununurra in Western Australia (lat. 15°42′ S, long. 128°36′ E) is drawn. The results compare strikingly well. The standard deviation from regression of the Australian data was as twice as

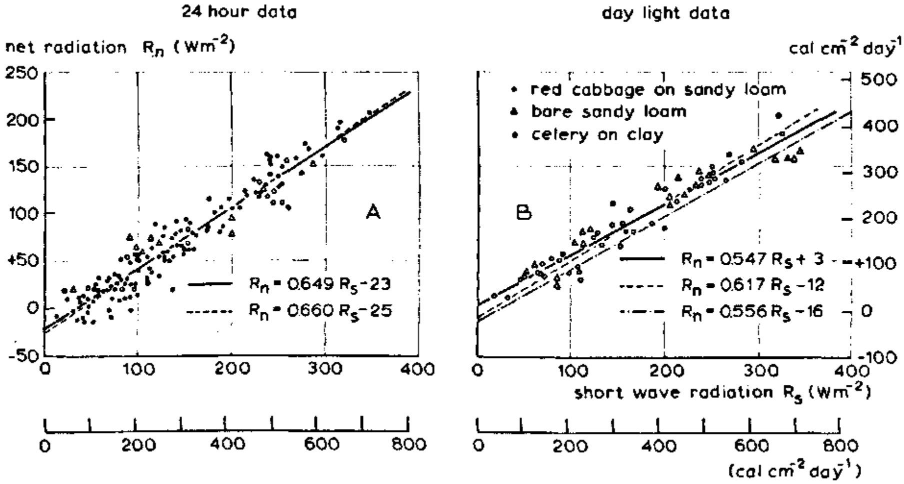


FIG. 15. *Regression of net radiation on shortwave radiation for three different pooled objects. For the given relations, see text.*

high (26 W.m^{-2}) as the present data. The previous mentioned authors reduced this by the introduction of atmospheric transmissivities (R_s/R_s^{top}) into eq. (34), getting a standard error of estimation of about 18 W.m^{-2}. Analyses by the same authors of 249 radiation data collected above non-irrigated grass at Centerton in the USA (lat. 39°30′ N, long. 75°10′ W) yielded the regression equation $R_n = 0.50\ R_s - 3.4$. According to the authors the differences in slope between the regression lines of Kununurra and Centerton are probably due to differences in reflection coefficient (0.18 and 0.22 respectively) and in temperature.

The radiation data on daylight basis are shown in table 7 for bare sandy loam in line 10′, for celery on clay in line 11′ and for the two pooled objects in line 12′. The regression equation for the latter is:

$$R_n = 0.547\ R_s + 3 \qquad (\text{W.m}^{-2}) \tag{34b}$$

which is shown in fig. 15B together with the 61 individual data. The standard deviation $S_{y,x}$ is about 13 W.m^{-2}, which compares favourably with data of FRITSCHEN (1967) who found for clear days a standard error of 16.5 W.m^{-2}. For varying cloud conditions relatively few expressions are known. DAVIES (1967) developed the relation: $R_n = 0.617\ R_s - 11.6$, also drawn in fig. 15B, for 14 stations throughout the world.

The differences in R_n data obtained with eq. (34b) versus the equation of Davies are about 5% at high values of R_s, and about 25% at very low values of R_s. The latter are of minor importance when estimating evaporation, however.

The criticism of IDSO et al. (1969) that the slope of the line obtained by Davies is inaccurate as a result of a lack of points in the region close to the origin, is exaggerated. Idso et al., by combining data of Fritschen, which were obtained at high radiation intensities, with data of Van Bavel, obtained at low radiation intensities, found a slope of about 0.80, which is true for conditions of clear sky only, whereas the results of Davies were obtained under varying conditions of cloud cover. Recently DAVIES and BUTTIMOR (1969) published another relationship as being valid for various crops, namely: $R_n = 0.556\ R_s - 16$, which is shown in fig. 15B and which has a standard deviation of 19 $W.m^{-2}$. The slope of this line is practically the same as that found for the experimental field (eq. 34b), but the intercepts of the two lines with the ordinate differ. This may be partly due to differences in standard error, because the last mentioned authors extrapolated the intercept which was obtained on a minute basis, to a daytime basis by multiplying the intercept with an approximate 11-hour average, the period the R_n values were positive.

An incorporation of net shortwave radiation by using instead of $R_n = aR_s - b$ the expression $R_n = a'\ (1 - \alpha)\ R_s - b$ does not improve the results (e.g. FRITSCHEN, 1967; DAVIES and BUTTIMOR, 1969), which is in agreement with findings of the present author.

MONTEITH and SZEICZ (1961, 1962) developed for clear days with a relatively constant incoming flux of thermal radiation, the expression $R_n = [(1 - \alpha)/(1 + \beta)]\ R_s - b$, where the so-called heating coefficient β is defined as $-\mathrm{d}R_l/\mathrm{d}R_n = (1 - a')/a'$. The value of b is equal to R_n when $R_s = 0$ and can be found by correlating R_n and $(1 - \alpha)R_s$. The authors report constant β values of 0.1 for well-watered crops completely covering the soil, of 0.1 to 0.2 for incomplete soil cover and of 0.3 to 0.4 for a very dry bare soil. If an average reflection coefficient α of say 0.20 is adopted, β values can be computed with the a values given in table 7 in lines 10′ and 11′, and the relationship $a = (1 - \alpha)/(1 + \beta)$, yielding $\beta = 0.68$ for bare sandy loam and $\beta = 0.35$ for celery on clay. DAVIES and BUTTIMOR (1969) found β values ranging from 0.17 to 0.51, but no significant differences in β between the various crops could be shown, nor a trend with increasing soil cover during the growing season. From these findings the conclusion can be drawn that neither α, nor β have to be necessarily included in regressions of R_n on R_s.

It is obvious that regression equations can be computed either on an hourly, a daylight or on a 24-hour day basis. For weekly mean day-data (table 7, lines 13, 14, 15, 16 and 16′), the standard deviation is about half that of the 24-hour data. Since the number of available weekly mean day-data of R_s was not sufficient, R_s data were derived from Den Helder, after which a linear transformation was carried out with the aid of line 4 of table 7. As a result of the lack of data, the slope of the pooled regression line may be somewhat in error. Therefore preference should be given to the regression equation based on 24 hours (eq. 34a), from which weekly means can be derived. However, it still seems that R_n data at the experimental field can with sufficient accuracy be derived directly from R_s data of Den Helder (see lines 13 and 14 of table 7).

4. Reflected shortwave radiation

A determination of the reflection coefficient of shortwave radiation (α) by means of computation is hardly possible, because α depends on a number of factors as for example the spectral reflectivity of the soil, which varies with its moisture content, and the amount of cover by a leaf canopy. The spectral reflectivity of the soil itself also depends upon the solar altitude which causes changes in spectral composition during the day and accordingly changes in α. Because the direct and diffuse components of shortwave radiation have different wave lengths, α is also dependent on their ratio. The influence of solar altitude and ratio of radiation types implies that similar surfaces at different latitudes may have different values of α. Therefore care should be taken in transferring the results found at one latitude to another one. In general there is a decrease in α from the higher to the lower latitudes (see OGUNTOYINBO, 1970). The structure of the surface is also of importance, as for example differences between ploughed or smooth soils, or differences in leaf angles of different types of vegetation. That all these factors have influence, as is shown by the variation in reflection data presented in the literature, is pointing to the necessity of measuring α directly. It must be noted that existing data mostly refer to either bare soils or to situations of crops fully covering the soil. Relatively few data exist on reflection under circumstances of increasing soil cover, especially for vegetable crops.

Usually α increases with wave length; bare soils show a more gradual increase, whereas vegetal covers generally show a small increase within the range of wave lengths of 0.4 μm to about 0.5 to 0.6 μm, a very sharp increase within the infrared region (0.6 – 0.7 μm) and a more or less constant course beyond the infrared region (IOFFE and REVUT, 1966). Therefore the visual reflection coefficient is generally smaller than the total reflection coefficient, which was a reason for MONTEITH (1959) to reserve the term albedo strictly to the visible range (about 0.4 – 0.7 μm) of the spectrum.

The reflection coefficient of the various plots was determined from continuous measurements taken on all kind of days throughout the growing season. With respect to the way of measuring α it has to be noted (BROWN et al., 1969) that the shading of the pyranometer by shields should not restrict the view angle (from the center of the sensing surface) to angles less than 160°, because otherwise an underestimation as well as a distortion of the diurnal pattern of α might occur.

The quantitive value of α can be defined as:

$$\alpha = \int_0^t R_s^{re} \mathrm{d}t \bigg/ \int_0^t R_s \mathrm{d}t \tag{35}$$

where R_s^{re} is the reflected shortwave radiation flux. A proper way of obtaining α is to determine during a certain period with a planimeter or integrimeter the respective areas under the recorded curves for R_s^{re} and R_s and take their ratio. Another method often being applied is to take readings of R_s^{re} and R_s at fixed intervals and average their ratios (see e.g. DAVIES and BUTTIMOR, 1969). This method will generally result in too high values, because at low solar altitudes in the morning and evening the highest ratios occur at periods of lowest radiation fluxes. Therefore STANHILL et al. (1966) determined α from the slope of the regression line between R_s^{re} and R_s, which method according to the authors assigns equal weight to periods of low and high radiation fluxes. This may be only partly true because the slope of the regression line is more sensible for high radiation values than for small ones. To get out of this dilemma and to avoid the time-consuming area determination with an integrimeter, hourly values of R_s and R_s^{re} were read from the charts of the recorder and the areas were determined by means of the trapezoidal rule. If the interval is divided in equal parts of $2\,k$ by the divisional points, so

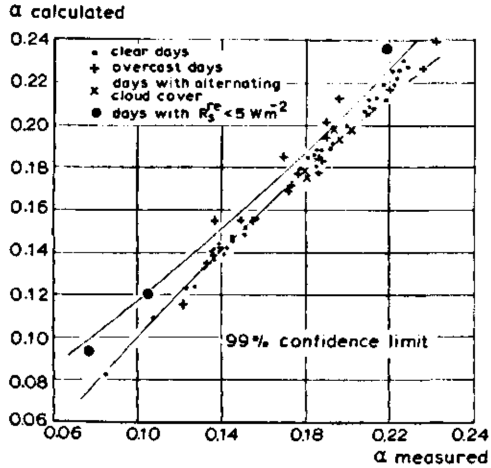
$$y_0 = 0,\ y_0 < y_1, < y_2 < \ldots . < y_{2k} = h$$

the trapezoidal rule can be expressed as:

$$\int_{x_0}^{x_1} \mathrm{f}(x)\,\mathrm{d}x \sim \frac{h}{2k}\left(y_0 + y_{2k} + 2 \sum_{1}^{2k-1} y_n\right) \tag{36}$$

with which the trapezia are summarized with parallel sides of length y_n and y_{n+1} and height $h/2k$. To check the accuracy obtained with this rather simple method, the calculated daylight totals were compared with the integrimeter daylight data, separately for clear days, for overcast days as well as for days with alternating clouds (fig. 16). For 71 daily totals, the results were compared by means of linear regression, yielding $\bar{\alpha}_{meas} = 0.175$, $\bar{\alpha}_{calc} = 0.177$, $a = 0.962$, $b = 0.0085$, $r = 0.99$, $S_{y,x} = 0.0057$, $S_a = 0.018$ and $S_b = 0.0032$. In the graph the 99% confidence limit ($t_{69} = \pm 2.654$) is drawn, showing that the accuracy obtained with this calculation method is high. Heavy overcast days with a R_s^{re} of about 5 W.m^{-2} were left out of the calculations, because at low radiation fluxes the reading reliability of the recorder charts was too low. Due to the various random errors previously mentioned, it can be said that data on α are roughly within an accuracy of 0.01.

FIG. 16. *Comparison of the surface reflection coefficient α for various types of days obtained by integrimeter and by calculation with the trapezoidal rule.*



In fig. 17 the variation of α with solar altitude and with solar time is shown for a dry clay, a wet clay and a dry sandy loam. The altitude of the sun (a) for the time of the day (h) is given by:

$$\sin a = \sin \phi \sin \delta + \cos \phi \cos \delta \cos h \tag{37}$$

where ϕ is latitude of the location of the observer, δ is seasonal declination of the sun (SMITHS. METEOROL. TABL., 1951, table 169) and h is the hour angle of the sun. The solar time is calculated from the equation

$$\textit{true solar time} = \textit{local mean solar time} + 4\,(l - l_s) + \\ + \textit{equation of time} \tag{38}$$

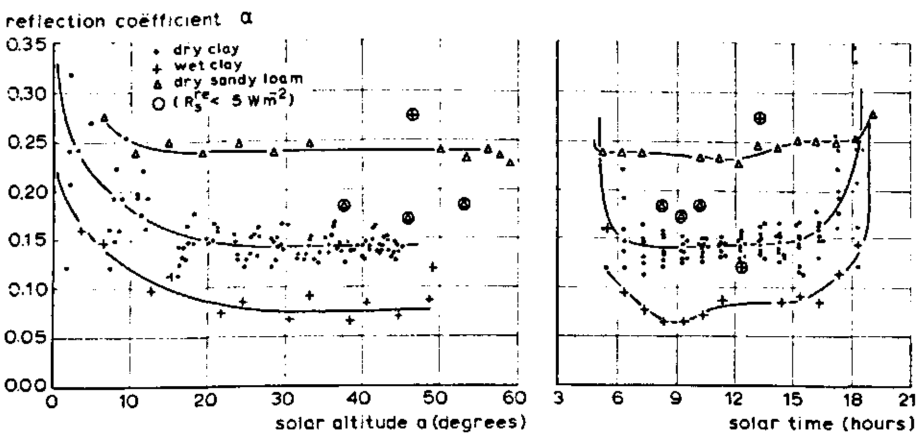


FIG. 17. *Diurnal variation of the reflection coefficient with solar altitude and with solar time during a number of days for dry clay, wet clay and dry sandy loam.*

where l and l_s are the degree of longitude of the site of the observer and the standard meridian respectively (see NAUTICAL ALMANAC of the AMERICAN EPHEMERIS, 1970). From fig. 17 it can be seen that α is higher at lower solar altitudes. From area determinations it was found that above about 20 degrees, α remains nearly constant around an average value of 0.14 to 0.15. This gives an indication for the period of the day over which measurements of α will yield the daily average. Both from fig. 17A and calculations of α for consecutive 3- and 4-hour periods, it appeared that in summer a few measurements between say 0900–1600 hours L.S.T. will represent a good daily average, in spring and autumn with lower solar altitudes one has to measure within a more restricted period of time, say between 1200–1400 hours L.S.T., to find representative daily averages.

Fig. 17 shows that wetting the soil reduces the reflection coefficient. For dry clay α is 0.14–0.15 and for wet clay α is 0.08–0.09. Sandy loam shows an average α of 0.17–0.18 if dry and an α of about 0.10–0.11 if wet. Due to the specific nature of clay sharp distinctions in α for dry and wet circumstances could be made. The sandy loam showed irregular variations between dry and wet. This caused the attempts which were made to relate changes in soil moisture content of the upper soil layer of the sandy loam to changes in α, to fail.

Fig. 18 gives the results of measured α values at increasing fractions of soil covered for a number of vegetable crops both for clay and sandy loam. The graphs refer to dry situations. The lines for the various crops were obtained by drawing by hand a fitting line through the points. The results of the measurements on the clay soil are more reliable than those on the sandy loam because of the reason mentioned above. Analogous to bare soil a wet vegetation has a lower reflection coefficient than a dry vegetation. Wet vegetations on the sandy loam show at soil cover fractions below about 0.5, α values similar to those of dry vegetations on the clay soil. The decrease in α for a wet vegetation is on sandy loam roughly 10 to 20% and on clay 5 to 10% of that for dry vegetations. The differences in reflection of bare soil ($S_c = 0$) in fig. 18 are due to random differences between the plots.

As the values for bare clay soil are low, the increase in α with increasing soil cover is more pronounced on clay than on sandy loam. Except for spinach the α values of the crops grown on sandy loam (fig. 18) show all a more or less similar pattern within a small band of α values. The α values at full soil cover seem to be slightly higher on sandy loam than on clay, which is probably caused by the better development and different leaf angles of the crops grown on sandy loam. It has been shown (ÅNGSTRÖM, 1925; DAVIES and BUTTIMOR, 1969) that the α of single leaves increased with time as they loose their turgidity. As crop structure, leaf arrangement and incomplete soil coverage weaken this trend, the α of a whole crop is lower than that of a leaf. The low α value for spinach at low

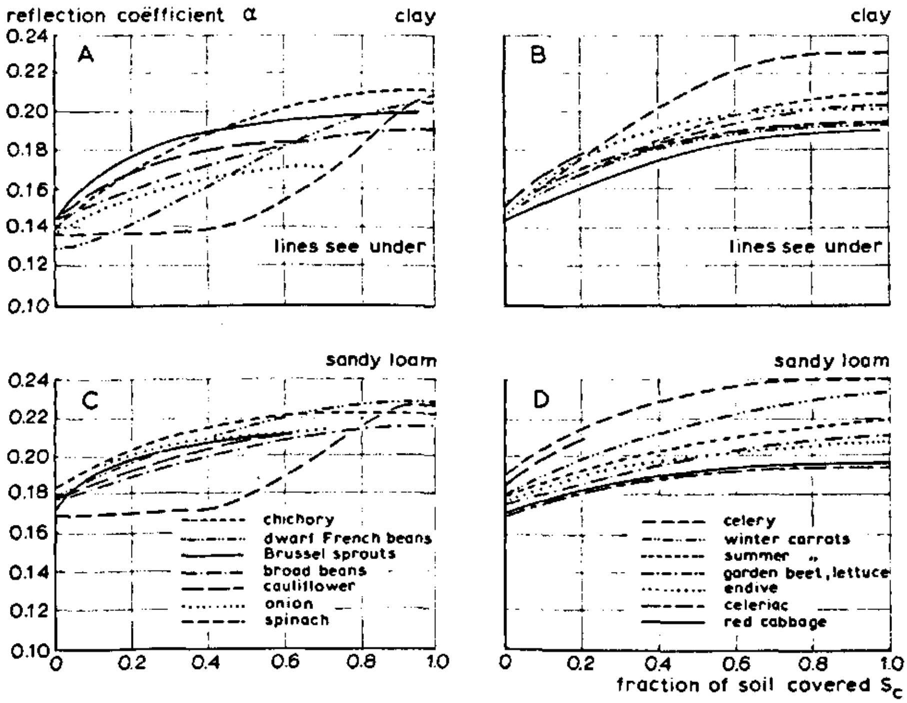


FIG. 18. *Dependence of the reflection coefficient on fraction of soil covered for various vegetable crops grown on clay and sandy loam.*

S_c values is probably due to the vertical habit of the young spinach leaves shading each other, so radiant energy is trapped. This may be the reason for the low α values of sugar beets found by MONTEITH (1959). Later on the leaves are growing more horizontally, resulting in a higher α value (see also FRITSCHEN, 1967). It appears that red cabbage shows a pattern not very different from the other investigated crops. This again affirms the earlier statement that total reflection is hardly influenced by reflection in the visible range.

c. EVAPORATION FROM BARE SOIL

Before dealing with evaporation from cropped surfaces, brief attention will be given to evaporation from bare soils.

As stated before, evaporation from any surface is determined by the amount of energy available, the vapour gradient and the supply of water. This implies that evaporation from bare soil can be studied either from the 'demand' or from the 'supply' point of view.

The first approach is based on the external atmospheric conditions which involves a detailed study of the energy balance at the surface. The second approach is based on the flux of water in the soil which involves a study of the physical transport properties of the soil.

For the choice of the method to be used, the various stages of desiccation which occur during the evaporation process of an initially saturated soil can be considered (e.g. WIEGAND and TAYLOR, 1960). During the first stage when the soil is wet the evaporation flux is independent of the moisture content of the soil, but dependent on the external meteorological conditions. In the second stage the evaporation flux decreases rapidly and is mainly determined by the limiting transport capacity of the (unsaturated) soil for liquid water. In the third stage the evaporation flux is small, falling slowly, and depending on the vapour diffusion through the surface layer which is controlled by the vapour diffusivity of and the vapour concentration in that layer.

Under natural conditions in the field, evaporation generally passes through all three stages, hence evaporation should preferably be studied considering both atmospheric and soil conditions. This has the advantage that the processes above as well as below the soil surface can be linked together. In the present research the approach based on transport in the soil was of necessity given less attention than the energy balance approach.

To obtain information on evaporation from bare soils with different groundwater tables, a detailed study was made of two clear days evaporation from the clay and sandy loam profiles of the 0.45 and 1.65 m groundwater plots. One of the clear days (March 28, 1968) belonged to a rather wet, the other clear day to a rather dry period (June 14, 1969). The transfer of heat in the soils during these days was also studied (see Chapter IV).

In the energy balance approach the study of evaporation is started with the energy balance equation:

$$R_n = LE + H + G \qquad (\mathrm{W.m^{-2}}) \tag{16}$$

The transport equations for LE and H are now written in a form somewhat different from those appearing in eq. (16a), where they did apply over a height increment dz at height z. Considering the transport from the surface $z = 0$ to a height z (here 2 m), the transport equations can be written as:

$$LE = -(L\rho_a \varepsilon/p_a)(e_2 - e_0)/r_a \qquad (\mathrm{W.m^{-2}}) \tag{39}$$

$$H = -(\rho_a c_p)(T_2 - T_0)/r_a \qquad (\mathrm{W.m^{-2}}) \tag{40}$$

where r_a is assumed to be the same both for heat ($r_a = 1/K_h$) and water vapour ($r_a = 1/K_v$). The soil heat flux G can be obtained with the temperature gradient method (see Chapter IV). Using temperature measurements at depths $z = 0$

and $z = 0.01$ m, and the mean thermal conductivity λ over these depths the expression for G is:

$$G = -\lambda_{z=0.005}(T_{0.01} - T_0)/0.01 \qquad (\mathrm{W.m^{-2}}) \qquad (41)$$

The various terms appearing in eqs. (16) and (39) to (41) were obtained in the following way.

Net radiation (R_n) was measured with the net radiometer above the sandy loam of the 1.65 m groundwater plot. From these measurements, R_n data of the other objects were derived by taking into account the 5% difference in reflection between clay and sandy loam. No corrections were applied for mutual differences in thermal radiation caused by differences in soil surface temperatures (about 1 to 2 K over one day, see Chapter IV), because for energy balances applied at hourly basis such corrections are roughly within the accuracy of the method.

The transport resistance values to the air (r_a) were obtained by means of table 6, taking for $f_1(l)$ the value at $l = 0$. Individual values of wind velocity (u) were derived from data of Den Helder. The fluctuations occurring at Den Helder were reduced by the ratio $\bar{u}_2$ (the experimental field)/$\bar{u}_{10}$ (Den Helder). It has to be emphasized that under these circumstances a high accuracy in r_a is not essential, because in the climate of the Netherlands the (radiation) term $(R_n - G)$ in eq. (27) is usually large during the day as compared with the (sensible heat) term $[\rho_a c_p(e_z^* - e_z)/r_a]$ (DE VRIES and VAN DUIN, 1953; TANNER and FUCHS, 1968).

The temperatures at the soil surface of the various objects were obtained by extrapolation of the Fourier coefficients of the measured soil temperature waves at the depths of 0.01, 0.02, 0.03 and 0.05 m, while the thermal conductivities were obtained both from measurements with the transient needle method and from an analysis with an electric analog (see Chapter IV). The temperatures (T_2) and vapour pressures (e_2) were derived from data recorded within the screen.

The mentioned procedures resulted in measured R_n data from which calculated H and G data could be subtracted, finally giving LE. The differences in the various components of the energy balance at the different objects will be discussed first. After that the rest term (e_0) will be discussed, which can finally be obtained with the aid of eq. (39). It must be remarked here that it is also possible to obtain e_0 from extrapolation when the Bowen ratio is measured at various heights above the ground (e.g. FRITSCHEN, 1965; FUCHS and TANNER, 1967).

In fig. 19 an example of the hourly distribution of the energy balance components over the clay soil of the 0.45 m groundwater plot for June 14, 1969 is given, as also the hourly distribution of the vapour pressure at the surface of this object. In order to obtain daily totals of the different fluxes involved, the trapezoidal rule (eq. 36) was applied. The main results are shown in table 8.

TABLE 8. Energy balance results on a 24-hour basis of the clay and sandy loam soils of the 0.45 m (in rather dry period), with some meteorological conditions

Components	Units	0.45 m sandy loam	
		March 28	June 14
R_n	$W.m^{-2}$	43	134
H	$W.m^{-2}$	−12	29
G	$W.m^{-2}$	24	2
LE (24 hours)	$W.m^{-2}$	31	104
E (24 hours)	mm H_2O	1.09	3.66
LE (nocturnal)	$W.m^{-2}$	−10	− 13
E (nocturnal)	mm H_2O	− 0.35	− 0.45
LE^* (24 hours) eq. 26	$W.m^{-2}$	31	97
E^* (24 hours)	mm H_2O	1.09	3.43
e_0	mbar	11.6	26.4
ψ_m (calculated)	bar	80	143
	pF	4.91	5.16
ψ_m (soil sampling)	bar	—	874
	pF	—	5.25
$\bar{T}_{0.00}$	°C	8.5	22.5
$\bar{T}_{0.01}$	°C	8.0	22.4
$\bar{e}_0^*$	mbar	11.1	27.2
$\bar{\lambda}_{0.005}$	$W.m^{-1}.K^{-1}$	0.481	0.21

Meteorological conditions:	T_2^{max} °C	T_2^{min} °C	T_2 °C	e_2^{max} mbar
March 28, 1968	18.4	1.0	10.4	9.6
June 14, 1969	22.6	12.1	18.0	19.2

From this table it is seen that except for the clay soil of the plot with a groundwater table at 1.65 m below soil surface (1.65 m clay), the sensible heat flux (H) on March 28 is negative on all objects, this being most pronounced on the relatively coldest object (0.45 m sandy loam). This means that energy from the air is used to warm up the soil. Because of the high surface temperature on June 14 the loss of sensible heat to the air, is for 0.45 m clay twice that of the 0.45 m sandy loam. Due to the high thermal conductivity of the 0.45 m sandy loam on March 28, the soil heat flux (G) for this object is 2 to 3 times as large as for the 0.45 m clay. The very low G values on June 14 for the 0.45 m sandy loam and 1.65 m clay are probably caused by temperature measurements at incorrect depths (subsidence) near the surface as well as to the mentioned extrapolation procedure. All the estimation errors made in determining R_n, H and G are finally reflected in LE. To check the results obtained, the maximum possible evaporation (LE^*) was calculated with aid of eq. (26). It is seen that on March 28 for the objects with shallow groundwater levels no reduction in evaporation m groundwater plots respectively, for March 28, 1968 (in rather wet period) and June 14, 1969

0.45 m clay		1.65 m sandy loam		1.65 m clay	
March 28	June 14	March 28	June 14	March 28	June 14
52	149	43	134	52	149
− 6	56	− 2	37	11	42
9	25	20	10	19	4
49	68	25	72	21	103
1.73	2.40	0.88	2.54	0.74	3.63
− 13	− 10	− 12	− 24	− 16	− 13
− 0.46	− 0.35	− 0.42	− 0.85	− 0.56	− 0.46
44	91	33	91	38	105
1.53	3.23	1.15	3.22	1.34	3.70
12.9	22.9	10.7	25.0	10.5	26.5
137	849	308	393	649	375
5.15	5.95	5.50	5.60	5.82	5.58
-	816	-	619	-	1207
-	5.92	-	5.80	-	6.09
9.6	26.6	10.0	23.5	12.0	24.4
9.1	24.7	9.6	22.8	10.6	24.1
11.9	34.8	12.2	28.9	14.0	30.6
0.197	0.138	0.619	0.159	0.142	0.134

e_2^{min} mbar	$\bar{e}_2$ mbar	$\bar{e}_2/\bar{e}_2^*$ -	$\bar{u}_2$ $m.s^{-1}$	$\bar{r}_a$ $s.m^{-1}$	E_{pan} mmH_2O	$W_{t, dir}$ -
5.8	7.8	0.66	2.8	190	1.5	S
14.1	17.4	0.88	2.5	210	5.2	NE

is present as distinct from the deep groundwater plots. The estimated *LE* value on this day for the 0.45 m clay seems to be somewhat high. As a result of the earlier mentioned low *G* values, this also holds for the 0.45 m sandy loam and the 1.65 m clay on June 14. Except for the 0.45 m sandy loam with its favourable water supply conditions, evaporation during this day is reduced on all plots.

During the nights of both days *LE* was negative on all objects, indicating that dew formation by condensation of water vapour occurred. The order of magnitude (about 0.5 mm) agrees with observations of MONTEITH (1957, 1963b). This indicates the necessity of using 24-hour data in evaporation studies, since by using only daylight data evaporation might be overestimated. Dew might promote germination of shallowly sown seeds. Water vapour may also be directly adsorbed by the soil when the relative humidity of the air exceeds the relative humidity of the soil. The last situation is of minor importance because water is then held at water pressures below −15 bar (pF 4.2), so far below those that are optimum for germination (see Chapter V).

TABLE 8. Energy balance results on a 24-hour basis of the clay and sandy loam soils of the 0.45 m (in rather dry period), with some meteorological conditions

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9	25	20	10	19	4
49	68	25	72	21	103
1.73	2.40	0.88	2.54	0.74	3.63
− 13	− 10	− 12	− 24	− 16	− 13
− 0.46	− 0.35	− 0.42	− 0.85	− 0.56	− 0.46
44	91	33	91	38	105
1.53	3.23	1.15	3.22	1.34	3.70
12.9	22.9	10.7	25.0	10.5	26.5
137	849	308	393	649	375
5.15	5.95	5.50	5.60	5.82	5.58
-	816	-	619	-	1207
-	5.92	-	5.80	-	6.09
9.6	26.6	10.0	23.5	12.0	24.4
9.1	24.7	9.6	22.8	10.6	24.1
11.9	34.8	12.2	28.9	14.0	30.6
0.197	0.138	0.619	0.159	0.142	0.134

e_2^{min} mbar	$\bar{e}_2$ mbar	$\bar{e}_2/\bar{e}_2^*$ –	$\bar{u}_2$ $m.s^{-1}$	$\bar{r}_a$ $s.m^{-1}$	E_{pan} mmH_2O	$W_{i,air}$ –
5.8	7.8	0.66	2.8	190	1.5	S
14.1	17.4	0.88	2.5	210	5.2	NE

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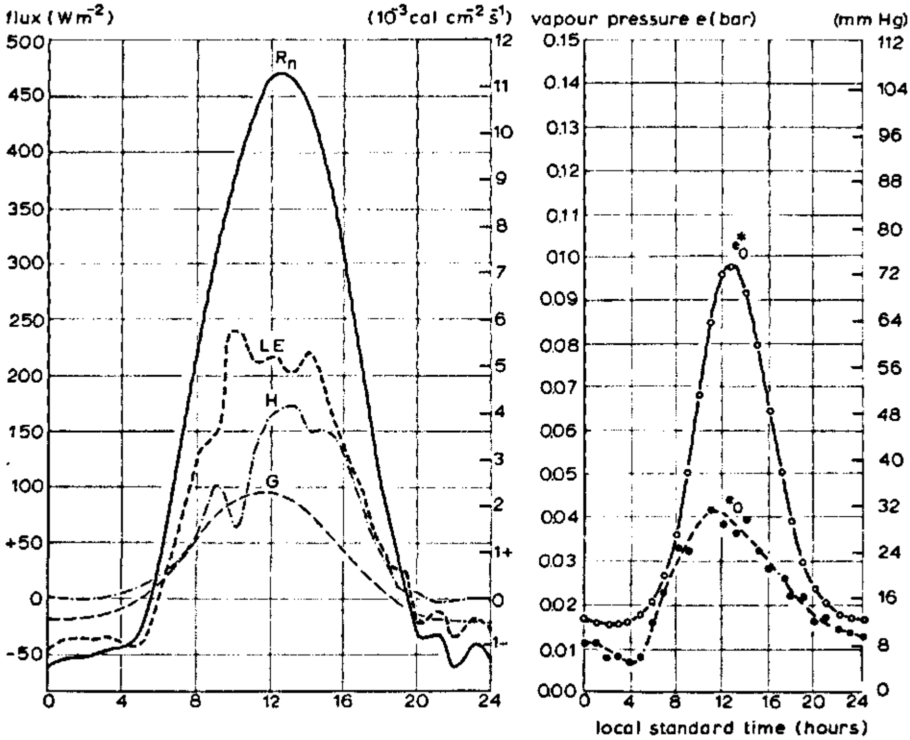


FIG. 19. *Hourly distribution of the energy balance components over and the vapour pressure at the surface of clay with 0.45 m groundwater depth, on June 14, 1969. R_n = net radiation; LE = latent heat; H = sensible heat; G = soil heat; e_0* = saturated vapour; e_0 = unsaturated vapour.*

By means of eq. (39) the vapour pressure at the surface ($\bar{e}_0$) was calculated and compared (table 8) with data obtained from sampling of the surface layer (as indicated in Chapter III). As there exists a logarithmic relationship between relative humidity (e_0/e_0^*) and matric pressure [$\psi_m = (\rho RT/M) \ln (e_0/e_0^*)$] at high humidities small errors in e_0 will result in large errors of ψ_m (see for example the 0.45 m sandy loam and clay plots on March 28). For drier circumstances there is a rather good agreement between calculation and sampling (see the ψ_m values in table 8). The height of the samples varied from about 5–15 mm, so the vapour pressure found by sampling refers to a depth lying slightly below surface.

It is to be noted that any energy balance approach needs very precise measurements of the various fluxes involved. Because in the present study part of the data had to be derived instead of measured (e.g. R_n, r_a, T_0, e_0), it must be emphasized that here only general differences could be shown. The lack of precise data on soil moisture content was also the reason that detailed studies on water transport through the soil could not be carried out for the days involved. However, from successive weekly determinations of soil moisture content it indirectly appeared that the order of magnitude of the evaporation data were estimated reasonably well. For example, the 0.45 m sandy loam proved to be in the equilibrium situation on June 3, 1969. From comparison with June 14, 1969 it was found that up to and inclusive that day 31.6 mm water was lost from the profile which gives a mean value of 3.2 $mm.day^{-1}$. The energy balance approach for the 0.45 m sandy loam on June 14, resulted in evaporation data ranging from 3.4–3.7 $mm.day^{-1}$. Calculations on capillary rise furthermore did show that on June 14 for the layers 0.05 to 0.20 m fluxes between 0.3 and 0.4 $mm.day^{-1}$ had to be accounted for.

Part of the data were used to quantify the importance of thermally induced moisture flow as compared with flow induced by potential gradients (Chapter IV).

d. Evaporation from cropped surfaces

1. Crop development

The two important quantities to be measured to characterize the development pattern of a crop, are crop height and the fraction of soil covered. The height of the crop (l) and the wind velocity (u) determine the roughness length (z_0) and zero plane displacement (d), and finally the aerodynamic resistance (r_a) (see table 6). The fraction of soil covered (S_c) is directly related to the vapour diffusion resistance r_s (eq. 25).

The crop height can be measured easily. Estimation of fraction of soil covered is more complicated. It can be done by taking photographs at regular intervals from a point above the crop surface, and determining the relevant parts of the photographs afterwards either by means of an integrimeter or by weighing. However, this tedious and awkward procedure may be avoided by making frequent eye view estimations of soil cover with the aid of a horizontal frame, and finally smoothing a curve through the estimated data. Use can be made of the rather good relationship which exists between the easy measurable crop height and the fraction of soil covered, so that possible wrong estimations of soil cover can be corrected afterwards. As an example the relationship for spinach and red cabbage grown on different soil profiles is given in fig. 20. Many investigators use as a measure of the photosynthetic size of the crop, the leaf area index (LAI) which is the total area of leaves per unit area of soil surface. This index was obtained by totalizing the area of all individual leaves of a few individual plants. It appeared that, except for the maturing stage of the crop, there is a close relationship between fraction of soil covered and leaf area index as shown for red cabbage in fig. 20.

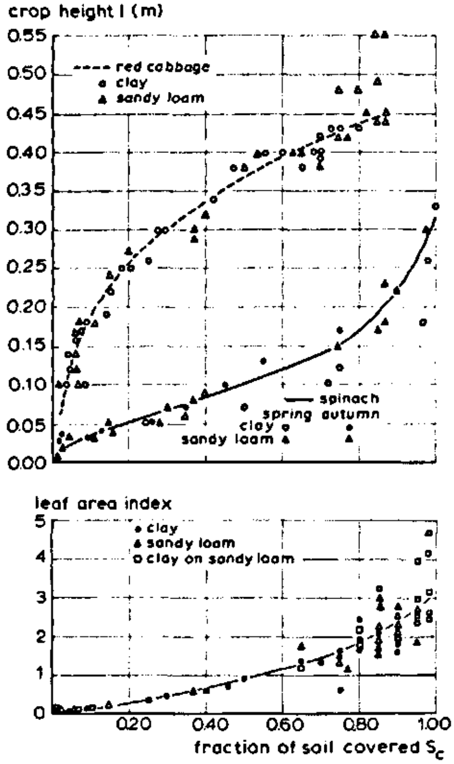


FIG. 20. *Relation of crop height and leaf area index with fraction of soil covered for red cabbage and spinach grown on clay and sandy loam.*

2. Diffusion resistances

Resistance (r_l). It is known from literature (e.g. GAASTRA, 1963; SLATYER, 1967) that stomatal aperture, which is of importance for evaporation as well as for photosynthesis, is predominantly affected by the visible part of the radiation (0.4 – 0.7 μm) flux (R'), CO_2 concentration in the intercellular spaces of the leaves, water pressure in the leaves and temperature. Under conditions of optimum water and nutrient supply, and normal outdoor CO_2 concentrations (about 300 ppm), stomatal aperture is mainly determined by R'. The value of R' is about 0.4–0.5 of the shortwave radiation flux (R_s). KUIPER (1961) performing experiments on bean leaves at temperatures above 10 °C found a S-shaped dependency of stomatal aperture on R′ with the stomata nearly closed at darkness and with their widest aperture at a R' value of about 80 $W.m^{-2}$.

Stomatal apertures can be related with stomatal diffusion resistances. Hence RIJTEMA (1965) derived a hyperbolic expression, based on Kuiper's data, for the relationship between stomatal diffusion resistance and R', which expression agrees with values measured in field experiments on maize, sorghum and tobacco by TURNER (1970) and on corn by SHAWCROFT and LEMON (1970). The latter authors, paying attention to the influence of water potential in the leaves of the crop, also show that the stomata do not respond until some critical water potential has been reached.

The integrated parameter of these diffusion resistances of the separate leaves, i.e. the diffusion resistance of the crop (r_l), can also be related to R'. For practical evaporation purposes, however, assuming that R' is about 0.4 R_s, the solar radiation flux can be used as a measure of R' by taking the mean shortwave radiation flux $[\bar{R}_s = (1/N \int_0^N R_s \, dN)]$ during a balance period as the factor controlling stomatal opening, fig. 21 (RIJTEMA, 1965). The critical value for a crop appears to be 265 $W.m^{-2}$, while in single leaves stomata already begin to close at a radiation flux of 174 $W.m^{-2}$. This difference is probably caused by mutual shading of the leaves in a leaf canopy. The reduction of evaporation due to the resistance value r_l is important in the early morning, evening and night. Over the year the influence of r_l is largest in spring and autumn and during cloudy periods in summer, in general in all periods with low radiation fluxes in which evaporation is usually low.

Resistance (r_c). In the section on evaporation from bare soil it has been shown that at increasing desiccation of the soil, evaporation is reduced by the restriction of vapour diffusion as imposed by the dry top layer. This means that especially in the initial stages of plant development, this resistance to diffusion has to be taken into account under dry soil conditions.

For spinach and cabbage, values of r_c were computed from eq. (27) for a number of periods under varying fractions of soil covered, with the evaporation flux (E) taken from water balance studies. The results are presented in fig. 22. Included are some data on wheat as computed by RIJTEMA and RYHINER (1966).

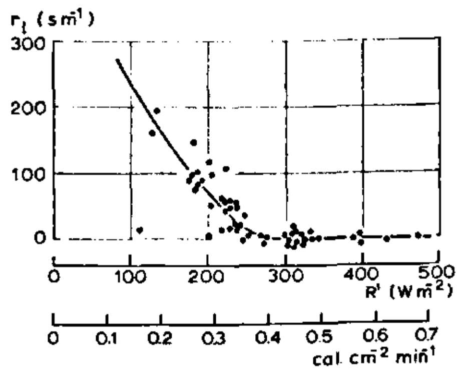


FIG. 21. *Dependence of diffusion resistance r_l on photosynthetic active shortwave radiation flux R' (after* RIJTEMA, *1965).*

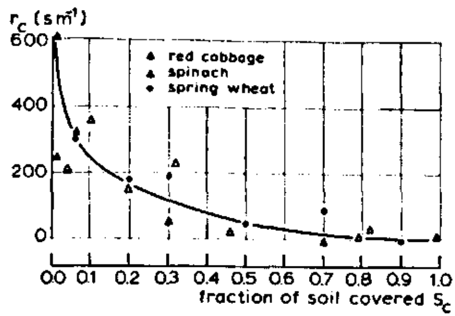


FIG. 22. *Dependence of diffusion resistance r_c on fraction of soil covered for red cabbage and spinach (data of spring wheat after* RIJTEMA *and* RYHINER, *1966).*

The scatter in the data follows from the influence of small amounts of precipitation in some balance periods. It appears from this graph, that the soil must be covered to about 70 to 80% before r_c decreases to approximately zero. The results agree with experiments of MILTHORPE (1960) who found little variation in evaporation from different crops with leaf area indexes larger than 2 (compare fig. 20, where fractions of soil covered for cabbage of 0.7–0.8 correspond with leaf area indexes larger than 2). MONTEITH (1965) reports a constant evaporation rate for broad-leaved crops with a leaf area index larger than 3 grown on dry soils and for more upright standing leaves a constant rate when the leaf area index is about 6. MARLATT (1961) found constant evaporation from mowed orchard-grass if the soil covered was above 50%. STANHILL (1958) obtained constant evaporation rates for irrigated carrots when the crop covered more than 40% of the soil. TANNER (1967) reports for irrigated dwarf French beans and potatoes that 55% of the row spacing had to be covered before evaporation rates became constant. SWAN et al. (1963) found for the same crops a percentage of 50. COWAN and MILTHORPE (1968), commenting on the differences found by the various authors for the same kind of crop, state that small leaf area indexes resulting from wide spacing (in combination with a more extensive root system) give lower evaporation rates than with close spacing.

To the present author's opinion an important part of the mentioned differences can be explained by considering the soil conditions in the various experiments. The evaporation from a wet soil generally encounters small diffusion resistances and hence variation in the area of soil covered has not much effect. Due to irrigation treatments in most of the experiments mentioned in literature it can be expected that the critical fractions of soil covered under these circumstances will be lower than under drier conditions.

It should be mentioned that for soils which have good transmissivities from lower to upper layers, one has to be careful when using fig. 22 for evaporation calculations, as will be shown in the section specially dealing with the evaporation of celery (section IIIB-4).

Resistance r_h. If water in the soil becomes limiting, an internal water deficit in the plant develops, the potential in the leaves decreases and the stomata begin to close, which results in an increase of r_h. For a full discussion on this subject, the reader is referred to SLATYER (1967, 1970) and CRAFTS (1968). This resistance can be derived from eq. (27) when the other terms are known. For an investigation on the significance of r_h, dry meteorological conditions as well as soils sensible for drought are required. In the three years of the experiments there were only very few dry periods. On the average a 'dry year' in the Netherlands occurs only once in ten years, and for example the r_h investigations of RIJTEMA (1965) on grass covered soils pertain therefore mainly to the very dry conditions in 1959. The dry period in the first growing stage of red cabbage was used for an analysis of the transport resistance for liquid flow in the plant as well as for an investigation on the geometry factor of the root system. When knowing these factors, more can be said about r_h. For a more complete discussion on this subject, the reader is referred to section IIIB-3c-4.

3. Interception

The amount of water which can adhere to the surfaces of the leaves of a crop depends on factors like intensity, amount and distribution of precipitation, evaporation flux, and shape, stand, size and nature of the leaves. Because cabbage is a broad-leaved crop it was expected that some differences in interception with that of a crop like spinach might occur.

The results of the measurements of interception of red cabbage are presented in fig. 23. The data represent means of 5 measurements at 5 different places with fractions of soil covered varying from 0.70 to 0.95. The scatter of the data is large due to the variation in the different environmental factors. A smooth line was drawn through the points and it appears from fig. 23 that at precipitation amounts smaller than 1 mm from one rain shower, 50 to 100% adhered to the leaves. At higher amounts (> 5 mm), only 15% remained on the leaves. In fig. 23 the interception curve for grass, which crop is comparable with for example a vegetable crop like spinach, as derived from data of RIJTEMA (1965) is also drawn. Taking the scatter in the various data into account, the curves do not show many differences. The slightly lower interception values of cabbage are probably due to the lower fraction of soil covered and to the specific nature of the surface of cabbage leaves. As a wax layer is present upon the surface of the leaves the water layer is not spread all over the surface but big droplets are formed. For calculation purposes the curve of cabbage was used for broad-leaved crops, and that of grass for small-leaved crops.

Interception is especially important in periods of reduced evaporation [i.e.

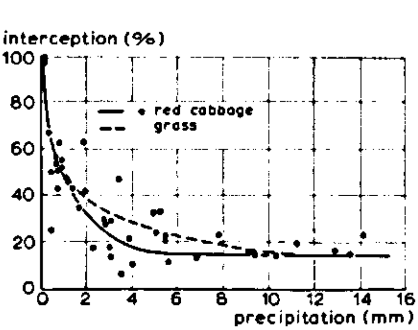


FIG. 23. *Dependence of interception on rainfall depth for red cabbage grown on three different profiles (the curve for grass derived from data of* RIJTEMA, *1965).*

when r_s is large, see eq. (29)]. The interception increases the total evaporation but reduces the evaporation of the crop, because part of the energy is used for the evaporation of intercepted water (E_i). It is to be noted that when a relatively large error is made in estimating E_i a relatively small error in the final calculation of evaporation is made. This may be illustrated with an example, which concerns the evaporation of cabbage during the period August 9 to August 16, 1967. The results are presented in table 9.

TABLE 9. Influence of the interception term (E_i) on the calculation of evaporation (eq. 29) at various reductions in evaporation; $E^* = 4.04$, real $E_i = 0.64$ (mm. day^{-1})

$(\delta + \gamma)/[\delta + \gamma(1 + r_s/r_a)]$	E mm.day^{-1}		
	$E_i = 0.64$	$E_i = 1.28$	$E_i = 0$
0.9	3.70	3.76	3.64
0.8	3.36	3.49	3.23
0.7	3.02	3.21	2.83
0.6	2.68	2.94	2.42
0.5	2.34	2.66	2.02

At a 100% error in the estimation of E_i ($E_i = 1.28$ instead of $E_i = 0.64$) a relative small error in E is made, the error being larger the larger the reductions in evaporation are. Neglecting interception completely ($E_i = 0$) will result in an underestimate of E, the more so the more the surface is dry.

For balance periods with precipitation, E_i was determined daily with the aid of fig. 23, after which a mean value of E_i was obtained for the whole balance period.

4. Water withdrawal by plant roots

Any reasonably accurate mathematical description of water uptake by crops with a non-uniform root system, is complicated. Therefore it has been become customary to describe water flow through the entire soil-plant-atmosphere system with an electric analog (VAN DEN HONERT, 1948). This model assumes that the water flux q (m^3.m^{-2}.s^{-1}) through the rooted soil zone and the root-stem-leaf-stomata path is proportional to the total water head difference Δh_{total} (m) and inversely proportional to the total resistance r (expressed in seconds) met in the system. Thus, considering a one dimensional steady state flow in a series parallel network, the liquid flow equation is:

$$q = \frac{1}{A}\frac{dV}{dt} = -\frac{\Delta h_{total}}{r_{total}} = -\frac{(h_{root\ surface} - h_{soil})}{r_{soil}} =$$

$$= - \frac{h_{leaf} - h_{root\ surface}}{r_{plant}} = - \frac{h_{leaf} - h_{soil}}{r_{soil} + r_{plant}} \qquad (m.s^{-1}) \qquad (42)$$

It should be noted, that the resistance r (expressed in seconds) to liquid flow in the soil-plant system is a function of water head gradients and must not be confused with the earlier mentioned diffusion resistances r ($s.m^{-1}$) which are functions of water vapour concentrations. A full discussion on the various plant factors involved has been given by SLATYER (1967) and KRAMER (1969).

Water flow to the roots has been described by models as developed for example by PHILIP (1957), GARDNER (1960), VISSER (1964), COWAN (1965), WHISLER et al. (1968), MOLZ and REMSON (1970). In the model of Gardner a single root is taken to be a hollow infinitively long cylinder of uniform radius v (m) with uniform water absorbing properties. From a soil which is initially at uniform equilibrium water content, the rate of water uptake q_r' per unit length of root ($m^3.m^{-1}.s^{-1}$) can be calculated if water under a head gradient is flowing radially toward the root. Under steady state conditions ($\partial\theta/\partial t = 0$) and in analogy of heat flowing from a distance v_2 (m) toward a linear sink of infinite length, one can write

$$q_r' = - \frac{h_{m,\ root\ surf.}(v_1) - h_{m,\ soil}(v_2)}{[\ln (v_2^2/v_1^2)]/4\pi k} \qquad (m^2.s^{-1}) \qquad (43)$$

where k, the hydraulic conductivity ($m.s^{-1}$), is a function of the matric head h_m (m).

A daily fluctuation in q_r' will exist and GARDNER (1968) therefore extended his model by assuming a sinusoidal variation of evaporation flux with time. In the present study a constant mean weekly q_r' has been taken. COWAN and MILTHORPE (1968) are of the opinion that in drying soil which shrinks there is water transport to the roots in the vapour state when the air gaps around the roots are not larger than the radius of the root. In this study any possible vapour transport has been left out of consideration.

The model of a single root can be extended to an entire root system (GARDNER, 1960; GARDNER and EHLIG, 1962). If l_r is the length of root per unit volume of soil ($m.m^{-3}$) and z_r is the rooting depth (m), the flux q_r ($m.s^{-1}$) can be written as:

$$q_r = l_r z_r q_r' \qquad (m.s^{-1}) \qquad (44)$$

Combination of the third term of eq. (42) with eqs. (43) and (44) gives:

$$q_r = - (h_{root\ surf.} - h_{soil})/(b/k) \qquad (m.s^{-1}) \qquad (45)$$

where b (m) is a constant with which the length and geometry of the root system are accorded for:

$$b = [\ln (v_2^2/v_1^2)]/4\pi l_r z_r \qquad \text{(m)} \tag{46}$$

In eq. (46) it is assumed that the mutual distance between the roots is sufficiently large to avoid interference.

The final equation expressing the water flux through the roots to the atmosphere is obtained by combining the fifth term of eq. (42) with eq. (45) yielding:

$$q = q_r = -(h_{leaf} - h_{soil})/(b/k + r_{pl}) \qquad (\text{m.s}^{-1}) \tag{47}$$

A detailed analysis of the importance of the effects of r_{soil} ($= b/k$) and r_{pl} with the aid of measured heads and fluxes will be necessary, because a reduction in evaporation flux may cause a serious reduction in the production rate of a crop. For such an analysis one needs to be informed on the water withdrawal patterns from the soil by the roots, both with depth and time. For a certain period these patterns may be obtained by applying the water conservation equation (14b) on a given volume of soil (RIJTEMA, 1965; ROSE and STERN, 1967). The flow through the roots is calculated as the measured total flow both through roots and soil, diminished with the calculated flow through the soil. Using the subscripts r, u, d and s for roots, upward, downward and soil respectively, eq. (14b) can for every depth z be written as:

$$\int_{t_1}^{t_2} q_r(z,t)\mathrm{d}t = \int_{t_1}^{t_2} \Big[q_u(0,t) - q_d(0,t) \Big] \mathrm{d}t - \int_0^z \int_{t_1}^{t_2} \frac{\partial \theta}{\partial t} \mathrm{d}z\mathrm{d}t +$$

$$- \int_{t_1}^{t_2} q_s(z,t)\mathrm{d}t \qquad \text{(m)} \tag{48}$$

To simplify calculations, it will be convenient to use time averaged values at depths z:

$$\bar{q}_r(z) = \frac{1}{(t_2 - t_1)} \int_{t_1}^{t_2} q_r(z)\mathrm{d}t \qquad (\text{m.s}^{-1}) \tag{49}$$

For the water uptake by the roots $\Upsilon_r = \mathrm{d}q_r/\mathrm{d}z$ (s^{-1}), the following expression is always valid:

$$\int_{t_1}^{t_2} q_r(z,t) = \int_0^z \int_{t_1}^{t_2} \Upsilon_r(z,t)\mathrm{d}z\mathrm{d}t \qquad \text{(m)} \tag{50}$$

For the first three terms on the right hand side of eq. (48), see section IIIB-3a. The vertical flux through the soil $q_s(z,t)$ can be calculated from eq. (14).

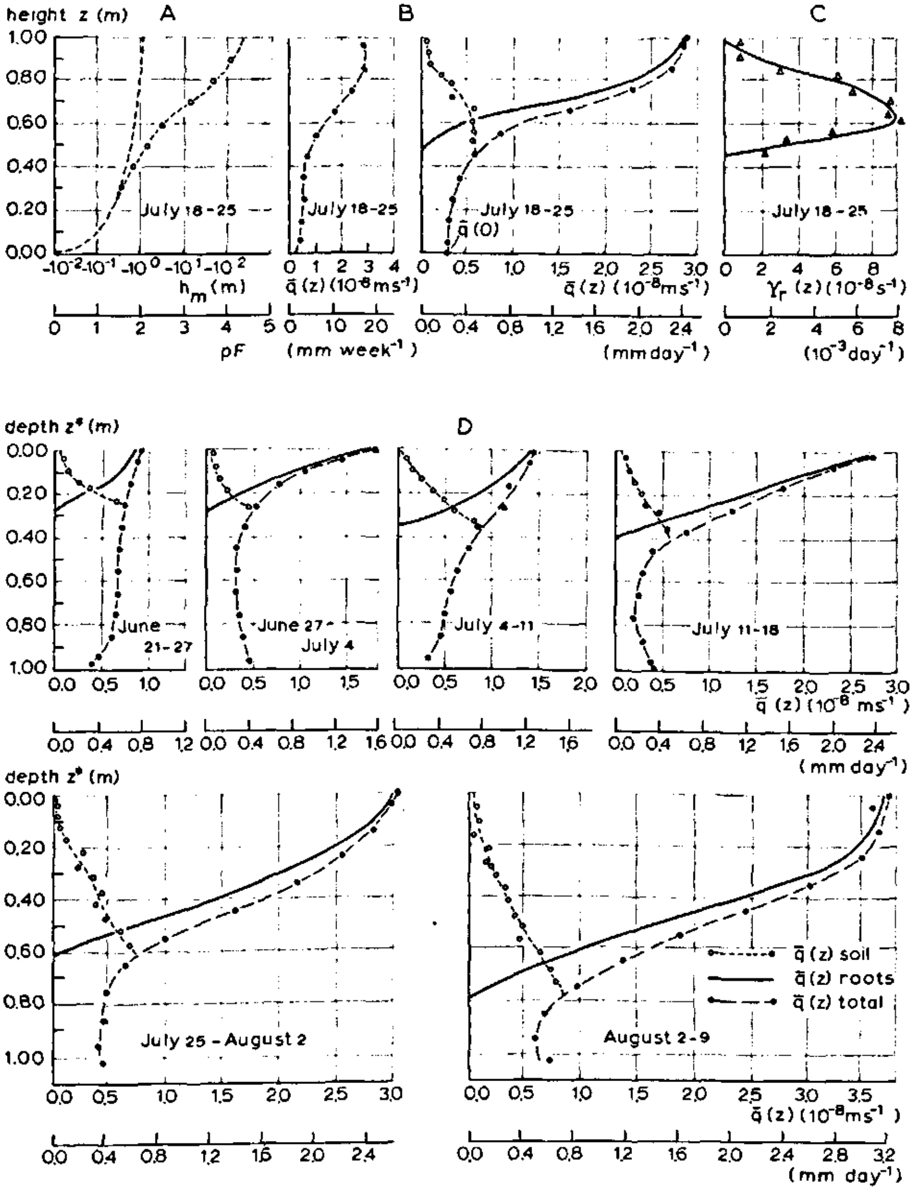


FIG. 24. *Profiles of A, matric head, B, time averaged cumulative water withdrawal of both crop and soil $\bar{q}_{total}$, of soil only $\bar{q}_{soil}$ and of roots only $\bar{q}_{roots}$, and C, time averaged extraction rate Υ_r for red cabbage on clay for the period July 18 through July 25, 1967; D, water withdrawal patterns for the other periods indicated.*

In table 10 an example of the calculation of water extraction by roots at depths z is given for red cabbage on clay for the period July 18 through July 25, 1967. The mean matric head in this period is given in fig. 24A, the time averaged cumulative withdrawal patterns of both crop and soil ($\bar{q}_{total}$), soil only ($\bar{q}_s$) and crop only ($\bar{q}_r$) are presented in fig. 24B, while the time averaged extraction rate at depths z, $\Upsilon_r(z)$, is given in fig. 24C. From figs. 24B and 24C it follows that the effective rooting depths can be found as the depths where the cumulative with-

TABLE 10. Calculation of the water withdrawal pattern of red cabbage on clay for $z = 0$ at 1 m below

z	$\rho^*(t_1)$	$\rho^*(t_2)$	$\bar{\rho}^*$	$\rho^*(t_1)-\rho^*(t_2) = \Delta M_i$	$\bar{q}_{total}(z,t) = \bar{q}(0,t) + \frac{\Sigma \Delta M_i}{\Delta t}$	$\bar{q}_{total}(z,t)$	$-\bar{h}_m$
m	kg.m^{-3}	kg.m^{-3}	kg.m^{-3}	mm	mm.week^{-1}	mm.day^{-1}	m
1.000						2.48	223.90
0.975							205.05
0.950					16.2	2.45	186.20
0.925							163.75
0.900	1514	1515	1514	−0.1		2.41	141.30
0.875							121.80
0.850					16.3	2.33	102.30
0.825							79.27
0.800	1573	1546	1559	2.7		2.15	56.23
0.775							42.21
0.750					13.6	1.95	28.18
0.725							21.84
0.700	1573	1534	1554	3.9		1.67	15.49
0.675							10.90
0.650					9.7	1.36	6.31
0.625							4.85
0.600	1554	1508	1531	4.6		0.97	3.39
0.575							2.74
0.550					5.1	0.73	2.09
0.525							1.73
0.500	1627	1610	1618	1.7		0.60	1.38
0.475							1.24
0.45					3.4	0.50	0.96
0.40	1708	1698	1703	1.0		0.39	0.74
0.35					2.4	0.35	
0.30	1706	1704	1705	0.2		0.33	0.33
0.25					2.2	0.30	
0.20	1662	1659	1660	0.3		0.27	0.20
0.15					1.9	0.26	
0.10	1605	1602	1604	0.3		0.25	0.10
0.50					1.6	0.24	
0.00	1558	1554	1556	0.4		0.17	

* all values in this column from fig. 27A

** all values in this column from fig. 9

drawal function of total flow and soil flow intercept. In this paper the maximum rooting depth (z^*_{max}) is taken to be the effective rooting depth z^*_{eff} plus 50 mm. From fig. 24C it can be seen that for the period in question the maximum extraction rate occurs at a depth of 0.30 to 0.35 m below surface.

As said before, the extraction patterns of the various periods (fig. 24B and 24D) will be used to give a quantitative analysis of the magnitude and variation of the resistances r_{soil} and r_{pl} with time. For this purpose one can divide for

urface $\bar{q}(0, t) = 1.2$ mm. week $^{-1}$, t_1 = July 18 and t_2 = July 25, 1967

$-\bar{h}_m)$ m	$\bar{k}$ mm.day^{-1}	$\frac{\Delta(-\bar{h}_m)}{\Delta z} - 1$	$\bar{q}_{soil}(z, t)$ mm.day^{-1}	$\bar{q}_{soil}(z, t)$ mm.day^{-1}	$\bar{q}_{roots}(z, t)$ mm.day^{-1}	$\Delta\bar{q}_{roots}$ mm.day^{-1}	$\frac{\Delta\bar{q}}{\Delta z} = \Upsilon_r$ day^{-1}
				0.02	2.46		
7.70	4.65×10^{-5}**	753	0.03			0.06	0.0012
				0.05	2.40		
4.90	6.3×10^{-5}	897	0.06			0.06	0.0012
				0.07	2.34		
9.00	9.5×10^{-5}	779	0.08			0.13	0.0026
				0.12	2.21		
6.06	1.72×10^{-4}	920	0.18			0.26	0.0052
				0.20	1.95		
8.06	4.25×10^{-4}	560	0.28			0.30	0.0060
				0.30	1.65		
2.69	1.05×10^{-3}	253	0.30			0.38	0.0076
				0.40	1.27		
9.18	2.70×10^{-3}	183	0.49			0.38	0.0076
				0.47	0.89		
2.93	8.20×10^{-3}	57	0.47			0.41	0.0082
				0.49	0.48		
1.30	1.90×10^{-2}	25	0.48			0.25	0.0050
				0.50	0.23		
0.71	3.60×10^{-2}	13	0.47			0.14	0.0028
				0.51	0.09		
0.42	5.40×10^{-2}	7	0.54			0.09	0.0018
				0.50	0.00		

each period the root zone in a number of layers, say of 50 mm thickness and apply eq. (47) to each layer separately. In this way for each period a number of equations can be evaluated with known values of $\bar{q}_r(z)$, $\bar{h}_s(z)$ and $1/\bar{k}(z)$, from which the three unknown variables $\bar{h}_{leaf}$, $b(z)$ and $r_{pl}(z)$ have to be solved.

For a crop as grass with a root system which does not change too much with time, it can be assumed that $b(z)$ and $r_{pl}(z)$ are constant with time (RIJTEMA, 1965). This does not hold, however, for a cabbage crop where the root system changes with depth and time (see also RAWITZ, 1970). Therefore one can expect that for each period there must be a functional relationship of b and r_{pl} with depth. As a first approach, for each period an exponential variation of b, r_{pl} and the root mass W with depth was adopted. With the depth below soil surface z^* (m) taken positive downwards, it follows that:

$$b(z^*) = b(0) \exp(\alpha z^*) \tag{51}$$

$$r_{pl}(z^*) = r(0) \exp(\alpha z^*) \tag{52}$$

$$W(z^*) = W(0) \exp(-\alpha z^*) \tag{53}$$

To determine α it is assumed that $W(z^*_{max}) = 0.01\, W(0)$ and, as stated earlier, $z_{max} = z_{eff} + 0.05$. The value of z_{max} is obtained both from measurements of rooting depth in the field and from figs. 24B and 24D. This procedure was taken, because a determination of root mass in clay soil is hardly possible. Now, substitution of eq. (51) and (52) in eq. (47) leads for every depth z^* to:

$$\bar{q}_r(z^*) = -\frac{\bar{h}_l - \bar{h}_s(z^*)}{\bar{b}(0)\exp(\alpha z^*)/\bar{k}(z^*) + \bar{r}_{pl}(0)\exp(\alpha z^*)} \tag{47a}$$

Rearranging it to a simpler form gives:

$$\begin{aligned} -[\bar{h}_s(z^*)/\bar{q}_r(z^*)]\exp(-\alpha z^*) &= -[\bar{h}_l/\bar{q}_r(z^*)]\exp(-\alpha z^*) + \\ &\quad - \bar{b}(0)/\bar{k}(z^*) - \bar{r}_{pl}(0) \end{aligned} \tag{47b}$$

$$\begin{aligned} y \quad &= \quad a_1 \qquad x_1 \quad + \\ &\quad + a_2 \quad x_2 + a_0 \end{aligned} \tag{47c}$$

Application of eq. (47c) to three or more depths z^*, with increments of 0.05 m leads to a solution of the variables $\bar{h}_l$, $\bar{b}(0)$ and $\bar{r}_{pl}(0)$. Because of uncertainty of matric heads and fluxes in the top layer, calculations were started at 0.125 m below surface.

With this procedure a first impression was obtained of the order of magnitude of $\bar{h}_l$, $\bar{b}(0)$ and $\bar{r}_{pl}(0)$, as well as of the values of b and r_{pl} at various depths z^*. It soon appeared that according to the assumed exponential variation of r_{pl} with depth, r_{pl} increased too fast at the greater depths. At these depths the root mass decreases but young roots are also formed, giving a relatively large activity and permeability. Due to high k values in these regions, the soil resistance (b/k) will be low and possibly incorrect estimates of b will be of minor importance. Soil resistance is relatively low as compared with plant resistance. The reverse holds for the top layers. When the soil is dry, k will be small and in comparison with r_{pl}, the soil resistance (b/k) becomes relatively large. This is the reason that wrong estimates of r_{pl} have for dry top layers only a slight influence on the total resistance of the soil-plant system.

Keeping this in mind, in a second attempt an iterative procedure was followed, taking the $\bar{h}_l$ values and b values obtained with the first approach as a basis. This yielded r_{pl} values through which smooth curves were drawn. From the curves, corrected r_{pl} values were read and used to determine new b values and vice versa. This procedure was repeated several times.

The variation of r_{pl} and b with depth during the various periods is shown in fig. 25. It is seen that during the first three weeks of growth there is an almost exponential relationship of r_{pl} with depth. After this period, the r_{pl} data below 0.25 m increase less fast as a result of the development of new roots. The depth and extent of branching of the roots is nicely marked by the b curves. In the first 3 to 4 weeks root development was rather poor. After this period, however, the distance to which roots were extending changed in less than a month from 0.35 to 0.80 m.

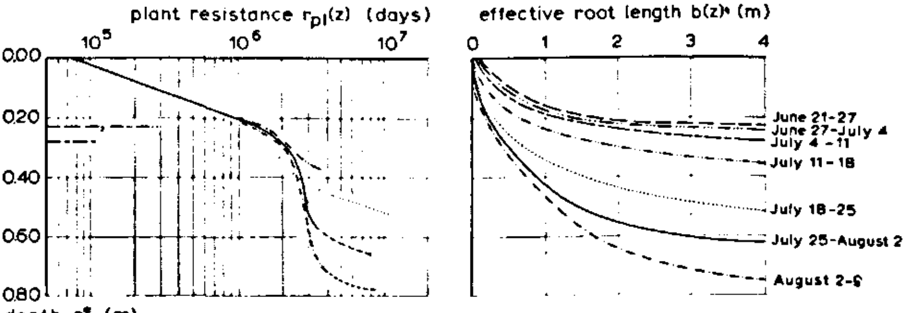


FIG. 25. *Variation in plant resistance and effective root length with depth for a number of consecutive periods.*

With aid of the curves for r_{pl} and b in fig. 25 root extraction rates at different depths were calculated for the various periods and compared with data obtained from the water balance studies. The results for the last 4 weeks are shown in fig. 26. Except for the upper soil layer, the calculated values compare well with the experimental data. It has already been pointed out that the 'experimental'

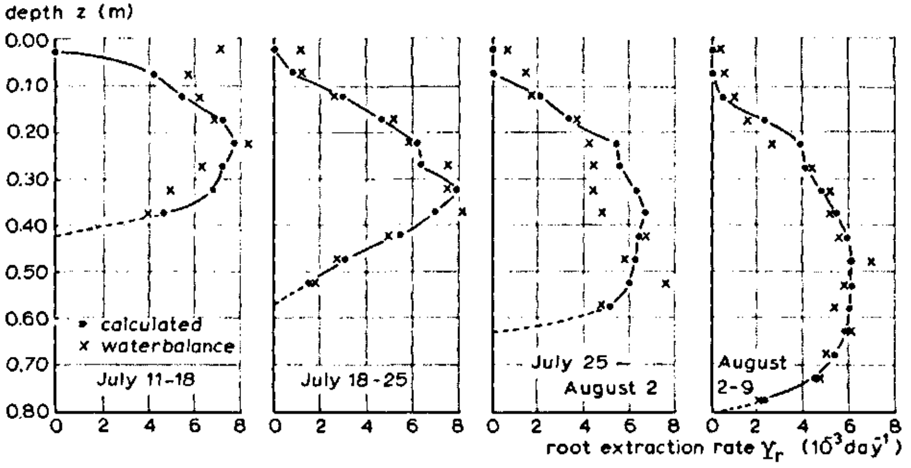


FIG. 26. *Comparison of time averaged root extraction rates for red cabbage calculated with the r_{pl} and b curves of fig. 25 with rates obtained from water balance studies over four consecutive weeks.*

data in the top layer were obtained by means of extrapolation. Due to small amounts of rain, the matric heads in the top layer may be changing from time to time, probably resulting in errors. Fig. 26 shows that water uptake by the roots in a top layer of about 0.05–0.10 m is about neglectible.

The overall resistance can be computed when taking into account the various resistances in each part of the pathway by means of the following expressions:

$$\bar{r}_{pl} = \frac{1}{n} \frac{\sum_{n=1}^{n=n} \Upsilon_{r,n} r_{pl,n}}{\sum_{n=1}^{n=n} \Upsilon_{r,n}} \tag{54}$$

$$\bar{r}_{soil} = \frac{1}{n} \frac{\sum_{n=1}^{n=n} \Upsilon_{r,n} b_n / k_n}{\sum_{n=1}^{n=n} \Upsilon_{r,n}} \tag{55}$$

The results of the calculations for $\bar{r}_{pl}$ (eq. 54), $\bar{r}_{soil}$ (eq. 55) and $\bar{b}$ are presented in table 11.

In literature the potential is often expressed per unit volume, then r_{pl} has the dimension of for example bar.day.cm^{-1} (equivalent with 1019 days) and b the dimension of bar (equivalent with about 10.19 m). GARDNER and EHLIG (1963) report from pot experiments with Birdsfoot Trefoil a r_{pl} of 11,200 days and a b

TABLE 11. Results of the calculations on the mean plant resistance ($\bar{r}_{pl}$), mean soil resistance ($\bar{r}_{soil} = \bar{b}/\bar{k}$) and mean effective root length (b)

Line	Period	$\bar{r}_{pl}$ days	$\bar{r}_{soil}$ days	$\bar{b}$ mm
1	June 21 to 27	27,160	37,240	24.2
2	June 27 to July 4	18,290	33,210	19.8
3	July 4 to 11	31,860	33,420	13.0
4	July 11 to 18	31,410	21,430	5.4
5	July 18 to 25	41,220	10,150	2.7
6	July 25 to Aug. 2	37,490	7,920	1.6
7	Aug. 2 to 9	36,340	5,490	1.2

of 0.4 mm. RIJTEMA (1965) gives for grass grown on clay a r_{pl} value of 10,618 days and a b value of 3.8 mm. RIJTEMA and RYHINER (1966) found for summer wheat grown on sand, values of 30,570 days and 2.2 mm respectively. For potatoes on sand r_{pl} was equal to 10,322 days and b was 3.1 mm (ENDRÖDI and RIJTEMA, 1969).

The r_{pl} data for cabbage (table 11) are in reasonable agreement with the mentioned data. The b data of the initial growing stage (lines 1, 2, 3, table 11) differ a factor ten from the b values reported above. This is mainly caused by the non-homogeneous and poor root development in the early stages of growth. When root development increases with depth, b values (lines 4–7, table 11) decrease to values also reported for other crops. The given data for red cabbage demonstrate that possible effects of low matric heads in the soil on the evaporation flux in the early stages of growth, will start at a lower matric head than in the later stages of growth.

The accuracy of the calculation of the overall resistances of plant and soil is largely dependent on the accuracy of the determination of Υ_r (eqs. 54 and 55). Moreover the distribution in Υ_r with depth is important in the determination of the final resistances. For example in consecutive dry periods the water uptake in the top layers is about zero, but this may be quite different in a dry period occurring shortly after a rainy one. This may cause relatively important changes in r_{pl} and minor changes in b.

RIJTEMA (1965) derived the following expression for the effect of soil moisture conditions, overall resistance of the soil plant system and climatological conditions on the diffusion resistance r_h:

$$\bar{r}_h = \mathrm{f}(\bar{h}_l) = \mathrm{f}\,[\bar{E}_p(\bar{r}_{pl} + \bar{b}/\bar{k}) + \bar{h}_{soil}] \tag{56}$$

From substitutions of the various data mentioned, it appears that r_h can range from 0 to 600 s.m^{-1}. The high r_{pl} value of summer wheat may be the reason that even under a good supply of water evaporation of the crop at high E_p values can be severely reduced, as shown in fig. 14.

5. Calculated and measured evaporation

The evaporation results obtained with four crops during the three years of the experiments will now be discussed briefly.

Spinach. The first experiments with spinach were performed in the spring of 1967 on the 0.90 m groundwater plot only. The problem immediately met during the dry period the experiment started was the poor germination and emergence on the clay and on the clay on sandy loam, in sharp contrast with the good and rapid development on the sandy loam. During the growing period this last profile showed potential evaporation rates amounting from 1.4 mm.day^{-1} during the period of April 28 to May 11, to about 3 mm.day^{-1} during the period of May 11 to May 31. The average matric pressure in the root zone was about -0.3 bar (pF 2.5) which is to be regarded as optimum for the production of spinach (FEDDES, 1969b). The average matric pressure on the top 0.10 m of clay soils was below -3.1 bar (pF 3.5) which is far too low for a good germination (see Chapter V).

The experiments with spinach were repeated in the autumn of 1968 on the profile of the 0.45, 0.90 and 1.20 m groundwater plots. The crop started well on all the plots after 20 mm precipitation falling soon after sowing. After seedling emergence, however, frequent rain storms totally destroyed the crop on the 0.45 m groundwater plot (fraction of soil covered at the end of the growing season about 0.04). On the sandy loam profiles of both the 0.90 and 1.20 m groundwater plot the crop reached a S_c value of about 0.50. The same values occurred on the clay soils of the 0.90 m plot. On the 1.20 m plot on the other hand the soil covered reached 100% on the clay indicating the favourable behaviour of the deeply drained clay soil in autumn as compared with sandy loam soils and the shallowly drained clay soils. Evaporation on the various plots varied between 0.5 to 1.5 mm.day^{-1}.

Red cabbage. An example of the evaporation calculations for red cabbage on the clay and on the clay on sandy loam profile is presented in fig. 27A and B. It appears that calculations of real evaporation (eq. 29) for clay on a weekly basis compare reasonably well with the values obtained by means of the water balance. The calculations were performed including the resistance r_h. For clay on sandy loam, however, not enough information about this resistance was available as the rooting depth was too small. From comparison with the calculated potential

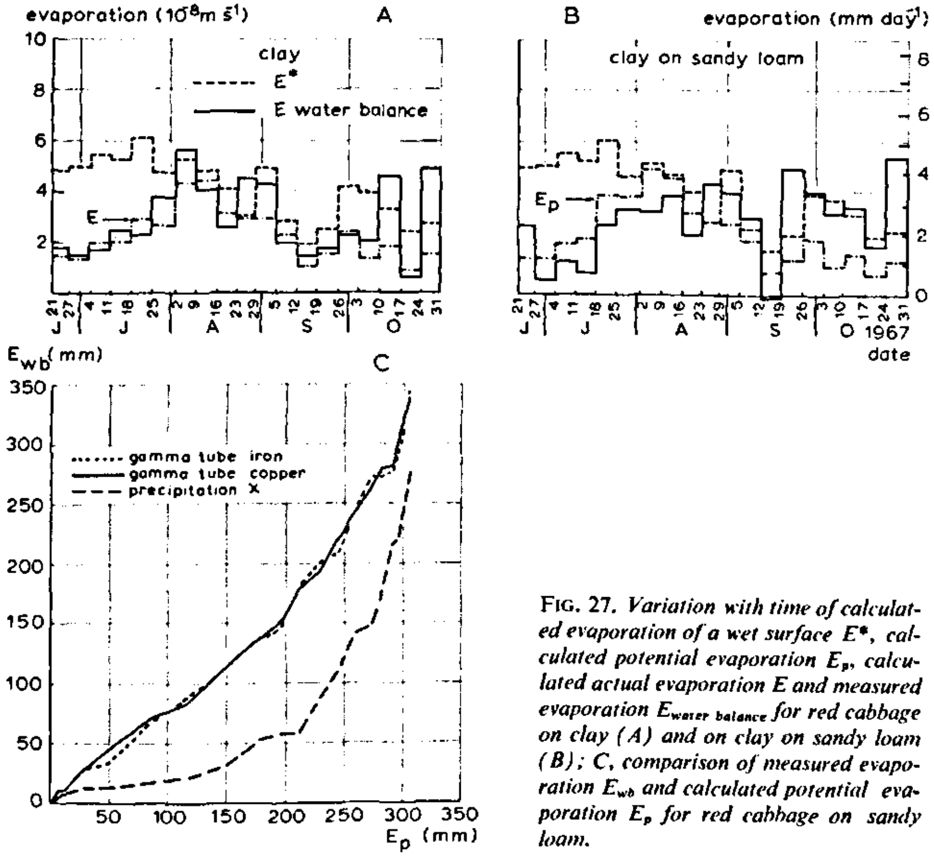


FIG. 27. *Variation with time of calculated evaporation of a wet surface E^*, calculated potential evaporation E_p, calculated actual evaporation E and measured evaporation $E_{water\ balance}$ for red cabbage on clay (A) and on clay on sandy loam (B); C, comparison of measured evaporation E_{wb} and calculated potential evaporation E_p for red cabbage on sandy loam.*

evaporation (E_p) it is seen that also on this profile actual evaporation is reduced in the initial stages of growth. From the discussion presented in section B3-d4 of this Chapter, it is clear that this must be due to the restricted root growth on the clay on sandy loam which apparently causes high plant (r_{pl}) and soil (b/k) resistances.

There are some deviations between calculated and measured evaporation from both profiles, occurring in the same periods. In the periods of September 12 to 19, October 10 to 17 and October 24 to 31, heavy rainfall caused a rise of the water tables even above some of the tensiometers. From inspection after the experiments it appeared that leakage along the tensiometers had taken place, which disturbed the water balance measurements. In the period of August 23 to 29 the measured evaporation (E_{wb}) increased to above the wet evaporation (E^*). This was probably due to the extraordinary meteorological conditions of relatively high solar radiation (178 W.m^{-2}) *in* combination with low wind velocities ($\bar{u}$ = 1.38 m.s^{-1}) and high humidities (0.91), causing wrong readings

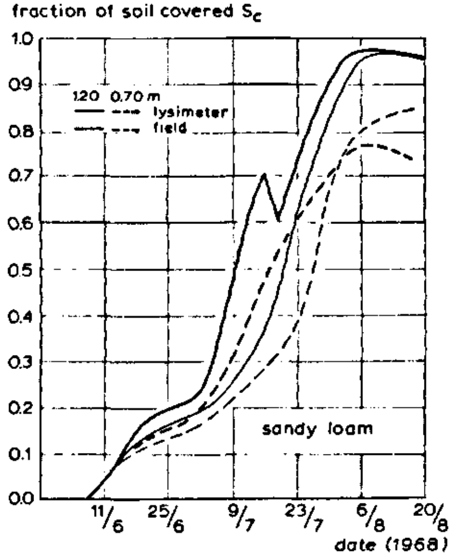


FIG. 29. *Fraction of soil covered by dwarf French beans on the sandy loam profiles of the lysimeters and of the fields at groundwater depths of 0.70 m and 1.20 m below surface.*

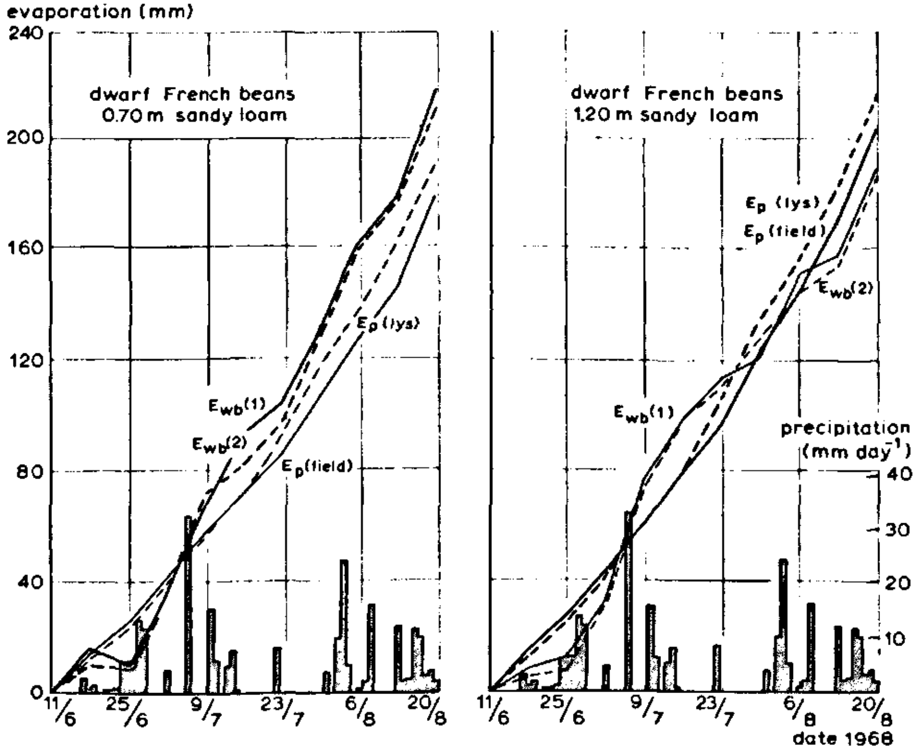


FIG. 30. *Measured cumulative evaporation E_{wb} (1 = gamma tubes in lysimeter; 2 = tubes in field) and calculated cumulative potential evaporation E_p of dwarf French beans grown on the lysimeters and fields of the sandy loam profiles of the 0.70 m and 1.20 m groundwater plots.*

Similar results were obtained on the clay and on the clay on sandy loam profiles of the 0.70 and 1.20 m groundwater plots.

Celery. Because celery makes high demands upon structure, water content and nitrogen content of the soil, this crop was chosen for the last year (1969) of the experiments.

On April 9 celery was sown on the field by machine in rectangular patterns of 0.22 × 0.22 m at a shallow depth of about 10 mm. In the lysimeters and between the gamma tubes the seed was sown by hand. About 4 weeks after sowing the seedlings emerged. It soon appeared that the best emergence occurred in the tracks made by the tractor wheels. Although the precipitation conditions were favourable, the fact that the seed was lying deeper in the looser upper soil layer between these tracks and in the lysimeters finally resulted in poor emergence. On June 3 celery plants taken from the field were transplanted in the lysimeters and between the gamma tubes. Because of the partial failure of the transplants settled in the rather dry month of May, transplanting was repeated on June 25. Additional water was supplied. As a total failure of the crop by the very dry conditions in the top soil threatened to occur, it was decided as a last measure to sprinkle the entire experimental field on July 1. Because the crop (especially on the sandy loam soils) showed a yellow colour which pointed to nitrogen deficiency, an additional amount of nitrogen was applied together with the irrigation treatment. Probably the earlier applied amount of nitrogen did not become available to the plants owing to the dry conditions. One week after irrigation the crop started to grow very rapidly and height of and soil coverage by the crop quickly increased. After the first harvest, a second irrigation treatment was carried out on July 28 in combination with an additional gift of nitrogen.

The various treatments are reflected in the curves of fig. 31, where the evaporation results of the sandy loam objects are shown. During 1969 one set of the gamma tubes was installed inside the lysimeter, the other set remaining outside the lysimeter. Because soil coverage on the lysimeter and between the gamma tubes in the field was almost equal and evaporation data obtained with the different tubes did show only small differences, the evaporation data obtained with these tubes were averaged. Both calculated potential evaporation and measured evaporation on the 0.70 m sandy loam plot were higher than on the sandy loam of the 1.20 m groundwater plot. This was due to considerable capillary rise on the 0.70 m plot, 39 mm between April 29 and June 23 as against 11 mm in the 1.20 m groundwater plot. After the crop after 45 days had reached a soil coverage of 40 to 50% not much difference remained between E_p and E_{wb} of the lysimeter, the curves running almost parallel. It now was taken that E_p calculated for the field was also correct. As the field had a much lower percentage

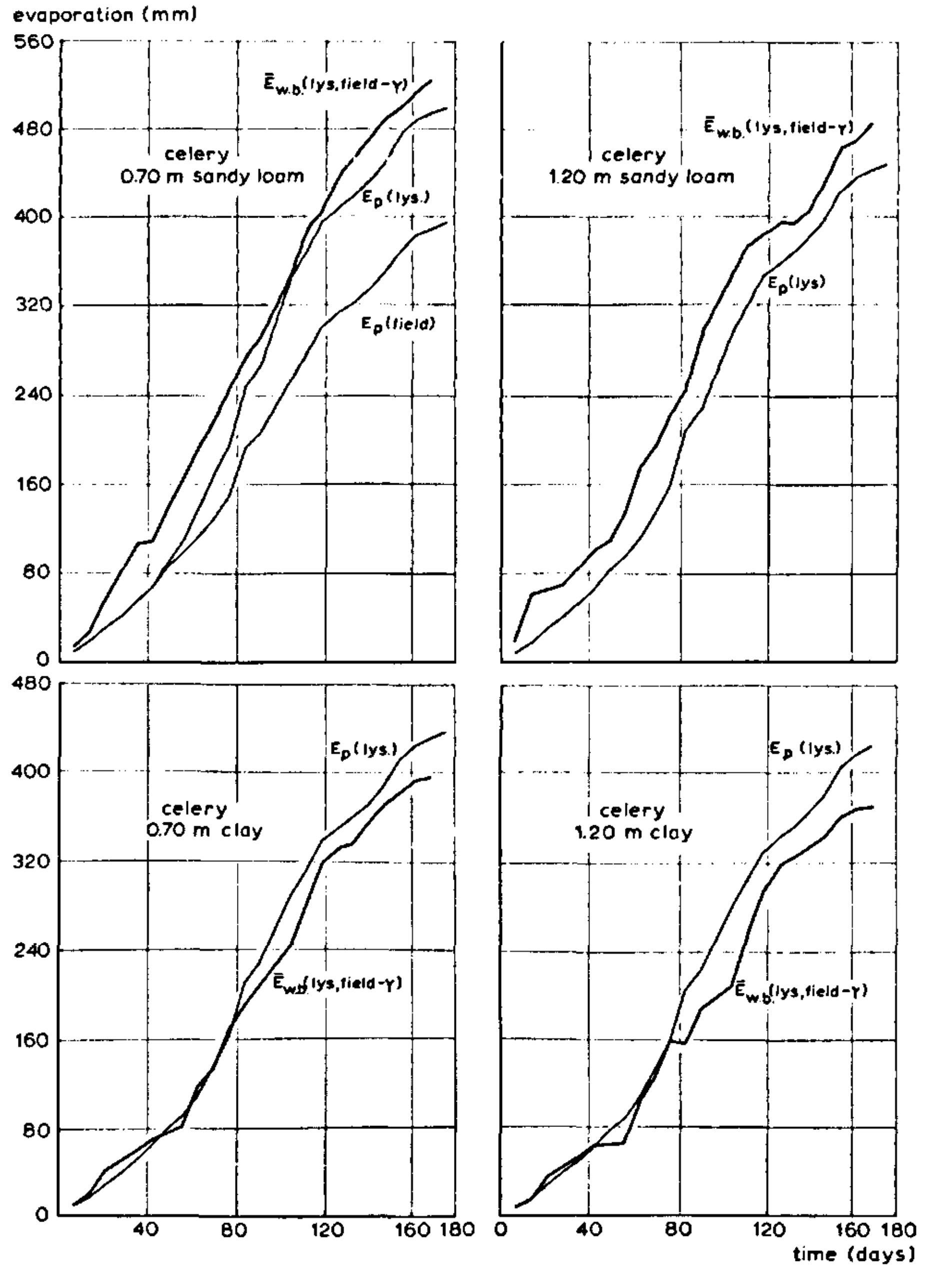


FIG. 31. *Cumulative evaporation data of celery on the sandy loam and clay profiles of the 0.70 m and 1.20 m groundwater plots. E_p = potential evaporation; $\bar{E}_{wb}$ (lys., field–γ) is the mean evaporation obtained from the set of the gamma tubes installed in the lysimeter and the set of gamma tubes installed in the field (see also text).*

of soil covered, water use on the field was far below that on the lysimeter (fig. 31).

It is striking that the 0.70 and 1.20 m clay soils and clay on sandy loam soils, did not show differences in evaporation, neither for the different profiles nor for the different water table depths (fig. 31). The reason for this is the shallow root system of celery, on all objects going to only 0.25 m below surface. This explains that on these heavy profiles the groundwater tables were still not high enough to ensure an optimum water supply from the groundwater. It also explains why the celery could not readily compete with weeds, why nitrogen did not become available in sufficient amounts in dry periods and why the top soil had to be abundantly supplied with water and nutrients. The necessity to supply additional water was clearly illustrated in the middle of August when, owing to the very humid meteorological conditions, a tremendous amount of new roots was formed at the surface of the soil all over the field.

From the evaporation data for celery additional information on evaporation from bare soil can be obtained. In fig. 32 the wet evaporation (E^*) of the 0.70 m sandy loam, the potential evaporation (E_p) of all the other objects, the precipitation (χ), and the measured data of the water balance of the different plots are shown. It appears that in times of precipitation the evaporation of the 0.70 m and 1.20 m sandy loam soils follow the evaporation (E^*) of an almost bare wet soil. In periods of little precipitation, evaporation of the 0.70 m sandy loam still follows the E^*. Apparently the capillary rise from the 0.70 m groundwater table is still sufficient for this soil to have it behave as a wet soil. In periods of drought this does not hold for the 1.20 m sandy loam, where evaporation follows the potential evaporation (E_p) of a dry soil. The clay soils behaved as dry soils both in dry and in rather wet periods, obviously because the rain did not completely moisten the cultivated top layer but percolated mostly to the subsoil. Summarizing, it can be stated that evaporation from bare soil both for clay soils and deeply drained sandy loam soils can be computed from eq. (27), with additional use of fig. 22 to estimate r_c. For bare sandy loam soils with a high groundwater table and also for deeply drained sandy loam soils in wet periods, evaporation (E^*) should be computed with the aid of eq. (26).

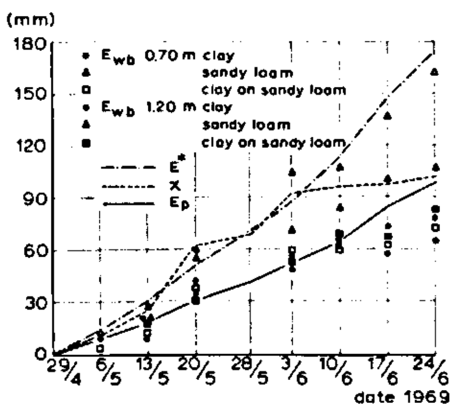


FIG. 32. *Cumulative evaporation data of celery on the three profiles of both the 0.70 m and 1.20 m groundwater plots.* E_{wb} *= measured evaporation;* E^* *= evaporation of a wet surface;* χ *= precipitation;* E_p *= potential evaporation.*

IV. HEAT

1. INTRODUCTION

Soil temperature is one of the most important factors for vegetable crop production, as it influences germination, seedling emergence, early growth of plants, maturation and yield.

The thermal regime that exists within a soil profile in situ is the result of the energy balance at the surface and the thermal properties of the soil, e.g. the thermal conductivity and the thermal capacity. Measures like changing the depth of the water table and improving the soil will change the thermal properties and therefore influence the thermal regime. To evaluate the consequences these measures will have, and to create the most favourable environmental conditions for the crop in this regard, a detailed knowledge of heat transfer in soil is required, with which the thermal regime at any depth and time under given climatological conditions can be described.

2. HEAT CONDUCTION IN SOIL

To a large extent heat transfer in vertical direction in soil can be considered as consisting of a molecular conduction of heat, the transport of heat by convection, by distillation and by radiation being generally of minor importance. The magnitude of the heat flux in a homogeneous soil is proportional to the temperature gradient and the thermal conductivity:

$$q_h(z,t) = -\lambda \nabla T(z,t) \tag{57}$$

where:

q_h = heat flux ($W.m^{-2}$)
z = depth (m)
t = time (s)
T = temperature (K)
λ = thermal conductivity of the soil ($W.m^{-1}.K^{-1}$)

The negative sign was introduced because the heat flux (q_h) is positive in the direction of falling temperatures and ∇T is negative. Eq. (57) is known as the one-dimensional form of Fourier's law of heat conduction.

The principle of continuity requires that the difference in heat flux into and out of an elementary soil element with thickness dz equals the rate of storage:

$$- \nabla q_h(z,t) = \rho c \frac{\partial T(z,t)}{\partial t} \tag{58}$$

where:
ρ = specific mass (kg.m^{-3})
c = specific heat per unit mass (J.kg.K^{-1})
ρc = heat capacity of the soil per unit volume (J.m^{-3}.K^{-1})

Applying eq. (58) requires that no heat exchange in other than vertical (z) direction and no heat generation takes place. Substitution of eq. (57) into eq. (58) gives:

$$\nabla \left[\lambda \nabla T(z,t) \right] = \rho c \frac{\partial T(z,t)}{\partial t} \tag{59}$$

3. FIELD AND LABORATORY EXPERIMENTS

Soil temperatures in the field were measured with parallel circuits of copper-constantan couples at depths of 0.01, 0.02, 0.03, 0.05, 0.10, 0.30 and 1.20 m, the output of which was written down by two multipoint recorders. The accuracy of temperature readings was within 0.1 to 0.2 °C. The measurements were taken on the clay and sandy loam; in spring on the 0.45 and 1.65 m groundwater plot, in summer and autumn on the 1.20 m plot.

On some clear days both in wet and dry periods, the variation with depth of the thermal conductivity of the soil in situ was determined with the so-called transient needle method, which will be discussed below. For this purpose line probes were inserted horizontally, at depths of 0.01, 0.02, 0.03, 0.04, 0.05, 0.075, 0.10, 0.15, 0.20 and 0.30 m, into the wall of a small trench. The measurements in the top soil were repeated several times, until only small deviations in the individual data remained. The applied current was 0.1–0.2A, the measuring time did not last longer than 100 seconds and the difference in temperature between the needle and its surroundings was always smaller than 2 °C. Simultaneously, moisture content and dry bulk density were obtained by taking samples of known volume at depths of 0.05, 0.15, 0.25, 0.35 and 0.45 m.

The obtained variations of the thermal properties with depth were checked in the laboratory with an electric analog, the details and theory of which are discussed in section 5 of this Chapter.

Undisturbed soil samples of both profiles were taken in plastic cylinders of

0.10 m diameter and 0.50 m length, following the method of WIJNSMA and WIT (1970). These cylinders were cut into pieces of 0.10 m length. Measurements of thermal conductivity were carried out with the needle method on each separate sample at successive stages of desiccation. With the data obtained, a number of calibration curves of conductivity versus moisture content were constructed for various dry bulk densities. With the aid of these calibration curves, it was thought that the course of conductivity with depth could be determined indirectly from core samplings.

4. THERMAL PROPERTIES OF SOIL

As can be seen from eq. (59) heat transfer in soil is completely governed by two independent thermal parameters, the thermal conductivity and the volumetric heat capacity. They depend mainly on the mineral composition, the dry density and the moisture content of the soil will therefore vary with depth and time. The methods of determination of these properties will be briefly discussed.

a. THERMAL CAPACITY

For a non-homogeneous material such as soil, the heat capacity per unit volume of soil (ρc) equals the sum of the volumetric heat capacities of the different components. Denoting the volume fractions of the solid material, of the water and of the air as x_s, x_w and x_a respectively, and the volumetric heat capacities of these components as $\rho_s c_s$, $\rho_w c_w$ and $\rho_a c_a$, one can write:

$$\rho c = x_s \rho_s c_s + x_w \rho_w c_w + x_a \rho_a c_a \tag{60}$$

According to measurements of KERSTEN (1949), as quoted by DE VRIES (1966), ρ_s is for mineral soils about 2700 kg.m^{-3}, while c_s varies linearly from 670 $\pm$ 40 J.kg^{-1}.K^{-1} at 255 K ($-18\,^\circ$C) to 790 $\pm$ 40 J.kg^{-1}.K^{-1} at 333 K (60 $^\circ$C). This results in an average value of $\rho_s c_s$ for mineral soils of 1.92×10^6 J.m^{-3}.K^{-1}.

For organic soils ρ_s is as an average 1500 kg.m^{-3}, while for c_s a value of 1.92×10^3 J.kg^{-1}.K^{-1} can be used [DE VRIES (1966) referring to measurements of Lang, Ulrich, Bracht and DE VRIES and DE WIT (1954)]. Thus for organic soils $\rho_s c_s = 2.88 \times 10^6$ J.m^{-3}.K^{-1}.

The volumetric heat capacity for water $\rho_w c_w = 4.18 \times 10^6$ J.m^{-3}.K^{-1}. As $\rho_a c_a$ is in the order of magnitude of 1255 J.m^{-3}.K^{-1}, it usually can be ignored. So, denoting the volume fractions of soil minerals by x_{sm} and of organic matter by x_{so}, one can write:

$$\rho c = (1.92\, x_{sm} + 2.88\, x_{so} + 4.18\, x_w)\, 10^6 \qquad (\mathrm{J.m^{-3}.K^{-1}}) \qquad (61)$$

where $x_{sm} + x_{so} + x_w + x_a = 1$.

From mechanical and chemical analyses the particle size distribution and organic matter content of the soil can be obtained. The various volume fractions of the soil can be obtained from sampling at various depths, and ρc can be calculated with the aid of eq. (61), according to DE VRIES (1966) giving an accuracy varying from 5 to 10%. In this way the variation of volumetric heat capacity with depth can be found. In fig. 33 an example of the linear variation of ρc with moisture content is given for two dry bulk densities of the sandy loam.

Measurements of ρc are based on calorimetric methods. Lang, Ulrich, Patten and KERSTEN (1949) used the method of 'mixtures', which consists of mixing a given mass of soil with known temperature and a given mass of liquid with known temperature and specific heat. By measuring the resulting temperature, the specific heat of the soil can be computed. BOWERS and HANKS (1962) as well as DE VRIES and DE WIT (1954) quoting this method, prefer an alternative method based on Newton's law of cooling. This law states, that the rate at which a body with an initial uniform temperature cools is directly proportional to the temperature difference between the body and its surroundings, provided this difference is small. WIERENGA (1968) deduced the heat capacity from the

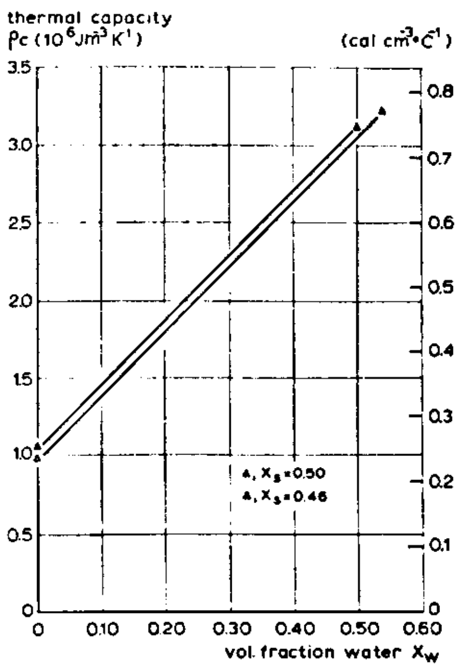


FIG. 33. *Dependence of thermal capacity on volume fraction of water for two volume fractions of solid material x_s.*

temperature rise of a known mass of soil and water in a calorimeter, as caused by a known electric power.

Methods of measurement of ρc in the field are discussed in the next section.

b. Thermal conductivity

An estimation of thermal conductivity of the soil from its structure, in the same way as heat capacity, by considering the conductivity as a function of volume fractions and specific conductivities of the various components, is difficult. There are theoretical and empirical solutions, however. De Vries (1952a) has given a survey of the theoretical work, and developed an approximate solution based on theories of Maxwell, Burger and Eucken in analogy to the calculation of the dielectric constant of granular material, as a function of the electric conductivities and the volume fractions of the components. With this model, one can estimate the thermal conductivity of a soil, if kind, shape and arrangement of the particles and moisture content are known (see also van Wijk, 1966). It seems that the errors in De Vries' calculations were not exceeding 10%.

Measurements of λ in the field or in the laboratory are based on specific solutions of eq. (59), depending on the method used, the geometry and the boundary conditions.

The heat flux is mostly initiated by heating (electrically or by other means e.g. water) a probe (of flat, spherical or cylindrical shape) situated in the soil. Measurements are made either of the power consumption, the rate of heating or cooling of the probe, or the rise in temperature of the surrounding soil.

The surface of the probe may be kept for a short or for a long time at a temperature different from that of the surrounding soil. Long heating of the probe is undesirable because it provokes migration of soil moisture away from the heated body and therefore causes a loss in latent heat. Such a heat loss will also occur if the temperature of the probe relative to that of the surrounding soil is too high. Therefore, the shortly applied temperature differences between probe and soil should not reach more than 2 K.

The shape of the probe is also important. Flat probes have for example the disadvantage of sideway heat losses. Chudnovskii (1948) gave an extensive review on determining thermal properties of soils with plane, cylindrical and other types of probes, both with steady and unsteady state methods, including the often laborious mathematical treatments. With most of the methods the thermal diffusivity $\lambda/\rho c$ is obtained, which in combination with a separate determination of ρc ,yields λ, see for example Krischer and Esdorn (1955).

For a simultaneous determination of ρc and λ in the top layer of the soil, CHUDNOVSKII (1948) used an icebox instrument which is placed on the surface of the soil. Based on the same principle, VAN WIJK and BRUIJN (1964) made use of a heat flux plate also placed on the top of the soil and thermocouples in the soil. From a heat impulse of known intensity which causes a temperature increase in the surface layer, λ and ρc can be found. VAN WIJK (1964, 1967) developed the 'contact'-method, by which a block of Perspex of known temperature is brought into contact with the soil surface (see SCHNEIDER, 1969). This method is in principle the same as the junction method of CHUDNOVSKII (1948) who brings a soil sample in contact with a standard material of paraffin. STIGTER (1969), discussing the method of Van Wijk, concludes that accurate results could only be obtained on dry sandy soils and homogeneous solids.

In the present study, the direct measurement of λ in the field as well as in the laboratory was realized with a line source heater in the form of a needle, first used by Stålhane and Pyk in 1931 as quoted by DE VRIES (1952a). The device consists of an electrically transient heated probe with a temperature measuring device and is constructed in such a way that it represents as much as possible an infinite line source of heat (e.g. CHUDNOVSKII, 1948; DE VRIES, 1952a, b; VAN DRUNEN, 1949; JANSE and BOREL, 1965; HAARMAN, 1969; MOENCH and EVANS, 1970).

GOLOVANOV (1969) showed that, in order to prevent convection of heat, the time of measurement on soil samples should not exceed 100 seconds with soil samples of a diameter of 40 mm.

An example of the results of laboratory measurements of λ for sandy loam at different moisture contents for two dry bulk densities is presented in fig. 34A. In fig. 34B a comparison between measurements and calculations according to

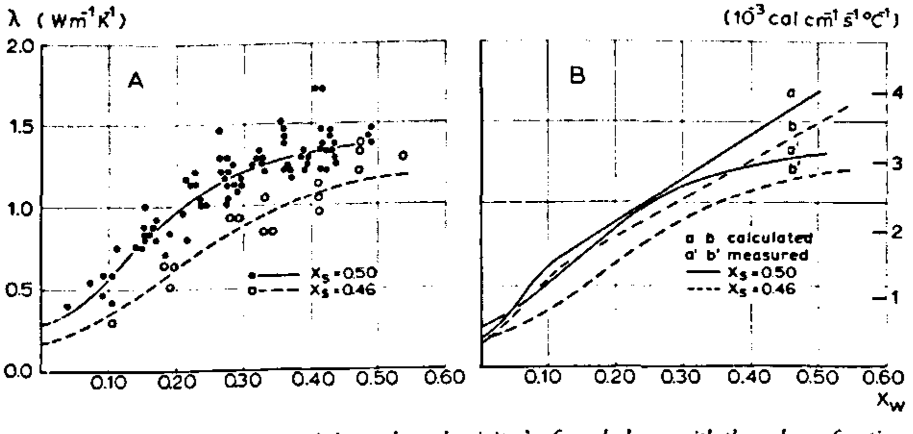


FIG. 34. *A, variation of measured thermal conductivity λ of sandy loam with the volume fraction of water x_w for two volume fractions of solid material x_s; B, comparison of thermal conductivity of sandy loam calculated by the De Vries' method (a, b) with those measured with the transient needle method (a', b') for two volume fractions of solid material.*

DE VRIES (1952a, b) is given. It is seen that in general the calculations result in too large λ values. Such type of calculations are among other things based on the assumption that the (randomly oriented) granules are ellipsoids of revolution, i.e. that the soil particles can be considered as spheroids. In agreement with DE VRIES (1966), for the depolarization factor of the ellipsoid in the direction of the a axis (g_a) the value 0.144 for quartz sand was adopted. Part of the deviation between measured and calculated λ values may be explained by a possible wrong choice of this depolarization factor. For example, calculations for clay using $g_a = 0.125$ resulted in λ curves more similar to the measured curves of the sandy loam. Therefore it may be necessary to derive from diffusion experiments actual g_a values for a soil, as was done by De Vries. There is little doubt, however, that the theory describes the shape and orientation of the particles of these sedimented soils in a sufficient way.

Because clay soil is suspect to shrinking at desiccation and to swelling at wetting, dry bulk density changes during these processes. In relatively small soil columns neither calculations nor measurements in the laboratory will give reliable results unless corrections for changes in ρ_d are applied. The scatter in the laboratory λ data on clay was that large, that preference was given to measure λ in the soil in situ with the transient needle method.

Soil temperature also influences the value of λ. For dry soils an increase in temperature from 273–323 K (0°–50°C) gives an increase in λ of about 20%. For moist soils the dependency is much less and can be neglected for practical purposes (CHUDNOVSKII, 1948; IOFFE and REVUT, 1959).

There are three known methods to find the heat flux $q_h(z,t)$ in the soil. A quantity which is of importance in the determination of evaporation via energy balance and combination methods.

The first method is known as the temperature gradient method. If λ is known and measurements of soil temperature at several depths are available, the heat flux can be computed according to eq. (57) as the product of λ and the vertical temperature gradient ∇T. In wet soils too low values are found when additional transport of heat via the vapour phase occurs (e.g. WIERENGA, 1968).

A second method of determining $q_h(z,t)$ is the measurement by means of heat flux plates buried horizontally just below soil surface. Heat flux plates are thin plates usually of plastic material, having a λ which is approximately equal to that of soil. The temperature is measured at both sides of the plate, and the temperature gradient is obtained by dividing the temperature differences by the thickness of the plate. The heat flux is again obtained with eq. (57). Careful calibration of the heat flux plates in situ is necessary. Errors are due to disturbance of the soil and in wet soils to interference of moisture flow (e.g. FUCHS and TANNER, 1968).

The third method, known as the integral method, requires knowledge of temperature as well as volumetric heat capacities with depth. From the change in heat content during a certain time interval a mean value for q_h is obtained (e.g. SLATYER and MC.ILROY, 1961).

The above mentioned methods are more deeply discussed by for example CARSON (1961), VAN WIJK and SCHNEIDER (1967).

c. THERMAL DIFFUSIVITY

The rate of temperature equalization in the soil is determined by the thermal diffusivity $a = \lambda/\rho c$. The dimension of a is $m^2.s^{-1}$ which is the same dimension as that of a diffusion coefficient. A large diffusivity causes rapid changes in temperature and a quick and deep penetration of the heat wave into the soil. An example of the thermal diffusivities of the investigated sandy loam in relation to porosity and soil moisture content is shown in fig. 35. It is seen that maximum values of a occur roughly at moisture fractions x_w of 0.25 to 0.30 for $x_s = 0.50$, and at x_w of 0.30 to 0.40 for $x_s = 0.46$.

Besides determinations of λ, with the cylindrical probe method a direct measurement of a is also possible by measuring the temperature of the soil at the probe-soil interface or more radially distant from the probe. Because of the contact resistance between probe and medium, a high degree of accuracy cannot be expected (DE VRIES and PECK, 1958a) especially when measuring in soils (DE VRIES and PECK, 1958b).

Determination of a in the field from the analysis of temperature measurements at various depths is practised frequently, by assuming constancy of the thermal parameters with depth. Due to the periodic heating of the surface of the

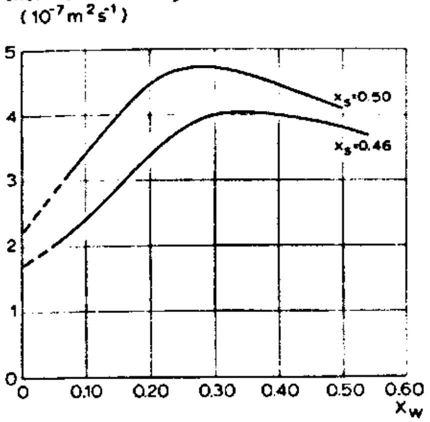


FIG. 35. *Thermal diffusivity of sandy loam versus volume fraction of water x_w for two volume fractions of solid material x_s, as derived from figs. 33 and 34 ($a = \lambda/\rho c$).*

soil by incident radiation, the surface temperature during days or periods of days without cloud cover can be approximated by a sinusoidal periodic function of time. Assuming that period and amplitude of the heat flux before and during the period considered is constant, the temperature at the surface can be expressed as:

$$T(0,t) = T_a + A_0 \sin \omega t \tag{62}$$

where:

A_0 = amplitude at the surface (K)
ω = angular frequency (rad.s^{-1})

If $T(z,0) = T_a$, then eq. (59) can according to CARSLAW and JAEGER (1959) be solved as:

$$T\left(z,t\right) = T_a + A_0 \exp\left(-\frac{z}{D}\right) \sin\left(\omega t - \frac{z}{D}\right) \tag{63}$$

where:

$D = \sqrt{\dfrac{2\lambda}{\rho c \omega}} = \sqrt{\dfrac{2a}{\omega}}$ is called the damping depth (m)

Comparison of eqs. (62) and (63) shows that at any depth z the amplitude $A_z = A_0 \exp(-z/D)$ and the phase shift $\varphi_z = -z/D$.

The values of D and therefore of a, may be found by plotting log A_z against depth, or phase shift (for example of the maxima) against depth. Usually different results are obtained from these two methods, indicating that the thermal properties vary with depth.

If the temperature at the soil surface is a more general function of time, the sine term can be replaced by a Fourier series (e.g. CHUDNOVSKII, 1948; CARSON, 1961, 1963; BROOKS et al., 1966; VAN WIJK, 1966) and the theory holds for the identical terms of these series. For non-periodicity of the temperature wave, (linear) corrections can be applied. A further discussion of eq. (63) is given in section IV5-b.

Van Wijk proposes another solution of eq. (59) by expressing the Laplace transform of the temperature as a function of depth. The present author applied this theory to temperature measurements in the field. The results showed that this method is very sensible for the condition of a homogeneous initial temperature of the soil. Deviations from this condition of 0.1 K had a large influence on the results of the calculations. HADAS (1968) compared these two methods when determining a under laboratory conditions. For a dry soil he found lower values with the sine wave solution than with the Laplace solution. For a wet soil he found higher values with the sine wave solution if the period of the applied heat wave lasted at least 16 min.

5. SOLUTION OF THE HEAT CONDUCTION EQUATION

a. GENERAL

The temperature field in the soil is determined by the variation of temperature at the soil surface and the variation of the thermal parameters with depth and time [eq. (59)].

Classical theories often assume a vertical uniform soil and independency of the thermal properties with depth and time. As was shown λ and ρc are functions of mineral composition, density and moisture content of the soil (figs. 34 and 35), however, and these usually vary considerably with time and depth, so λ and ρc are also functions of time and depth. LETTAU (1954) proposed a formula where the diffusivity is taken as the product of a depth function and a periodic function of time. Fortunately, the dependency of time can be neglected for the daily temperature variation, as was shown by IOFFE and REVUT (1959). The latter authors reported that for daily periods the variation in the thermal properties of soils did not exceed 10%. This implies that these properties may be regarded as constant when solving practical problems. This simplifies the mathematical problem of solving eq. (59) for depth dependent thermal properties with the boundary condition of harmonic variation of surface temperature with time.

b. ANALYTICAL APPROACH

For cases with a special functional relationship of λ and ρc with depth, analytical solutions exist. CHUDNOVSKII (1948), STERK (1956), DE VRIES (1956, 1957) and VAN WIJK (1966) treated the case for a semi-infinite medium (air), with ρc constant with height and λ increasing linearly with height. Eq. (59) then takes the form:

$$\nabla \left[a_0 \left(1 + mz\right) \nabla T \right] = \frac{\delta T}{\delta t} \tag{64}$$

and the solution is given in terms of Bessel and Hankel functions.

If λ and ρc are exponential functions of depth the following form has to be solved:

$$\nabla\{[\lambda_\infty - (\lambda_\infty - \lambda_0)\,e^{-\mu z}]\,\nabla T\} = \{(\rho c)_\infty - [(\rho c)_\infty - (\rho c)_0]\,e^{-mz}\}\frac{\partial T}{\partial t} \quad (65)$$

For exponential variation of λ and constancy of ρc with depth, one has to olve:

$$\nabla\{[a_\infty - (a_\infty - a_0)\,e^{-\mu z}]\,\nabla T\} = \frac{\partial T}{\partial t} \quad (66)$$

In case of a layered soil with each layer having different thermal properties, modifications of eqs. (64), (65) and (66) have to be applied to each layer separately with the condition that there is equality of temperature as well as heat flux at the boundaries of the various layers. An extensive treatment of eq. (64) and eq. (65) has been given by CHUDNOVSKII (1948).

Analytical solutions are not possible when λ and ρc are arbitrary functions of depth. Then, other solutions have to be applied (see section IV5-c and d).

In order to evaluate the influence of groundwater depth and type of profile on the thermal regime of the soil, the observational data taken on four individual, clear days were studied in more detail. Two of the selected days, April 17, 1968 and June 14, 1969 belonged to a rather dry period, the other two days, March 28, 1968 and May 14, 1969 to a rather wet and a very wet period respectively.

As a first approach the measured soil temperatures at the various depths were subjected to a Fourier harmonic analysis. Expressing the temperatures at each depth as a sum of cosine and sine components, and taking 12 temperatures at equidistant, fixed intervals of two hours, the following equation can be evolved (e.g. WHITTAKER and ROBINSON, 1958):

$$\begin{aligned} T(z,t) = {} & a_0 + a_1 \cos \omega t + a_2 \cos 2\omega t + \ldots + a_5 \cos 5\omega t + \\ & + a_6 \cos 6\omega t \\ & + b_1 \sin \omega t + b_2 \sin 2\omega t + \ldots + b_5 \sin 5\omega t \end{aligned} \quad (67)$$

where a_0 is the mean temperature (K), a_k and b_k the amplitudes (K) of the cosine and sine waves respectively and t time (s).

Using the property that in one period equal absolute values of cos $k\omega t$ and sin $k\omega t$ occur, the 12 unknown coefficients a_k and b_k can be determined by applying schemes of successive sums and differences of the 12 temperature values.

Instead of representing the temperature as a dual co-ordinate wave (eq. 67), a more convenient expression in the form of a single sine wave can be found. For that purpose the following auxiliary relationship, which is always valid, can be introduced:

$$a_k \cos k\omega t + b_k \sin k\omega t = (a_k^2 + b_k^2)^{1/2} \left[\frac{a_k}{(a_k^2 + b_k^2)^{1/2}} \cos k\omega t + \right.$$

$$\left. + \frac{b_k}{(a_k^2 + b_k^2)^{1/2}} \sin k\omega t \right] \tag{68}$$

where k is the wave number of the harmonics.

Denoting the phase angle of the single sine wave as φ_k (radians), it can be calculated by writing

$$\sin \varphi_k = \frac{a_k}{(a_k^2 + b_k^2)^{1/2}} \quad (68a), \qquad \cos \varphi_k = \frac{b_k}{(a_k^2 + b_k^2)^{1/2}} \quad (68b),$$

$$A_k = (a_k^2 + b_k^2)^{1/2} \quad (68c)$$

where A is the amplitude of the single sine wave. Substitution of eqs. (68a), (68b) and (68c) in eq. (68) yields the following expression:

$$T(z,t) = a_0 + \sum_{k=1}^{5} A_k \sin(k\omega t + \varphi_k) + a_6 \cos 6\,\omega t \tag{69}$$

Because on the mentioned selected days the variation of temperatures with time was not strictly periodic, an improvement in the approximation was made by considering a linear trend between the midnight temperature values. The interpolated temperature $T(z,t)$ is then given by

$$\frac{T(z,t) - T(z,0)}{T(z,2\pi) - T(z,0)} = \frac{t_t - t_0}{t_{2\pi} - t_0} \qquad \text{or} \tag{70}$$

$$T(z,t) = T(z,0) + [T(z,2\pi) - T(z,0)] \frac{t_t - t_0}{t_{2\pi} - t_0} \tag{70a}$$

The procedure taken was to subtract from the 12 measured temperature values the pertaining values of the right hand side of eq. (70a) and next to subject the reduced temperatures to the Fourier analysis (eq. 69). The final equation describing $T(z,t)$ over the period of observation is then formulated as:

$$T(z,t) = T(z,0) + [T(z,2\pi) - T(z,0)] \frac{\omega t}{2\pi} + a_0 +$$

$$+ \sum_{k=1}^{5} A_k \sin(k\omega t + \varphi_k) + a_6 \cos 6\omega t \tag{71}$$

and

$$T = T(z,0) + \frac{1}{2}\left[T(z,2\pi) - T(z,0)\right] + a_0 \tag{71a}$$

An example of the results of Fourier analyses of soil temperature data at a variety of discrete depths, is given in table 12.

Generally spoken, the accuracy of a Fourier analysis will be the higher, the more harmonics are taken. The accuracy of the temperature measurements was in the range of 0.1 to 0.2 K, which restricts the number of reliable harmonics. It can be seen from table 12 that for all depths but 0.01 m, A_4 and A_5 can be neglected. For the depths of 0.05, 0.10 and 0.60 m even A_3 can be neglected. For the clear days taken, the sum of the first three harmonics fitted the measured temperature data reasonably well. An example is given in fig. 43.

As pointed out earlier, the thermal diffusivity of a homogeneous soil can be derived via the constant decrease of the logarithm of amplitude with depth, which should equal the retardation of phase with depth. In fig. 36, A_1 is plotted on half-logarithmic paper and φ (radians) on normal graph paper, both against depth for the clay and sandy loam of the 0.45 m groundwater plot and for clay on the 1.65 m plot, for April 17, 1968. The graphs show that, except for the top 0.03 m, there is a distinct deviation from the above mentioned linear relationship. To a lesser extent this also holds for the curves concerning May 14, 1968, where it was expected that the wet circumstances would make the profiles more uniform in water content and therefore more uniform in thermal properties. The uniformity of the 0.03 m top layers will have been the result of the shallow cultivation before starting the sowing experiments. This uniformity made it possible to obtain the Fourier coefficients at the surface by extrapolation from the values at the 0.03, 0.02 and 0.01 m depths, allowing the unmeasured temperature waves at the surfaces to be reconstructed.

Because the procedure to derive the thermal diffusivity of soils from the amplitude and phase relationships with depth failed, two other lines of attack were followed to try to establish functional relationships of thermal diffusivities with depth. Both methods are based on the principle that a discontinuous field can approach the continuous thermal field by expanding the partial differential equation (59) into a set of finite difference equations. This expansion can be realized by means of either an electric analog approach, or a mathematical, i.c. numerical approach.

TABLE 12. Results of Fourier analyses based on two-hourly soil temperature observations at depths z, sandy loam, 0.45 m groundwater plot, April 17, 1968

z	$\bar{T}$	$d\bar{T}/dt$ (mean trend)	A_1	A_2	A_3	A_4	A_5	φ_1	φ_2	φ_3	φ_4	φ_5
m	°C	°C/2π rad	°C	°C	°C	°C	°C	rad	rad	rad	rad	rad
0.01	14.27	0.3	6.74	2.50	0.53	0.36	0.45	3.984	1.104	5.184	4.304	3.292
0.02	13.74	0.8	5.84	2.05	0.45	0.22	0.36	3.880	0.867	4.975	4.494	3.090
0.03	13.32	1.2	5.09	1.70	0.35	0.14	0.31	3.790	0.686	4.854	4.610	2.901
0.05	12.56	1.7	3.91	1.30	0.24	0.14	0.14	3.625	0.402	4.609	−1.170	2.917
0.10	11.23	2.0	2.28	0.62	0.12	0.08	0.05	3.222	6.093	3.729	−1.456	2.031
0.60	6.86	0.8	0.09	0.02	0.02	0.01	0.01	0.613	3.665	6.283	4.189	3.512

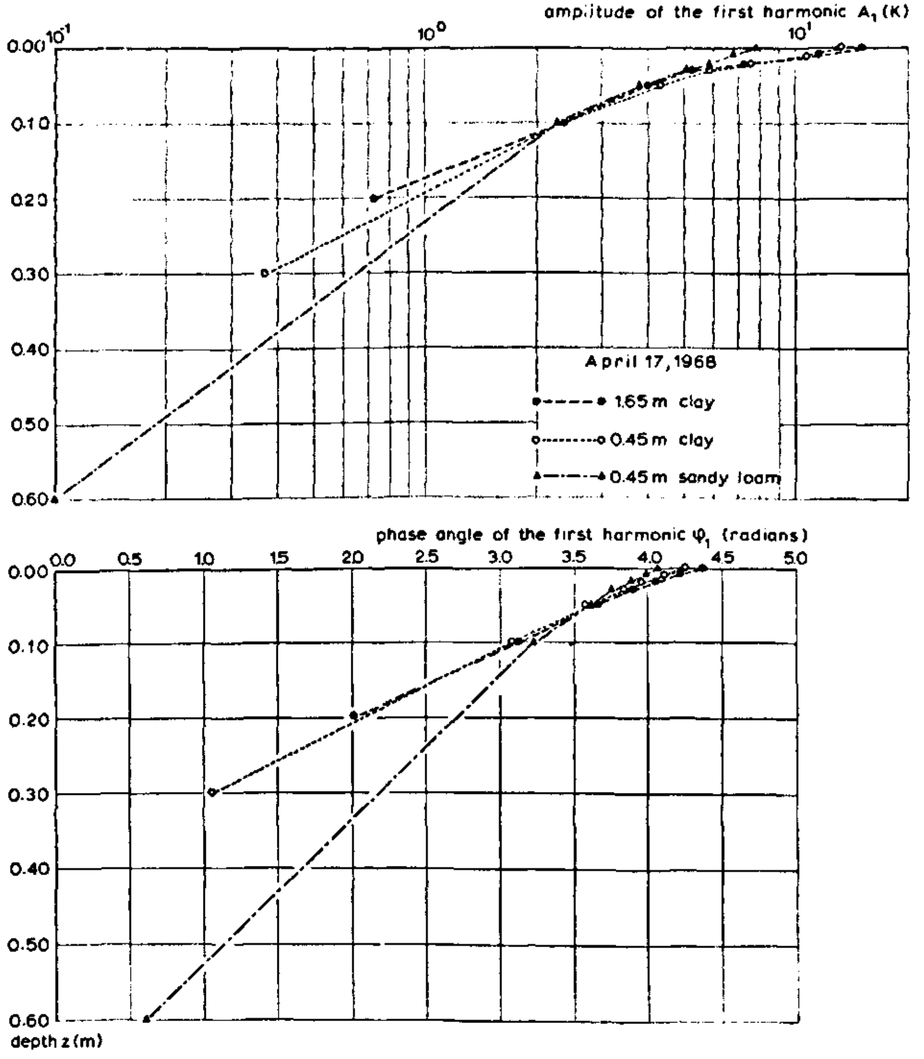


FIG. 36. *Example of the variation of the amplitude and phase angle of the first harmonic of the daily soil temperature wave as obtained by Fourier analysis with depth for the clay and sandy loam of the 0.45 m groundwater plot and the clay of the 1.65 m plot, April 17, 1968.*

c. ELECTRIC ANALOG APPROACH

The electric analog approach is based upon the analogy between the flow of heat in a thermal field and that of electricity in an electrical field, which can be illustrated by comparing the partial differential equations governing the two flow systems. The one dimensional equation governing electrical flow in an electrical resistance – capacitance circuit can be derived from fig. 37.

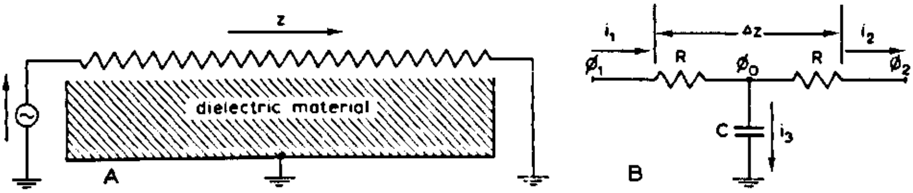


FIG. 37. *A, one-dimensional electrical resistance – capacitance system; B, differential element (after KARPLUS, 1958).*

According to Ohm's law, the current i (A) is proportional to the voltage gradient $\nabla\phi$ (V.m^{-1}), and the reciprocal value of the resistivity R (ohm.m^{-1})

$$i = -\frac{1}{R}\nabla\phi \tag{72}$$

The principle of continuity requires that the difference of incoming and outcoming current through the differential element equals the rate of storage:

$$i_1 - i_2 = i_3 = -\nabla i\,\Delta z = C\,\Delta z\frac{\partial\phi}{\partial t} \tag{73}$$

where C is the capacitivity (farad.m^{-1}). Substitution of eq. (72) into eq. (73) gives

$$\nabla\left(\frac{1}{R}\nabla\phi\right) = C\frac{\partial\phi}{\partial t} \tag{74}$$

Comparison of eqs. (72), (73) and (74) with eqs. (56), (57) and (58) shows that ϕ plays a role analogous to T, i to q_h, $1/R$ to λ and C to ρc.

Considering the above mentioned equations, the electric analog equation for the system shown in fig. 37, becomes

$$\frac{\phi_1 - \phi_0}{R} - \frac{\phi_0 - \phi_2}{R} = C\frac{\mathrm{d}\phi_0}{\mathrm{d}t} \tag{73a}$$

where it is seen that the left side of eq. (74) is discretized, while the time variable is kept in continuous form. Hence, the temperature field can be simulated by means of a rectangular one-dimensional network of lumped resistance – capacitance elements, where each parameter and variable at a junction in the electric analog, corresponds with the parameter and variable at a specific location in the thermal field. This type of network is known as the model of BEUKEN (1937). LIEBMANN (1956) developed another model containing only resistors. In his model both place and time derivatives are discretized.

Conversion constants (scale factors) relate the corresponding parameters and variables in the two systems. The corresponding waves were displayed on an oscilloscope as voltage versus time plots. The technical realization of the Beuken model as used in this study was carried out by BOELS (1969).

The determination procedure of apparent thermal conductivities, can be shortly summarized as follows:

- The volumetric heat capacities at different depths, obtained by means of volumetric sampling and successive computation with eq. (61), are inserted as electrical capacitors in the model.
- The voltage and current excitations corresponding with the first harmonic of the temperature wave at 0.01 m, are applied to the network entrance.
- The amplitudes and phase angles at discrete depths are known from the Fourier analysis of the measured temperatures, and consequently the rates of decrease in amplitude and increase in phase lag with respect to the temperature wave at 0.01 m depth.
- Then the resistors are adjusted by trial and error until at the points corresponding with each depth, the rates of decrease in amplitude and increase in phase lag of the first harmonic correspond with the values obtained by the Fourier analysis. The same procedure is carried out for the second harmonic.
- Finally, the found resistances are converted into apparent thermal conductivities.

A few examples of the results obtained with this electric analog method for April 17, 1968 are represented in fig. 38. The scatter of the data can be explained to some extent.

The mean volumetric heat capacities obtained may differ from actual values because of heterogeneity of the soils and a probable underestimation inherent to the computation method. WIERENGA (1968) found for a silt loam computed ρc values which were on the average 13% lower than the experimentally determined heat capacities. Instrumental errors may also have had some influence. The recording and reading errors from the temperature registrating device were in the range of 0.1 to 0.2°C. The accuracy of an oscilloscope is limited to approximately 3% of the full scale value (KARPLUS, 1958). Another important reason introducing uncertainties in the electric analog method are the relatively widely spaced temperature observations at the larger depths. The consequences of this will be discussed in more detail in the next section.

It turned out that the apparent λ values obtained from the adjustment of the amplitude of each first harmonic, were significantly lower than those obtained from adjustment of the phase shift of it (T-test; $n = 78$, $T = 46$; if $T \geqslant 24$ the differences are significant at 1 per cent level). The differences were on the average in the range of 0.29 to 0.79 $W.m^{-1}.K^{-1}$ (Student's t-distribution; $n = 66$, significancy at 1 per cent level). For the second harmonic no significant differences were found.

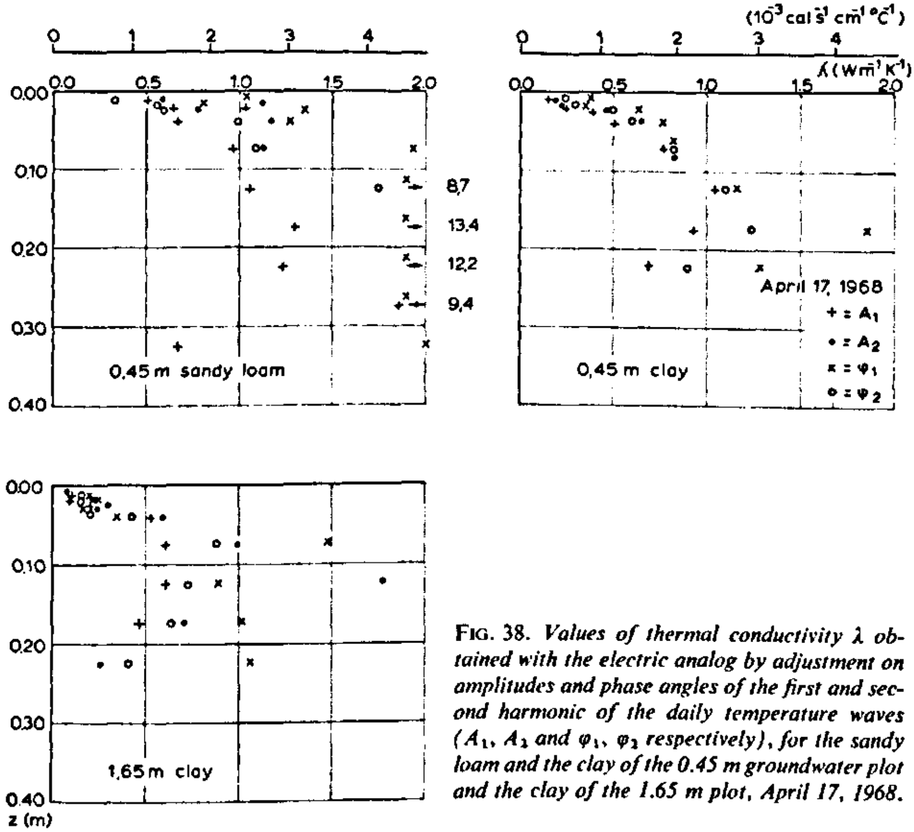


FIG. 38. *Values of thermal conductivity λ obtained with the electric analog by adjustment on amplitudes and phase angles of the first and second harmonic of the daily temperature waves (A_1, A_2 and φ_1, φ_2 respectively), for the sandy loam and the clay of the 0.45 m groundwater plot and the clay of the 1.65 m plot, April 17, 1968.*

After due consideration it has been assumed that the procedure of averaging the λ values obtained from the amplitude and phase relationships of the first harmonics will be acceptable and will at least give an indication about the order of magnitude.

It is interesting to note that the same phenomenon of obtaining different soil physical constants from amplitude and from phase relationships, also occurs in a quite different field of application, namely the determination of hydrological constants from the transmission of tidal waves in groundwater (STEGGEWENTZ, 1933). WESSELING (1960) considered the propagation of tides in aquifers and found that the hydrological properties calculated for a non-compressible aquifer were significantly lower than when the elasticity of the aquifer was taken into account. The similarity of these results and those obtained with the analog method is that the elasticity of soil and water in the aquifer works analogous but in an opposite way as a corresponding additional negative heat generation occurring in the soil during the transmission of heat. As already mentioned, this latter process may be caused by a movement of water in the vapour phase under action of thermal gradients, tending to yield higher thermal properties then when not considering this heat loss. The importance of such a process will be discussed in more detail in section IV5-e.

d. NUMERICAL APPROACH

The numerical approach discretizes the partial differential equation (59) in the following way. Let a co-ordinate net or grid be drawn over a scale diagram of the temperature field under consideration, as shown in fig. 39. In this net the index along the space ordinate is denoted i, the index along the time abscissa is denoted j. Considering the node point $T_{i,j}$, frequently used approximations for the first derivative with respect to time in terms of the adjacent node temperatures are:

$$\left(\frac{\partial T}{\partial t}\right)_{i,j;i,j+1} \cong \frac{T_{i,j+1} - T_{i,j}}{\Delta t_j} \tag{75a}$$

$$\left(\frac{\partial T}{\partial t}\right)_{i,j-1;i,j} \cong \frac{T_{i,j} - T_{i,j-1}}{\Delta t_j} \tag{75b}$$

where Δt_j is the uniform time interval. The forward time difference eq. (75a) and backward time difference eq. (75b) approximations can be considered to represent the temperature change midway between the node points indicated by the indices, with the assumption that the change is constant between these points.

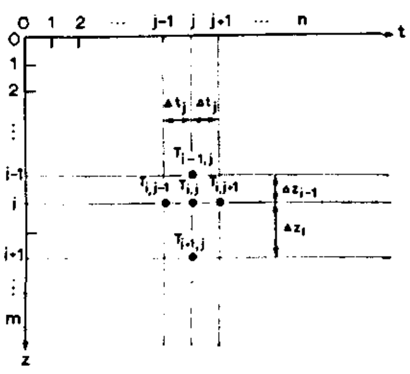


FIG. 39. *Grid of rectangular mesh with time arms of equal and space arms of unequal lengths.*

The approximation for the second derivative with respect to space, being the rate of change of the first derivative to space, is obtained by subtracting the backward space difference from the forward space difference and dividing by the net spacing:

$$\left(\frac{\partial^2 T}{\partial z^2}\right)_{i,j} \cong \frac{1}{\Delta z_i}\left(\frac{T_{i-1,j} - T_{i,j}}{\Delta z_i} - \frac{T_{i,j} - T_{i+1,j}}{\Delta z_i}\right) \tag{76}$$

where Δz_i is the uniform space interval and assuming that $\partial^2 T/\partial z^2$ remains constant between the net points $(i-1,j)$ and $(i+1,j)$. The one-dimensional difference equation is then obtained by combination of either eq. (75a) with eq. (76), SCHMIDT (1942), or by combination of eq. (75b) with eq. (76), LIEBMANN (1955). The first combination is subject to the stability criterion $a(\Delta t)/(\Delta z)^2 \leqslant 0.5$. WIERENGA (1968) applied this method to compute the temperature field from known boundary conditions and known (constant) apparent thermal diffusivity, as well as estimated the latter from observed temperature variations at various depths. Except for depths smaller than 0.10 m, he received fairly good results in predicting soil temperatures in a soil with approximately uniform water content. CRANK and NICHOLSON (1947) give a more stable approximation, but their method involves more terms and requires an iterative procedure. The combination method used by Liebmann, has a very stable nature permitting the choice of relatively large time intervals. The disadvantage of this approach is, however, that the solution requires a relaxation technique, using at each point the relation between four unknown and one known temperature value.

Another approximation can be used by assuming that the temperature distribution over two line segments of time is linear, so the first derivative with respect to time may be represented by the mean of eqs. (75a) and (75b). Combination of this with eq. (76) then gives:

$$\frac{T_{i,j+1} - T_{i,j-1}}{2\Delta t_j} = a\,\frac{\dfrac{T_{i+1,j} - T_{i,j}}{\Delta z_i} - \dfrac{T_{i,j} - T_{i-1,j}}{\Delta z_{i-1}}}{\dfrac{\Delta z_i + \Delta z_{i-1}}{2}} \tag{77}$$

where the thermal diffusivity (a) can be solved from a system of linear equations.

Calculations have been carried out using this method. As a first approach for Δt_j an interval of 2 hours was taken. Later on intervals of 1 hour were used. The space interval chosen was not uniform, but taken according to the depths of the temperature measurements.

It appeared from the output of this numerical method that sometimes unrealistic a values, as for example zero, were obtained. This may have resulted from the already mentioned recording and reading errors of the temperature data. To correct for this type of errors the temperature data read from the slips of the recording device were replaced by reconstructed, smoothed temperatures computed from the Fourier analysis by taking the sum of the first three harmonics.

Even when the temperatures are exactly correct, another error may be introduced, i.c. the error inherent in discretizing the continuous temperature field to a discontinuous field. To get more accurate interpolations between two known points, approximations by polynomials of a power determined by the net points available, can be used. Interpolation of temperature values with depth by means of such a polynominal resulted in temperatures near the ground-water table approaching almost the boiling point of water. Moreover the use of a polynomial through a great number of net points is a less attractive method.

A more satisfactory interpolation technique can be found in applicating spline functions. The term spline originates from long thin strips of wood (splines) used by draftsmen to fair in a smoothed curve between specified points. The mathematical spline approximates the draftsman's spline by a piecewise different cubic polynomial between each pair of adjacent points. The spline is continuous and has both a continuous first and second derivative (e.g. AHLBERG et al., 1967; REINSCH, 1967).

If a function y is known at $x_0 < x_1 < x_2 < x_3$, a polynomial of third degree

$$y = a_0 + b_0 x + c_0 x^2 + d_0 x^3 \tag{78}$$

can be drawn through the points y_0 and y_1.

A second cubic curve can be drawn through y_1 and y_2. Then the equation for their common point y_1 is:

$$a_0 + b_0 x_1 + c_0 x_1^2 + d_0 x_1^3 = a_1 + b_1 x_1 + c_1 x_1^2 + d_1 x_1^3 \tag{79}$$

Applying the condition that in point y_1 both the first and second derivative must be equal, the relationships derived by differentiation of eq. (79) are:

$$b_0 + 2c_0 x_1 + 3d_0 x_1^2 = b_1 + 2c_1 x_1 + 3d_1 x_1^2 \tag{80}$$

$$2c_0 x_1 + 6d_0 x_1 = 2c_1 + 6d_1 x_1 \tag{81}$$

From this set of equations the coefficients and the first and second derivatives can be determined. Thus, from the interpolation of temperature data by spline functions in i and j direction, the related derivatives can be obtained.

Computations with the finite difference method (eq. 77) and with spline functions (eqs. 78–81) were carried out on a CDC 3200 digital computer with FORTRAN IV programs used for the difference scheme and the spline functions (the latter according to REINSCH, 1967).

From the large amount of data obtained, one example was taken (table 13), which shows the results for the clay profile of the 1.65 m groundwater plot, during part of the day of March 28, 1968. It appears from this table that because of large scatter in the computed a values the numerical approaches give rather poor results. For lower depths a similar scatter in a values was observed.

TABLE 13. Example of the computation results of the apparent thermal diffusivity a (10^{-7} $m^2.s^{-1}$) by numerical analyses with the finite difference method (diff.) and with spline functions (spline), clay, 1.65 m groundwater plot, part of the day of March 28, 1968, with t_j 2 hours respectively 1 hour, with measured temperatures (T_m) and with temperatures reconstructed from Fourier analysis (T_r). For the determination of a at z = 0.02 m, use was made of temperatures measured at 0.01, 0.02 and 0.03 m depth

z m	Method	Time hours 6	8	10	12	14	16	18
0.02	diff. with $\Delta t_j = 2$ and T_m	0.12	1.58	0.71	0.23	0.06	0.08	1.97
	diff. with $\Delta t_j = 1$ and T_m	0.05	1.84	0.75	0.24	0.04	0.24	2.31
	spline with $\Delta t_j = 1$ and T_m	0.08	1.39	0.56	0.17	0.02	0.15	1.37
	diff. with $\Delta t_j = 1$ and T_r	0.09	5.59	0.64	0.25	0.04	0.32	2.38
	spline with $\Delta t_j = 1$ and T_r	0.03	0.42	0.97	0.20	0.02	0.17	11.97

A type of error that has to be taken into account is the error in reading the temperature from the chart of the recorder. The order of magnitude of this error can be shown by the following example. Considering the soil temperatures at the 0.01 (i–1), 0.02 (i) and 0.03 (i + 1) m depths at 15 (j – 1), 16 (j) and 17 (j + 1) hours for 1.65 m clay, March 28, 1968, where $T_{i,j-1}$ = 292.2 K, $T_{i,j}$ = 292.1 K, $T_{i,j+1}$ = 290.3 K, $T_{i+1,j}$ = 290.4 K and $T_{i-1,j}$ = 294.9 K. From computation with eq. (77) it follows that a is 0.24×10^{-7} $m^2.s^{-1}$ (see table 13).

Assuming a reading error of 0.2 K and taking values of 292.3 K and 290.2 K for $T_{i,j}$ and $T_{i+1,j}$ respectively, an a value of 0.53×10^{-7} $m^2.s^{-1}$ is computed, showing an error of estimate of more than 100%. This type of error is the most dangerous at lower depths as there changes in temperature with time and depth are small.

Another error that may have been introduced is the error resulting from small deviations in Δz. Suppose that Δz_{i-1} is 9 mm instead of 10 mm (then Δz_i = 11 mm), an a value of 0.17×10^{-7} $m^2.s^{-1}$ is computed, yielding a difference with the 'real' a value (0.24×10^{-7}) of about 30%. Assuming Δz_{i-1} is 8 mm instead of 10 mm, an a value of 0.13×10^{-7} $m^2.s^{-1}$ is found, the error being about 46%.

An important reason for the scatter at the lower depths could be the widely spaced intervals of the temperature observations, which cause considerable errors in the second space derivatives. For example at the 0.20 m depth, $\partial^2 T/\partial z^2$ values were computed from the 0.10 and 1.20 m depths, yielding a values which were a factor 10 too high. Moving the temperatures from the 1.20 m depth to the 0.60 m depth (which is not unrealistic as between these depths there is little difference in temperature) more acceptable a values were obtained, indicating that Δz_i had been chosen too large. Application of spline functions in z direction should improve the second space derivatives. However, it turned out that due to the large z_i's, physically impossible curves were drawn through the points. Therefore the temperature observations should be spaced more closely to obtain realistic values. For the relatively close spaced observations in the top layers, interpolation by spline functions should be preferred above the finite difference method, because it gives better approximations of the related derivatives.

The instrumental errors as also the reading and certain round off errors are partly neutralized by the use of spline functions as well as by the use of the reconstructed temperatures by means of the Fourier analysis.

Another, already mentioned, reason for scatter of apparent a values for the dry top layer, is a possible disturbing influence of water transport in the vapour phase as induced by temperature gradients. This is accompanied by a transport of, principally latent, heat which influences the temperature distribution and will result in higher a values than could be expected from thermal conduction alone. If this vapour transport plays an important role, a daily pattern should occur in the a values of the top layer, as a result of the daily variation in temperature and soil moisture content. From the data such a pattern can hardly be derived. In the next section it will be illustrated that, at least at the temperature gradients measured in the present research, the influence of thermally induced vapour transport can be neglected as a possible explanation of the scatter in the data.

e. Influence of thermally induced vapour flow

The flow of water through the soil occurs in the liquid phase, in the vapour phase or in both.

The general equation describing *liquid* flow in the soil as induced by gradients of water head, gravitational head and temperature can be written as (Philip and de Vries, 1957):

$$q_l/\rho_l = -D_{\theta l}\nabla\theta_l - D_{Tl}\nabla T - k \qquad (\mathrm{m.s^{-1}}) \qquad (82)$$

where l refers to the liquid phase, $D_{\theta l}$ is the isothermal liquid diffusivity ($m^2.s^{-1}$) and D_{Tl} is the thermal liquid diffusivity ($m^2.s^{-1}.K^{-1}$). ROSE (1967a, b) showed from field experiments that liquid flow induced by thermal gradients even under conditions of temperature gradients of 10 °C.cm^{-1} was relatively unimportant. Hence eq. (82) reduces for the soils under investigation to

$$q = -k\,(\nabla h_m + 1) \qquad (m.s^{-1}) \tag{8b}$$

The general molecular diffusion equation describing the flow of water *vapour* in air free of convection, can be written as:

$$q_v = -D_{H_2O}\nabla C \qquad (kg.m^{-2}.s^{-1}) \tag{83}$$

where D_{H_2O} is the molecular diffusion coefficient of water vapour in air ($m^2.s^{-1}$) and C is the water vapour concentration ($kg.m^{-3}$). Under equal conditions of vapour gradient and temperature gradient, the water vapour flow in the soil is less than in air because of the available cross-sectional area and increased path length, hence instead of D_{H_2O} a reduced diffusion coefflcient ($\bar{D}_{H_2O}$) has to be used. CURRIE (1960) found from diffusion experiments with dry porous granular materials that the empirical equation $\bar{D}_{H_2O} = a\varepsilon_g^b\, D_{H_2O}$ (where ε_g is the air filled porosity in m^3 air.m^{-3} soil) fitted all the tested materials well. Taking for the dry top layers of the soils of the experimental field the average total porosity 0.67, the equation of Currie for Woburn and Highfield soil crumbs with $a = 0.83$ and $b = 1.7$ can be used. Hence isothermal diffusion of water vapour in the top layers of the investigated soils can be described as:

$$q_v = -0.83\,\varepsilon_g^{1.7}\, D_{H_2O}\,(0.622\,\rho_a/p_a)\,\nabla e \tag{84}$$

Substitution of the appropriate values (see the list of used symbols) for D_{H_2O}, ρ_a and p_a reduces this expression to:

$$q_v = -16.0 \times 10^{-6}\,\varepsilon_g^{1.7}\,\nabla e \qquad (kg.m^{-2}.s^{-1}) \tag{84}$$

with vapour pressure e in bar. However, many investigators found from experiments that the measured vapour fluxes generally exceed fluxes calculated with a molecular diffusion type of equation such as eq. (84), with sometimes up to a factor 10. This means that flow mechanisms other than molecular diffusion also play a role.

PHILIP and DE VRIES (1957) have therefore extended the molecular diffusion theory to account for these differences. The reader is also referred to the elucidating papers of ROSE (1966, 1967a, 1967b). Philip and De Vries separated the total vapour flow in isothermally and thermally induced vapour flow by considering the effect of relative humidity ($h_r = e/e^*$) as follows. For ∇e in eq. (84) one may write:

$$\nabla e = \nabla h_r e^* = e^* \nabla h_r + h_r \nabla e^* \qquad (\text{bar.m}^{-1}) \tag{85}$$

where h_r is a function of moisture content (θ_l) only and the saturated vapour pressure e^* is a function of temperature (T) only. Hence eq. (85) may be written as:

$$\nabla e = e^* (\partial h_r/\partial \theta_l) \nabla \theta_l + h_r (\partial e^*/\partial T) \nabla T \tag{85a}$$

Substitution of eq. (85a) into eq. (84) gives the vapour flow equation

$$\begin{aligned} q_v = &\underbrace{- 16.0 \times 10^{-6} \varepsilon_g^{1.7} e^*(\partial h_r/\partial \theta_l) \nabla \theta_l}_{\text{isothermally induced flux}} + \\ &\underbrace{- 16.0 \times 10^{-6} \varepsilon_g^{1.7} h_r(\partial e^*/\partial T) \nabla T}_{\text{thermally induced flux}} \qquad (\text{kg.m}^{-2}.\text{s}^{-1}) \end{aligned} \tag{86}$$

Now Philip and De Vries consider two factors which may explain the differences in vapour fluxes found from measurements and those calculated with the molecular diffusion theory.

The first one is that the macroscopic temperature gradient in the medium (∇T) is generally exceeded by the microscopic temperature gradient across the air-filled pores $(\nabla T)_a$ (see also ROSE, 1966). Following the same symbols as used by ROSE (1967a):

$$(\nabla T)_a = \zeta \nabla T \tag{87}$$

The second reason is that vapour diffusion in fairly dry soils is not restricted to the gas-filled pores only (ε_g), but is aided by liquid islands which cause condensing on the upstream side and re-evaporation on the downstream side (see ROSE, 1966) and therefore vapour diffusion is dependent on the total pore space ($\varepsilon_g + \theta_l$). This holds not only for the isothermally induced flux, but also for the thermally induced flux (ROSE, 1967a). For this enhancement of vapour flux the expression holds:

$$\varepsilon_g + \theta_l = \xi(a\varepsilon_g^{\,b}) \tag{88}$$

The combined effect of the two factors then gives the expression with which the molecular diffusion flux has to be multiplied to yield calculation results which agree with measurements:

$$\xi \, \zeta = \frac{\varepsilon_g + \theta_l}{a\varepsilon_g^{\,b}} \times \frac{(\nabla T)_a}{\nabla T} \tag{89}$$

The enhancement factor ξ was calculated by substitution of the relevant values in eq. (88), while the enhancement factor ζ was derived from a combination of own data and those of table 2 of PHILIP and DE VRIES (1957). Taking all the given considerations into account, the total vapour flux can be found from eq. (86) and eq. (89), yielding

$$q_v = -\xi\,(isoth.\ ind.\, flux) - \xi\,\zeta\,(therm.\ ind.\, flux) \tag{90}$$

Using the theory of irreversible thermodynamics, CARY (1966) showed that when the vapour pressure gradient in soil is determined by temperature only and not by changes in moisture content, the thermally induced vapour flux under steady state conditions can be described as:

$$q_v = -\beta[1.56 \times 10^{-5}(T-273)^2 + 2.72 \times 10^{-3}](\mathrm{d}T/\mathrm{d}z) \quad (\mathrm{mm.hr^{-1}}) \tag{91}$$

where β is a dimensionless factor accounting for pore geometry and T is the temperature of the soil.

Another possible intensification of the vapour flux as calculated with molecular diffusion theories, is due to the influence of air turbulence (FARRELL et al., 1965; SCOTTER and RAATS, 1969). These authors show that wind induced atmospheric pressures at the soil surface may induce mass flow of vapour transport into and out of the upper layers of the soil. This enhancement due to air turbulence can be in the order of a factor 2 to 4 for soils with particle sizes of 4 to 6 mm (SCOTTER and RAATS, 1969), to even a factor 100 for coarse mulches (FARRELL et al., 1965).

Because thermally induced vapour flow becomes most important under drier soil conditions, the data obtained from the 0.45 m clay plot for June 14, 1969 (as presented in table 8) were used to calculate a numerical example which may give an insight into the relative importance of the various factors involved. For the top layer of the 0.45 m plot the following conditions did apply: $\varepsilon_a = 0.628$; $\theta_l = 0.031$; $\overline{T}_{0.00} = 299.6$ K, $\overline{T}_{0.01} = 297.7$ K, $\overline{T}_{0.02} = 297.0$ K, $\overline{T}_{0.03} = 296.3$ K; $\bar{e}_0 = 22.9 \times 10^{-3}$ bar.

The enhancement factor ξ can be found from eq. (88) by

$$\xi = (\varepsilon_a + \theta_l)/(a\varepsilon_a^b) = 0.659/[0.83 \times (0.628)^{1.7}] = 1.75.$$

From table 2 of PHILIP and DE VRIES (1957) it follows that for the prevailing porosity and moisture content the enhancement factor ζ is about 1.37. For a proper evaluation of the results of the calculations with eq. (90), one needs to be informed about the depth below surface where evaporation takes place, i.e. the depth where $h_r \approx 1$. In general, the evaporation front will not move below a few centimeters into the soil (see GARDNER and HANKS, 1966; ROSE, 1967a). As the depth of the evaporation zone was not exactly known, calculations were performed assuming different depths of evaporation zone under the same existing temperature regime. The results of these calculations are presented in table 14.

TABLE 14. Calculation by eq. (90) of the order of magnitude of the isothermal vapour flux and the thermally induced vapour flux, assuming different depths of the evaporation zone, for the 0.45 m clay plot, June 14, 1969.

Depth evaporation zone mm	Isothermally induced vapour flux $mm.day^{-1}$	Thermally induced vapour flux $mm.day^{-1}$	Net vapour flux $mm.day^{-1}$
2.5	4.99	−0.98	4.01
5.0	2.47	−0.67	1.80
7.5	1.63	−0.52	1.11
10.0	1.22	−0.44	0.78
12.5	0.97	−0.38	0.59
15.0	0.80	−0.35	0.45
20.0	0.68	−0.32	0.36

In the calculations z was taken positive upwards. From eq. (86) it follows that the direction of the isothermally induced vapour flux, as determined by ∇h_r, is positive upwards for the 24-hour period taken. During such a period the direction of the thermally induced vapour flow will generally be upward at night and mainly downward during the day in accordance with the temperature profiles.

Because on June 14, 1969, there was an average thermal gradient from the surface downwards, the direction of the thermally induced vapour flux for the 24-hour period, as determined by ∇e^*, is also downwards. From the data in table 14 it can be seen that thermally induced vapour flow becomes a more important component of net vapour flux when the evaporation zone is located deeper. For example, if this zone is located at 2.5 mm depth, thermally induced flow is about 20% of isothermal flow. With the zone at 20 mm depth this percentage is about 50.

Thermally induced vapour flow can also be calculated from eq. (91). The magnitude of the empirical constant β should, however, be determined experimentally (CARY, 1965, 1966). For different moisture conditions and porosities the latter author found for vapour flow in moist soils β values ranging from 1.7 to 5. The thermally induced vapour fluxes calculated with a β value equal to the product $\xi\, \zeta = 2.4$ [see last term of eq. (90) and eq. (91)] amounted to values about three times those as calculated with eq. (90). Moreover CARY (1966) states that for Columbia loam soil, eq. (91) was only applicable when the moisture content of this soil was above the 10 to 12% level. Values of thermally induced vapour flux comparable with those found with the theory of Philip and De Vries could only be obtained with the theory of Cary by taking β equal to 1. This is in full agreement with the findings from HANKS et al. (1967) in laboratory studies on soils with relative low moisture contents.

By means of sampling the surface layer (see section IIIA-2), as well as by sampling the deeper layers with Kopecky cores, the variation in matric head with depth could be found. From interpolation between the data obtained it appeared that the depth where relative humidity (h_r) reached about unity ($h_r = 0.9885$ at 15.6 bar or pF 4.2) was about 10 mm below surface. At this depth the calculated isothermally induced and thermally induced vapour flux were 1.22 and -0.44 mm.day^{-1} respectively, so at this depth thermally induced flow was about 36% of isothermal flow. For less extremely dry conditions, the contribution of thermal flow will be much less.

The calculations were carried out assuming steady state conditions and a linear relationship over the dry top layer of h_r and e^* with depth. It is more probable, however, that the relationship is for example parabolic or exponential. Moreover, evaporation does not take place from a plane but over a depth range (e.g. FUCHS and TANNER, 1967). The performed calculations will at least give a reasonable impression of the order of magnitude, however, and it is seen that the calculated net vapour flux by means of molecular diffusion theories (0.78 mm.day^{-1}; see table 14) is about one third of the evaporation flux (2.40 mm.day^{-1}; see table 8) deduced from energy balance considerations. The reason for the larger evaporation flux found is probably due to wind-induced pressure fluctuations at the soil surface, transferring water vapour into and out of the surface layer (FARRELL et al., 1965; CARY, 1966; SCOTTER and RAATS, 1969). This explanation seems to be acceptable at least for the crumbled upper clay top layer and at the wind velocities occurring at the experimental field (about 3–4 m.s^{-1} during daylight hours).

Under the prevailing conditions the temperature gradients for the lower soil layers are generally less steep than for the upper soil layers. This means that ∇e^* for the deeper layers is relatively low, hence thermally induced vapour flow in these regions will be of less importance.

Reviewing all the above mentioned considerations, it can be concluded that for very dry situations in the top layer of a soil, the values of thermally induced vapour flow may amount to about 35% of those of isothermal vapour flow. In most other cases it will be far less. Moreover, the effect of thermally induced downward vapour flow will be counteracted by matric head induced upward liquid flow. The latter influence will be the higher, the more vapour condenses in the lower layers. HANKS et al. (1967) found that thermally induced downward vapour flow for a top soil layer over a period of 40 days amounted to about 10% of the net upward flow.

As thermally induced liquid flow may also be considered as being relatively unimportant, the conclusion can be reached that thermally induced vapour transport can hardly be considered as one of the main reasons for the scatter in the thermal diffusivity data as found from the numerical analysis.

6. TEMPERATURE REGIME OF SOIL

The most direct and probably best method to determine a, is to measure ρc by means of a calorimetric method, to measure λ with the transient needle method and finally to estimate a at each depth by considering the ratio of λ and ρc. As pointed out earlier the ρc values were obtained from duplicate samples and computation with eq. (61), while the λ values were measured in situ by means of the needle method.

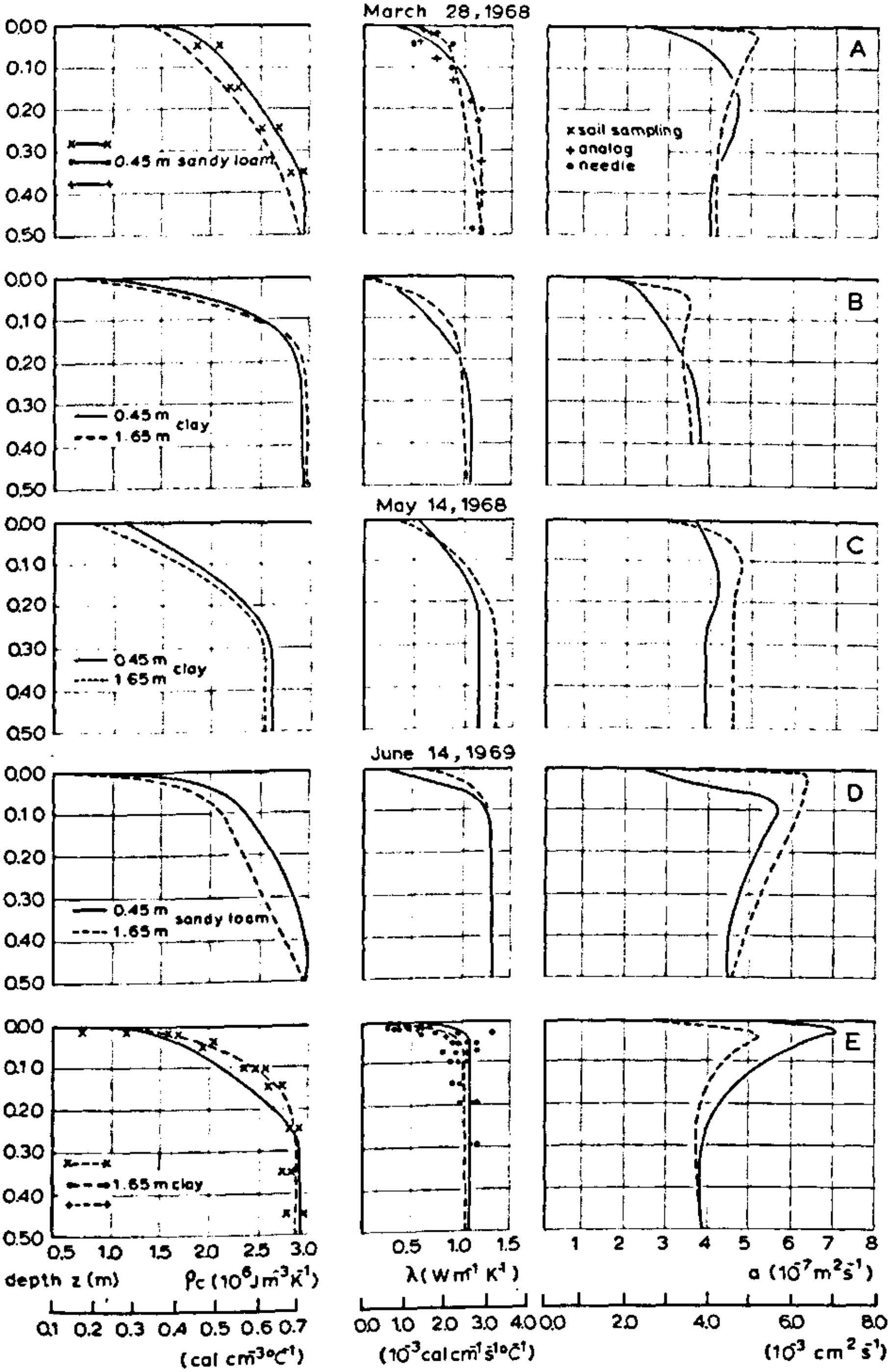
In fig. 40 some results on the variation of thermal properties with depth for the 0.45 and 1.65 m clay and sandy loam groundwater plots are presented. March 28, and May 14, were days in rather wet periods, June 14, 1969 a day in a rather dry period. The differences in the duplicates of the ρc data are mainly due to the natural inhomogeneity of the soil. Therefore it was decided to draw by hand a smooth curve through the means of the data.

The λ measurements which did not accord with the theory of a linear relationship of temperature against log time were ignored, as the scatter in these cases could be ascribed to poor contact between the needle and its surroundings. The remaining measurements give a clear picture of the large heterogeneity of the soil and therefore the same procedure as applied to the ρc values by drawing smooth curves through the means of the data was followed. In the λ graphs the mean λ values, obtained from the electric analog using the amplitude and phase relationships of the first harmonics, have been drawn in fig. 40A and E.

The thermal diffusivity at each depth now could be obtained by dividing λ by ρc.

Despite the various complications mentioned earlier, it can be seen from fig. 40 that at higher groundwater levels thermal diffusivity is generally lower, especially in the top 0.10 to 0.15 m layer. This can be explained by the fact that at higher groundwater levels because of higher soil moisture contents λ tends to increase, but ρc even more, which results in a decrease in a. Thus a shallower

FIG. 40. *Variation of thermal capacity ρc, thermal conductivity λ and thermal diffusivity a with depth for the clay and sandy loam of the 0.45 m and 1.65 m groundwater plots on March 28 and May 14, 1968 in a rather wet period, and for June 14, 1969 in a rather dry period. Thermal capacity was obtained from soil sampling and consecutive computation with eq. 61. Thermal conductivity was obtained from measurements with the transient needle method and from averaging the λ values obtained from the amplitude and phase relationships of the first harmonics. Thermal diffusivity was obtained by considering the ratio of λ and ρc.*



March 28, 1968
May 14, 1968
June 14, 1969
A
B
C
D
E
0.45 m sandy loam
x soil sampling
+ analog
• needle
0.45 m
1.65 m
clay
sandy loam
1.65 m clay
depth z (m)
ρc (10^6 J m^-3 K^-1)
(cal cm^-3 °C^-1)
λ (W m^-1 K^-1)
(10^-3 cal cm^-1 s^-1 °C^-1)
a (10^-7 m^2 s^-1)
(10^-3 cm^2 s^-1)

FIG. 40.

groundwater table will generally cause a lower soil temperature. The difference in thermal diffusivity of the different groundwater plots is most sharply pronounced in a dry period, e.g. on June 14, 1969 (fig. 40 D and E). The differences tend to decrease in a wet period, as for example for March 28, 1968 (fig. 40A and B) and for May 14, 1968 (fig. 40C), which makes the plots more uniform in thermal properties.

The higher diffusivity in the top layer of the 0.45 m clay on June 14 (fig. 40D) as compared with the 1.65 m clay is an exception to the rule, as on all other selected days all objects with high groundwater levels had lower diffusivities in the top layers. The cause may become clear by considering the curves of λ and ρc (fig. 40D). It is seen that λ (0.45 m) > λ (1.65 m). However, despite the higher moisture content of the 0.45 m plot ρc (0.45 m) < ρc (1.65 m). On closer inspection it appeared that the dry bulk densities of the top layers of the 1.65 m plot were higher than those of the 0.45 m plot (fig. 41). This had as final result higher ρc values and lower a values for the 1.65 m plot.

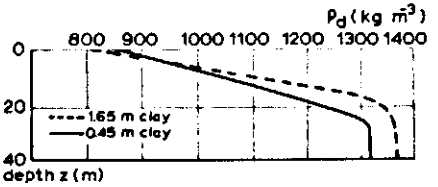


FIG. 41. *Variation of mean dry bulk density ρ_d (obtained from the sampling data of the emergence trials) with depth for the clay profiles of the 0.45 m and 1.65 m groundwater plots.*

It should be noted that differences in a values do not give direct information on differences in the absolute temperature regimes of soils. The final temperature distribution in a soil will be determined by the initial and boundary conditions of the system, i.c. the temperature distribution in the soil and the amount of available heat at the surface. Particularly the last term will be different for plots with a high respectively a low groundwater table, as a result of differences in evaporation between the plots (see table 8), which may result in considerable differences in the amounts of latent heat used.

In this respect it may be interesting to mention first the findings of other investigators, before presenting the present author's results. VAN DUIN (1960, 1963) computed the annual temperature wave at the surface of a sandy soil with groundwater tables at depths of 1.50 and 0.50 m below surface. Assuming the same amount of heat available at the surface, he calculated a difference of only 0.2°C due to differences in thermal properties. With the assumption of differences in amounts of heat available, as caused by a maximum difference in evaporation of 1 mm.day^{-1}, he came to a yearly mean temperature 2 to 2.4°C higher for the 1.50 m object under the assumption that shallow and deeply drained soils had the same yearly minimum temperature.

KOPECKY (1908) reports an increase in temperature in spring in the order of 1 to 2 °C as caused by tile drainage. KUNTZE (1958) mentions differences in temperatures of 7 °C between drained and undrained sticky clay soils, which seems to be very high. HERLYN (1964), reporting on temperature measurements in a peat soil covered with unripened mud, and groundwater tables varying from 0 to 0.50 m below the surface, found no statistically significant differences in temperatures for plots which had a maximum difference of 0.45 m in groundwater table depth.

Horticultural crops in glasshouses have relatively high heat requirements. VAN DER POST (1960) reviews the adaptation of growing methods of lettuce to soil properties. Lettuce has a relatively low heat requirement and is grown in spring in cold glasshouses on sandy and light sandy loam soils, which warm up quickly and have good aeration and water supplying capacities. On the other hand lettuce harvested in autumn and winter, is grown in heated glasshouses on clay soils which do not cool off as quickly as the previous mentioned soil types, and have a slower growing rate. VAN DAM and VAN DER KNAAP (1968) showed that on all kinds of soils the harvesting time of tomatoes grown in heated glasshouses was earlier when soil temperatures were higher. They could not derive a significant relationship between earliness and groundwater table depths for depths varying from 0.50 to 1 m below surface, indicating that the growers eliminate the harmful influence of high groundwater tables on soil temperature by increasing artificial heating. VAN WIJK (1969) compared the rate of warming up of a peat soil with a 0.50 m groundwater table and that of a clay soil with a groundwater table at 1 m below surface. From his data it can be derived that, if the same air temperature is maintained, the mean temperature of the peat soil to a depth of 0.50 m is 1 °C lower than the temperature of the clay soil. This means that with the same amount of energy it takes on the average 4 days longer to heat up the peat soil to the same temperature as the clay soil. KNOPPIEN and HILLE (1967) reported from measurements in a heated glasshouse at about 0.15 m depth below surface that a sticky clay soil was 0.5 °C warmer than a sandy loam soil. Also included in their experiments was a clay soil mixed with some sand to improve the structure. This soil type proved to be 1 °C colder than the pure clay soil, indicating the adverse effect of the sand mixture on soil temperature.

The results of the temperature measurements of the 0.45 and 1.65 m plots of the groundwater experimental field over the observation period March 18 to May 19, are listed in table 15.

From this table it can be seen that the mean daily temperatures of the plots with the higher groundwater tables are 1 to 2 °C lower than the temperatures of the plots with the deeper groundwater tables. This holds both for clay and sandy

TABLE 15. Soil temperatures measured at different depths on clay and sandy loam of the 0.45 m and 1.65 m groundwater plot over the observation period March 18 to May 19, 1968 (63 days)

Depth m	Temperature °C															
	1.65 clay			0.45 clay			1.65 sandy loam			0.45 sandy loam			$\overline{1.65-0.45}$		$\overline{1.65-0.45}$	$\overline{1.65-0.45}$
	max.	min.	mean	max.	min.	mean	max.	min.	mean	max.	min.	mean	clay	sandy loam	clay−sandy loam	clay−sandy loam
0.01	21.4	4.9	13.2	19.8	3.9	11.8	18.5	6.2	12.4	16.2	4.5	10.4	1.4	2.0	0.8	1.4
0.02	17.8	6.0	11.9	16.7	4.7	10.7	17.6	6.6	12.1	14.9	4.9	9.9	1.2	2.2	−0.2	0.8
0.03	15.7	6.6	11.2	15.1	5.4	10.2	16.4	7.1	11.7	14.1	5.4	9.7	1.0	2.0	−0.5	0.5
0.10	12.1	7.9	10.0	10.9	7.1	9.0				10.8	6.9	8.8	1.0			0.2

loam. Generally, it can also be stated that with the same groundwater depth the clay is warmer than the sandy loam (see column 0.45 clay – 0.45 sandy loam). This does, however, not hold for the 1.65 clay and 1.65 loam at the depths of 0.02 and 0.03 m. This exception may be due to small errors in the exact measurement depths, as it was noted that when heavy rainstorms occurred some of the top soil of the sandy loam was eroded away. Furthermore cultivated top layers are liable to subsidence. Temperature measurements under a cover of red cabbage in the 1.20 m groundwater plot over the observation period May 31 to November 11, 1968, showed that the temperatures at depths of 0.10 and 0.20 m were on the average 1.1 °C higher on the clay soil than on the sandy loam. Similar results were obtained from measurements in 1969.

In fig. 42 the course of daily mean temperatures of the 0.10 and 0.03 m depths during the spring of 1968 are plotted. From the measurements at the 0.10 m depths it can be seen that there is little difference between the clay and sandy loam profiles with the same groundwater level of 0.45 m and that on the investigated soils the difference in groundwater table depth plays a more important role than the differences in type of profile. The same holds for the 0.03 m depths. Generally it can be said that the sequence of cold to warm is: 0.45 sandy loam → 0.45 clay → 1.65 sandy loam → 1.65 clay.

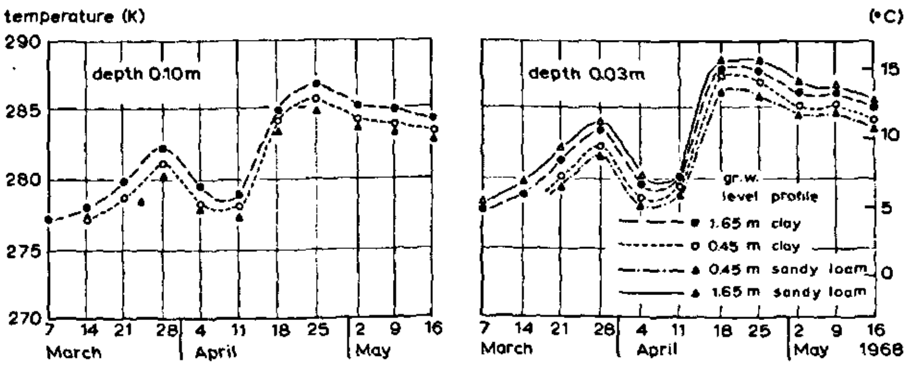


FIG. 42. *Course of the daily mean soil temperature at 0.10 m and 0.03 m below surface during spring of 1968 for the indicated soil profiles and groundwater table depths.*

The maximum and minimum temperatures of the various treatments can be compared by looking at fig. 36, table 16 and fig. 43. In fig. 43A the temperature differences in a (clay) profile as caused by a difference in groundwater table is shown. This holds even for a day in a wet period (May 14, 1968). Fig. 43B gives an idea of the temperature differences in a dry period (June 14, 1969) on a clay and sandy loam plot having the same high groundwater table.

It can be concluded that in spring the maxima and minima in the top soil are

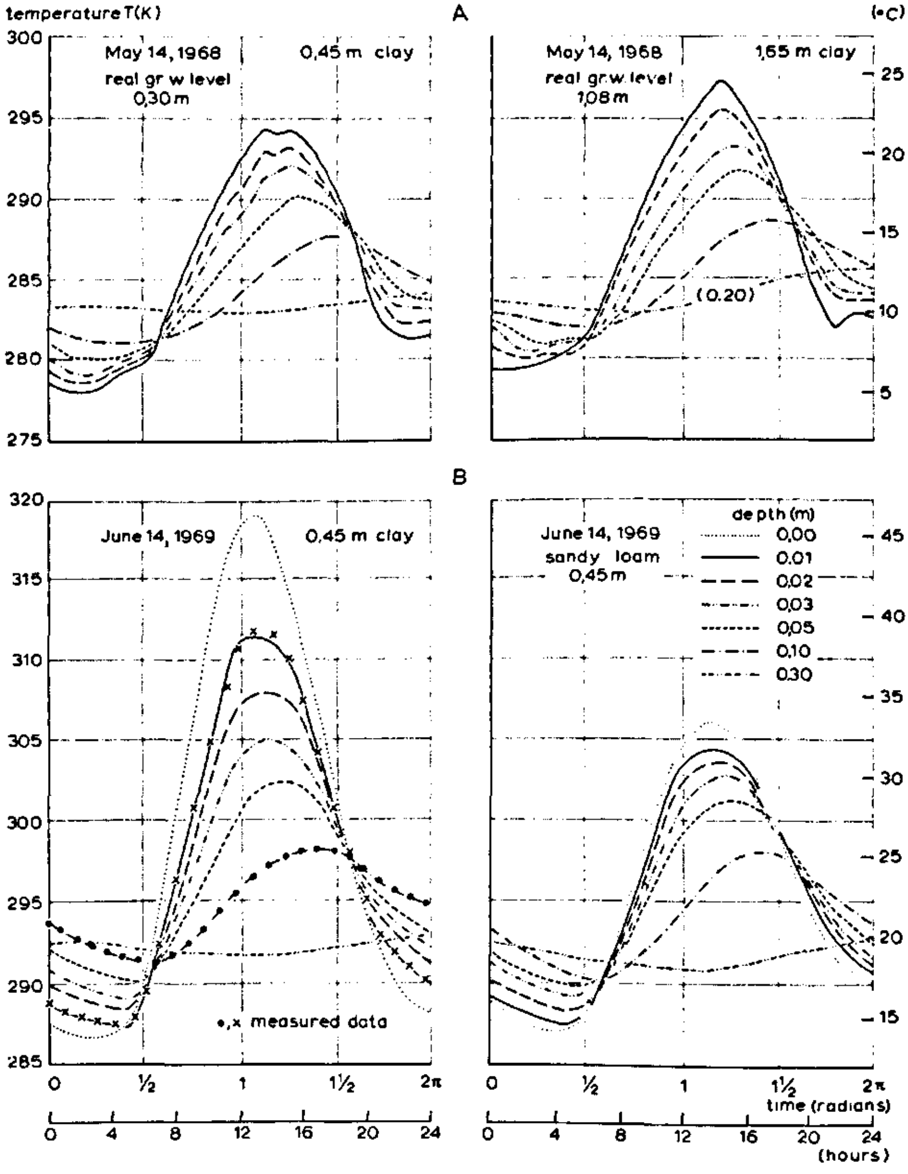


FIG. 43. *Diurnal variation of measured soil temperature at various depths showing A, temperature differences in a clay profile as caused by a difference in groundwater table and B, temperature differences between a clay and sandy loam profile having the same high groundwater table. The temperatures at the surface were found from extrapolation to the surface of each of the amplitudes and phases of the first three harmonics at the 0.03, 0.02 and 0.01 m depths and successive computation of the sum of the first three harmonics in its Fourier expansions. For the 0.01 and 0.10 m depths of the 0.45 m clay, June 14, 1969, it is seen that the curves for soil temperature as calculated by taking the sum of the first three harmonics fits the measured temperature data rather well (left hand fig. B, 0.01 and 0.10 m depths).*

higher when the soil is deeply drained and that the amplitudes decrease faster with depth in soils with a deep groundwater table. This decrease is more marked on the clay than on the sandy loam (see fig. 36). In accordance with this, the risk of (night-) frost is larger on a deep than on shallowly drained soil. The danger again being more pronounced on the clay than on the sandy loam soil. Measurements during a night-frost period confirmed this. Later in this period it appeared that the night-frost penetrated deeper into the clay soil. The temperatures of the plots with the high and with the low groundwater table did not show much difference, however, obviously as a result of rainfall antecedent to the frost period.

It is particularly interesting that the minimum temperatures of the shallowly drained plots are lower than the temperatures of the deeply drained plots. This is, however, generally true when, as in the cases under investigation, for the temperature-time curves in equal soils at two different groundwater depths, here 1.65 and 0.45 m below surface, $(\overline{T}_{1.65} - \overline{T}_{0.45}) > (A_{1.65} - A_{0.45})$ holds.

V. EFFECTS OF WATER AND HEAT ON SEEDLING EMERGENCE AND CROP PRODUCTION

1. INTRODUCTION

The suitability of a soil for horticultural use depends to a large extent upon its physical properties, as they determine the final effect of the external climatological conditions on crop growth. The Chapters III and IV dealt with the determination of the basic physical properties and the related transport processes of heat and water. These processes as occurring under different environmental conditions, were discussed for the three profiles each having its own water regime.

A study of physical processes is interesting from a scientific point of view, but for practical purposes the observed differences in behaviour of soil profiles will have to be expressed in terms of final effects on growth and production of crops. It is the purpose of this Chapter to evaluate some of the mentioned results with regard to their effect on two stages of plant development: the stage of germination to seedling emergence and the stage of seedling emergence to maturity.

In the first stage the most important process is respiration, yielding energy to change solids stored in the seed into soluble foods which can be transported to radicle and plumule to be utilized for cell division, elongation and differentiation. One of the prerequisites is water absorption. In this stage there is virtually no evaporation, nor dry matter production by photosynthesis. Growth chiefly depends on moisture content and temperature of the top surface layer of the soil.

In the second stage (from seedling emergence to maturity), the leaf area gradually increases. Owing to this, evaporation and photosynthesis are increasing too. The water uptake of the plant from the soil is then practically controlled by evaporation. In this stage, growth not only depends on moisture content and temperature of the soil, but also on the temperature of the air and mainly on leaf area and net radiation.

There is little need to emphasize the importance of an effective germination. Particularly in the case of vegetable crops, the establishment of a uniform crop stand and a uniform time of maturity is a strict economic necessity. Failure as well as retardation in germination are detrimental to an economically efficient production.

It is well-known that limiting available water reduces plant production, which especially holds true for vegetable crops where production is more aimed at quality and fresh weight than at an increase in dry matter. It is of interest to know the maximum production to be obtained under the prevailing weather conditions with an optimum water supply. Therefore attention will be paid to the calculation of maximum as compared with measured production rates. Special attention will be given to the influence of groundwater table depth.

2. FIELD AND LABORATORY EXPERIMENTS

Emergence experiments were conducted in the spring of 1968 and 1969, on the 0.45 m and 1.65 m groundwater plots, both on clay and sandy loam. Four different kinds of seeds were used: radish (*Raphanus sativus* L., var. *radicula*, cv. *Cherry Bell*), spinach (*Spinacia oleracea* L., cv. *Spinoza*), broad bean (*Vicia faba* L., cv. *Trio*) and garden beet (*Beta vulgaris* L., cv. *Juweel*). The separately determined germination percentages amounted to 94, 94, 99 and 93 respectively. In 1968 as well as in 1969 the seeds were sown on seven successive dates. In 1968: March 4, 18 and 29, April 8, 16, 22 and 29; in 1969: March 11 and 25, April 8, 22 and 29, May 7 and 13. At each sowing date 200 seeds of each crop were sown by hand at a depth of 3 cm with a row distance of 0.10 to 0.15 m on 4 areas of 1.5 m^2 (so in total 16 objects per sowing date). Soil temperatures were registered at the sowing depth of 3 cm, according to the procedure described in section IV3. Each week the soil moisture content of the various plots was determined by means of vertical soil sampling with Kopecky rings (height 5 cm) at depths of 0.05, 0.15, 0.25 and 0.30 m. Emergence was considered to have taken place when the cotyledons had developed. The number of seeds that emerged were picked and counted each day.

In order to check the results and conclusions obtained from the field experiments, another experiment with radish seeds under controlled soil moisture and temperature conditions was carried out in the laboratory. For that purpose undisturbed soil samples were taken from the surface of the two profiles in the field in rings of 10 cm diameter and 2.5 cm height. These samples were brought in equilibrium with various matric pressures by means of the apparatuses and methods used for the determination of soil moisture retention curves, already described in section III2. The matric pressures employed were: -0.0098, -0.031, -0.098, -0.196, -0.49, -0.98 and -2.47 bar (pF 1.0, 1.5, 2.0, 2.3, 2.7, 3.0 and 3.4 respectively). Of the 7 pressure treatments, the first 5 were carried out in duplicate for each of the two profiles. In a first experiment the seeds were sown in the top of the samples, in a repeat in the turned over underside of the same samples. The temperature of the samples was in the range of

20 to 21 °C and was recorded by means of ferro-constantan thermocouples.

In each sample 50 radish seeds for uniformity passed through a 2 mm mesh, (the batch tested for germination giving a germination percentage of 98), were sown at a depth of 0.25 to 0.50 cm. Owing to this smaller sowing depth as compared with the 3 cm sowing depth in the field (necessary because the necessary large diameters of the undisturbed samples entailed a small height of the samples), it was to be expected that the heat sums necessary for a certain emergence percentage would differ from the values obtained in the field experiments. As emergence criterion a normal development of both radicle and cotyledons was taken.

For production experiments crops like spinach, red cabbage, dwarf French beans and celery were grown (details have been given in section IIIB-2). Periodical measurements on fresh and dry weight production, crop height, soil cover and rooting depth were carried out on the 0.30, 0.60, 0.90, 1.20 and 1.50 m groundwater plots. In addition to this, also determinations of the distribution of dry matter between shoots and roots on some of the sandy loam plots.

On the 0.45, 0.75, 1.35 and 1.65 m groundwater plots, only final production was measured.

3. GERMINATION AND SEEDLING EMERGENCE

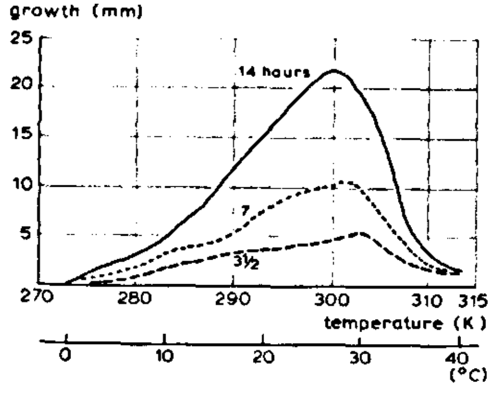
After reviewing known effects of soil temperature and soil moisture content on germination and seedling emergence, the results obtained from the field and laboratory experiments will be discussed.

a. Effect of soil temperature

It has been known for a long time that temperature has a remarkable effect on the germination of seeds (see for example Sachs, 1860; Haberlandt, 1874; Kotowski, 1926; Edwards, 1932). In general three cardinal temperatures for germination have been defined: the minimum and maximum temperatures below and above which no germination occurs, and the optimum temperature giving the highest germination rate.

Instead of definite cardinal points it is more correct to speak of temperature ranges because the temperatures found are influenced by such external conditions as the stage of after-ripening in the seeds, the testing technique and the germination criteria used. Moreover it is very difficult to determine the minimum and maximum temperatures, as they are reached asymptotically (fig. 44 after Talma, 1918). For a review of these and similar considerations and the

FIG. 44. *Dependence of growth of roots of pepper weed (Lepidium) at increasing periods of exposure to various constant temperatures (after* TALMA, *1918).*



early literature on the effects of temperature on various kinds of crops, see HAGAN (1952a, b).

Temperature investigations on the germination and emergence of six range grasses were reported by MCGINNIES (1960), on nineteen native and cultivated herbaceous species in Southern Alberta by DUBETZ et al. (1962), on thirteen kinds of vegetables by HARRINGTON (1962), on cauliflower by HEPTON (1957), on celery by GUZMAN (1964) and on tomato to JAWORSKI and VALLI (1964).

For the relationship between temperature and growth the temperature coefficient (Q_{10}), defined as the factor by which the growth rate in a certain temperature range is multiplied when the temperature is increased with 10°C, is often used. From fig. 44 it follows that within the daily temperature range of the soil a linear relationship between growth rate and temperature can be used with good approximation for growth processes like germination, e-mergence and initial stem elongation. In the range between minimum and optimum temperature the growth rate at any time is proportional to temperature and at a fixed temperature level growth rate is proportional to time. The combined relationship for a definite emergence percentage can therefore be represented by:

$$F = (\bar{T} - T_{min}) \times t \qquad (92)$$

where $\bar{T}$ and T_{min} are the mean and minimum soil temperature (K), t is time (seconds) and F (K.s) is a constant heat sum. In this way a rather simple means to relate temperature and emergence has been obtained. In practice these heat sums are often used to schedule plantings and to predict maturity, to select crop varieties appropriate to different areas or to schedule work in fruit orchards (DETHIER and VITTUM, 1963). Although applicable for early growth, it has been shown (BIERHUIZEN and FEDDES, 1969) that predictions of yield and harvest time have to be treated with great care since radiation then plays a pre-dominant role. In such type of studies, the environmental air temperature is mostly used. In the case of emergence of seeds it is advisable, however, to register the temperature in the direct environment of the seed, c.q. to measure at sowing depth. It is clear that temperatures at other depths will give different heat sums for a particular stage, especially as the temperature fluctuations in the top layer are generally very large.

Under field conditions soil temperature will vary during the day, and $\bar{T}$ has to be found from the expression:

$$\bar{T}(z) = \frac{1}{(t_2 - t_1)} \int_{t_1}^{t_2} T(z)\,\mathrm{d}t \qquad (93)$$

where $t_2 - t_1$ is 24 hours. One of the most correct methods of solving eq. (93) is to use an integrimeter. Then, in cases when during part of the day the temperature is below T_{min} or above T_{max}, the integrated temperature sum below T_{min} and/or above T_{max}, which must be disregarded, can be easily omitted.

For days with soil temperatures continuously above T_{min} and below T_{max} an approximation of eq. (93) can be given by:

$$\bar{T} = (T_{max} + T_{min})/2 \qquad (94)$$

which relationship will give particularly good results when the temperature varies sinusoidally with time. For a few other analytic approximations of eq. (93), see ARNOLD (1959).

b. EFFECT OF SOIL MOISTURE

The other essential condition for germination and emergence, namely soil moisture content, has also been early recognized. However, most studies were dealing to only a slight extent with quantitative aspects (see DONEEN and MACGILLAVRY, 1943). These authors, using soil as a germination medium. performed experiments with seeds of different vegetables. They found that, with the exception of celery seeds, total emergence was not affected in the range between field capacity and permanent wilting percentage. However, they also reported that the rate of emergence decreased with decreasing matric pressures. AYERS (1952) working on the emergence of onion seeds in soil, found both a decrease in rate and final percentage of emergence with decreasing matric pressures. HUNTER and ERICKSON (1952) found for five soil types minimum values of matric pressures of −12.7, −8.0, −6.7 and −3.5 bar necessary for emergence of corn, rice, soybeans and sugar beet seeds respectively. MCGINNIES (1960) used osmotic solutions of different concentrations in order to get various (osmotic) pressures. Reviewing the results of various workers who used this type of method, and studying pressure effects on range grasses, he concluded that increasing concentrations delayed the rate and reduced total germination percentage. TAYLOR (1964) reports no statistical significant differences on germination of radish seed at concentrations giving absolute osmotic pressures below 15 bar.

Obviously many investigators assumed that osmotic pressures, which are numerically equivalent with matric pressures, have a similar influence on germination by lowering the free energy of water in the same quantitative way. Germination experiments of a basic nature by COLLIS-GEORGE and SANDS (1959, 1962) on sintered discs showed that this is only true if seeds have semi-permeable membranes capable of preventing movement of solutes into the seed. With permeable membranes, solutes enter the seeds by diffusion and when equilibrium has been reached, the osmotic component is no longer effective in decreasing the free energy of water. The last mentioned authors pointed out, as they detected retardation of the germination rate already at matric pressures of -0.238 bar, that the delay in germination rate which was found by various authors at the very low osmotic pressures of -5.97 to -7.85 bar is not caused by a physiological drought, but by a final toxidity of the solution. TAYLOR (1964) reported that the same germination percentage of radish seed occurred at lower pressures in solutions than in soil.

The interesting experiments of COLLIS-GEORGE and HECTOR (1966) and of COLLIS-GEORGE and WILLIAMS (1968) did show that matric pressure influences the germination of seeds in various ways. It influences the free energy of the soil water, the mechanical strength of the solid matrix, the area of seed – liquid contact, and the aeration and hydraulic conductivity of the soil. In the range between 0 – 400 cm H_2O (0 to 0.393 bar), it is thought by them that the influence of matric pressure must be completely attributed to its contribution to the effective stress in the solid framework surrounding the seed and not to its influence on the free energy regime of the soil water. HADAS (1970) reports from his experiments that there was no influence on germination at effective stresses of the order of 10^7 dyne.cm^{-2} (10 bar). The area of water contact between soil and seed is regarded as being of less importance when matric pressure decreases (SEDGLEY, 1963; COLLIS–GEORGE and HECTOR, 1966). This means that the area of water contact would under field conditions be of only minor importance.

Considering aeration HACK (1963), who worked with tomato seeds in compost with high matric pressures, found good emergence at -0.027 bar which means an air filled pore space of 20 vol % of air, but poor emergence at -0.013 bar, corresponding with 8 vol. % of air. DASBERG (1968) states that the bulk of oxygen is taken up by the seed by means of diffusion. He found a decrease in total emergence of wheat and of a range grass (*Oryzopsis*), when the oxygen diffusion rate to Pt electrodes was below 20×10^{-8} g.cm^{-2}.min^{-1} (1.33 ×

10^{-9} kg.m^{-2}.s^{-1}). HANKS and THORP (1956) observed reduction in seedling emergence at oxygen diffusion rates below 75 to 100 × 10^{-8} g.cm^{-2}.min^{-1} (5.00 – 6.67 × 10^{-9} kg.m^{-2}.s^{-1}). DASBERG (1968) did show that irrespective of the medium, germination and emergence were dependent on the amount of water taken up by the seeds, which is determined by the matric pressure of the soil. It seems (HADAS, 1970) that for good germination and emergence a relative water uptake around 100% of the initial seed weight is necessary. Dasberg found that water at distances exceeding 10 mm was not taken up by 2 mm seeds, which would mean that outside that range the hydraulic conductivity is of no importance.

c. COMBINED EFFECT OF SOIL TEMPERATURE AND SOIL MOISTURE

The combined effect of temperature and soil moisture content on seedling emergence was studied in field experiments. For each sowing treatment on each plot graphs were made of percentage of emergence versus time in days after sowing. Fig. 45 shows the emergence rates of radish sown on March 18, and April 8, 1968, respectively. It appears that emergence is highly correlated with precipitation. Another common feature is that seeds emerge earlier in sandy loam than in clay. The seed-soil contact as well as the contact with the subsoil is much better in the sandy loam than in the crumble-structured top layer of the structured clay soil. Obviously the water uptake by the seedlings is favoured in the sandy loam, where a high hydraulic conductivity at high matric pressures ensures an ample water supply. These are reasons that on all sowing dates the sandy loam plots with the high groundwater table showed the highest emergence rate as well as the highest total emergence percentage. From graphs like fig. 45 the time in days for 50% emergence (t_{50}) was obtained.

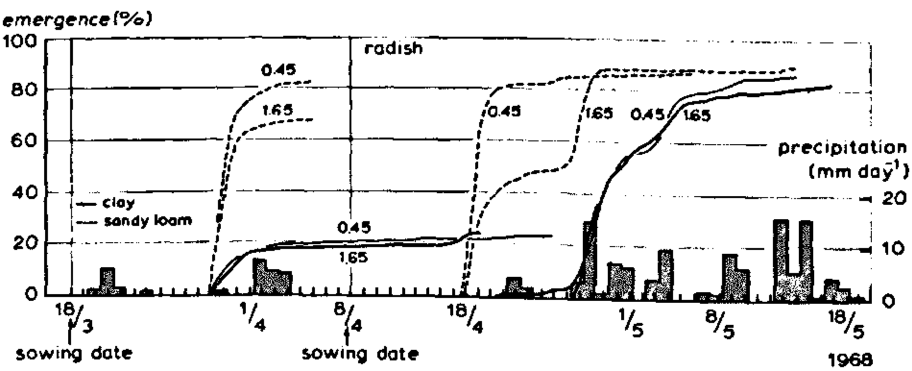


FIG. 45. *Course of emergence of radish sown in the clay and sandy loam profiles of the 0.45 and 1.65 m groundwater plots on March 18, and April 8, 1968 respectively.*

As a first approach to determine the heat sum required for 50% emergence, for each treatment heat sums were calculated by multiplying $\bar{T}$ with t_{50}. It appeared that in general all mean heat sums to 50% emergence on sandy loam were lower than those on clay, and that the heat sums of the 0.45 m groundwater plots were lower than those of the 1.65 m plots. Except for the 0.45 m sandy loam, extremely high heat sums were required during a rainless period from the 5th till the 21st of April in the 4th and 5th emergence experiment, resulting in a water shortage in the top 5 cm soil layer.

For a calculation of the minimum temperature for emergence (T_{min}) only those treatments were used in which no limitation of water could be expected to occur. According to the heat sum concept a linear relationship should be present when the mean soil temperature ($\bar{T}$) is plotted versus the reciprocal of the period to emergence ($1/t_{50}$). In this way the intercept with the temperature axis directly gives T_{min} and the slope the heat sum (fig. 46). The correlation

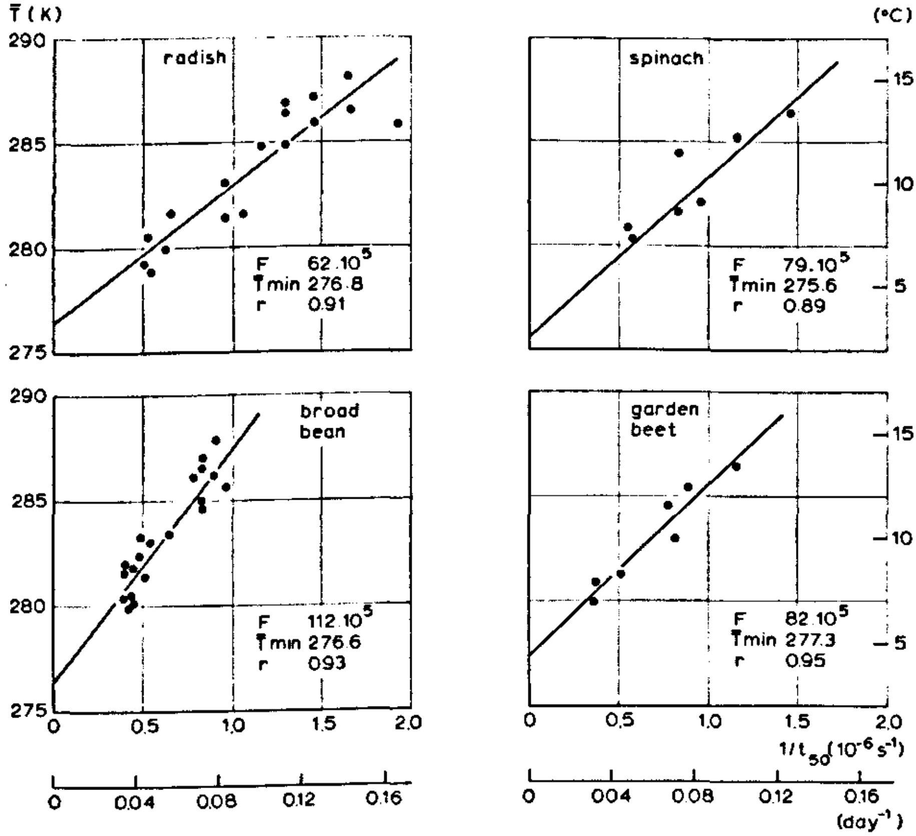


FIG. 46. *The mean soil temperature $\bar{T}$ at 0.03 m depth plotted versus the reciprocal of time required for 50% emergence (t_{50}) in order to derive the minimum temperature for emergence T_{min} for the four vegetable crops. F = heat sum in K.s; r = correlation coefficient.*

coefficient of $\bar{T}$ and $1/t_{50}$ was relatively high, namely in the order of 0.9. Under the condition of no water limitation, the minimum temperatures for emergence of respectively radish, spinach, broad bean and garden beet were 3.8°, 2.6°, 3.3° and 4.3°C and the heat sums 72, 91, 130 and 95 degree.days. The results of the minimum temperatures for emergence agree with findings of KOTOWSKI (1926) who mentions a minimum of 4°C for radish and spinach, and of 4 to 8°C for beet. In horticultural practise it is known that spinach emerges already at fairly low temperatures. HARRINGTON (1962) reports for spinach a development of 83% normal seedlings after 63 days at a temperature of 0°C and of 96% normal seedlings after 27 days at a temperature of 5°C. It is also generally known that radish emerges quickly and broad beans rather slowly, which agrees with the respective heat sums of 72 and 130 obtained by the present author.

Now T_{min} was known, in a second approach the effect of soil moisture on emergence could be evaluated by calculating the heat sums of all treatments, taking into account the minimum temperature for emergence, hence $(\bar{T} - T_{min}) \times t_{50}$. An average soil moisture content for the first five days after sowing was for each treatment calculated from the soil moisture data and the precipitation records. The period of the first five days was taken to be the most important one for swelling, initial water uptake and recommencing of metabolism of the seeds. This is in accordance with HUNTER and ERICKSON (1952) who also believed that the five-day seed-soil contact is the best indicator for the minimum moisture content required for emergence. As soon as the primary roots have developed water from deeper layers will come within reach and the seed will become less dependent on the water content of the top soil layer.

Soil moisture measurements were usually performed on the sowing date. It was assumed that precipitation amounts were stored in the top 5 cm layer up to field capacity and that evaporation losses were compensated for by upward water flow.

In fig. 47 the heat sum required for 50% emergence in each treatment is presented against calculated matric pressure. It is obvious that below -0.49 bar (above pF 2.7) all seeds mentioned require a sharply increasing heat sum to bring them to 50% emergence. This result is in accordance with the known retardation of elongation at lower matric pressures. The variation in heat sum needed for 50% emergence at matric pressures below -0.49 bar is large, which may be ascribed to the large fluctuations in soil moisture content in the top 5 cm layer, and the difficulties to obtain exact measurements of the soil moisture content at this depth. The circled points are from the 5th sowing period which was very dry. The broken circled points are belonging to the 3rd germination experiment in which rainfall on the 5th day after sowing date obviously disturbed the calculation of the mean soil moisture content during the first five days, resulting in a too high mean matric pressure value for this period.

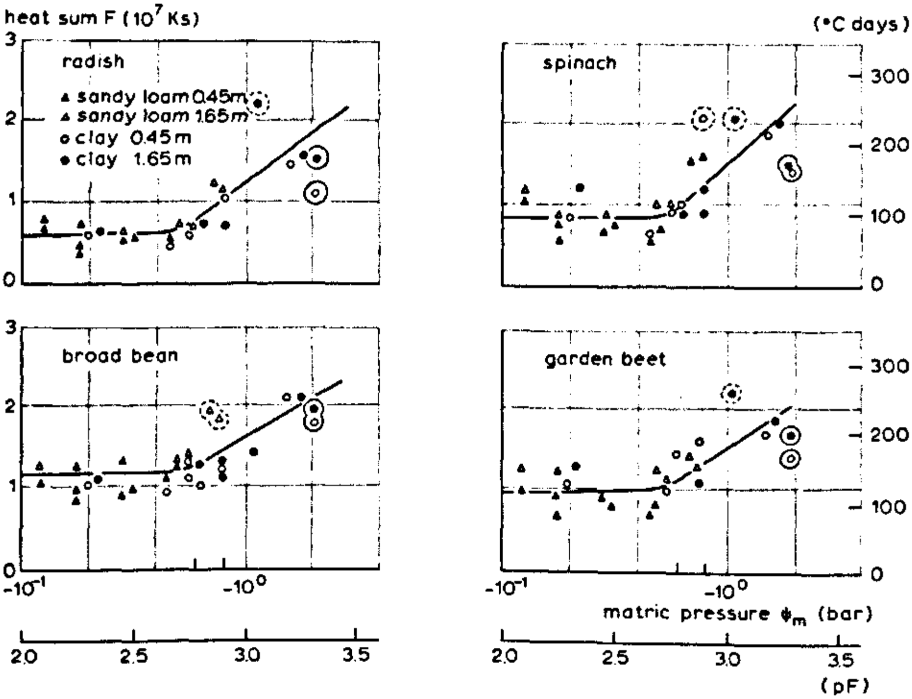


FIG. 47. *Effect of matric pressure of the soil on the heat sum required for 50% emergence for four vegetable crops sown in the clay and sandy loam profiles of the 0.45 and 1.65 m groundwater plots. Circled points are from the 5th sowing period which was very dry, broken circled points are from the 3rd sowing period with rainfall on the 5th day after sowing.*

The results of the laboratory experiments are given in fig. 48A. It is clear that at matric pressures between −0.0098 and −0.031 bar (pF 1.0 and 1.5 respectively), the soils are too wet for good emergence. The seeds start to germinate, but their development is too poor to yield normal seedlings. From −0.098 to −0.49 bar (pF 2.0 to 2.7) an emergence percentage of 90% or higher occurred. At lower pressures −0.98 to −2.47 bar (pF 3.0 to 3.4) total emergence percentages again decreased. This confirms the results obtained from the field experiments that −0.49 bar (pF 2.7) is the critical value for emergence as regards the dry side. From fig. 48A the mean total emergence percentage occurring after 7 days at the various pressure treatments were derived, see fig. 48B. In this graph also the separate total emergence values of each experiment are given. It appears that in the second experiment total emergence percentages were generally lower than in the first experiment. The reason for this can be that the seeds in the repeats were sown in the turned over underside of the soil samples and therefore in denser soil layers. Also a slightly larger sowing depth may have had some influence on the total emergence percentage. By application of horizontal sectioning of fig. 48A at 50% emergence, the time needed to reach this

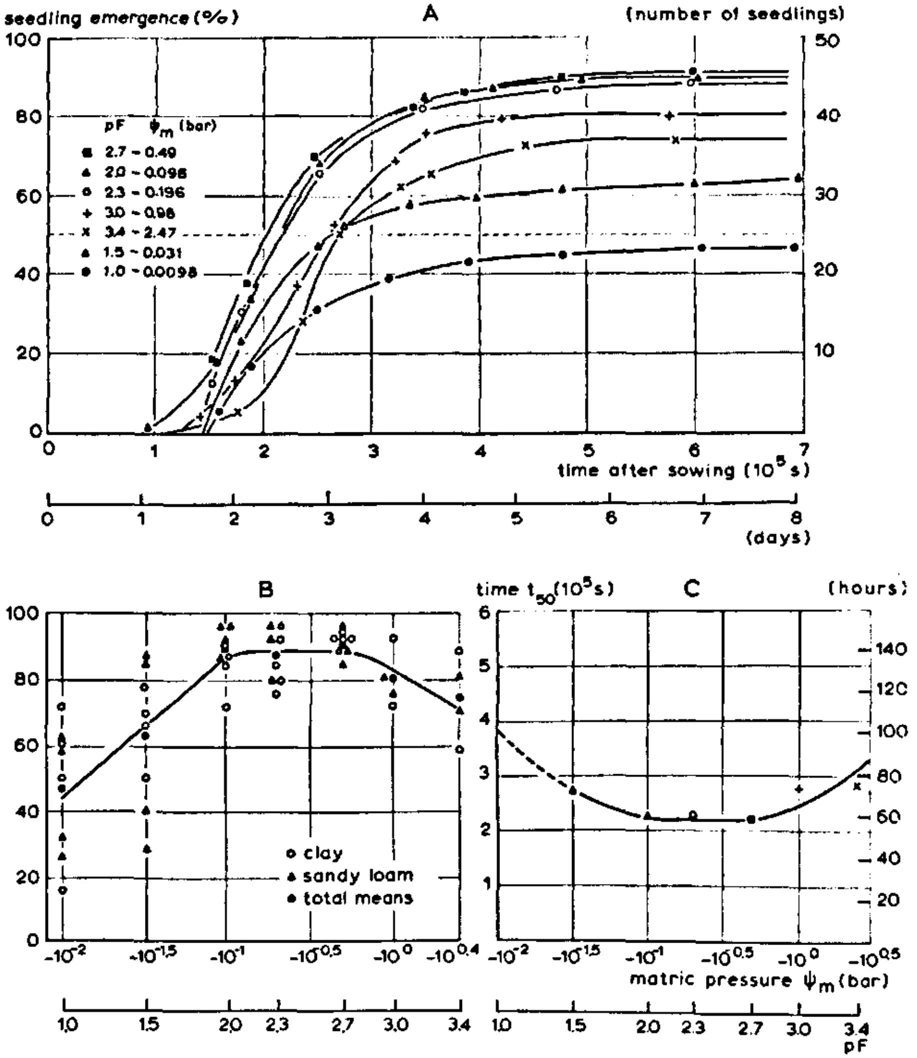


FIG. 48. *A, mean seedling emergence of radish in clay and sandy loam at seven controlled matric pressures versus time after sowing; B, mean total emergence percentage occurring seven days after sowing at the various pressure treatments for clay and sandy loam, as derived by vertical sectioning of fig. 48A; C, time required for 50% emergence at the various pressure treatments, as derived from horizontal sectioning of fig. 48A.*

emergence stage at various matric pressures can be obtained. The result is shown in fig. 48C indicating that the time for emergence increases rapidly at matric pressures above —0.098 bar and below —0.49 bar.

For the sandy loam, with $\rho_d = 1320$ kg.m^{-3}, BAKKER (1971) determined the oxygen diffusion rate as a function of water content. At field capacity (—0.098

bar, pF 2.0) the soil contains 37 volume per cent water and showed an oxygen diffusion rate of 1.33×10^{-9} kg.m^{-2}.s^{-1}, which is just the minimum value necessary for emergence as found by DASBERG (1968). This supports once more the conclusion that *for this soil* -0.098 bar *(pF 2.0) is the critical value for* emergence as regards the wet side.

It appears from the field and laboratory experiments that heat sums can give a relatively accurate prediction for emergence, if soil moisture content is taken into account.

For a fast and adequate seedling emergence both a high temperature and a sufficient moisture content are necessary.

Under field conditions this combination is seldom reached: higher temperatures are generally related with lower moisture contents and deeper groundwater tables. To get out of this dilemma one can maintain a relative deep groundwater table, which gives a relatively high temperature, and keeps by means of sprinkler irrigation the sowing bed at the desired moisture content between -0.098 bar (pF 2.0) and -0.49 bar (pF 2.7). This offers the additional advantage of keeping the temperature of the seed bed low in periods when soil temperature would exceed the optimum temperature for germination and emergence. Even on the experimental field in the temperate climate of the Netherlands near the sea coast, temperatures of 40°C and higher were in mid summer no exception in the top layer of clay soils. Such temperatures are far too high for germination and emergence of seeds of for example cabbage and spinach, which have optimum emergence temperatures of 8 to 11°C, beet and cauliflower of 11 to 18°C as well as asparagus, carrot, celery, endive, lettuce, onion, pea, radish and tomato all with optimum temperatures between 18 to 25°C.

A few important conclusions on the effects of drainage upon germination and emergence can be drawn. It is known that yields of spring crops decrease when sowing dates are later, the depressions increasing progressively with the shift in sowing date (WIND, 1960). It is also known that owing to adequate drainage, soils can be cultivated and sown approximately 5 to 14 days earlier (WESSELING and VAN WIJK, 1957; WIND, 1963). An additional advantage of drainage is the shortening of the germination and emergence period as a result of the higher soil temperatures. This can be elucidated with the following example. For a crop like spinach, which has a minimum emergence temperature of about 3°C and an optimum temperature of 11°C, a difference in soil temperature of 1.5°C means a shortening of the emergence period of say 19% $[100/(11 - 3) \times 1.5]$. This can be expressed in days as follows (see eq. 92). If a heat sum of 180 degree. days is necessary for 100% germination and emergence, this means that in a shallowly drained soil with a mean temperature of 7.5°C a time t of 180/

/(7.5 − 3) = 40 days is needed. If due to drainage the temperature is raised to 9.0°C the time will be 180/(9.0 − 3) = 30 days, which gives a 10-day gain in emergence as compared with the shallowly drained soil, assuming that enough water for germination and emergence remains present.

4. CROP PRODUCTION

a. Calculation of potential production

Photosynthesis is the process with which radiant energy is converted into chemical energy by the reduction of CO_2 (in the presence of H_2O) to carbohydrates (CH_2O). This process occurs in the chloroplasts of green plants. Because energy is needed for the maintainance of the structure and the living faculties of the plants, some of the stored carbohydrates are oxidized to deliver the required energy by the process of respiration. The difference between photosynthesis and respiration is called net photosynthesis. In general three main types of processes are of importance for net photosynthesis, namely diffusion and photochemical and biochemical processes (e.g. GAASTRA, 1963).

The *diffusion* processes take care of the transport of CO_2 from the external air through the stomata and then to the chloroplasts situated in the leaf mesophyll. Within the normal temperature range these processes mainly depend on the difference in CO_2 concentration outside the leaf (C_z) and inside the chloroplasts (C_0). The CO_2 diffusion encounters the following resistances. First the aerodynamic resistance between bulk air and effective leaf surface (r_a), which is assumed to be equal for all gases because air exchange is governed by turbulent mixing (MONTEITH, 1963). Secondly the resistance for molecular diffusion transport through the stomata (r_s'). Finally the resistance (especially associated with net photosynthesis) of the mesophyll cells (r_m') for CO_2 diffusion, which transport is supposed to take place in the liquid phase. Thus, except for the last resistance, the same resistances are encountered for CO_2 as for water vapour transport. The molecular diffusion resistances for gaseous CO_2 transport (r_s') are related to the diffusion resistances for water vapour (r_s) by the ratio of their diffusion coefficients: $r_s'/r_s \approx D_{H_2O}/D_{CO_2} = 1.8$. Hence under steady state conditions the photosynthetic flux of CO_2 in anology to eq. (23) can be written as

$$P = -(C_0 - C_z)_{CO_2}/(r_a + 1.8r_s + r_m') \qquad (\mathrm{kg.m^{-2}.s^{-1}}) \qquad (95)$$

As stated, in connection with eq. (25), r_s represents for a single leaf the stomatal resistance, for a crop surface, however, it represents the resistances of both crop and soil surface.

/(7.5 − 3) = 40 days is needed. If due to drainage the temperature is raised to 9.0 °C the time will be 180/(9.0 − 3) = 30 days, which gives a 10-day gain in emergence as compared with the shallowly drained soil, assuming that enough water for germination and emergence remains present.

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Radiant energy may indirectly influence the diffusion process by closing of stomata at low radiant fluxes, increasing the value of r_s.

The *photochemical* processes involved in photosynthesis are responsible for the reduction of CO_2 to CH_2O under influence of radiant energy. The solar radiation active in these processes (R') lies within the spectral range of 0.4–0.7 μm. The value of R' is about 0.5 of the shortwave radiation flux R_s.

The *biochemical* processes responsible for the chemical reduction of CO_2 are strongly temperature dependent.

The three groups of processes can be illustrated with fig. 49 (after GAASTRA, 1963), where photosynthesis (P) of a single cucumber leaf is shown in relation to photosynthetically active radiation flux (R') and the temperature at a limiting (0.03%) and at a non-limiting or 'saturated' (0.13%) CO_2 concentration. From this graph it is seen that at low values of R', photosynthetic flux P is limited by the photochemical process (curve A). At normal CO_2 concentrations (0.03%) and high values of R', P increases rapidly when the external CO_2 concentration is raised (curve B). This means that under the conditions of curve A, CO_2 diffusion is limiting photosynthesis. A change of temperature from 293 K to 303 K has hardly any effect (curve A). At high values of R' and CO_2 the same temperature increase has a strong effect (curve C) because biochemical processes then were limiting.

GAASTRA (1963) also showed that in principle photosynthesis can be more increased by increasing the flux of the diffusion process (which is commonly

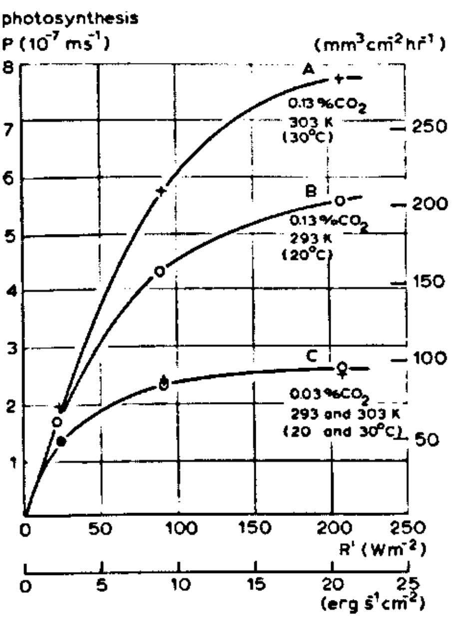


FIG. 49. *Photosynthetic flux of a cucumber leaf in relation to photosynthetically active radiation flux R' for a limiting (0.03%) and a 'saturated' (0.13%) CO_2 concentration at two temperatures (after* GAASTRA, *1963).*

done in greenhouses by increasing CO_2 concentrations of the air) than by increasing the radiation flux.

Generally the relationship derived for leaves also holds for leaf canopies, with differences, however. For instance, as a result of the mutual shading of leaves, the saturation level of radiant fluxes lies much higher than for single horizontal leaves. Furthermore a leaf canopy is not equivalent with a large horizontal leaf as it consists of leaves spatially arranged at various angles, causing a lower reflection and by means of a better radiation distribution a more efficient uptake of the incident radiation flux. The uptake efficiency depends on factors like number, size and spatial arrangement of the leaves, transmission and reflection, solar altitude and cloudiness (DE WIT, 1965). Based on radiation and leaf distribution functions, the latter author combined the variables mentioned in a model which makes it possible to calculate for any date and place the daily potential photosynthesis of a crop. This approach gives the investigator interested in actual crop problems a valuable standard with which he can weigh various measures, as for instance change in groundwater table depth, in soil profile, etc. The procedure of De Wit can be summarized as follows. For the Netherlands (52°N latitude) the variation in average photosynthetically active radiation flux on clear days ($\bar{R}'_c = 0.5\ \bar{R}^c_s$) over the year is shown in fig. 50A. The average photosynthetically active radiation flux on overcast days ($\bar{R}'_o$) is assumed to be $0.2\bar{R}'_c$. With these data and certain radiation and leaf distribution functions he evolved, De

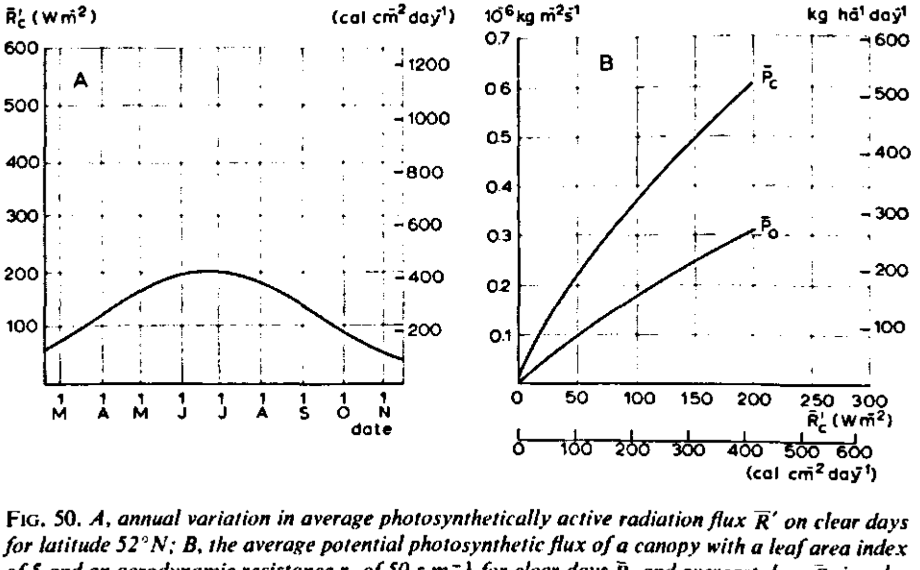


FIG. 50. *A, annual variation in average photosynthetically active radiation flux $\bar{R}'$ on clear days for latitude 52°N; B, the average potential photosynthetic flux of a canopy with a leaf area index of 5 and an aerodynamic resistance r_a of 50 s.m^{-1} for clear days $\bar{P}_c$ and overcast days $\bar{P}_o$ in relation to average photosynthetically active radiation $\bar{R}'$ for latitude 52°N (derived from data of* DE WIT, *1965).*

Wit calculated the average potential photosynthetic flux of a canopy under standard conditions for both clear and overcast days. The results are summarized in fig. 50B. From this graph it can be seen that the average photosynthetic flux on overcast days $\bar{P}_o$(kg.m^{-2}.s^{-1}) is not 0.2 times the average flux on clear days $\bar{P}_c$, but about 0.5 $\bar{P}_c$, which is due to the better radiation distribution on these days.

The average photosynthetically active radiation flux on an arbitrary day ($\bar{R}'$) is taken to be equal to 0.5 R_s. Reading at this value in fig. 50B gives the values of $\bar{P}_c$ and $\bar{P}_o$. The mean fraction of time the sky under the actual conditions is overcast, $\bar{\Lambda}$ is found from

$$\bar{\Lambda} = (\bar{R}'_c - \bar{R}')/(0.8\bar{R}'_c) \tag{96}$$

The mean potential photosynthetic flux ($\bar{P}_{pot}$) is then obtained by:

$$\bar{P}_{pot} = \bar{\Lambda}\bar{P}_o + (1 - \bar{\Lambda})\,\bar{P}_c \qquad (\mathrm{kg.m^{-2}.s^{-1}}) \tag{97}$$

This mean potential dry matter production rate ($\bar{P}_{pot}$) refers to conditions with a leaf area index of 5 (which means a closed green crop), optimum water supply and $r_a = 50$ s.m^{-1}.

RIJTEMA and ENDRÖDI (1970) have calculated with the aid of eq. (95) photosynthetic fluxes (P_1) at other values of r_a relative to the photosynthetic fluxes (P_2) at $r_a = 50$ s.m^{-1}. Then $P_1/P_2 = [\Delta_1 C_{CO_2}/(r_a + 1.8r_s + r'_m)]/[\Delta_2 C_{CO_2}/(50 + 1.8r_s + r'_m)]$ and assuming $\Delta_1 C_{CO_2} = \Delta_2 C_{CO_2}$ gives $P_1/P_2 = (50 + 1.8r_s + r'_m)/(r_a + 1.8r_s + r'_m)$. Reading from fig. 12 of DE WIT (1965) P values at different r_a values, they came for the conclusion that if r_s for the conditions of optimum water supply may be taken to be zero, the r'_m value of the standard crop is 440 s.m^{-1}. Hence for r_a values different from 50 s.m^{-1}, the $\bar{P}_{pot}$ value has to be multiplied with the ratio of the sum of the resistances of the standard crop (50 + 440 = 490) and the sum of the resistances of the actual crop (r_a + $1.8r_s$ + 440). This gives:

$$\bar{P} = [490/(r_a + 1.8r_s + 440)]\,\bar{P}_{pot} \tag{98}$$

where r_s is the resistance of the leaf canopy and is obtained from evaporation data.

Under natural field conditions potential photosynthesis will never be reached. One of the reasons is that during the beginning of the growing period photosynthesis is only performed by the area fraction of soil covered (S_c). Another reason is the loss in dry matter production by respiratory processes. Whereas photosynthesis reaches fairly early at increasing temperatures a rather flat topped maximum (and decreases at still higher temperatures), respiration remains increasing to a point where all physiological processes of the plant stop. Net photosynthesis will therefore rather quickly increase with increasing temperatures to a rather sharp maximum, then more slowly decreasing again. For a discussion on the effects of air temperature, source and sink relationships, and ageing of leaves on net photosynthesis, see BIERHUIZEN (1970). The composite effect of all this can be summarized by the introduction of a reduction factor (α_{ph}). Introduction of S_c and α_{ph} in eq. (98) then yields for the actual photosynthetic flux (RIJTEMA and ENDRÖDI, 1970) with the stipulation of an optimum nutrient supply:

$$\bar{P} = 490/(r_a + 1.8 r_s + 440)\, \alpha_{ph} S_c \bar{P}_{pot} \qquad (\mathrm{kg.m^{-2}.s^{-1}}) \qquad (99)$$

where r_s, still determined from evaporation data, now only consists of the sum of the first two resistances (r_l) and (r_h) of eq. (25), as the effect of partial soil cover r_c is incorporated into S_c.

Various authors calculated α_{ph} or have given data from which it can be derived. So for instance SIBMA (1968), who analyzed a number of agricultural crops grown under optimum conditions of water and nutrient supply. He concluded that in the Netherlands a green crop surface from the time the soil is completely covered until maturity, produces on the average about 225 kg.ha^{-1}.day^{-1}. From fig. 25a as given in DE WIT (1965) it can be derived that for the main growing period from May to October the potential production rate is about 320 kg.ha^{-1}.day^{-1}. This means for Sibma's data an α_{ph} factor of about 0.7. RIJTEMA and ENDRÖDI (1970) calculated for potatoes an α_{ph} of 0.68. GAASTRA (1963), quoting data from field experiments by Thomas and Hill (1949), reports average daily α_{ph} values of alfalfa varying from 0.51 to 0.65 and for sugar beet varying from 0.67 to 0.71.

As discussed in section IIIB-3d-2 low radiation fluxes cause closing of the stomata (r_l increases) and hence both evaporation and photosynthesis are reduced. The same holds when the stomata close as a result of a low water head in the leaves, resulting in an increase of r_h. For extensive discussions on the influence of water potential on photosynthesis the reader is referred to SLATYER (1967, 1970), COWAN and MILTHORPE (1968), CRAFTS (1968) and KRAMER (1969). Generally, the influence of closing of stomata will be larger on evaporation than on photosynthesis, because CO_2 diffusion is controlled by the additional mesophyll resistance, r_m' (see e.g. GAASTRA, 1963), which has a large value (440 s.m^{-1} for the standard crop of DE WIT, 1965).

With the use of eq. (99), the maximum production rates of the crops red cabbage, dwarf French beans and celery were calculated. These rates were compared with the dry matter production rates obtained from periodical harvests. The relationships between real and calculated production rates were analyzed by linear regression of measured on calcateuld production, from which the reduction factors α_{ph} were derived.

It should be noted that the calculations refer to total dry matter production (shoots plus roots). Therefore, at the periodical harvests an estimation was made of the shoot/root ratio and after conclusion of the experiments the change of this ratio with time was determined. The ratios were estimated on the sandy loam where the roots could easily be clean-harvested. The ratios for the other profiles were approximated from the sandy loam data taking into account the measured differences in rooting depths. For red cabbage on sandy loam a constant (shoot plus root)/(shoot) ratio of 1.11 proved to be valid during the growing season. The same ratio did hold for the cabbage grown on clay on sandy loam, which crop had an effective rooting depth of 0.40 m similar to that in the sandy loam. In accordance with the larger effective rooting depth in the clay profile (0.60 m), a ratio of 1.13 was here adopted. For dwarf French beans with rooting depths about equal on all profiles (0.20 m), a (shoot plus root)/(shoot) ratio of 1.08 was found. For celery the rooting depth was also equal in the three profiles: about 0.25 m. As various cuttings were made the shoot/root ratio varied during the growing season. However, it appeared that total root production was about 24% of total shoot production.

The results of the calculations for the three crops, as presented in fig. 51, will now be discussed successively.

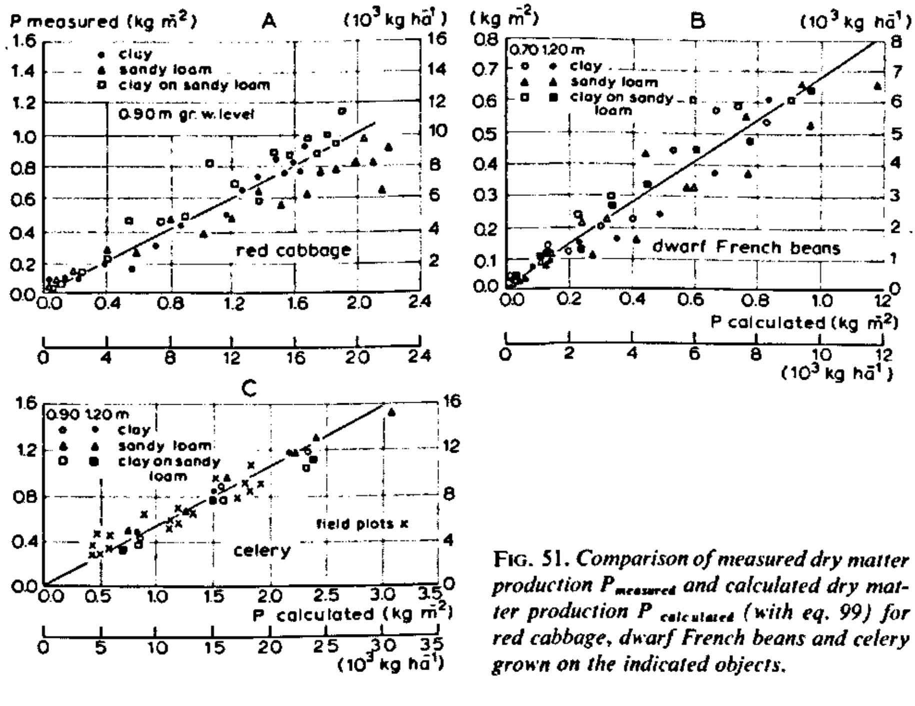


FIG. 51. *Comparison of measured dry matter production $P_{measured}$ and calculated dry matter production $P_{calculated}$ (with eq. 99) for red cabbage, dwarf French beans and celery grown on the indicated objects.*

b. CALCULATION OF REAL PRODUCTION

Red cabbage. In order not to destroy too many plants with the weekly harvests, each time only one plant of each object was harvested. With a heterogeneous crop like cabbage, a relatively large variation in dry matter production is then to be expected. This variation is reflected in the scatter of the yield data in fig. 51A. The obtained reduction factors α_{ph} of the single as well as of the combined objects, are summarized in table 16.

The calculated α_{ph} values are lower than those found by the earlier mentioned investigators for other crops. This may be due to the following reasons. From fig. 51A it is seen that in the beginning period the production rate of the cabbage on clay is delayed as compared with the other profiles. In this period the evaporation of the cabbage on clay was reduced because of lack of water and this consequently led to the development of an internal water deficit and an increase in the surface resistance r_s. ČATSKY (1965) found that a decrease in water potential was directly proportional to the resulting decrease in photosynthetic rate. He also remarked that this slowing down of photosynthesis developed faster in older than in younger leaves.

TABLE 16. Slope (α_{ph}), correlation coefficient (r) and standard deviation (0.1 kg.m^{-2}) from regression ($S_{y,x}$ and $S_{\alpha ph}$) for regression of measured production (P) on calculated production ($P_{calc.}$) of red cabbage grown on the three profiles of the 0.90 m groundwater plot (1967)

Profile	α_{ph}	r	$S_{y,x}$	$S_{\alpha ph}$
Clay	0.51	0.99	0.543	0.012
Sandy loam	0.40	0.97	0.858	0.014
Clay on sandy loam	0.56	0.98	0.875	0.017
Pooled objects*	0.51	0.96	0.950	0.012

* without the last 6 data of fig. 2A for the sandy loam

For cabbage grown on sandy loam the contradictory is true. The development is fast in the beginning, but slows down around the end of August and almost ceases in autumn (fig. 51A). This feature is also mentioned by VAN DER VALK and NICOLAI (1969) for red winter cabbage and by VAN DER VALK and SCHONEVELD (1963) for savoy cabbage. They report for instance that the exterior leaves of the latter crop went yellow-green to yellow on the sandy loam profiles, but remained green on the clay profiles. The yellowing of leaves (first seen in older leaves) is a condition associated with severe proteolysis without concurrent synthesis, and the decrease in chlorophyll content is a clear visible symptom of nitrogen deficiency (STEWARD, 1963), The high rate of photosynthesis of red cabbage on sandy loam in the early stages of growth is accompanied by an uptake of considerable quantities of nitrogen. This in combination with the leaching processes during wet periods, makes that available nitrogen on the sandy loam profiles is soon lacking. The effects of N-deficiency are premature ripening of the heads and earlier ageing of the leaves. Similar effects were reported for early cauliflower grown in dry periods on clay soils (VAN DER VALK and SCHONEVELD, 1963). The higher α_{ph} values found for cabbage grown on clay on sandy loam are probably due to the combination of a better nitrogen availability in the top soil with a favourable water availability in the subsoil.

Another important reason for the relatively low α_{ph} factors found for red cabbage may have been an air deficiency in the later stages of growth. Heavy rains in autumn caused the groundwater tables to rise above the, rather deeply lying, zones with largest root activity (see IIIB-3d-4). The profiles of the 0.90 m groundwater plot are, depending on the meteorological conditions, both sensible for desiccation as well as for water excess. This feature will be discussed in section d of this Chapter.

Dwarf French beans. For this crop an average reduction factor (α_{ph}) of 0.67 was found for the combined objects (fig. 51B). This value is in agreement with the previous reported literature data. This indicates that the environmental circumstances were fairly optimum, which was the more easily the case as it is known for dwarf French beans that in the stages before flowering the soil in the root zone can dry out to -1.4 bar (pF 3.6) without the crop showing any reduction in final yield. During the flowering and further growth stages of beans, a slight desiccation of the soil strongly reduces yields (see FEDDES, 1969b). The desiccation limit for these stages appears to be -0.2 bar (pF 2.3). Because of the frequent rainfall occurring in 1968 during and after the flowering stage, the matric pressure limit of -0.2 bar was not reached. Under these wet conditions it could be expected that poor aeration would have damaged the crop and in this way have reduced production. From measurements of O_2 and CO_2 concentrations at depths of 0.125, 0.25 and 0.40 m on August 19, 1968 in the clay and sandy loam of the 1.20 m groundwater plot, it appeared that the lowest O_2 concentration measured on the sandy loam was 11.5% (CO_2 value 4.3%). This low O_2 concentration was caused by ponding after a sequence of heavy rains (see fig. 31). This could have been the cause of the low α_{ph} value of 0.52 found for the 1.20 m sandy loam (fig. 51B). DASBERG and BAKKER (1971), however, have shown in pot experiments that reducing the O_2 concentrations at the soil surface from 21 to 11% had no significant reducing effect on bean growth. On the other hand the authors pointed out that dry matter production was strongly affected by dry bulk density. From density measurements of the top layers at the experimental field it could be shown that the ρ_d values of the 1.20 m sandy loam were higher than those of the 0.70 m plot, which may explain the low α_{ph} value found for the 1.20 m sandy loam.

Celery. The results of the regression analysis for celery are presented in table 17 and fig. 51C. For the growing season as a whole for the combined objects a mean α_{ph} of 0.52 was obtained. The standard deviation from regression $S_{y,x}$ (0.1 kg.m^{-2}) was about twice as high for the fields as for the lysimeters. This was caused by the very irregular soil coverage on the field plots, which feature was already discussed in section IIIB-3d-5.

TABLE 17. Results of the regression analysis for celery grown on the three profiles of the 0.70 m and 1.20 m groundwater plots (1969). The figures in brackets are values valid after a recovery period of 1 to 2 weeks after the cut

	α_{ph}	r	$S_{y,x}$	$S_{\alpha ph}$
6 lysimeters	0.51	0.99	0.654	0.009
6 field plots	0.53	0.94	1.049	0.020
12 pooled objects	0.52	0.97	0.867	0.009
1st cut lysimeters	0.54	0.91	0.473	0.0221
2nd cut lysimeters	0.53 (0.61)	0.86 (0.86)	0.263 (0.256)	0.0131 (0.0148)
3rd cut lysimeters	0.42 (0.48)	0.57 (0.85)	0.306 (0.199)	0.0159 (0.0119)

The low reduction factor α_{ph} can be partly explained by considering the decreasing α_{ph} values of the separate cuttings (see table 17, 1st, 2nd and 3rd cut lysimeters).

Just as red cabbage, the celery on the sandy loam objects, contrary to celery on the other profiles, showed nitrogen deficiency symptoms (yellowing of the leaves) during the very wet second part of August 1969. Yellowing occurred as well on the clay as on the clay on sandy loam high groundwater plots. In the beginning of October, the leaves of celery on all profiles of all plots became yellow, resulting in a low α_{ph} value of 0.42. The final effect was that the overall α_{ph} value was low.

The yield calculations were performed for maximum production over the total growing period. However, during the periods just after cutting, some time (about 1 to 2 weeks) is needed for the formation of new leaves and photosynthesis in this period must be very low. If these periods are neglected in the calculations, as is scientifically correct, the α_{ph} of the second cut increases from 0.53 to 0.61, and the α_{ph} of the third cut increases from 0.42 to 0.48 (table 17).

It must be emphasized that to obtain real production the calculations ought to embrace the whole growing season.

c. CALCULATION OF WATER USE EFFICIENCY

Under water limiting conditions it is important for a grower to know what is the minimum amount of water needed to ensure a maximum production of a certain crop, i.e. to know the efficiency of water use. As a measure of efficiency the amount of water used per unit of dry matter produced (kg.kg^{-1}), called the 'transpiration ratio' (E/P), generally is taken. This ratio will give within a certain climatic region a reasonable idea of the water requirement of a crop. Application of the ratio obtained for this crop to other climatic regions gives poor results, however. DE WIT (1958) has pointed out that the differences in radiation (which determines both evaporation and photosynthesis) between different climatic regions must be taken into account. In first instance he concluded that in arid regions, where shortwave solar radiation flux is high and not restricting evaporation, E/P would be more or less proportional to the mean evaporation flux of open water $\bar{E}_0$ (m.day^{-1}) over the growing season. However, as the scatter in the data remained large De Wit showed that, based on a proper statistical evaluation of the data, not E/P should be plotted against $\bar{E}_0$, but P should be plotted against $E/\bar{E}_0$ (m.m^{-1}.day). For low radiation fluxes as occurring in temperate climates, he found that E/P remains about constant with radiation and that for these regions the length of the growing season should be included by plotting P against total amount of evaporated water during that period E (m). De Wit also showed that these (linear) relationships are hardly affected by small variations in water potential and nutrient level, because the resulting changes in increase of leaf area affect both evaporation and photosynthesis in the same way.

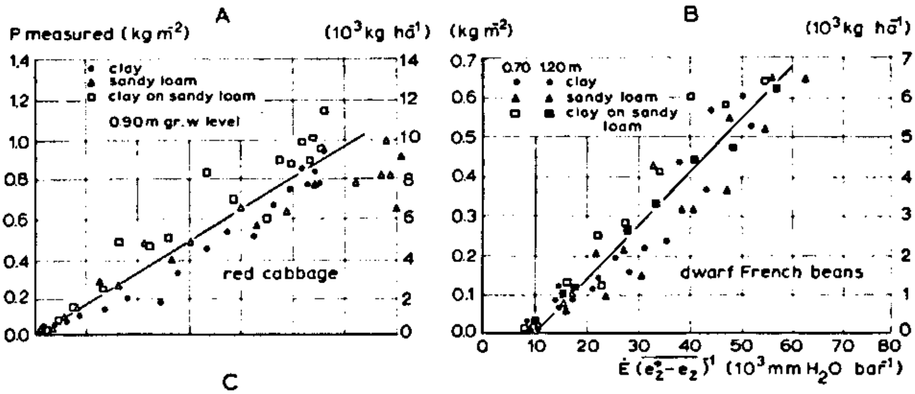
A review of the relationships between nutrition, production and water use efficiency has been given by VIETS (1962). Improvement of the E/P ratio by reducing E through application of antitranspirants has been reviewed by GALE and HAGAN (1966).

BIERHUIZEN and SLATYER (1965) have shown that evaporation from leaves is finally determined by the mean vapour pressure deficit and that net photosynthesis is finally determined by the mean carbon dioxide concentration difference. In the field, where more or less constant CO_2 concentrations prevail (see BROWN and ROSENBERG, 1970), the observed variation in E/P between different climatic regions then will be mainly due to differences in mean vapour pressure deficit $(\overline{e_z^* - e_z})$, rather than to radiation differences as mentioned by De Wit.

In the present study the procedure of Bierhuizen and Slatyer was applied to red cabbage, dwarf French beans and celery, but instead of plotting E/P against $(\overline{e_z^* - e_z})$, production P was plotted against $E/(\overline{e_z^* - e_z})$. Because interception of precipitation reduces evaporation of the crop this effect was accounted for by the approximation (see RIJTEMA and ENDRÖDI, 1970):

$$\dot{E} = \frac{E - E_i}{E^* - E_i} E^* \qquad (\mathrm{kg.m^{-2}.s^{-1}}) \tag{100}$$

where $\dot{E}$ is the evaporation flux of a non-wetted crop surface. The results of the calculations are shown in fig. 52. The scatter of the data for red cabbage is mainly caused by the periodical harvests of only one plant per week. It is seen from fig. 52A that the ceasing of growth on the sandy loam is not a result of water deficiency, but of another factor (e.g. nitrogen deficiency) which limits growth. For the cabbage grown on sandy loam production is limited at about 103. 10^3 mm $H_2O.bar^{-1}$. On the other profiles production was not yet limited. The same was true for dwarf French beans and celery. The slopes of the lines in fig. 52 give the water use efficiency expressed in kg.bar.mm H_2O^{-1}. It follows that of the three crops the water use efficiency of dwarf French beans is highest, followed by celery and then red cabbage.



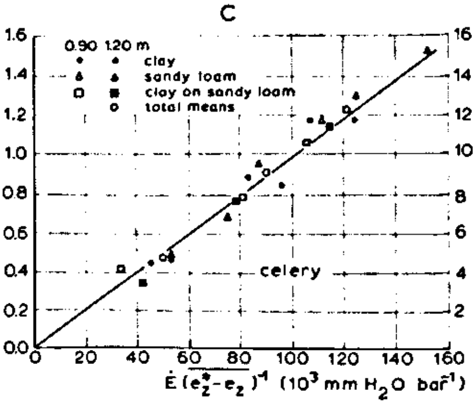


FIG. 52. *Measured dry matter production $P_{measured}$ versus the ratio evaporation over mean vapour pressure deficit for red cabbage, dwarf French beans and celery grown on the indicated objects.*

d. INFLUENCE OF GROUNDWATER TABLE DEPTH

1. Optimum amount of available water

Much is known about the influence of groundwater table depth on production of different crops. The results of the investigations differ widely from each other, however. This is partly due to differences in experimental procedures, but undoubtedly the main reason is the variation in environmental conditions, like weather regime and soil type for instance. Extensive reviews on influence of groundwater depth have been given by WESSELING and VAN WIJK (1957), VAN 'T WOUDT and HAGAN (1957), and VAN DER MOLEN (1969).

VISSER (1958) presents for seven soil types the average relative yields which are obtained at average groundwater table depths occurring during the growing season. Yield depressions at high water tables are due to lack of aeration, yield depressions at deep water tables are due to lack of water. The optima of the curves shift to higher water tables in dry years and to deeper water tables in wet years. The effects of high water tables on aeration have been discussed by WESSELING (1957), while the effects on crop yield have been treated by SIEBEN (1964). The harmful influence of flooding upon yield of grass has been dealt with by BAKKER (1967).

In the Netherlands a number of groundwater level experimental fields have been established to investigate the influence of constant groundwater tables upon yield of crops. For arable crops the reader is referred to HOOGHOUDT (1952) and VAN HOORN (1959), for grass to MINDERHOUD (1960) as well as HOOGERKAMP and WOLDRING (1965), and for fruit trees to SEGEREN and VISSER (1969).

Because of changes in weather conditions the relationship between yield and groundwater level differs from year to year and it seems of little use to make this comparison for years on end.

It was shown that for a certain production of dry matter a certain amount of water has to be evaporated. If one knows the maximum amount of water that has to be evaporated to ensure maximum production, one can consider if, under the prevailing circumstances, this required amount of water is available. The availability of water in any growing period is partly determined by the amount of precipitation and partly by the amount of water that can be delivered by the soil.

Now, as mentioned in section IIIA-6 fresh yield of most horticultural crops is already being reduced at water pressures lower than -0.4 bar (pF 2.6), which means that the amount of soil water available for evaporation should be calculated taking this limit into account. Doing this, the amounts of water available in the soil at different groundwater table depths were calculated for different lengths of growing period (table 5).

The hydrological properties of the three profiles were determined on the 0.90 m plots. It appeared, however, that the composition of the sandy loam layer in the clay on sandy loam plot did show a local aberration from the composition elsewhere on the field. Therefore the hydrological properties of the 'normal' sandy loam were used in the calculations for the other clay on sandy loam profiles.

Adding to the amounts of water available in the soil the amounts of fallen rain, the total amount of water available for evaporation during the growing period is obtained.

For red cabbage the fresh yields measured on all profiles of all groundwater plots were compared with the total amounts of water available at each object (fig. 53). Fresh yields instead of dry matter yields were used as a measure for production, because they are of most interest for the market gardener. It should be remembered, however, that it is from a scientific point of view better to compare dry matter production with the amount of available water, since at the high groundwater table plots often a lower dry matter percentage was found than at deeper water tables.

In this type of relationships the scatter of the data may be expected to be rather large, because the distribution of rainfall over the period in question is not included in the parameter of total amount of available water. So for instance in the first half of the growing period of red cabbage (June 21 to August 29, 1967) total rainfall was only 58 mm. A lesser amount falls in this period on the average only once in 5 to 10 years. This means that this particular period can be considered to have been very dry. In the second half of the period (August 30 to October 31) total rainfall was 220 mm, which amount is in this period on the average exceeded only once in 10 years and this period can be considered to have been very wet. From such amounts undoubtedly surplus rainfall will have been drained away and will not have become available for evaporation.

Such features are reflected in the circled points of fig. 53, which all concern the 0.90 m groundwater plot. Slight changes in the 0.90 m groundwater table depth induced either dry or wet symptoms; compare table 5 where it is seen that with relatively small changes in the 0.90 m groundwater table depth the amounts of available water change considerably.

From fig. 53A it can be seen that for a total amount of available water of about 365 mm a maximum fresh yield production is obtained. Left of this maximum, production is reduced by water shortage and to the right excess of water reduces production. Fresh yield production in relation to mean groundwater table depth is shown in fig. 54. The scatter in the data in this graph is uncomfortably large, and from such graphs optimum groundwater table depths can only be estimated very roughly.

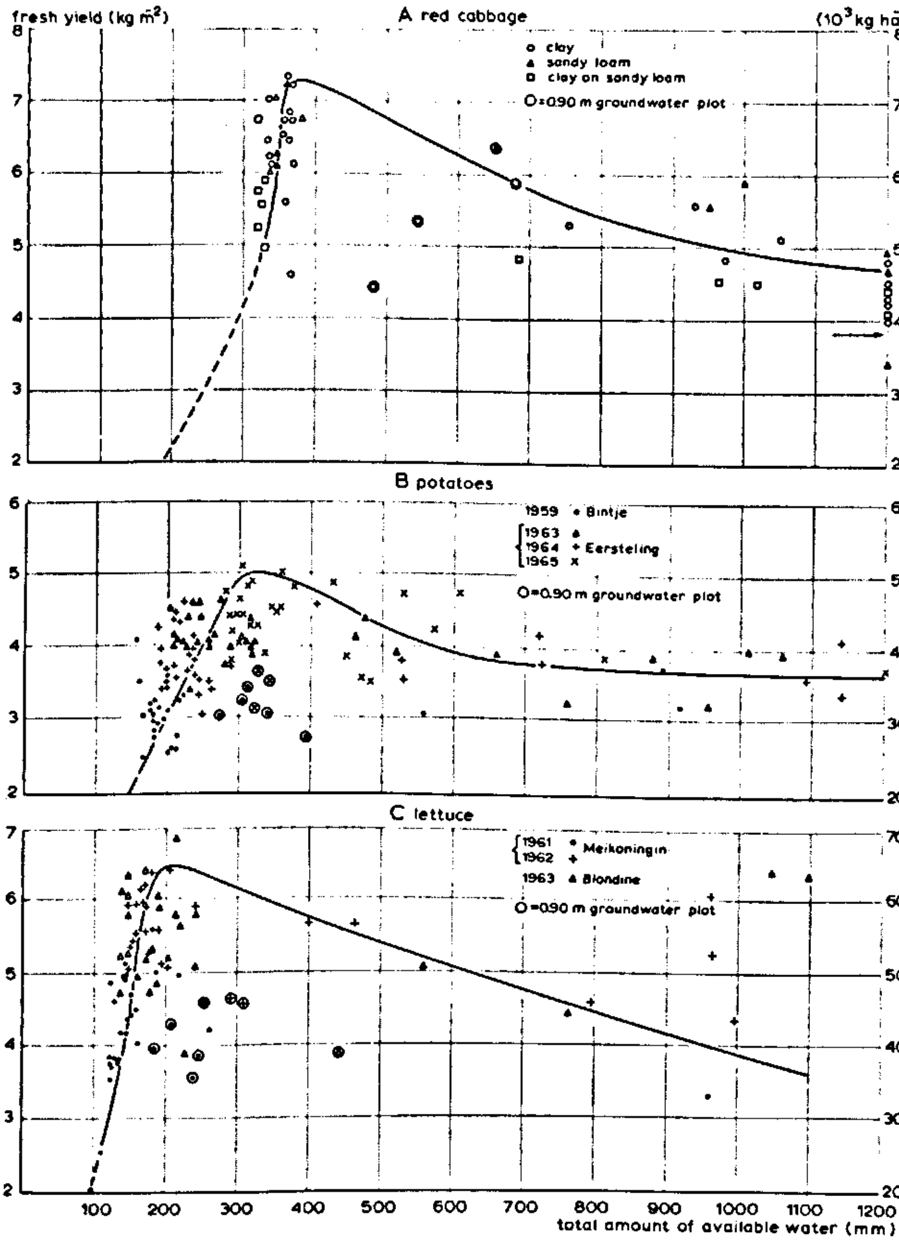


FIG. 53. *Dependence of fresh yield of red cabbage, potatoes and lettuce, as measured on the three profiles of all groundwater plots, on total amount of available water.*

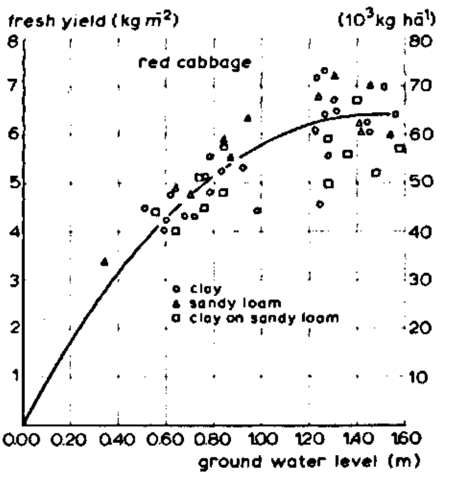


FIG. 54. *Dependence of fresh yield of red cabbage, as measured on the three profiles of all groundwater plots, on mean groundwater table depth over the growing season.*

VAN DER VALK and NICOLAI (1969) have published curves of yield versus mean groundwater table depth for potatoes grown on the experimental field in 1959 (Bintje) and in 1963 to 1965 (Eersteling). It appeared that in the extremely dry year of 1959 on the clay and the clay on sandy loam maximum yields were obtained at a water table depth of about 0.35 to 0.40 m below surface. For the sandy loam maximum yields were obtained at a water table depth of about 0.70 m below surface. For 1963 the maxima shifted to deeper water tables: for the clay and clay on sandy loam to about 0.80 m and for the sandy loam to about 1.05 m below surface. In fig. 53B fresh tuber yields of potatoes are plotted against the total amounts of available water. The scatter in the data is partly due to the few measured groundwater table depths on which the calculated amounts of available water are based. As has been done in fig. 53A the deviating yields of the 0.90 m groundwater plot are circled. The previously mentioned authors have commented that the regular increase of yields on the sandy loam occurring during the consecutive years, may have been caused by a gradual improvement of the nutrient status of the profile.

The scatter in the data given in fig. 53B was not improved by plotting dry matter production against the ratio of total available water to mean open water evaporation. Nor was an improvement obtained by plotting relative production (P/P_{pot}) versus the total amount of available water.

As an estimate, it seems from fig. 53B that about 310 mm water is needed for maximum potato production under the prevailing climatic conditions. This agrees with the water use necessary to obtain maximum dry matter production as found from seven years of experimental data (RIJTEMA and ENDRÖDI, 1970). VAN DUIN and SCHOLTE UBING (1955) found from field experiments with potatoes grown on sandy soil over the period 1940 to 1950, that no depression in yield occurred when total potential evaporation minus total available water over the entire growing season lay between 0–100 mm. This corresponds for the climatic conditions in the Netherlands with a total available amount of water lying between potential evaporation of the crop (E_p) and 0.75 E_p, where E_p was calculated according to PENMAN and SCHOFIELD (1951). It can be derived from correlation of cumulative sunken pan evaporation at the experimental field, with calculated cumulative open water evaporation at Den Helder, that the average total sunken pan evaporation during the growing seasons of the years 1959 and 1963 to 1965 was about 407 mm. If potential evaporation over the growing period is taken to be equal to open water evaporation, then 0.75 E_p = 305 mm which is rather close to the value of 310 mm derived from fig. 53B.

Finally the experimental data of VAN DER VALK and NICOLAI (1969) concerning lettuce grown in 1961 and 1962 (*Meikoningin*) and in 1963 (*Blondine*) were analyzed, the results are shown in fig. 53C. Taking into account the remarks made above on the scatter of the data, it is seen that maximum yields of lettuce occur at about 200 mm of total available water.

2. Lowest admissible water table depth

Having obtained the optimum amounts of available water for the crops, the maximum groundwater table depth over a period in which the critical value of the total amount of rainfall is known, can be calculated for each crop and each profile. This was done with the aid of the frequency distribution over 103 years of the rainfall depth of k consecutive days at Den Helder (KNMI tables, 1958). For the treatment of precipitation frequencies, see STOL (1965/66).

For red cabbage the beginning of the growing season was taken at June 1 and the growing period was assumed to be 130 days (k). For potatoes these data were April 16 and k 90 days and for lettuce April 1 and k 60 days respectively. The total rainfall amount which in a period of k consecutive days is on the average once in j years less than a critical value χ, can be derived from the mentioned KNMI tables. Subtracting the critical value of rainfall amount from the total optimum amount of available water, yields the amount of available soil moisture required for maximum production. Then with the aid of table 5 the lowest groundwater table depth to be admitted for maximum production can be obtained by means of interpolation. The results are presented in table 18.

A striking feature of table 18 is that per profile, despite differences in crops, in lengths of growing periods and in frequency distributions of amount of rainfall, the deepest admissable water table depths do not differ very much. The reason for this is that the amounts of available water change considerably with slight changes in groundwater table depths around 0.90 m below surface.

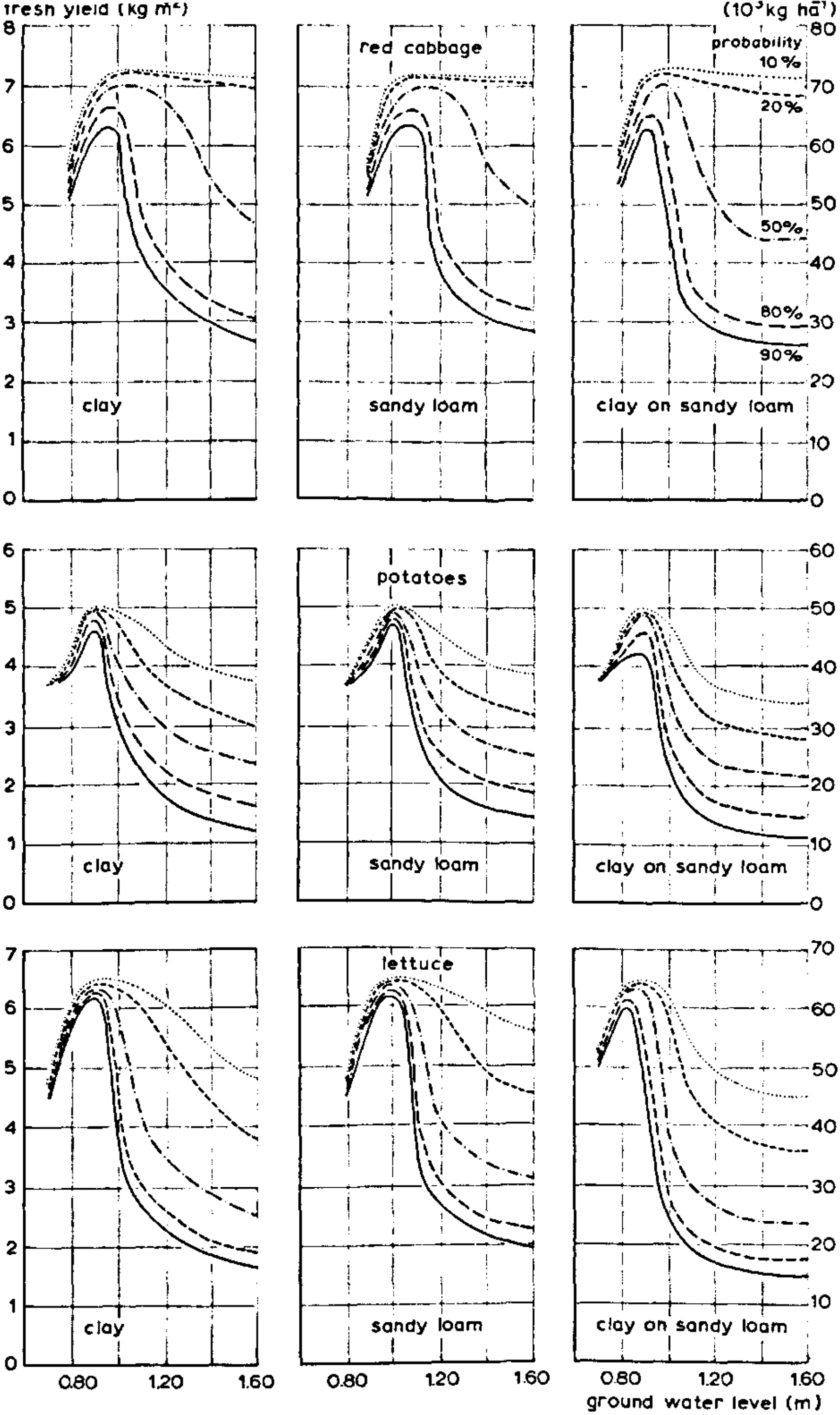
the aid of available rainfall data, of the last hundred years say, the total amounts of water which per profile were available to the crop in each year at each groundwater table depth. Knowing these data and using fig. 53, estimates of the yields which in theory would have been obtained in each year on each object can be made. From these theoretical production data over the last 100 years, mean values can be taken, and the optimum water table depth for a specific crop grown on a specific profile could have been derived.

To avoid this time-consuming procedure, another line of attack was taken, which can be shortly summarized as follows:

- The amount of soil water available for a crop in a certain profile at a certain groundwater table depth during a fixed growing period is known from calculations as mentioned in section IIIA-6 (see also table 5). From fig. 53 the yield corresponding with this amount of water can be obtained, this being the yield in a situation of zero rainfall.
- The interval on the ordinate of fig. 53 between this yield and maximum yield is divided in (preferably) equal parts, giving a number of yield classes.
- For each of these yield classes, (one or two) corresponding classes of total amounts of available water can be read from fig. 53. By subtracting the known amount of available soil water from the lower and upper total amounts of available water, the (one or two) classes of additional water to be supplied by rain, are obtained.
- With the aid of the rainfall depth frequency distribution of k consecutive days the frequencies of occurrence of a rainfall amount within the specific classes of additional water, can be obtained.
- These frequencies are the frequencies of occurrence of yields within the corresponding yield classes.
- By adding the frequencies of the various yields within the specific yield classes, the cumulative relative frequency curve of yield for each groundwater table depth can be obtained.
- By application of horizontal sectioning of this family of curves, the relation between yield and groundwater table depth can be given for specific frequencies of exceedance.
- When, for similar environmental conditions, these curves are used for prediction purposes the frequencies can be taken to be probabilities of exceedance.

Some of the results of this procedure are presented in fig. 55, where curves for several probability levels of exceedance are drawn.

FIG. 55. *Dependence of fresh yield of red cabbage, potatoes and lettuce grown on clay, sandy loam and clay on sandy loam respectively, on groundwater table depth over the growing season at five probability levels of exceedance.*


fresh yield (kg m⁻²)
(10³kg ha⁻¹)
red cabbage
probability
10%
20%
50%
80%
90%
clay
sandy loam
clay on sandy loam
potatoes
clay
sandy loam
clay on sandy loam
lettuce
clay
sandy loam
clay on sandy loam
0.80
1.20
1.60
ground water level (m)

It can for example be seen that for potatoes grown on clay on sandy loam at a groundwater table depth of 1.20 m over a period of 90 days starting from April 16, a yield of 3.1 $kg.m^{-2}$ (31 $ton.ha^{-1}$) or more can be expected with a probability of 20%, or once in 5 years. It also appears that there is a probability of 50%, or once in 2 years, that at a groundwater table depth of 0.90 m the yield will be 4.9 $kg.m^{-2}$ (49 $ton.ha^{-1}$) or more. This depth is the optimum water table depth for all probability levels for potatoes grown on the clay on sandy loam profile. The curves of these probability levels are running almost horizontally at groundwater table depths larger than 1.20 m, indicating that below 1.20 m marked differences in yields are not to be expected.

The other given results show that the optimum water table depth for red cabbage on clay is about 1.00 to 1.10 m, on sandy loam 1.10 to 1.20 m and on clay on sandy loam 0.90 to 1.00 m; for potatoes on clay (and clay on sandy loam) 0.90 m and on sandy loam 1.00 m; for lettuce on clay 0.90 m, on sandy loam 1.00 m and on clay on sandy loam 0.80 to 0.90 m.

Looking for the same crop at for example the 50% probability curves, it is clear that at deeper groundwater tables, the sandy loam profile carries less risk in production than the clay profile. This last profile in its turn, carries less risk (for red cabbage much less risk) than the clay on sandy loam profile.

From the slopes of the curves towards deeper groundwater table depths, it is seen that the clay on sandy loam profile is more susceptible for water table drawdowns below the optimum level than the clay and the sandy loam profile.

The sharper the top of the probability curves, the more difficult the choice of the groundwater depth to be taken will be. In general, taking into account the irregularities of rainfall distribution and the shape of the curves around their top, it is to be advised to maintain groundwater levels at depths that are somewhat lower than those corresponding with the top of the probability curves (see for example the 50% probability curve of red cabbage on clay).

It is thought that with this approach there is a possibility to solve the problem of determining the total amount of water required for maximum crop production and, where relevant, to decide on the optimum groundwater table depth then necessary.

SUMMARY AND CONCLUSIONS

To a large extent the results of a farmer's efforts to get higher crop yields will be determined by the prevailing environmental conditions, i.e. by the existing complex of physical, chemical and biological factors. The possibilities of an efficient use of these factors are enlarged by our knowledge of their effects on agricultural production. Such knowledge is in first instance mostly gained from practical experience or field experiments often giving, however, only empirical information of local value.

To apply local results with success to analogous problems in other regions a fundamental treatment of research data is necessary, by which the occurring phenomena and processes can be analyzed more objectively.

In this paper attention has mainly been paid to the effects of the physical environmental conditions on plant development. As local research object, part of the economically important horticultural region of Geestmerambacht was taken, situated in the Netherlands in the northern part of the province of North-Holland.

This area is partly covered with heavy sea clay soils under insufficient water management, resulting in a one sided cropping pattern. This together with a fragmented parcelling and a poor accessibility makes cultivation of profitable intensively managed crops difficult.

To investigate, in connection with an integrated land consolidation program for the region, the possibilities of improving the soil profile and the water management a groundwater level experimental field was laid out (Chapter II). On the experimental field ten different constant groundwater table depths were realized in three soil profiles, namely the original heavy clay soil and two improved clay soils: a sandy loam and a clay on sandy loam. Previous investigations on this field had the purpose of finding relationships between yield of a number of vegetable crops and mean optimum groundwater table depth during the growing season. Due to changes in weather conditions the results differed from year to year. It was therefore decided to apply a more fundamental treatment to the encountered problems by considering by means of mass and energy balance approaches what happens to the mass and energy fluxes in the soil-plant-atmosphere system.

Because the experimental field did not lend itself to water balance studies, a special type of non-weighable lysimeter was developed in which the same water table could be maintained as in the surrounding field, including the fluctuations. Lysimeters were installed in the profiles of the 0.90 m and 1.20 m groundwater plots (Chapter II).

Before dealing with the influence of physical conditions on crop growth, attention was paid to the evaluation of the physical properties of the system by which the transport processes are determined. Chapter III deals with the factor water, Chapter IV with the factor heat.

In Chapter IIIA the transport of water in the unsaturated zone has been treated. To describe the state of water in the soil the energy concept has been used and the various terms involved have been discussed. The soil moisture retention curves were determined in the laboratory. Changes of soil moisture content with depth and time in the field were followed mainly by the gamma transmission method as well as by sampling with undisturbed soil cores. With the gamma transmission method an accuracy in the range of 0.5 vol.% could be obtained by expressing the mass densities of the soil components as an equivalent 'electron' density of water and by applying corrections for distance deviations and non-parallelism of the gamma access tubes. Because of large variations in the dry bulk density of the various layers, a relationship between moisture content, matric pressure and dry bulk density was developed, from which moisture retention curves at various dry bulk densities could be derived. This provided a method to be less dependent on the differences in dry bulk densities between the individual samples, especially in the gravimetric sampling procedure.

Hydraulic conductivity data were determined both from laboratory experiments as well as from measurements in the field in dry periods. The results compared favourably well with each other. Having evaluated the hydrological properties of the soil, relationships between flux, matric head and height above the groundwater table could be given. Also the influence of surface evaporation on drawdown of the groundwater table, assuming certain boundary conditions, was theoretically evaluated. It appeared that the calculated groundwater depths agreed reasonably well with groundwater depths found in similar soils in arid zones. Moreover it could be shown that the evaporation flux decreased roughly with the square root of time, as reported in literature for different boundary conditions.

To obtain the amount of moisture available for evaporation over a certain growing period, the various quantities of moisture available in and coming available from below the root zone and from capillary rise from the groundwater table were calculated. The calculations were performed taking into account the admissable pressure at which soil water begins to limit plant growth. As appeared from literature studies by the present author, this value is for most vegetable crops about −0.4 bar (pF 2.6).

In Chapter IIIB the transport of water to the atmosphere has been treated. Evaporation was calculated by means of a combination method based on rather easily measurable meteorological as well as soil and crop factors. The calculations were compared with data obtained from water balance studies. For the aerodynamic resistance of the crop, values were calculated for various crop heights and wind velocities.

As net radiation is one of the most important terms determining evaporation from a surface, and direct measurements of net radiation are often not available, generally empirical formulae are used. Some of these formulae were used in this study and the calculation results were compared by means of linear regression techniques with measured net radiation data. It appeared that for calculated shortwave radiation high correlations were obtained with various formulae. For thermal radiation, however, calculations over short periods of a week gave poor results. Therefore a few expressions based on the high correlationship between net and shortwave radiation were derived for various lengths of time and analyzed by linear regression. The relation obtained on a 24-hour basis compared strikingly well with similarly computed relations found in Western-Australia. Also equations computed on a daylight basis compared favourably well with literature data. It further appeared that for calculations of evaporation on a weekly basis, radiation data measured at relatively distant meteorological stations can sometimes be used with sufficient accuracy.

Reflection coefficients of various surfaces were determined from continuous measurements taken on all kind of days. A discussion was presented on the various methods determining this coefficient. Results of reflection coefficients at increasing fractions of soil covered for a number of vegetable crops were given. Except for spinach, they showed all a more or less similar pattern within a small band of data. It was shown that wetting the surface reduces it, both for bare soil as well as for a cropped surface.

Evaporation from bare soils with different groundwater tables was studied on a hourly basis by means of an energy balance approach on clear days in a rather wet and a rather dry period. Differences in the energy balance components between the various objects were established and calculated vapour pressures at the surface were compared with data obtained from sampling of the surface layer, giving a rather good agreement.

A discussion on the diffusion resistances encountered by evaporation from cropped surfaces has been presented. It appeared that the soil must be covered to about 70 to 80% before constant evaporation rates are obtained. If the soil is rather wet, however, variation in the area of soil covered has not much effect.

Measurements of interception of red cabbage agreed with literature data concerning interception of grass. It was shown that interception is important in periods of reduced evaporation, as it then influences evaporation most.

An analysis of the transport resistance for liquid flow in the plant as well as an investigation on the geometry factor of the root system for red cabbage was presented. The variation of these factors with depth could be shown, and root extraction rates at different depths were calculated and compared with data obtained from water balance studies. The plant resistance data were in reasonable agreement with literature data. Because of a non-homogeneous and poor root development in the early stages of growth, the geometry data of the initial growing stages differed a factor ten from data found in literature. After root development increased with depth, geometry data decreased to values also reported for other crops.

From comparison of evaporation calculated by the combination method and evaporation obtained from water balance studies, satisfying results were obtained. The problems encountered in the first growing stages with seedling emergence and initial growth were emphasized. It appeared from the calculations that except for a few periods during the first growing stages only a small reduction in evaporation by water shortages under the prevailing circumstances was present.

Chapter IV deals with the flow of heat in soil. To evaluate the thermal properties of the soil, calculations were carried out on measurements taken in the laboratory as well as in the field. The various determination methods as found in literature were discussed. Thermal capacity was calculated by considering it as a function of volume fractions and specific heat capacities of the various components.

Calculations of thermal conductivity based on the assumption that soil particles can be considered as spheroids gave poor results, possibly because of a wrong choice of the depolarization factor. Measurements of thermal conductivity in the laboratory with the transient needle method yielded good results for sandy loam. For the shrinking and swelling clay, however, a large scatter in the data occurred because of change in dry bulk density which could hardly be corrected for. Therefore preference was given to measure thermal conductivity in the soil in situ.

Thermal diffusivity of the sandy loam could be deduced from laboratory data. Determination of thermal diffusivity in the field from the analysis of temperature measurements at various depths, by considering the amplitude and phase relationships with depth, failed, indicating that the thermal properties did vary with depth. To establish functional relationships of thermal diffusivity with depth, an electric analog as well as a numerical approach, both based on the principle that a discontinuous field can approach the continuous thermal field by expanding the partial differential equation into a set of finite difference equations, was used.

The scatter in the data obtained with the electric analog method was large due to various reasons like for example recording and reading errors from the temperature recorders, relatively widely spaced temperature observations at the larger depths and heterogeneity of the soils. With the numerical approach also disappointing results were obtained. Improvement of recording and reading errors by replacing measured temperature data by reconstructed smoothed temperatures computed from Fourier analysis, did not give better results. Improvement of the error inherent in discretizing the continuous temperature field to a discontinuous field by application of interpolation by means of spline functions, neither did. From extensive calculations on a possible disturbing influence of water transport in the vapour phase as induced by thermal gradients, it could be shown that this process could be neglected as a possible explanation of the scatter in the thermal diffusivity data.

The best method to determine thermal diffusivity proved to be deriving thermal capacity from soil sample data, measuring thermal conductivity with the transient needle method in situ, and estimating thermal diffusivity at each depth by considering the ratio of thermal conductivity and capacity. It was found that at higher groundwater levels thermal diffusivity is generally lower, especially in the top 0.10 to 0.15 m layer, indicating lower soil temperatures at higher groundwater levels. From temperature measurements it appeared that the mean daily temperatures of the plots with the higher groundwater tables were 1 to 2°C lower than the temperature of the plots with the deeper groundwater tables. With the same groundwater depth, clay proved to be warmer than sandy loam. It could be concluded that on the investigated soils the difference in groundwater level was playing a more important role than the difference in type of profile. The maxima and minima in the top soil were higher and the amplitudes decreased with depth faster in soils which had a deep groundwater table. The decrease was more marked in the clay than in the sandy loam soil.

In Chapter V the combined effects of water and heat on seedling emergence and crop production were treated. In the stage of germination and seedling emergence, the soil moisture content and soil temperature are the most important factors. In the stage from seedling emergence to maturity growth depends also on air temperature, but mainly on leaf area and net radiation.

The known effects of soil temperature and soil moisture content on germination and seedling emergence were reviewed. To relate temperature and emergence the heat sum concept was used. This approach is in practice frequently applied to schedule plantings, to predict maturity, to select crop varieties appropriate to different areas, etc. In such type of studies mostly the environmental air temperature is used. In the case of emergence of seeds it is advisable, however, to register the temperature in the direct environment of the seed, c.q. to measure soil temperature at sowing depth. The results of investigations on the influence of soil moisture content on germination and emergence as reported from laboratory experiments in literature, generally differ widely, mainly because of differences in applied experimental conditions as for example the use of a solution instead of soil as a germination medium.

The combined effect of soil temperature and moisture content on seedling emergence was studied with four different kinds of vegetable seeds in field experiments in a clay and a sandy loam profile, both with a shallow and a deep groundwater table. Various sowing dates were applied. It appeared that emergence was highly correlated with rainfall, and that because of the favourable hydrological properties of the soil seeds emerged earlier in sandy loam than in clay. It was found that on all sowing dates the sandy loam plots with the shallow groundwater table showed the highest emergence rate as well as the highest total emergence percentage. The mean heat sums required for 50% emergence were lower on sandy loam than on clay, and the heat sums of the shallow groundwater plots were lower than those of the deep groundwater plots.

The minimum temperatures for emergence of the various seeds were calculated using only those treatments in which no limitation of water could be expected. The effect of soil moisture on emergence could be evaluated by calculating the heat sums of all treatments, taking into account the minimum temperature for emergence. As indicator for the minimum moisture content required for emergence, the first five days after sowing were used, for which the average soil moisture content was calculated from sampling data and precipitation records.

It was found that the heat sum required for 50% emergence increased sharply below a matric pressure of -0.49 bar (above pF 2.7) of the soil. Laboratory experiments with radish seed under controlled soil moisture and temperature conditions confirmed the results obtained from the field experiments that -0.49 bar (pF 2.7) is the critical value for emergence as regards the dry side. From laboratory experiments it could be shown that -0.098 bar (pF 2.0) is the critical value for emergence as regards the wet side. The time needed for emergence increases rapidly at matric pressure above -0.098 bar and below -0.49 bar. It appeared from these experiments that heat sums can give a relatively accurate prediction for emergence, if soil moisture content is taken into account.

For a fast and adequate seedling emergence both a high temperature and a sufficient moisture content are necessary. Under field conditions this combination is seldom reached because higher soil temperatures are generally related with lower moisture contents and deeper groundwater tables. To get out of this dilemma one can maintain a relatively deep groundwater table, which gives a relatively high temperature, and keep by means of sprinkler irrigation the sowing bed at the desired moisture content. This offers the additional advantage of keeping the temperature of the seed bed low in periods when soil temperature would exceed the optimum temperature for germination and emergence. It is to be noted that temperatures of 40 °C and higher (which are far too high for germination and emergence of various seeds) in the top layer of clay soils in the Netherlands in mid summer are no exception.

From literature it is known that yields of spring crops decrease when sowing dates are later and that owing to adequate drainage, soils can be cultivated and sown approximately 5 to 14 days earlier. An additional advantage of drainage is the shortening of the germination and emergence period as a result of the higher soil temperatures. From calculations it was shown that due to drainage sometimes a 10-day gain in emergence can be obtained as compared with a shallowly drained soil. The effect of drainage (or any other measure) on seedling emergence will be the largest when soil temperatures in the range close to the minimum temperature for germination and emergence are increased.

If available water is limiting, plant production will be reduced, which especially holds true for vegetable crop production where production is more aimed at quality and fresh weight than at an increase in dry matter. To weigh the influence of various measures, as for instance change in groundwater table depth or soil profile, on the production of crops, potential (gross) production rates which could have been obtained under the prevailing weather conditions with an optimum water and nutrient supply were calculated for the crops red cabbage, dwarf French beans and celery. Taking into account effects of soil cover and diffusion resistances of the crop, maximum production rates for the prevailing environmental conditions were computed. These rates were compared with the dry matter production rates obtained from periodical harvests. The relationships between real and calculated production rates were analyzed by linear regression of measured on calculated production, from which reduction factors (α_{ph}) were derived. From various data reported in literature it appears that losses in dry matter production by respiratory processes is about 30 per cent, so that maximum dry matter production is about 0.7 ($= \alpha_{ph}$) of the potential production.

For dwarf French beans an average α_{ph} of 0.67 was found for the various objects, indicating that the environmental conditions were fairly optimum.

For red cabbage rather low α_{ph} values were found: 0.51 to 0.56 for clay and clay on sandy loam respectively and 0.40 for sandy loam. The main reason for these low α_{ph} values seemed to be nitrogen deficiency. This deficiency occurred mainly in dry periods on the clay profiles and in wet periods (in autumn) on the sandy loam profiles. Moreover air deficiency induced by heavy rains in the later stages of growth may have been of importance.

For celery also a low average α_{ph} of 0.52 for the various objects was obtained, again mainly because of nitrogen deficiency. Part of the discrepancy could be explained by the applied calculation procedure.

A discussion on the various existing methods in determining the water use efficiency of a crop was presented. By considering total dry matter production and the ratio total evaporation over mean vapour pressure deficit of the three crops mentioned, it could be shown that the water use efficiency of dwarf French beans is highest, followed by celery and then red cabbage. It was found that under the prevailing environmental conditions other than water shortage, red cabbage on sandy loam was rather soon limited in its production. On the other profiles limitation of production did not set in.

The influence of groundwater table depth on production of crops has always been widely investigated in the Netherlands, especially on groundwater level experimental fields. Because of changes in weather conditions the relationship between yield and groundwater level differs from year to year and it seems of little use to make this comparison for years on end.

Because for a certain dry matter production a certain amount of water has to be evaporated, it is better to consider for a growing period the amount of water available for evaporation, partly determined by the amount of precipitation and partly by the amount of water that can be delivered by the soil. Taking into account the admissable pressure −0.4 bar (pF 2.6) at which soil water begins to limit growth of most vegetable crops, the amounts of water available in the soil at different lengths of growing period were calculated.

Fresh yields measured on all profiles of all groundwater plots were plotted against the total amounts of water available at each object for red cabbage, potatoes and lettuce. It appeared that maximum fresh yield productions were obtained at total amounts of available water of 365 mm for red cabbage, of 310 mm for potatoes (which agrees rather well with data derived from literature) and of 200 mm for lettuce.

Having obtained the optimum amounts of available water for maximum production of the crops, the maximum groundwater table depth, over a period in which the critical value of the total amount of rainfall was known from frequency distributions of rainfall depth, could be calculated for each crop and each profile. It was shown that per profile despite differences in crops, in lengths of growing periods and in frequency distributions of amount of rainfall, the deepest admissable water table depths did not differ very much. The reason for this is that the amounts of available water in the investigated profiles change considerably with slight changes in groundwater table depths around 0.90 m below surface. Generally speaking, the crops on the sandy loam admit deeper groundwater levels than on the clay and on the clay on sandy loam profile.

In a similar way also highest admissable groundwater table depths to ensure maximum production could be calculated. It was found that for early crops like potatoes and lettuce rather high groundwater tables can be admitted. For late crops like red cabbage deeper water tables are required. This is a logical consequence of the fact that early crops are in the Netherlands grown in periods with generally less rainfall than late crops are.

Finally optimum water table depths for maximum productions could be calculated, the procedure of which has been explained. The results showed that the optimum water table depth for red cabbage on clay is about 1.00 to 1.10 m, on sandy loam 1.10 to 1.20 m and on clay on sandy loam 0.90 to 1.00 m. For potatoes the optimum groundwater table depths are 0.90 m, 1.00 m and 0.90 m respectively; for lettuce 0.90 m, 1.00 m and 0.80 to 0.90 m respectively. It appeared that at deeper groundwater tables, the sandy loam carries less risk in production than the clay. This last profile in its turn carries less risk (for red cabbage much less risk) than the clay on sandy loam profile. It was shown that the latter is more susceptible for water table drawdowns below the optimum level than the other two profiles.

In general it is to be advised to maintain groundwater levels at depths that are somewhat lower than the optimum depth.

It is thought that with this approach there is a possibility to solve the problem of determining the total amount of water required for maximum crop production and, where relevant, to decide on the optimum groundwater table depth then necessary.

Aangezien bij de bemonstering met Kopecky ringen een grote spreiding in de droog volumegewichten werd gevonden en daardoor exacte verschillen in vochtgehalten in onvoldoende mate waren te bepalen, werd een betrekking tussen het vochtgehalte, de druk tengevolge van matrixkrachten en het droog volumegewicht ontwikkeld, waarmee de vochtkarakteristiek behorende bij een bepaald droog volumegewicht kon worden afgeleid. Op deze wijze konden de resultaten van opeenvolgende bemonsteringen worden vereffend.

Het capillair geleidingsvermogen van de drie profielen werd zowel in het laboratorium als, tijdens droge perioden, in het veld bepaald. De resultaten van de twee bepalingswijzen bleken in verheugende mate met elkaar overeen te stemmen.

Met de zo verkregen gegevens konden voor verschillende stroomsnelheden verbanden tussen de hoogte boven het freatisch vlak en de equivalenthoogte tengevolge van de matrixkrachten worden gelegd.

Uiteraard heeft de verdamping aan het oppervlak invloed op de hoogte van de grondwaterspiegel. Onder aanname van bepaalde randvoorwaarden kon de invloed van verdamping van kale grond op de zakking van de grondwaterspiegel kwantitatief worden vastgesteld. Het bleek dat de berekende grondwaterstandsdiepten redelijk overeenstemmen met die welke in gelijkwaardige gronden in aride gebieden voorkomen. Bovendien kon worden aangetoond dat de verdampingssnelheid bij een zakkende grondwaterspiegel ongeveer evenredig met de wortel van de tijd afneemt, hetgeen ook voor bodemprofielen zonder grondwaterspiegel, maar met een homogene aanvangsverdeling van het bodemvocht, in de literatuur wordt vermeld.

Teneinde de hoeveelheid water die in een bepaalde groeiperiode voor de verdamping beschikbaar is vast te kunnen stellen, werden de verschillende hoeveelheden beschikbaar bodemvocht aanwezig in de wortelzone, in de diepere lagen onder de wortelzone, en via capillaire werking opstijgend uit het grondwater, berekend. Bij de berekeningen werd uitgegaan van een voor de maximale produktie toelaatbare druk van het bodemvocht. Zoals uit een literatuuronderzoek bleek, ligt de uitdrogingsgrens voor de meeste groentegewassen bij —0,4 bar (pF 2,6).

De verdamping van de gewassen werd berekend met behulp van een combinatiemethode, die gebaseerd is op betrekkelijk eenvoudig te meten meteorologische, bodem- en gewasfactoren. De resultaten van de berekeningen werden vergeleken met gegevens verkregen via waterbalansstudies. Voor de aerodynamische weerstand van het gewas werden waarden voor verschillende gewashoogten en windsnelheden berekend.

Daar de netto straling een van de belangrijkste grootheden is die de verdamping vanaf een oppervlak bepalen, en rechtstreekse metingen dikwijls niet ter beschikking staan, worden in het algemeen empirische formules toegepast. Van enkele van deze formules werd gebruik gemaakt en de resultaten van de berekeningen werden met behulp van lineaire regressietechnieken vergeleken met de gemeten netto straling.

De berekening van de kortgolvige straling werd met verscheidene formules uitgevoerd, waarbij hoge correlatiecoëfficiënten werden verkregen. De berekeningen van de thermische straling over periodelengten van een week vertoonden echter een grote spreiding. Teneinde hieraan te ontkomen werden met behulp van lineaire regressie enige vergelijkingen, gebaseerd op de hoge correlatie tussen netto en kortgolvige straling, voor verschillende periodelengten afgeleid.

Het verband gevonden via berekeningen op 24-uur basis stemde bijzonder goed overeen met gelijksoortige in Australië berekende relaties. Ook de verbanden die op daglichtbasis werden bepaald kwamen goed overeen met in de literatuur vermelde resultaten. Voor de verdampingsberekeningen op weekbasis bleek dat vaak met voldoende nauwkeurigheid gebruik kan worden gemaakt van stralingscijfers die gemeten zijn op relatief ver af gelegen weerstations.

Voor de diverse oppervlakken werden reflectiecoëfficiënten bepaald door middel van continue metingen op verschillende typen dagen. De verschillende methoden om deze coëfficiënten te bepalen werden besproken. Ten aanzien van de reflectiecoëfficiënten van een aantal groentegewassen bij toenemende bodembedekking, bleek dat ze, met uitzondering van die voor spinazie, alle een min of meer gelijk en vrij dicht aansluitend patroon vertoonden. Zowel bij kale als bij een met een gewas bedekte grond gaven natte oppervlakken lagere reflectiecoëfficiënten dan droge oppervlakken.

De verdamping van kale gronden werd voor verschillende grondwaterstandsdiepten met behulp van een energiebalansbenadering op heldere dagen en op uurbasis berekend. Dit voor tamelijk natte en tamelijk droge perioden. Verschillen in de componenten van de energiebalans van de verschillende objecten werden aangetoond. Berekende dampspanningen aan het bodemoppervlak werden vergeleken met waarden welke waren verkregen via bemonsteren van de bovenlaag. De resultaten van beide benaderingen stemden goed overeen.

Ten aanzien van de diffusieweerstanden die bij verdamping van begroeide oppervlakken optreden, bleek dat de bodem voor 70 à 80% moet zijn bedekt voordat de verdampingssnelheid constant wordt. Bij natte gronden heeft een verandering in het percentage bodembedekking relatief weinig invloed.

Interceptiemetingen aan rode kool stemden, de aard van het gewas in aanmerking genomen, goed overeen met in de literatuur vermelde interceptiemetingen. Aangetoond werd dat de invloed van interceptie op de berekende verdamping vooral van belang is na droge perioden en onder condities van hoge potentiële verdamping.

Voor het gewas rode kool werd de weerstand voor het watertransport in de plant en de geometriefactor van het wortelsysteem geanalyseerd. De variaties van deze factoren met de diepte werden bepaald en de wateronttrekkingssnelheden door de wortel werden berekend en vergeleken met resultaten uit waterbalansstudies. De gevonden transportweerstanden kwamen redelijk overeen met in de literatuur vermelde resultaten. Wat betreft de geometriefactor in de beginstadia van de groei, werd een afwijking met een factor tien van in de literatuur gegeven waarden gevonden. De oorzaak hiervan lag in de weinig homogene en spaarzame wortelverdeling van de geteelde rode kool in deze stadia van ontwikkeling. Nadat de wortelontwikkeling was toegenomen, nam de geometriefactor af tot waarden zoals die voor andere gewassen worden opgegeven.

De met de combinatiemethode berekende verdampingen bleken bevredigend overeen te komen met via de waterbalans bepaalde verdampingen. Uit de berekeningen bleek dat, uitgezonderd in een paar perioden tijdens het begin van de groei, weinig reduktie in verdamping tengevolge van watertekorten aanwezig was.

In hoofdstuk III werd verder aandacht geschonken aan vochtproblemen die optreden bij kieming, opkomst en aanslag van het gewas.

In hoofdstuk IV werd het transport van warmte in de grond behandeld. De gegevens betreffende de warmteeigenschappen van de grond werden verkregen uit veldmetingen, metingen in het laboratorium en berekeningen.

De warmtecapaciteit werd berekend door deze te beschouwen als een functie van de volumefractie en de soortelijke warmte van de verschillende componenten. Berekeningen van het warmtegeleidingsvermogen, gebaseerd op de aanname dat bodemdeeltjes beschouwd kunnen worden als omwentelingsellipsoiden, gaven, misschien als gevolg van een verkeerde aanname van de depolarisatiefactor, slechte resultaten. Voor de zavel gaven metingen van het warmtegeleidingsvermogen met de niet-stationaire naaldmethode goede resultaten. Voor de krimpende en zwellende klei werd een grote spreiding in de meetwaarden gevonden. Dit was voornamelijk het gevolg van veranderingen in het droog volumegewicht, waarin moeilijk correcties zijn aan te brengen. De voorkeur werd daarom gegeven aan het meten van het warmtegeleidingsvermogen van de grond in situ.

Wat betreft de temperatuurvereffeningscoëfficiënt, deze kan op verschillende wijzen worden verkregen. Bepaling van deze coëfficiënt in het veld via een analyse van de temperatuurmetingen op de verschillende diepten, door de amplitude- en faseveranderingen met de diepte te beschouwen, faalden. Dit duidt er op dat de met de diepte veranderende warmteeigenschappen van een profiel een zodanige procedure niet toelaten. Teneinde nu functionele verbanden van de coëfficiënt met de diepte vast te stellen, werd gebruik gemaakt van zowel een elektrisch analogon als van numerieke oplossingsmethoden. Beide type benaderingen zijn er op gebaseerd dat het continue temperatuursbeloop door een discontinu beloop kan worden benaderd door voor de partiële differentiaalvergelijking een serie differentievergelijkingen te schrijven.

De spreiding in de resultaten verkregen via het analogon was aanzienlijk. Dit als gevolg van onder meer registratie- en afleesfouten, de relatief grote afstanden tussen de temperatuurmetingen op de grotere diepten en de heterogeniteit van de profielen. Met de numerieke oplossingsmethoden werden eveneens teleurstellende resultaten verkregen. Correctie van de registratie- en afleesfouten door de gemeten temperatuurwaarden te vervangen door vereffende temperaturen, zoals berekend met behulp van een Fourier analyse, gaf geen betere resultaten. Een poging tot vermindering van de fout inherent aan het discretiseren van het continue temperatuursbeloop in een discontinu veld, door interpolatie met behulp van 'spline' functies had ook geen resultaat.

Een mogelijke verklaring van de spreiding in de gevonden temperatuurvereffeningscoëfficiënten had de invloed van door temperatuurgradiënten geïnduceerd watertransport in de dampfase kunnen zijn. Deze veronderstelling bleek na uitvoerige berekeningen niet houdbaar.

Uiteindelijk bleek de beste methode ter bepaling van de temperatuurvereffeningscoëfficiënt te zijn, het afleiden van de warmtecapaciteit uit volumebemonsteringen, het warmtegeleidingsvermogen in situ te meten met de nietstationaire naaldmethode en dan het quotiënt te bepalen van warmtegeleidingsvermogen en warmtecapaciteit. Uit de resultaten kwam naar voren dat de temperatuurvereffeningscoëfficiënt in het algemeen lager is bij hogere grondwaterstanden, en wel in het bijzonder in de bovenste laag van 0,10–0,15 m. Deze uitkomsten werden bevestigd via temperatuurmetingen, waarbij bleek dat de gemiddelde dagelijkse temperaturen in de vakken met de hogere grondwaterstanden 1 à 2°C lager waren dan in de vakken met de diepere grondwaterstanden. Bovendien bleek bij eenzelfde grondwaterstandsdiepte de klei warmer te zijn dan de zavel.

Het temperatuursbeloop wordt tenminste voor de onderzochte profielen, meer beïnvloed door de verschillen in grondwaterstand dan in bodemprofiel. Bij een diepe grondwaterstand liggen de maxima en minima in de bovenlaag hoger en nemen de temperatuuramplituden sneller af met de diepte. Deze afname was geprononceerder in de klei dan in de zavel.

Na op deze wijze een basis te hebben gelegd ten aanzien van de fysische processen van het met de plantegroei samenhangende transport van water en warmte, werd in hoofdstuk V gepoogd hun invloed op de opkomst en produktie van een aantal tuinbouwgewassen te analyseren.

Ten aanzien van de groei van een gewas is een onderscheid te maken in het stadium van kieming tot opkomst en het stadium van opkomst tot afrijpen. In het eerste stadium zijn bodemvocht en bodemtemperatuur de belangrijkste milieufactoren voor de ontwikkeling. In het tweede stadium hangt de groei tevens af van de luchttemperatuur, maar vooral van het bladoppervlak en de netto straling.

Na een kort overzicht te hebben gegeven van uit de literatuur bekende effecten van bodemtemperatuur en bodemvocht op kieming en opkomst, werd ingegaan op het gebruik van warmtesommen, teneinde temperatuur en opkomst met elkaar in verband te brengen. In de praktijk wordt veelvuldig gebruik gemaakt van warmtesommen om bijvoorbeeld planttijdstippen vast te stellen, oogsttijdstippen te voorspellen en geschikte gewasselecties voor bepaalde gebieden te vinden. Voor deze doeleinden wordt vaak uitgegaan van de luchttemperatuur. Voor kieming en opkomst is het echter aan te raden de temperatuur in de nabijheid van het kiemende zaad als basis te nemen, met andere woorden de bodemtemperatuur op zaaidiepte.

Wat betreft de invloed van bodemvocht op kieming en opkomst, verschillen de in de literatuur vermelde gegevens aanzienlijk. Dit vloeit voort uit het feit dat meestal is uitgegaan van laboratoriumproeven, waarbij de proefomstandigheden vaak sterke verschillen vertoonden. Zo zijn bijvoorbeeld de resultaten van kieming in grond en die op een voedingsoplossing niet vergelijkbaar.

In het hier behandelde onderzoek werd aan de hand van zaaiproeven in het veld met zaden van vier verschillende groentegewassen op een klei- en een zavelprofiel, zowel bij een hoge als een lage grondwaterstand, het gecombineerde effect van bodemtemperatuur en vochtgehalte op de kieming en opkomst bestudeerd. Uiteraard bleek dat de opkomst sterk samenhing met de hoeveelheid neerslag, maar door de gunstige hydrologische eigenschappen van de zavel kwam op dat profiel een eerdere opkomst dan op de klei tot stand. Het zavelprofiel met de hoogste grondwaterstand gaf steeds de snelste opkomst en het hoogste opkomstpercentage te zien. De gemiddelde warmtesommen nodig voor 50% opkomst waren op de zavel kleiner dan op de klei en op de proefvakken met hoge grondwaterstand lager dan op de vakken met een diepe grondwaterstand.

De voor opkomst benodigde minimum temperaturen werden berekend uit de gegevens van die objecten waarin geen watertekorten konden zijn opgetreden. Door nu, met inachtneming van deze minimum temperaturen, de warmtesommen voor alle objecten te berekenen, kon de invloed van het bodemvocht op de opkomst worden gekwantificeerd. Als bodemvochtgehalte werd genomen het gemiddelde vochtgehalte over de eerste vijf dagen na zaaien, berekend uit volumebemonsteringen en neerslagcijfers.

Het bleek dat de warmtesom benodigd voor 50% opkomst sterk toenam bij vochtgehalten behorende bij drukken tengevolge van de matrixkrachten gelegen beneden —0,49 bar (hetgeen overeenkomt met pF waarden boven 2,7). Ook uit laboratoriumproeven bleek dat, wat betreft kieming en opkomst, deze waarden de kritische waarden zijn. Ten aanzien van vochtoverschotten werd de kritische waarde van —0,098 bar (pF 2,0) gevonden. Boven —0,098 bar en beneden —0,49 bar neemt de tijd benodigd voor opkomst sterk toe.

Uit het onderzoek werd duidelijk dat men met behulp van warmtesommen een vrij goede voorspelling van het tijdstip van opkomst van een gewas kan geven, indien het vochtgehalte van de grond in aanmerking wordt genomen.

Voor een snelle en goede kieming en opkomst zijn zowel een hoge bodemtemperatuur als een voldoende vochtgehalte noodzakelijk. Onder veldomstandigheden wordt deze combinatie zelden bereikt, daar hoge bodemtemperaturen in het algemeen samengaan met lage vochtgehalten en diepe grondwaterspiegels. Teneinde uit dit dilemma te geraken, kan een relatief diepe grondwaterstand worden gehandhaafd die voor een relatief hoge bodemtemperatuur zorgdraagt, en kan door middel van beregening het zaaibed op het vereiste vochtgehalte worden gehouden. Dit laatste biedt het bijkomende voordeel dat de temperatuur van het zaaibed laag kan worden gehouden in perioden dat de bodemtemperatuur de optimum temperatuur voor kieming en opkomst zou overschrijden. In dit verband moet worden opgemerkt dat zelfs in Nederland temperaturen van 40°C en hoger (veel te hoog voor kieming en opkomst) in de toplaag van kleigronden in midden zomer geen uitzondering zijn.

Uit de literatuur is bekend dat de opbrengsten afnemen wanneer de zaaidata door onbewerkbaarheid van de grond later komen te liggen. Bij een goede drainage van deze gronden kan ongeveer 5 tot 14 dagen eerder worden bewerkt en ingezaaid. Een bijkomend voordeel van deze drainage is dat, als gevolg van de hogere bodemtemperaturen, de periode die nodig is voor de kieming en opkomst wordt verkort. Uit berekeningen kon worden aangetoond dat in vergelijking met een ondiep ontwaterde grond soms een winst van 10 dagen in de opkomst verkregen kan worden. Het vervroegend effect van drainage (of een andere binnen zekere grenzen temperatuurverhogende maatregel) is het grootst wanneer de oorspronkelijke bodemtemperaturen rondom de minimum temperatuur voor kieming en opkomst schommelen.

Teneinde de invloed van verschillende cultuurtechnische maatregelen, zoals in dit geval een wijziging van grondwaterstandsdiepte en van bodemprofiel, op de produktie van gewassen vast te kunnen stellen, werden potentiële (bruto) produkties berekend voor de gewassen rode kool, stamslabonen en selderij, zoals die onder de heersende weersomstandigheden verkregen zouden kunnen zijn met een optimale water- en voedingsstoffenvoorziening. Door rekening te houden met de effecten van bodembedekking en van diffusieweerstanden van het gewas, werden maximale produkties berekend voor de heersende condities. Deze produkties werden vergeleken met werkelijke droge stofprodukties verkregen door middel van periodieke oogsten. De relaties tussen werkelijke en berekende produkties werden vergeleken met behulp van lineaire regressie van gemeten op berekende produktie, waaruit reduktiefactoren werden afgeleid.

Uit diverse literatuurgegevens blijken de verliezen in droge stofproduktie tengevolge van ademhalingsprocessen ongeveer 30% te zijn, zodat de maximum produktie ongeveer 0,7 maal de potentiële produktie is.

In het hier beschreven onderzoek werd voor stamslabonen een gemiddelde reductiefactor van 0,67 gevonden voor de verschillende objecten, wat er op duidt dat het groeimilieu redelijk gunstig was. Voor rode kool werden lagere reductiefactoren gevonden, namelijk 0,51 tot 0,56 respectievelijk op de klei en op de klei op zavel, en 0,40 op de zavel. De voornaamste reden van deze lage waarden bleek stikstofgebrek te zijn. Dit gebrek trad op de klei voornamelijk in droge perioden op en op de zavelprofielen in natte perioden. Bovendien is het mogelijk dat luchtgebrek in de diepere lagen van de wortelzone na zware regens in de latere groeistadia een van de oorzaken kan zijn geweest. Voor selderij werd eveneens een lage gemiddelde reductiefactor gevonden (0,52), ook weer grotendeels tengevolge van stikstofgebrek. Verder ligt een deel van de verklaring in de toegepaste berekeningsprocedure voor dit gewas, dat in verschillende sneden wordt geoogst.

Teneinde het nuttig effect van het waterverbruik bij de produktie van een gewas te karakteriseren is een aantal methoden in gebruik. Aan de voor- en nadelen van deze verschillende methoden werd een korte bespreking gewijd.

Door voor de drie genoemde gewassen de totale droge stofproduktie te vergelijken met de verhouding van de totale verdamping tot het gemiddelde waterdampverzadigingsdeficiet, kon worden aangetoond dat het nuttig effect van het watergebruik bij stamslabonen het grootst is, gevolgd door respectievelijk selderij en rode kool. Het bleek dat onder de heersende uitwendige omstandigheden voor rode kool op zavel reeds spoedig een beperking in de produktie werd bereikt. Op de andere profielen werd voor de gewassen geen beperking in de produktie gevonden.

De invloed van de grondwaterstandsdiepte op de gewasproduktie is in Nederland reeds lang, en wel in het bijzonder op grondwaterstandenproefvelden, aan uitvoerige onderzoekingen onderworpen geweest. Daar de weersomstandigheden in elk proefjaar uniek zijn, is het niet aan te bevelen een direct verband te leggen tussen opbrengst en gemiddelde grondwaterstand gedurende het groeiseizoen. Het lijkt juister, aangezien voor een zekere droge stofproduktie een bepaalde hoeveelheid water benodigd is, om de gedurende de groeiperiode beschikbare hoeveelheid water te correleren met de produktie. Een beperking in deze beschikbare hoeveelheid zal de gewasopbrengst reduceren en wel in het bijzonder de geldelijke opbrengst van tuinbouwgewassen die meer berust op kwaliteit en vers gewicht dan op droge stof.

Met inachtname van de toelaatbare druk van --0,4 bar (pF 2,6) waarbij het bodemvocht de groei van de meeste groentegewassen begint te beperken, werden de hoeveelheden water die gedurende verschillende lengten van het groeiseizoen in het profiel beschikbaar zijn, berekend. Verse opbrengsten van rode kool, aardappels en sla werden gemeten op de drie profielen van alle grondwaterstandsvakken en uitgezet tegen de totale hoeveelheden water die beschikbaar waren in ieder object. Het bleek dat rode kool een maximum vers gewichtopbrengst bereikte bij een totale hoeveelheid beschikbaar water van circa 365 mm, aardappels bij circa 310 mm (hetgeen goed overeenstemt met uit de literatuur af te leiden gegevens) en sla bij circa 200 mm.

Nadat de optimale hoeveelheden beschikbaar water voor een maximale vers gewichtproduktie van de drie gewassen waren bepaald, werd de maximaal toelaatbare grondwaterstandsdiepte over de groeiperiode waarin een bepaalde kritische hoeveelheid neerslag viel, berekend voor elk gewas en elk profiel. Aangetoond kon worden dat per profiel, ondanks verschillen in gewas, lengte van groeiperiode en in frequentieverdeling van de neerslagsommen, de maximaal toelaatbare grondwaterstandsdiepten niet ver uit elkaar liepen. Dit vloeit voort uit het feit dat in de onderzochte profielen de hoeveelheid beschikbaar water sterk verandert bij kleine veranderingen in grondwaterstandsdiepte rond 0,90 m beneden maaiveld. In het algemeen gesproken kunnen de gewassen op de zavel een diepere grondwaterstand dan op de klei- en op het klei op zavelprofiel verdragen.

Op eenzelfde wijze konden ook de hoogst toelaatbare grondwaterstandsdiepten voor een maximale produktie worden berekend. Het bleek dat vroege gewassen, zoals aardappels en sla, relatief hoge grondwaterstanden verdragen. Voor late gewassen, zoals rode kool, zijn relatief diepe waterstanden vereist. Dit is een logisch gevolg van het feit dat in Nederland vroege gewassen groeien in perioden waarin in het algemeen minder regen valt, dan later in het seizoen.

Tenslotte konden optimale grondwaterstandsdiepten voor een maximale produktie worden berekend door middel van een in de tekst beschreven procedure.

De resultaten van de berekeningen lieten zien dat voor rode kool de optimum grondwaterstandsdiepte op klei ongeveer 1,00 à 1,10 m is, op zavel 1,10 à 1,20 m en op klei op zavel 0,90 à 1,00 m. Voor aardappels zijn de optimale grondwaterstandsdiepten respectievelijk 0,90 m, 1,00 m en 0,90 m; voor sla respectievelijk 0,90 m, 1,00 m en 0,80 à 0,90 m.

Uit de met de bovenvermelde procedure verkregen waarschijnlijkheidsdiagrammen van vers gewichtopbrengst tegen grondwaterstandsdiepte, waarbij rekening werd gehouden met de in Den Helder voorkomende neerslagverdeling over 103 jaar, bleek dat de zavel bij diepe grondwaterstanden minder risicogevoelig is dan de klei. Dit laatste profiel is, op zijn beurt, weer minder risicodragend dan het klei op zavelprofiel. Bovendien werd aangetoond dat de vers gewichtopbrengsten op het klei op zavelprofiel gevoeliger zijn voor een verlaging van de grondwaterstand beneden het optimum dan de vers gewichtopbrengsten op de twee andere profielen.

In verband met de moeilijkheid om in de praktijk de nadelige invloed op de produktie van plotselinge neerslagoverschotten op te vangen, is het in de praktijk raadzaam de grondwaterstand gedurende het groeiseizoen iets beneden de optimale diepte te fixeren. Eventuele daaruit tijdelijk voortvloeiende vochttekorten zijn immers gemakkelijk door additionele watergiften te verhelpen.

Uiteindelijk werd de hoop uitgesproken dat het met behulp van de hierboven beschreven benadering mogelijk zal zijn ook voor andere profielen en gewassen de voor een maximale produktie benodigde totale hoeveelheid water vast te kunnen stellen en, waar dit van toepassing is, de hiervoor benodigde optimale grondwaterstandsdiepte te bepalen.

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LIST OF USED SYMBOLS

The letters a, b, c, d, m, n, p, q, η, μ, are also used for any given constant. Some of the symbols used in a few consecutive equations falling outside the main line of argument are defined in the text only. Given are SI units, for conversion see table 20. Values of physical properties apply to conditions of 293 K and/or 1.013 bar. For units of water potential and their equivalents see also table 21.

SYMBOL	INTERPRETATION	UNITS	DIMENSION
A	Area	m^2	L^2
A_0, A_z	Amplitude of the temperature wave at the surface resp. at depth z	K	T
a	Thermal diffusivity ($a = \lambda/\rho c$)	$m^2.s^{-1}$	$L^2.t^{-1}$
b	Root system factor defined by eq. (46)	m	L
C, C_{CO_2}	Concentration of water vapour resp. carbon dioxide concentration	$kg.m^{-3}$	$M.L^{-3}$
C_0, C_0^*	Unsaturated resp. saturated water vapour concentration at height $(z_0 + d)$ and temperature T_0	$kg.m^{-3}$	$M.L^{-3}$
C_z, C_z^*	Unsaturated resp. saturated water vapour concentration in the air at height z and temperature T_z	$kg.m^{-3}$	$M.L^{-3}$
c	Specific heat per unit mass	$J.kg^{-1}.K^{-1}$	$L^2.t^{-2}.T^{-1}$
c_a, c_p, c_s, c_w	Specific heat per unit mass of air, the same at constant pressure, per unit mass of solid material, resp. per unit mass of water ($c_p = 1004$)	$J.kg^{-1}.K^{-1}$	$L^2.t^{-2}.T^{-1}$
D	Damping depth ($D = \sqrt{2\lambda/\rho c\omega}$)	m	L
D_{CO_2}, D_{H_2O}	Molecular diffusion coefficient of carbon dioxide in air ($D_{CO_2} = 14 \times 10^{-6}$) resp. of water vapour in air ($D_{H_2O} = 26 \times 10^{-6}$)	$m^2.s^{-1}$	$L^2.t^{-1}$
d	Zero plane displacement	m	L
E	Actual evaporation flux; mass flux volume flux	$kg.m^{-2}.s^{-1}$ $m^3.m^{-2}.s^{-1}$	$M.L^{-2}.t^{-1}$ $L.t^{-1}$
E_i	Actual evaporation flux of intercepted precipitation; mass (volume)	$kg.m^{-2}.s^{-1}$ ($m.s^{-1}$)	$M.L^{-2}.t^{-1}$ ($L.t^{-1}$)
E_p	Potential evaporation flux; mass (volume)	$kg.m^{-2}.s^{-1}$ ($m.s^{-1}$)	$M.L^{-2}.t^{-1}$ ($L.t^{-1}$)
E^*	Evaporation flux of a wet surface; mass (volume)	$kg.m^{-2}.s^{-1}$ ($m.s^{-1}$)	$M.L.^{-2}.t^{-1}$ ($L.t^{-1}$)
e	Base of natural logarithms (e = 2.71828...)	1	–
e	Water vapour pressure	bar	$M.L^{-1}.t^{-2}$
e_0, e_0^*	Unsaturated resp. saturated water vapour pressure at height $(z_0 + d)$ and temperature T_0	bar	$M.L^{-1}.t^{-2}$

SYMBOL	INTERPRETATION	UNITS	DIMENSION
e_z, e_z^*	Unsaturated resp. saturated water vapour pressure in the air at height z and temperature T_z	bar	$M.L^{-1}.t^{-2}$
F	Heat sum	K.s	T.t
f	Function	–	–
G	Heat flux into the soil	$W.m^{-2}$	$M.t^{-3}$
g	Acceleration due to gravity ($g = 9.813$)	$m.s^{-2}$	$L.t^{-2}$
H	Heat flux into the air	$W.m^{-2}$	$M.t^{-3}$
h_{total}	Total energy of water expressed per unit weight, thus equivalent with head ($h_{total} = \Psi_{total}/g = h + h_g = h_m + h_p + h_s + h_g$)	m	L
h, h_g, h_m, h_p, h_s	Equivalents of hydraulic head, gravitational head, matric head, pressure head resp. solute head	m	L
h_a	Equivalent of matric head at air entry point	m	L
h_{leaf}, $h_{root\ surf}$, h_{soil}	Equivalents of water head at the evaporating surfaces within the leaves, at the root surfaces resp. in the soil	m	L
K	Von Karman's constant ($K = 0.428$)	1	–
K_h, K_v	Eddy transfer coefficients for heat resp. water vapour	$m^2.s^{-1}$	$L^2.t^{-1}$
k, k_0	Unsaturated resp. saturated hydraulic conductivity expressed per unit weight	$m.s^{-1}$	$L.t^{-1}$
k^*, k_0^*	Unsaturated resp. saturated hydraulic conductivity expressed per unit volume	$m^3.s.kg^{-1}$	$M^{-1}.L^3.t$
L	Latent heat of vaporization of water per unit mass ($L = 2.4518 \times 10^6$)	$J.kg^{-1}$	$L^2.t^{-2}$
l_r	Length of roots per unit volume of soil	$m.m^{-3}$	L^{-2}
M	Amount of water extracted in unit time per unit of horizontal area from the soil above the groundwater table $M = -\int_{t_1}^{t_2}\int_0^z \frac{\partial\theta}{\partial t}\,dz\,dt$	m	L
N, n	Maximum possible resp. actual duration of sunshine per day	s	t
P, P_{pot}; P_c, P_o	Actual resp. potential photosynthetic flux; photosynthetic flux on clear days resp. on overcast days	$kg.m^{-2}.s^{-1}$	$M.L^{-2}.t^{-1}$
p_a	Atmospheric pressure ($p_a = 1.013$)	bar	$M.L^{-1}.t^{-2}$
q	Volume flux of water passing through a unit horizontal area per unit time	$m^3.m^{-2}.s^{-1}$	$L.t^{-1}$
q_d, q_u	Downward resp. upward flux of water through a phreatic surface at constant depth	$m.s^{-1}$	$L.t^{-1}$
q_r, q_s	Flux of water through the roots resp. the soil	$m.s^{-1}$	$L.t^{-1}$
q_r'	Rate of water uptake per unit length of root	$m^3.m^{-1}.s^{-1}$	$L^2.t^{-1}$

SYMBOL	INTERPRETATION	UNITS	DIMENSION
q_v	Water vapour flux through the soil; mass (volume)	$kg.m^{-2}.s^{-1}$ ($m.s^{-1}$)	$M.L^{-2}.t^{-1}$ ($L.t^{-1}$)
q_h	Vertical heat flux through the soil	$W.m^{-2}$	$M.t^{-3}$
R_n, R_s, R_t	Net, shortwave resp. thermal radiation flux	$W.m^{-2}$	$M.t^{-3}$
R_s^c	Shortwave radiation flux under clear sky conditions	$W.m^{-2}$	$M.t^{-3}$
R_s^{top}	Extra-terrestrial shortwave radiation flux at the top of the atmosphere	$W.m^{-2}$	$M.t^{-3}$
R_s^{re}	Shortwave radiation flux reflected at the surface	$W.m^{-2}$	$M.t^{-3}$
R', R_c', R_o'	Solar radiation flux involved in photosynthesis (0.4 to 0.7 μm) on actual days, clear days resp. overcast days	$W.m^{-2}$	$M.t^{-3}$
r_a	Diffusion resistance to water vapour of the air layer surrounding the leaves	$s.m^{-1}$	$t.L^{-1}$
r_s	Diffusion resistance to water vapour of both crop and soil surface ($r_s = r_l + r_c + r_h$)	$s.m^{-1}$	$t.L^{-1}$
r_c, r_h, r_l	Diffusion resistances to water vapour dependent on fraction of soil covered, on water head equivalent in the leaf tissues resp. on radiation flux	$s.m^{-1}$	$t.L^{-1}$
r_{pl}, r_{soil}	Resistances to liquid flow in the plant resp. the soil	s	t
r_a'	Diffusion resistance to carbon dioxide of the air layer surrounding the leaves	$s.m^{-1}$	$t.L^{-1}$
r_m'	Diffusion resistance to carbon dioxide of mesophyll cells	$s.m^{-1}$	$t.L^{-1}$
r_s'	Diffusion resistance to carbon dioxide of both crop and soil surface	$s.m^{-1}$	$t.L^{-1}$
S_c	Fraction of soil covered	1	–
T	Temperature (293.15 K = 20°C)	K	T
T_0, T_z	Temperature at the surface resp. at vertical distance z from reference level	K	T
t	Time	s	t
u	Horizontal wind velocity	$m.s^{-1}$	$L.t^{-1}$
u_*	Friction velocity	$m.s^{-1}$	$L.t^{-1}$
V	Volume	m^3	L^3
x_a, x_s, x_{sm}, x_{so}, x_w	Volume fractions of air, solid material, minerals, organic matter resp. water in the soil	1	–
z	Vertical distance from reference level, as indicated in text	m	L
z^*	Depth below the surface (section IIIB-3d-4)	m	L
z_0	Roughness length	m	L
z_r	Rooting depth	m	L
α	Reflection coefficient of surface	1	–
α_{ph}	Reduction factor for dry matter production by a crop	1	–

SYMBOL	INTERPRETATION	UNITS	DIMENSION
β	Bowen's ratio ($\beta = H/LE$); Factor accounting for pore geometry (eq. 91)	1	–
γ	Psychrometric constant ($\gamma = c_p p_a / L\varepsilon = 0.67 \times 10^{-3}$)	$bar.K^{-1}$	$M.L^{-1}.t^{-2}.T^{-1}$
δ	Slope of the saturation vapour pressure curve ($\delta = de/dT$)	$bar.K^{-1}$	$M.L^{-1}.t^{-2}.T^{-1}$
ε	Ratio molecular weight of water vapour and dry air ($\varepsilon = 0.622$)	1	–
ε_{ea}	Long wave emissivity of the earth	1	–
ε_g	Volume of gas-filled pores per unit volume of soil in bulk	$m^3.m^{-3}$	–
ζ	Ratio of average temperature gradient in gas-filled pores and the overall temperature gradient	1	–
θ, θ_l	Volume of water per unit volume of soil in bulk	$m^3.m^{-3}$	–
θ/ρ_d	Weight of water per unit weight of dry soil	$kg.kg^{-1}$	–
Λ	Fraction of time the sky is overcast	1	–
λ	Thermal conductivity of soil	$W.m^{-1}.K^{-1}$	$M.L.t^{-3}.T^{-1}$
ν_1, ν_2	Radii	m	L
ξ	Factor arising from vapour flux enhancement mechanisms	1	–
ρ	Density	$kg.m^{-3}$	$M.L.^{-3}$
ρ^*	Equivalent 'electron' density of water	$kg.m^{-3}$	$M.L^{-3}$
ρ_a	Density of moist air ($\rho_a = 1.2047$)	$kg.m^{-3}$	$M.L^{-3}$
ρ_d, ρ_o, ρ_w	Bulk density of dry mineral soil, organic matter resp. water ($\rho_w = 998.23$)	$kg.m^{-3}$	$M.L^{-3}$
ρ_s	Specific mass of soil	$kg.m^{-3}$	$M.L^{-3}$
ρc	Heat capacity per unit volume	$J.m^{-3}.K^{-1}$	$M.L^{-1}.t^{-2}.T^{-1}$
Υ_r	Water uptake by roots ($\Upsilon_r = dq_r/dz$)	s^{-1}	t^{-1}
φ	Phase shift or angle	rad	–
χ, χ_i	Flux of precipitation resp. intercepted precipitation; mass (volume)	$kg.m^{-2}.s^{-1}$ ($m.s^{-1}$)	$M.L^{-2}.t^{-1}$ ($L.t^{-1}$)
Ψ_{total}	Total water potential expressed as energy per unit mass ($\Psi_{total} = \Psi + \Psi_a + + \Psi_m + \Psi_p + \Psi_s + \Psi_g$)	$J.kg^{-1}$	$L^2.t^{-2}$
Ψ, Ψ_g, Ψ_m, Ψ_p, Ψ_s	Hydraulic, gravitational, matric, pressure resp. solute potential	$J.kg^{-1}$	$L^2.t^{-2}$
ψ_{total}	Total energy of water expressed per unit volume thus equivalent with pressure ($\psi_{total} = \rho\Psi_{total} = \psi + \psi_a = = \psi_m + \psi_p + \psi_s + \psi_g$)	bar	$M.L^{-1}.t^{-2}$
ψ, ψ_g, ψ_m, ψ_p, ψ_s	Equivalent of hydraulic pressure, equivalent of gravitational pressure, equivalent of matric pressure, pressure, resp. equivalent of solute pressure	bar	$M.L^{-1}.t^{-2}$
ω	Radial frequency	s^{-1}	t^{-1}
∇	Gradient ($\nabla = \partial/\partial z$ or $\nabla = d/dz$, as defined in the text)	m^{-1}	L^{-1}

TABLE 20. Conversion of units

Quantity	Dimension	S.I. unit	S.I. symbol	c.g.s. unit	c.g.s. symbol	Other units	To convert FROM S.I. multiply by	To convert TO S.I. multiply by
Length	L	meter	m					
				centimeter	cm		1 $\times 10^{2}$	1 $\times 10^{-2}$
						inch	3.937 $\times 10^{1}$	0.25400051 $\times 10^{-1}$
						feet	3.2808333	0.30480061
Mass	M	kilogram	kg					
				gram	g		1 $\times 10^{3}$	1 $\times 10^{-3}$
Time	t	seconds	s					
				seconds	s		1	1
						days	1.1574074 $\times 10^{-5}$	8.6400 $\times 10^{4}$
						hours	2.777777 $\times 10^{-4}$	3.600 $\times 10^{3}$
						minutes	0.166666 $\times 10^{-1}$	60
Force	$M.L.t^{-2}$	newton	$N(kg.m.s^{-2})$					
				dyne ($=g.cm.s^{-2}$)			1 $\times 10^{5}$	1 $\times 10^{-5}$
Pressure	$M.L^{-1}.t^{-2}$	$N.m^{-2}$						
				$dyne.cm^{-2}$			10	1 $\times 10^{-1}$
						bar	1 $\times 10^{-5}$	1 $\times 10^{5}$
						mm Hg	0.7501 $\times 10^{-2}$	1.3332 $\times 10^{2}$
						atmosphere	0.9872 $\times 10^{-5}$	1.0130 $\times 10^{5}$
Energy	$M.L^{2}.t^{-2}$	joule ($=N.m$)	J					
				erg ($=dyne.cm$)			1 $\times 10^{7}$	1 $\times 10^{-7}$
						$calorie_{th}$	0.23901	4.1840
						kg H_2O (equivalent)*	4.0786 $\times 10^{-7}$	2.4518 $\times 10^{6}$
Power	$M.L^{2}.t^{-3}$	Watt ($=J.s^{-1}$)	W					
				$erg.s^{-1}$			1 $\times 10^{7}$	1 $\times 10^{-7}$
Energy flux	$M.t^{-3}$	$W.m^{-2}$						
				$erg.cm^{-2}.s^{-1}$			1 $\times 10^{3}$	1 $\times 10^{-3}$
						$cal.cm^{-2}.day^{-1}$	0.20650 $\times 10^{1}$	4.8426 $\times 10^{-1}$
						$cal.cm^{-2}.hr^{-1}$	0.86042 $\times 10^{-1}$	1.1622 $\times 10^{1}$
						$cal.cm^{-2}.min^{-1}$	0.14340 $\times 10^{-2}$	6.9733 $\times 10^{2}$
						$cal.cm^{-2}.s^{-1}$	0.23901 $\times 10^{-4}$	4.1840 $\times 10^{4}$
						mm $H_2O.day^{-1}$ (equivalent)** kg $H_2O.m^{-2}.day^{-1}$ (equivalent)**	0.352 $\times 10^{-1}$	2.838 $\times 10^{1}$

* Latent heat consumption in evaporating 1 kg H_2O at 293 K

** Power consumed per m^2 in evaporating H_2O at a flux of 1 $mm.day^{-1}$ at 293 K

TABLE 21. Units of water potential and their equivalents at 293 K and $g = 9.813$ m.s^{-2} (after TAYLOR, 1968; modified)

Energy units		Equivalent units if density of water is 10^3kg.m^{-3}		
per unit mass (potential Ψ)		per unit volume (pressure $\rho\Psi$)		per unit weight (head $\Psi.g^{-1}$)
erg. g^{-1}	joule. kg^{-1}	bar	atm	m water
1	10^{-4}	10^{-6}	9.872×10^{-7}	1.019×10^{-5}
10^4	1	10^{-2}	9.872×10^{-3}	1.019×10^{-1}
10^6	10^2	1	9.872×10^{-1}	1.019×10^{1}
10^3	10^{-1}	10^{-3}	9.872×10^{-4}	1.019×10^{-2}
1.013×10^6	1.013×10^2	1.013	1	1.032×10^{1}
9.813×10^4	9.813	9.813×10^{-2}	9.687×10^{-2}	1