INSTITUUT VOOR CULTUURTECHNIEK EN WATERHUISHOUDING

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The moisture flow technique for determining the water-balance

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The application of the technique to field conditions

The water-balance of any given area is determined by the many varying properties of soil and crop. Sub-surface run-off depends on the permeability of the sub-soil and the thickness of the permeable layer. Storage of water in the soil is determined by the distribution and magnitude of the pore spaces. Capillary movement is a function of the same pore-space distribution and of the capillary conductivity. Evaporation is determined by a number of properties such as radiation, windspeed, leaf area, soil moisture tension, and so on.

If one ponders on what the lysimeter - the commonest way of studying the water-balance - can reveal about the water-balance of an area with varying hydrological properties, it will be clear that this approach is of limited applicability. The conditions in the soil block of a lysimeter are unnatural, the variation in conditions very limited. The extrapolation of lysimeter findings to the varying conditions in the field is therefore restricted.

It also seems very uncertain as to whether the lysimeter technique, based on the measurement of volume or weight, may ever be adapted for use on large areas. Moisture sampling is very time-consuming, electric devices for moisture determination with resistance blocks are of limited reliability and are further hampered by the irregularity of the moisture distribution in the soil. It is understandable therefore that researchers seek other ways for the investigation of the water-balance.

The fundamentals of the moisture flow technique

The moisture stream, flowing from the atmosphere through the soil to a drainage canal or from the soil into the atmosphere may be depicted by a flow diagram. The water passes through four distinct sections of the flow path, the saturated zone in the soil, the unsaturated zone, the plant and the air. The water level in a drainage canal represents the potential at the lower end of the flow path, the evaporating capacity of the atmosphere represents the potential at the upper end.* If the flow is of a steady nature - with only the quantity of flow and the pressure-head as variable quantities - then, for each separate point along the path, one would require only one single determination of pressure or flow.

The amount of flow is a function of the pressure gradient. For every section it can be expressed as

$$Q = k\ f\left(\frac{dh}{dl}\right)F$$

where, Q = constant quantity of flow
F = cross-sectional area through which water flows
dh = difference in pressure head over a short length dl of the flow path
f = arbitrary function
k = hydraulic conductivity

The value of Q for a certain section is the same for the next section if no storage takes place. The relation may be re-written as:

$$\frac{f_1(dl_1)}{F_1}\ Q = k_1 f_1(dh_1) = k_2 f_2(dh_2)\ \frac{f_1(dl_1)}{f_2(dl_2)}\ \frac{F_2}{F_1} = k_3 f_3 dh_3\ \frac{f_1(dl_1)F_3}{f_3(dl_3)F_1} = \ldots\ldots\ldots$$

The subscripts indicate different sections of the flow path.

Because $\frac{f_1(dl_1)}{f_2(dl_2)} = c_2 \quad \frac{f_1(dl_1)}{f_3(dl_3)} = c_3 \ldots\ldots \frac{F_2}{F_1} = d_2 \quad \frac{F_3}{F_1} = d_3 \ldots\ldots\ldots$

with c and d constant, it follows that

$$k_1 f_1(dh_1) = c_2 d_2\ k_2\ f_2(dh_2) = c_3 d_3\ k_3 f_3(dh_3) = \ldots\ldots\ldots C = \text{constant}$$

* The evaporating capacity of the atmosphere is strictly speaking no potential, but a potential, multiplied with a number of constants. It may however be used in the same sense as a potential in this investigation into the number of required constants.

If C is known, then a value for k determines the value of f(dh). If f(dh) is known - and this means that h at the beginning and end of each section should be determined - then the k-values can be calculated.

If the course of the stream-flow is divided into the sections corresponding to the atmosphere, the plant, the capillary zone and the groundwater zone, then, for these four sections, the determination of five potentials and one absolute quantity of the flow will suffice. There are only six independent variables.

The stream-flow function may be complicated and may require a number of physical soil constants for its solution. However, these are properties that generally do not change appreciably with time: they may be determined beforehand and only once. Knowledge of these soil constants is a first requirement; a second is that the stream-flow function should be known. In the solution to be discussed later, there is sufficient opportunity to determine functions and constants and to check them on constancy of value.

If the flow is of a non-steady type the problem does not change greatly, but it will be clear that extra information is required. In order to get an absolute measure of storage capacity it is necessary to compare two quantities of water with the two corresponding levels of storage. It may be expected that for a larger number of corresponding quantities of water and levels of storage the storage capacity will show a simple relation with these two values. This furnishes a control.

An other control which ensures that the value found for the storage capacity is correct, arises from the periodic nature of the changes in the water-balance. If the sum of the quantities of storage and depletion over some length of time - for instance a year - becomes zero, then the levels of storage must at the beginning and end of this length of time be the same.

The variations in storage are in most cases related to the same potentials which govern the streamflow. For instance, if the groundwater table is nearer to the soil surface, the amount of stored water is increased. The pressure gradient and therefore the amount of flow may remain the same as before this increase in storage. Together the water levels in the soil and in the ditch not only give an idea of the amount of flow, but also of the amount of stored water. In this storage-study therefore, it is not necessary to determine new variables. However, it will be necessary to determine special storage constants, for instance the storage capacity of the soil. If storage is also included in the study, it is not necessary to determine more variables, but only to measure more constants which may describe the conditions of the flow.

Five potentials therefore not only suffice for the establishment of the quantities of flow, but also to determine the changes in storage. These five potentials should be measured at five points along the path of the stream-flow at the beginning and end of each section. If potentials are measured at more than this minimum number of points, then there is a possibility for adjustment of errors of determination. Also one may select the points of measurement in such a way that the determinations of quantity of flow and of storage are most profitable.

Application to the water-balance determination

The determination of the water-balance by streamflow technique has certain profitable aspects as well as some disadvantages. In the sequence, rainfall - capillary penetration - sub-surface run-off, the stream bypasses the plant. In this sequence interception in the canopy represents the storage in the section of the plant, but, in this section, the stream-flow is of no importance. In the sequence, sub-soil irrigation - upward capillary flow - flow through stems and leaves - vertical vapour transport in the atmosphere, the flow resistance in the plant has to be taken into account, but in this sequence, storage in the plant is negligible.

In the atmosphere storage is of no interest. With rainfall there is no flow resistance and with evaporation this resistance can accurately be included in one of the formulae for potential evaporation. For small crops, very often, interception is not determined. This is a limited disadvantage, because the intercepted water will seldomly stay on the leaves for a long time and over a longer period it may readily be included with the rain that reaches the soil or with the water that evaporates.

The water-balance determination may be simplified to a marked extent. If the sections corresponding to plant and atmosphere can be omitted, only three potentials are required. For the sequence of the upward flow one may restrict oneself to the determination of the water level in the ditch, of the groundwater level and of the potential evaporation. The rainfall is the fourth observation that is required for the downward flow and for the absolute basis from which the series of stream-flow relations can be calculated in terms of actual flowing quantities.

The formula for stream-flow and storage

For the formula for flow a choice can be made between steady and non-steady conditions. Several formulae of each kind are given in literature. Both types of formula have limitations. Because the steady flow formula is simplest, it is the most attractive. Where necessary the formula for tile drainage spacing is used.

Groundwater storage is so linked up with steady capillary storage, that these two are best treated together.

Capillary flow can, up to now, be described only in a formula for the steady case. Because even this formula is somewhat complicated, a nomograph was constructed to check the results of the graphical analyses.

Storage may be read from the same nomograph, which gives the relation between moisture content, groundwater depth and quantity of capillary stream-flow. The differences in moisture content in the profile for different depths of the water-table and different amounts of capillary flow give a more practical insight in these investigations than the absolute values of moisture content. For a variation of 1 mm. of flow a day or a water-level rise of 10 cm., this difference in storage may be calculated from the nomograph and used for comparison with the graphical solution, provided pF-curves for the different layers of the profile are available.

Besides the steady capillary storage, an unsteady storage has to be distinguished. This unsteady capillary storage corresponds to the water stored directly after a rain-shower during the period this amount of water is not yet distributed as capillary moisture according a steady state moisture profile. After some time, it will be distributed through the profile in conformity with the moisture content given by steady capillary flow. As far as present experience goes, in a normal agricultural soil this unsteady situation lasts only a part of a day. The rainfall of the preceding day is the most useful parameter. The accuracy of the solution of the water-balance for time intervals of 14 days is not materially decreased if this term is disregarded.

The formula for capillary flow only covers the flow through the soil. The flow through the plant has also to be described by a formula based on some physical conception if one wishes to check the empirical quantity of flow by a fundamental formula. Here lies the main difficulty. There does not yet exist any formula which gives the relation between potential and actual evaporation as a function of the moisture tension in the root-zone. This problem has still to be tackled entirely empirically.

The graphical method

The graphical solution is based on the identity of run-off, storage and evaporation on the one hand and rainfall on the other. In 3-dimensions one may, for 14 days' periods, plot rainfall R, groundwater depth minus depth of the water in the ditch z and the rise of the watertable within the 14 days Δz. In this figure (figure 0) a curved surface has to be constructed which, at the intersection with planes parallel to the R-z plane, corresponds to the steady run-off S. Steady flow is defined by Δz = zero. At points of intersection with planes parallel to the R-Δz plane, the storage B is depicted. Now the difference introduced by the rainfall R is equal to the sum of the real evaporation Er and the unsteady capillary storage C. Evaporation is mainly a function of the capillary potential in the root-zone. If this potential is not known, it may be replaced by the rainfall during the fortnight. Rainfall as an evaporation parameter, however, makes the solution of the water-balance much more difficult because several very disturbing correlations are introduced in this way. The non-steady storage is strongly related to the rainfall on the day preceding the reading of the watertable depth. Because of the slight influence of this non-steady storage, this term of the balance is best omitted if no figure for capillary potential is available.

Now the space, in which the points for each observation may be assumed to be dispersed, is divided into small intervals parallel to the R-Δz plane as well as to the R-z plane in such a way, that within each interval, sufficient readings are present. For each interval lines of average are drawn through the points as freehand curves and a check is made that the lines in both directions do not cross over each other, but intersect. In this case they are part of the same plane. If the formulae were sufficiently well established, the curves might be fitted by least squares. The plane is determined for groups of observations with more or less the same value for the potential evaporation.

If run-off curves are constructed for different values of potential evaporation this must give a discharge curve of identical shape for each level of evaporation. In the space diagram, where as a first approximation rainfall is taken as a measure for sub-surface run-off, the curves will show a vertical displacement for different values of potential evaporation. This vertical displacement accounts for the differences in real evaporation, which should be subtracted from the amount of rainfall to find the real run-off. The displacement compared with the curve for the winter months with low evaporation gives the first approximation for the real evaporation. The curve gives the first approximation for the sub-surface run-off.

Run-off

In figure 1, the first approximation for the run-off is given. By a reitterative procedure, in which the estimates for run-off, storage and evaporation are gradually improved, the accuracy may be increased. The last results are checked by comparing them with the drainage formula

$$\frac{S}{z} = A + Bz \qquad \text{where,} \quad A = \frac{8\ kd}{l^2}, \quad B = \frac{4k}{l^2}$$

S = run-off
z = pressure-head
A,B = constants for the same place in the terrain
k = permeability
d = thickness of aquifer
l = distance between drains or ditches

The test of the straight line relation by plotting $\frac{S}{z}$ against z as given by this formula is valuable though it is not very precise, because of inherent errors in the data and the limited variations in run-off. A better test is a comparison with run-off determined in the same area with lysimeters. This has been done for data from the Rottegatspolder. Figure 2 gives this result showing an error of 9 mm. for the run-off over 14 days.

Storage

The value for the storage capacity - the increase in moisture retention for a groundwater variation of 10 cm. - is a function of the groundwater depth and the excess of rainfall over real evaporation, whether positive or negative. For a restricted number of intervals of rainfall-excess and groundwater depths the storage data - obtained by subtracting run-off and evaporation from rainfall - are plotted against groundwater depth or rainfall-excess.

The curves for storage - to which the storage capacity is the tangent - are given in figure 3 for the fortnights with an evaporation from the free water surface of 4 mm. in 14 days with increasing groundwater depths. If from these graphs the tangents are determined and plotted against groundwater depth and rainfall-excess a result such as given in figure 4 is obtained.

To test this result, the storage for this profile, for the same streamflow intensity and various groundwater depths, was taken from the nomograph for capillary flow and the pF-curves for this profile. This comparison is given in figure 5. The curves found by the graphical analysis are sufficiently of the same type, but are obviously displaced to the right. It may be presumed that this is due to the cracks in this heavy acid clay, which do not appear in the soil samples taken for the pF-determinations because these samples are too small. The cracks may be several cm. wide and several dm. apart. Comparison of the two sets of curves leads to the assumption that the volume of these cracks will be 5% smaller in winter than in summer. This seems not unreasonable; there is no direct test for this, however. Because the determination of the volume of cracks does not appear to have an easy practical solution this might remain with heavy clays one of the difficulties for the future.

The calculated storage was tested by comparing it with the data for the moisture contents determined in the lysimeter area of the Rottegatspolder. Figure 6 gives this result; the error in this case is 14 mm.

<u>Evaporation</u>

If the corrected values for run-off and storage are subtracted from the rainfall, these data may be sorted out according to particular intervals of potential evaporation, rainfall and groundwater depth. There exists a strong correlation between potential evaporation and groundwater depth, which may only be overcome by collecting together the data from a few observation points within the same profile but with different water depths. This was not possible in this flat lysimeter area, so that no separation of the influence of water depth and potential evaporation could be made. Undulating land has, in this respect, an advantage over flat land.

The real evaporation plotted against potential evaporation for different depths of groundwater is depicted in figure 7. Because no theory on the exact shape and position of these lines exists and the accuracy and the distance between extremes does not permit any finer details, straight lines were drawn through the scatter-diagram. It is obvious that the curves must leave the origin at an inclination of about 45° and reach a maximum somewhere, corresponding with the highest possible evaporation, this maximum-curve being due to increased flow resistance in the plant and the soil.

In figure 8, the same data are arranged according to rainfall, with various values for evaporation from a free watersurface. Here the same applies, that the curves at some level of actual evaporation will approach a horizontal asymptote. They, however, do not start from the origin but from a point which indicates the capillary supply of water, in agreement with the depth of the groundwater level and the moisture tension in the root-zone.

What is not yet clear in this result is the influence of the depth of the watertable on the evaporation, because the capillary potential in the root-zone is not known and an indirect solution cannot be given with a sufficient accuracy. The intercepts of the lines in figure 8 with the vertical axis cannot be tested. It is clear that some relation between the intercepts - the evaporation in absence of rain - and the depth of the groundwater must exist. It is not known if influences of the plant also have to be taken into account.

The accuracy of the results was again checked with the results of the lysimeters of the Rottegatspolder. Figure 9 gives the result of this test. Here the mean error is 12,5 mm. for the evaporation over 35 days. This is about the same error as was found in the test of run-off and storage. If the error of the direct determination is put, approximately at 5 mm., it is clear that the error of the result of the stream-flow approach will remain above 10 mm. Some increase in accuracy may be achieved if a numerical adjustment, based on a good physical formula becomes available. Adjustment with casual empirical functions will not improve much upon the results now obtained and will probably give worse results. It is obvious that the moisture flow technique cannot claim to give the same accuracy as direct determinations. On the other hand it is less costly and time consuming.

Some final remarks

The stream-flow approach necessitates an explicit expression of what the process is expected to be, that distributes rainwater as storage, evaporation and run-off. Where water-table depths are observed over a number of years, frequently and regularly, these data may supplement, to a greater or lesser extent drainage trials, run-off measurements, groundwater depth trials and lysimeters. The work in the field is limited, the costs of the experiment are negligible, but the amount of work to be done at the desk is still very extensive.

Where the depth of the water-table is very great, where the variations in water-depth or evaporation are small, where the depth of the permeable layer is very variable or where, at peak run-off, the water flows over the land, or, at lower run-off values, through fissures in rocks, there the measurement of the groundwater depth, discussed here, may not be applicable. In these cases, the prospects of a simple application of the stream-flow technique with water depth determinations do not seem too bright. It is however possible to neglect under these circumstances, the problems of the saturated zone. With tensiometers at different depths it appears possible to determine how much water sinks below a depth at which no plant has any influence. This water will eventually help to stock the groundwater and add to the sub-surface run-off.

Quick and simple determinations regarding the width of the opening of the stomata in the leaves - already in use for advice on sprinkling irrigation - might provide a very promising supplement to the other determinations. They may promote a better understanding of water use by the plant or at least increase our knowledge where it is weakest. Also, potential determinations on lysimeters or self-recording moisture determinations with the new γ-ray apparatus in the field without a lysimeter, may add to our knowledge and enable us to develop the functional formulae of the water-balance, which will strongly promote the study of water relations in the field. The lysimeter will remain the cornerstone of water-balance studies; it should not, however, be looked upon as the only and entire foundation of this branch of science.

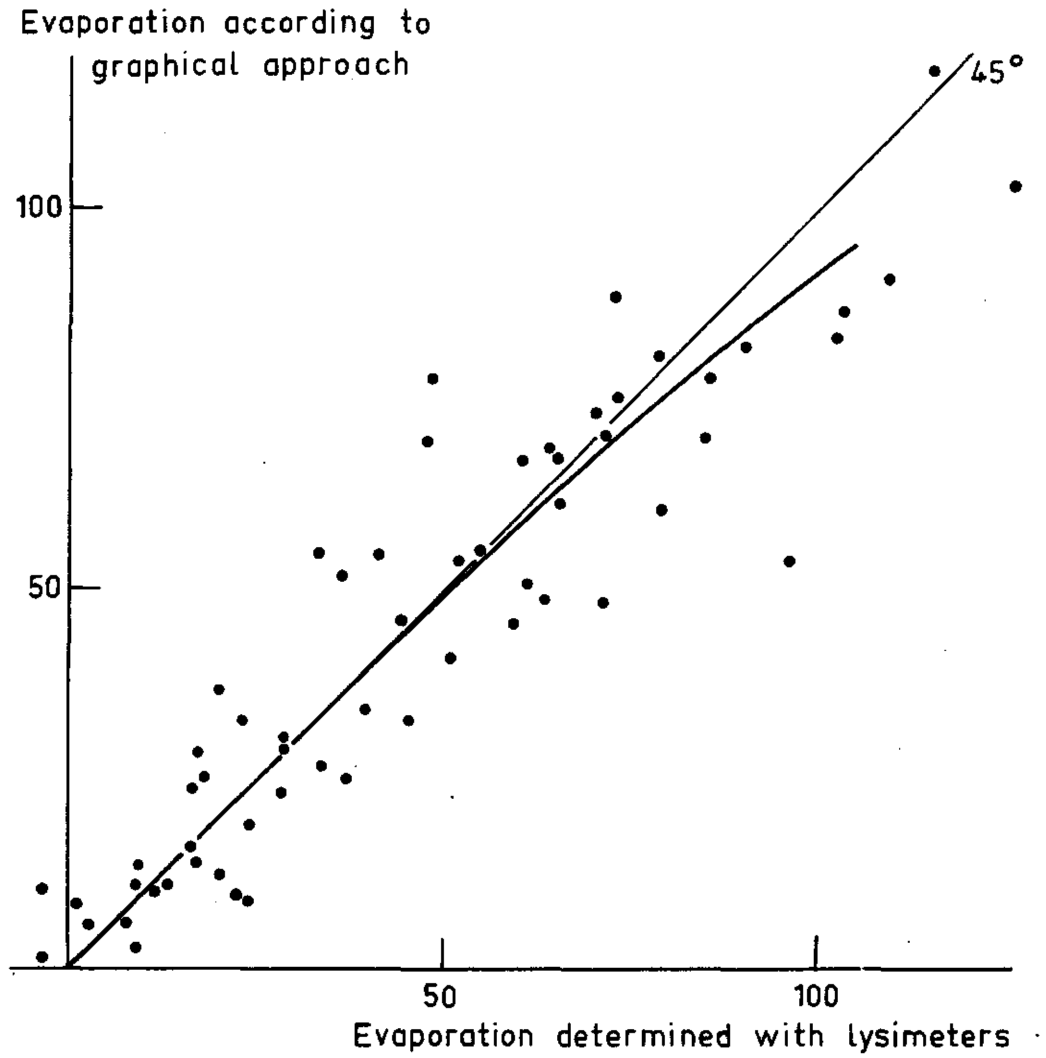

Fig. 9 The Rottegatspolder experiment yielded results for the evaporation from the lysimeter plots for time intervals of 35 days. From the graphical analysis the evaporation over the same intervals was calculated and plotted against the lysimeter results. The result is given in this scatter diagram. The mean error proved to be 12.5 mm. of moisture

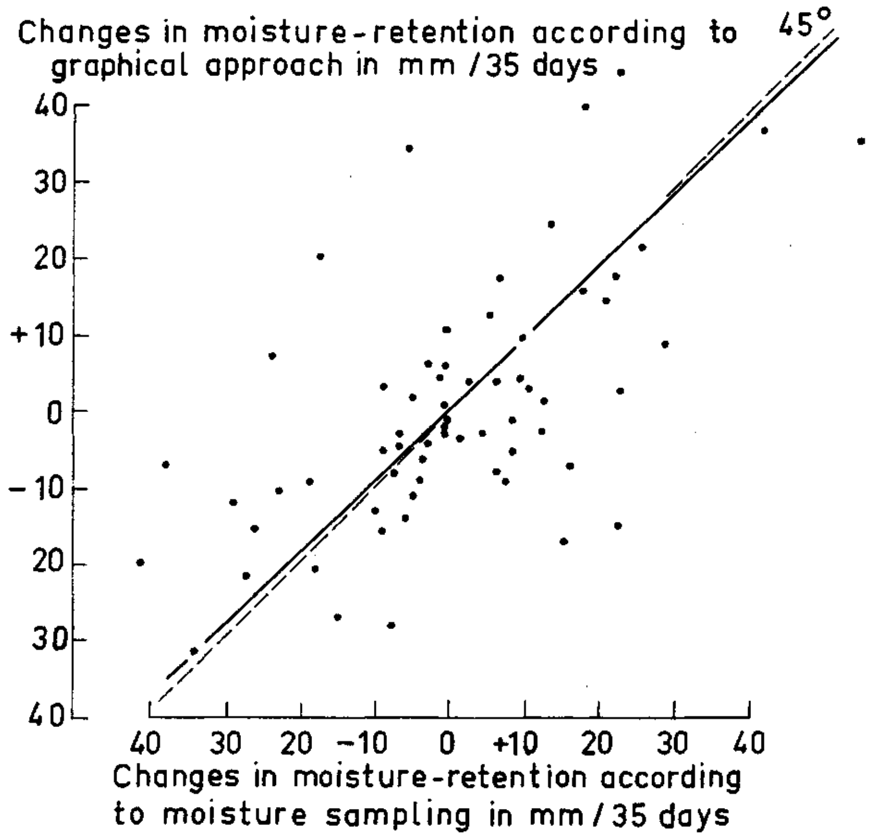

Fig. 6 In the Rottegatspolder lysimeter experiment the changes in moisture content of the profile are determined by soil sampling. These moisture content variations may also be calculated from the graphical analysis in fig. 4. The scatter diagram shows the extent to which the two sets of data agree

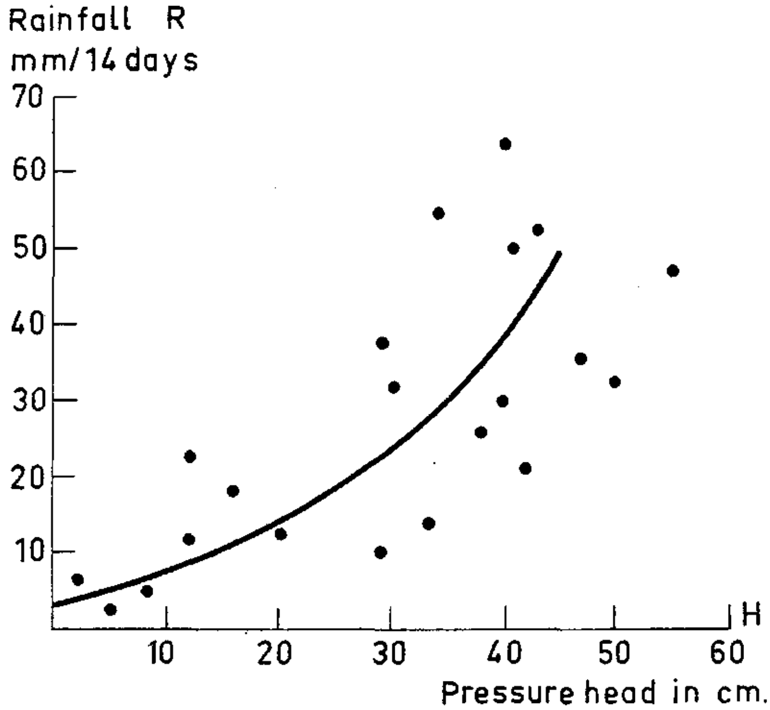

Fig. [illegible] Relation between rainfall intensity and the pressure head in a test-well with respect to the ditch-water level. The curve is valid for rainfall and evaporation over 14 days. The data were selected for intervals of [illegible] mm. of evaporation from open water and a fall of the water-table during these 14 days of [illegible] cm.

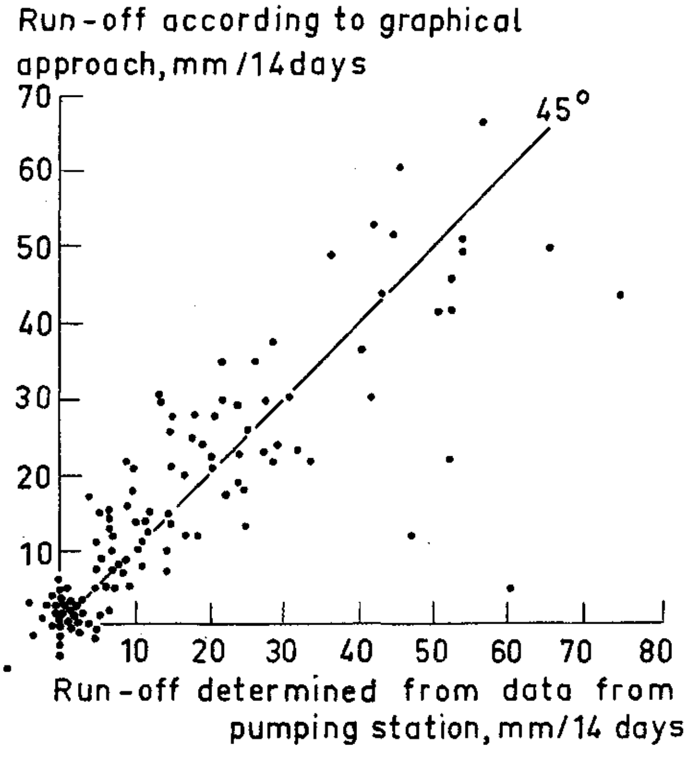

Fig. [illegible] Comparison between the discharge of the catchment area taken from the discharge data of the pumping station and the discharge determined by the graphical analysis

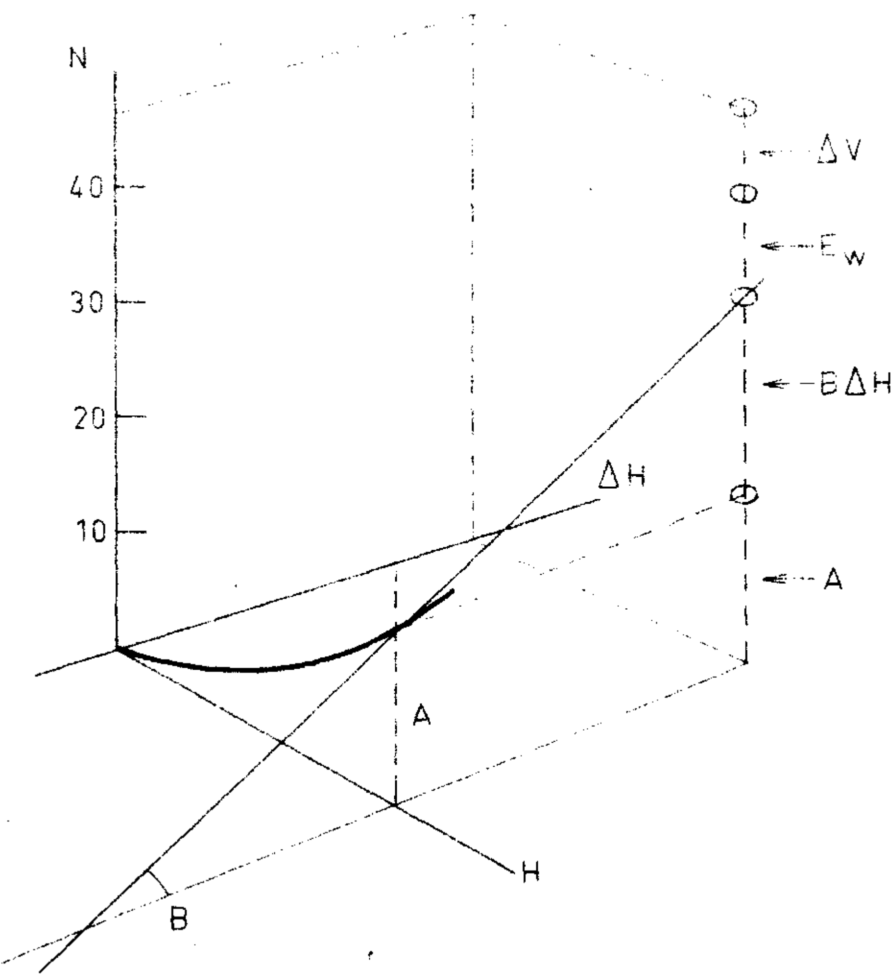

Fig. 6 The possibility of solving, in principle, the equation of the water-balance follows from the solid diagram. Rainfall R is split up into the discharge S, storage BΔz, actual evaporation Er and unsteady storage C. The parameters for these various quantities differ and are for A: z, for B: Δz, for Er: Eo, R and z, for C: time t. Because the parameters are not mutually dependent the equation can be solved

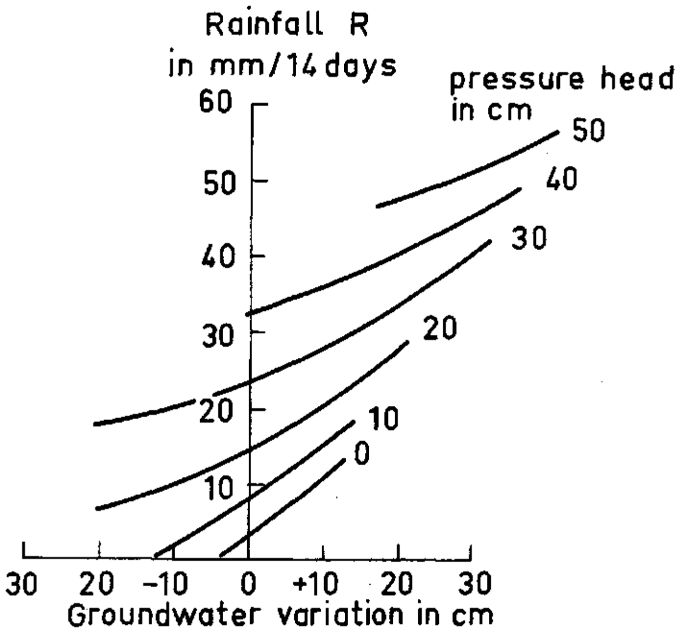

Fig. 3 Relation between change in groundwater level, rainfall and pressure head for periods with an average of 4 mm. evaporation from an open water surface for a time interval of 14 days. The inclination of the lines is a measure for the storage capacity

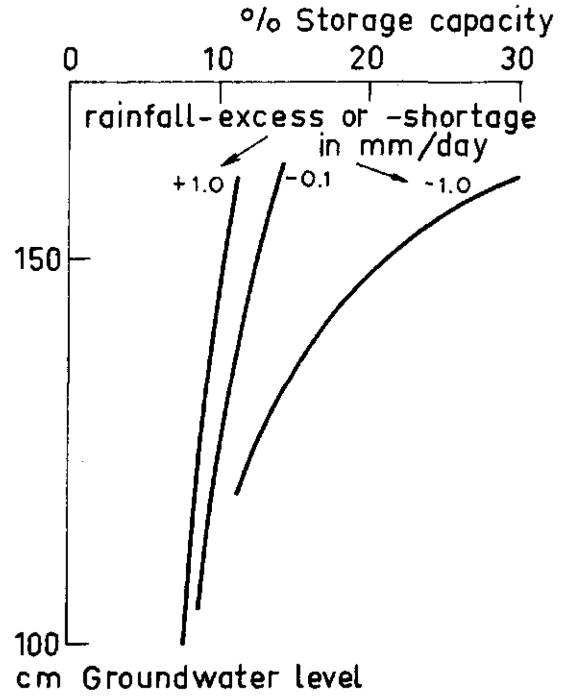

Fig. 4 The storage capacity varies according to the depth of the groundwater table and the direction and intensity of the capillary flow. The results of the analysis are given for zero flow, downward flow (+ 1.0) and upward flow (- 1.0)

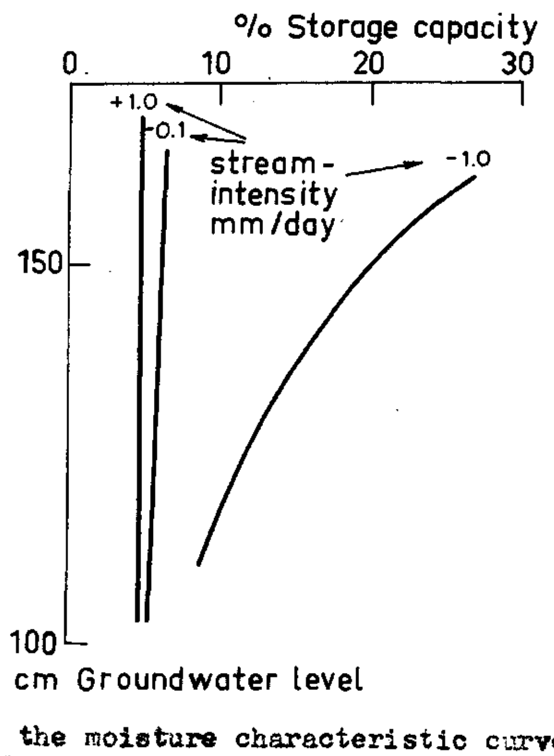

Fig. 5 From data of the moisture characteristic curve and the formula for steady capillary flow the air content of the soil may be calculated for the same capillary streamflow intensities as given in fig. 4. The difference in the position of the curves gives an indication of the influence of the non-capillary pore space of the coarse soil structure which is not taken into account in determining the moisture characteristic curve

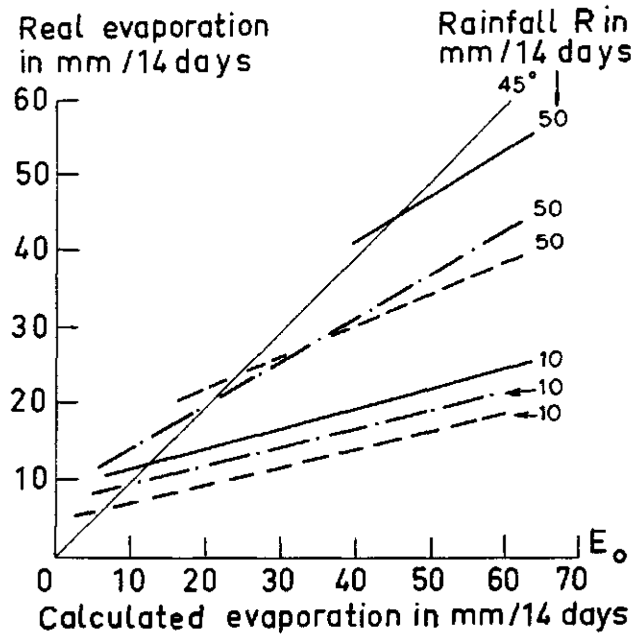

Fig.7 Actual evaporation is governed by the evaporating capacity of the atmosphere given by the Eo value and the rainfall during the period. The depth of the water-table - at the same time a measure for moisture depletion of the soil - also exerts an influence

water-table 155 cm. - full line
" 135 cm. - dashed line
" 115 cm. - dash-dot

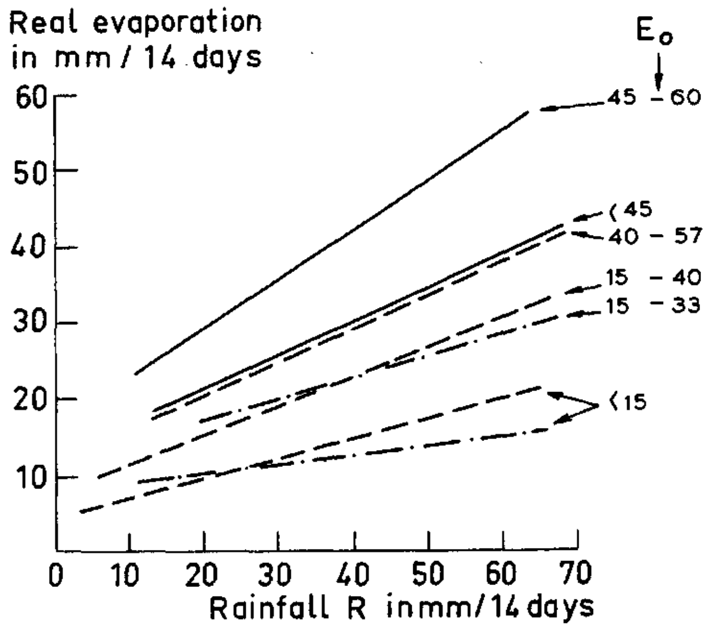

Fig.8 The four dimensional relation between actual evaporation, rainfall, evaporating capacity and ground-water depth, shown in fig. 7, may be presented in another projection, showing how strongly rainfall influences actual evaporation. The lines represent the mean of the interval of Eo, given in the graph, and the depth of the water-table indicated by the different lines, as mentioned in fig. 7

59
86